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Eugen Seibold
Wolfgang Berger

The Sea Floor

An Introduction to Marine Geology

Fourth Edition

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An Introduction to Marine Geology

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Eugen Seibold
Freiburg, Germany

Wolfgang Berger
Geosciences Research Division
Scripps Institution of Oceanography Geosciences
Research Division
La Jolla, California
USA

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Preface to the Third Edition

Man's understanding of how this planet is put together and how it evolved has changed radically during the last 30 years. This great revolution in geology – now usually subsumed under the concept of *Plate Tectonics* – brought the realization that convection within the Earth is responsible for the origin of today's ocean basins and continents, and that the grand features of the Earth's surface are the product of ongoing large-scale horizontal motions. Some of these notions were put forward earlier in this century (by A. Wegener, in 1912, and by A. Holmes, in 1929), but most of the new ideas were an outgrowth of the study of the ocean floor after World War II. In its impact on the earth sciences, the plate tectonics revolution is comparable to the upheaval wrought by the ideas of Charles Darwin (1809–1882), which started the intense discussion on the evolution of the biosphere that has recently heated up again. Darwin drew his inspiration from observations on island life made during the voyage of the *Beagle* (1831–1836), and his work gave strong impetus to the first global oceanographic expedition, the voyage of *HMS Challenger* (1872–1876). Ever since, oceanographic research has been intimately associated with fundamental advances in the knowledge of Earth. This should come as no surprise. After all, our planet's surface is mostly ocean.

This book is the result of our conviction that to study introductory geology and oceanography and environmental sciences, one needs a summary of the tectonics and morphology of the sea floor, of the geologic processes active in the deep sea and in shelf seas, and of the climatic record in deep-sea sediments.

Our aim is to give a brief survey of these topics. We have endeavored to write for all who might be interested in the subject, including those with but little background in the natural sciences. The decade of the 1980s was characterized by an increasing awareness of man's dependency on natural resources, including the ocean as a weather machine, a waste bin, and a source of energy and minerals. This trend, we believe, will persist as resources become ever more scarce and as the impact of human activities on natural cycles escalates in the coming decades. An important part of this awareness will be an appreciation for the elementary facts and concepts of marine geology, especially as they apply to processes within hydrosphere and atmosphere.

In what follows, we shall first give a brief overview of the effects of endogenic forces on the morphology of the sea floor. Several excellent summaries for the general reader are available for this topic, which is closely linked to the theory of continental drift, and has been a focus of geologic discussion for the last three decades. For the rest, we shall emphasize the exogenic processes, which determine the physical, chemical, and biological environment on the sea floor, and which are especially relevant to the intelligent use of the ocean and to an understanding of its role in the evolution of climate and life.

The results and ideas we report on are the product of the arduous labors of many dedicated marine geologists. We introduce some particularly distinguished scientists by portrait (Fig. 0.1). Of course, there are many more, and most of them are alive today. We have occasionally mentioned the authors of important contributions. However, we did not find it possible in a book like this to give credit systematically where it is due. We sincerely apologize to our colleagues for this unscholarly attitude, citing necessity in defense. For those who wish to pursue the subjects discussed in greater depth, we append suggested readings at the end of each chapter, as well as a list of key references.

For this second edition, we have extensively rewritten those parts of the first edition where substantial and fundamental progress has occurred in the fields of interest. Also, we have incorporated many of the suggestions for improvements that were communicated to us by several colleagues and reviewers. There are, however, limitations to the scope of subjects that can be treated in a short introduction such as this: we attempted neither a balanced nor an encyclopedic survey of all of marine geology with its many ramifications. We tried to keep highly technical information to a minimum, relegating certain necessary details to the Appendix.

Both authors wish to express their profound gratitude to collaborators and students who, over the years, have shared the excitement of discovery and the toil of research on numerous expeditions and in the laboratory. We also owe special thanks to the colleagues who helped us put this book together, by sending reprints and figures, or by offering advice.

Freiburg, Germany E. Seibold
La Jolla, CA, USA W.H. Berger
Spring 1993

Preface to the Fourth Edition of "The Seafloor"

This book is the fourth edition of the Seibold-Berger text on elementary marine geology mainly based on introductory lectures to students in Kiel and in La Jolla. W. Berger added materials concerning new developments in the field, some 30 years after E. Seibold determined the nature and range of subjects discussed in the second and third editions of the text. There are several things that set this text apart from many similar ones. Eugen Seibold (1918–2013), distinguished pioneer of marine geology (Fig. 1), emphasized observation of modern marine environments and the relationships between ongoing ocean processes and ancient marine rocks. His interest in ancient rocks and in sedimentation on Atlantic-type margins is reflected throughout in the book.

E. Seibold emphasized open questions, that is, the fact that much remains unknown in the (historically very young) fields of geology and especially of marine geology, notably at the cutting edge of exploration. As a consequence, he emphasized elementary findings that have proven their worth. He favored simple conceptualization, as in his classic paper on sediments in shelf seas (Fig. 2; Sect. 9.5.1). He clearly preferred concepts based on observation to nomenclature and to speculation. The term "new" did not carry special weight with him. On the contrary, if a newly introduced concept had not run a decade-long gauntlet of critique and survived, he remained doubtful of its viability. His basic philosophy is evident in all editions of *The Sea Floor*, including the present one. Also, it governs his book *The Memory of the Sea* (in German), and it emerged strongly in discussions, official or private. Also, it helped guide the synthesis reports of Leg 41 of the Deep Sea Drilling Project (off NW Africa), for which he was co-chief (together with the marine geologist Yves Lancelot).

Since the time of the early editions of this book, emphasis has grown in "earth system science," with forays into geophysics, geochemistry, oceanography, and indeed all of the

Fig. 1 Eugen Seibold, pioneer of marine geology (Photo courtesy of Dr. Ilse Seibold, Freiburg)

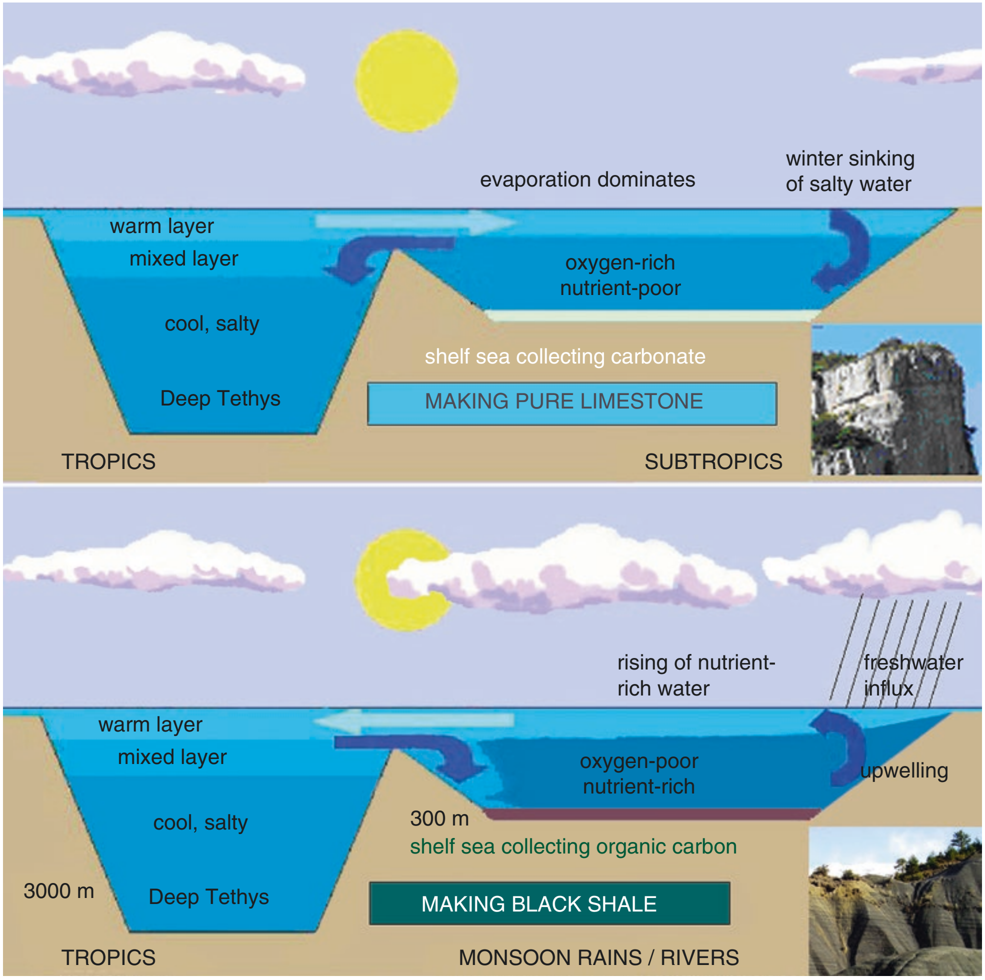

Fig. 2 Eugen Seibold's scheme of making carbonate or organic-rich shale in shelf seas, depending upon the contrast in circulation in arid and humid regions. Upper panel (**a**): arid conditions, schematic; compare Persian Gulf. Lower panel (**b**): humid conditions, schematic; compare Baltic Sea. Inset photos: Mesozoic shelf deposits, marine carbonate rocks (arid) and black shale (humid) in southeastern France (for oceanography, see Sect. 9.5) (After ideas of E. Seibold, published in 1971)

climate-related sciences including ecology (Figs. 3 and 4). It is an approach that Eugen Seibold urged and fostered. In his acceptance speech of the Blue Planet Prize (in 1994), he said this: "What is a marine geologist? A marine geologist investigates the present situation of the sea-floor and the processes which shape it. Furthermore, he tries to learn from the layers beneath the seafloor, i.e. he tries to learn from the past. With this knowledge from the present and the past, he has a responsibility to comment also on future developments if he is able to do so with scientific reasoning …" Evidently, he saw a marine geologist as a scientist who takes the ocean and climate change seriously. In this fourth edition, I have emphasized this approach. Space requirements calling for trade-off in space resulted in some cutting back of important items, notably the celebrating of contributions of some important pioneers he had identified.

Time scale matters in all of history. We now do have an excellent scale for the entire Cenozoic (i.e., the last 65 million years) largely through the untiring efforts of the Woods Hole biostratigrapher W. A. Berggren and his colleagues. A reliable scale is necessary to put rates of change in evolution and items of geologic history into perspective (Fig. 5). The established geologic time scale for pre-Cretaceous time is from reliable and traditional sources, being fundamental in geologic work (Fig. A3.1, in the Appendix).

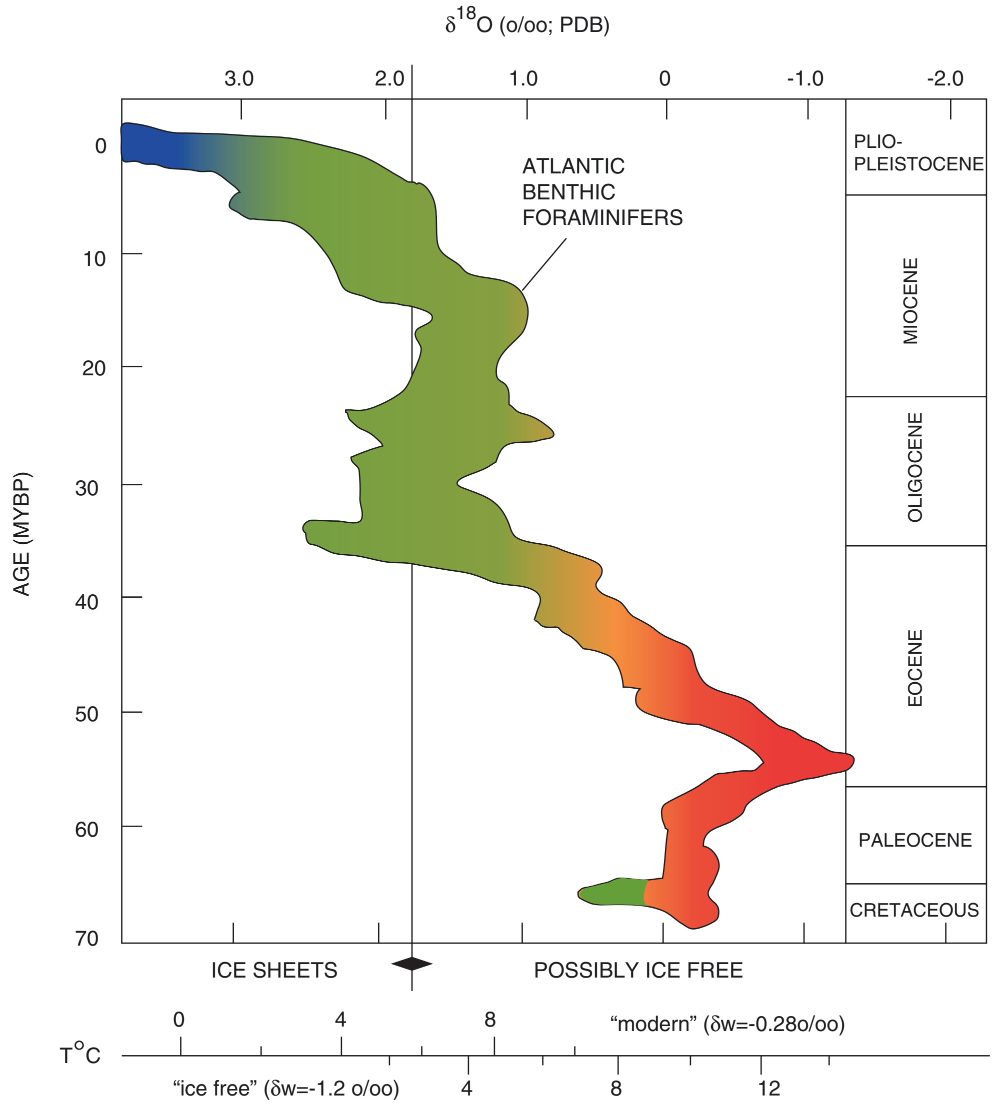

Fig. 3 A chief result from deep-sea drilling: the cooling of the planet since the early Eocene as seen in a temperature proxy on the deep seafloor (oxygen isotopes of benthic foraminifers; red, warm; yellowish green, intermediate; blue, cold (ice age)) (After K. G. Miller, R. G. Fairbanks, and G. S. Mountain, who compiled data from (Atlantic) DSDP sites (1987; Paleoceanography 2:1))

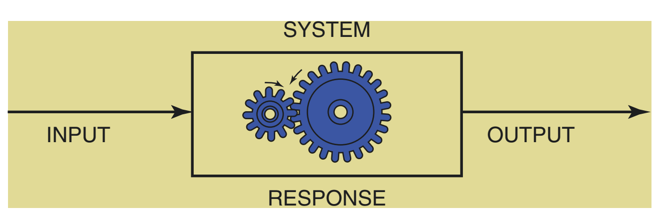

Fig. 4 The concept of earth system science. Input is exogenic and endogenic forcing; output is the recorded climate change and sedimentation (After J. Imbrie et al., 1982, in W. H. Berger and J. C. Crowell, (eds.) Climate and Earth History. Studies in Geophysics)

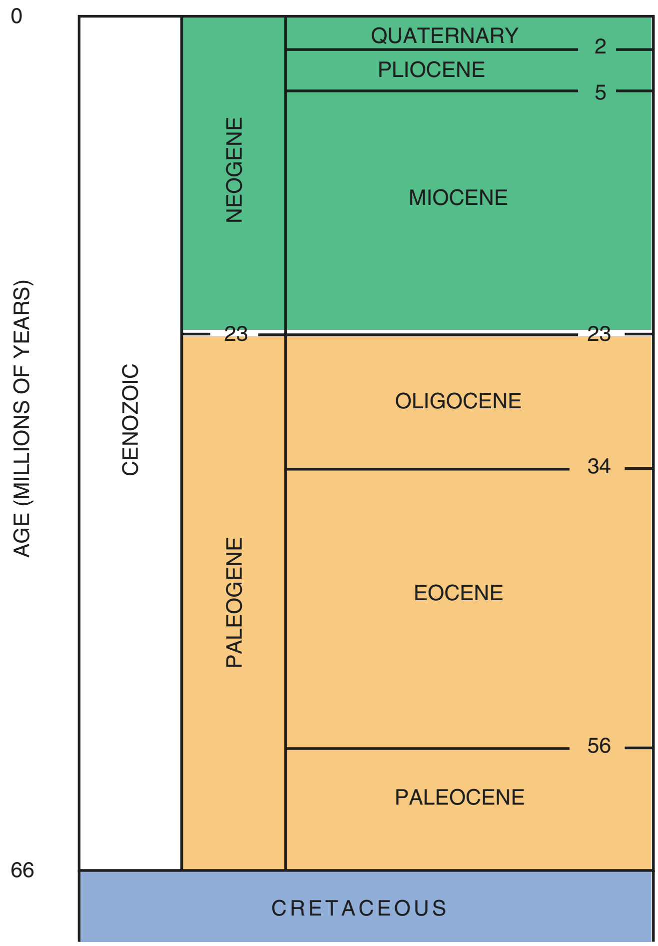

Fig. 5 Cenozoic time scale, simplified. Time scale of "the Age of Mammals," that is, the time since the Cretaceous, following the extinction of ammonites and dinosaurs. Numbers are published estimates of ages of stratigraphic boundaries in millions of years, mainly from the most recent ODP volumes. Note the great lengths of the Eocene and the Miocene, epochs that dominate the Paleogene and the Neogene periods, respectively. In turn, Cenozoic sediments and events dominate the marine geologic history of the modern ocean. Cretaceous deposits are found below somewhat less than one half of the seafloor

I am indebted to Eugen Seibold for many discussions and also to many other colleagues (including my mentors D. L. Eicher, Colorado, F. B. Phleger and F. L. Parker (La Jolla), and Gerold Wefer at the Marum Institute, University of Bremen) for advice or for offering (or reviewing) illustrations of important geological concepts. Authors are acknowledged in the appropriate figure captions. Many others, including pioneers in the field, contributed important ideas. Trying to mention them all here would run the risk of leaving off many important contributors. Many or most are listed in "suggested readings." In any case, it is well to realize that the selection of "pioneers" is quite arbitrary. Older pioneers of marine geology (starting in the nineteenth century) tend to be underrepresented, and the reverse is true for teachers and colleagues of the authors of this book.

Concerning this or any other textbook, it may be well to keep in mind what the famous Californian physicist Richard Feynman (1918–1988) said; that is, science begins with doubting traditional textbook assertions. Feynman made an interesting observation, but actually marine geology is too young a science to have a long list of textbooks for testing his statement. It seems that this particular field advanced not so much by raising doubts about what was being taught by the professionals but mainly by making new observations and measurements, commonly by using new methods, and by integrating with results from other disciplines (including physics). In this actual history of scientific research, much new information was delivered by geophysics ("physics applied to geology" in the words of erstwhile S.I.O. director Fred Spiess) and by deep-ocean coring and drilling (i.e., by engineering feats) in the second half of the last century. The ensuing results have changed our understanding of all aspects of seafloor lore. And yes, the advances did make old geology texts obsolete while building on established concepts that remained useful.

General Background and References

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Eugen Seibold (1918–2013)

Seibold was born in Stuttgart. He studied geology in Bonn and in Tübingen. Subsequently, he taught at the University of Tübingen but moved to Kiel in 1958 to study modern marine sedimentation and manage the Geological-Palaeontological Institut of the university as its director. From 1980, he accepted positions as the president of the DFG (the German National Science Foundation), as the vice-president of the European Science Foundation, and as the president of the International Union of Geological Sciences. He was president of the European Science Foundation from 1984 to 1990 and a member of various academies, including the Leopoldina (Akademie der Naturforscher, Halle, in Saxony-Anhalt) and the Académie des Sciences in Paris.

Seibold's many contributions to geology were well recognized – he was a recipient of internationally known awards (e.g., the Gustav Steinmann Medal, the Hans Stille Medal, the Leopold von Buch Plakette), as well as the Walter Kertz Medal in geophysics and the Blue Planet Prize of the Asahi Foundation. The Asahi Foundation's prize especially recognizes contributions of relevance to society. The prize was used, in part, to fund the Eugen and Ilse Seibold Prize, an award furthering Japanese-German scientific interaction.

Among outstanding paradigms within Seibold's many contributions (including geologic education), one might emphasize his insights regarding the role of exchange between marginal basins and the open sea in determining the deposits accumulating in shelf basins. He assigned an estuarine-type exchange to black shale sedimentation and an anti-estuarine type to carbonate deposits. Both types of sediment are prominent in the geologic record (and are conspicuous in the Jurassic of southern Germany, his original training ground). Significantly, black shales are commonly a source for hydrocarbon products, while carbonates often serve as reservoir rocks. Obviously, both rock types help define our time in human history. It is typical for Seibold that he thought we should know about their origin.

[Source of information: largely the Deutsche Forschungsgemeinschaft, DFG]

La Jolla, CA, USA W.H. Berger
December 2015

Contents

1 Introduction

1.1 The Great Geologic Revolutions of the Twentieth Century

1.1.1 General Background Information

There were an enormous number of striking geologic discoveries made and geologic theories created in the twentieth century. All of these bear importantly on marine geology. Four findings stand out: (1) *plate tectonics* (linked to continental drift and based largely on the geomorphology of the seafloor, geomagnetism surveys, heat flow patterns, and earthquakes at sea), (2) *Orbital Ice Age Theory* (informed by solar system astronomy and confirmed by the study of deep-sea sediment), (3) *stepwise Cenozoic cooling* (based on results from deep-sea drilling), and (4) *confirmation of the impact theory for the end of the Mesozoic* (clinched by stratigraphy of pelagic sediments on land). The respective widely recognized pioneers are (1) a number of largely US American and British geologists, geophysicists, and geomagnetists (e.g., Lamont's marine geologist Bruce Heezen (1924–1977), the US Navy's Robert Dietz (1914–1995) and Harry Hess (1906–1969), the UK geophysicist Fred Vine (Ph.D. 1965, Cambridge)), and also the German meteorologist Alfred Wegener (1880–1930); (2) Milutin Milankovitch (1879–1958), Serbian astronomer and civil engineer, and two astronomers (John Stockwell and Urbain Leverrier) delivering input to his calculations (The leading contemporaneous proponent of orbital forcing is André Berger, Belgian astronomer and climatologist.); (3) contemporaneous pioneers are the NZ-US marine geologist James P. Kennett (Ph.D. 1965, Wellington), the British geophysicist Nick Shackleton (1937–2006), the isotope chemist Sam Savin (Ph.D. 1967, Pasadena), and the geologist Robert Douglas (Ph.D. 1966, U.C. Los Angeles); and (4) the impact pioneers are the German-Swiss geologist and paleontologist Hanspeter Luterbacher, the Italian geologist Isabella Premoli Silva, and the Californian physicist Luis Alvarez (1911–1988) and his geologist son Walter (Berkeley) and their associates F. Asaro (1927–2014) and H. Michel (Berkeley). The crucial papers and books were published (1) in the 1920s and 1960s (*Continental Drift* and *Plate Tectonics*), (2) in the 1920s and 1980s (Orbital Ice Age Theory, proposed and verified), (3) in the 1970s (microfossils and oxygen isotopes), and (4) in the 1960s and 1980s (sudden end-of-Cretaceous mass extinction documented in pelagic sediments on land surveyed and iridium maximum found, respectively).

What, if anything, do the four great revolutions have in common?

They emphasize the control of geologic history by outside forcing, either "endogenic" (processes driven by mantle convection) or "astronomic" ones controlled by solar system phenomena and a call on "positive feedback" for amplification ("negative feedback" stabilizes).

The discovery of the great importance of outside forcing (including the unpredictable and highly variable factors of earthquakes, volcanism, and collisions in space) has resulted in some discrimination against regular surficial Earth processes, which have considerably damaged the perception that the pronouncements of the British lawyer Charles Lyell (1797–1875) regarding uniformitarianism (a long bastion of textbook geology) hold water. His central concept, that the present is the key to the past, became suspect as a dominating rule in the interpretation of the geologic record, especially in an ice age. The same is true for the reverse assertion that the past is the key to the present (or to the future). "Endogenic forces" are difficult to observe, being generated in the mantle of the planet. While regular astronomic forcing can be calculated with great precision, it is not necessarily well understood in its applications to a complicated Earth response.

In system analysis (which started as a concept in engineering), "negative feedback" drives a system toward original conditions during episodes of change. Negative feedback favors the *status quo*. The idea of negative feedback of a chemical system is the backbone of "Le Chatelier's Principle," proclaimed many decades ago by the French pioneer chemist Henry Louis Le Chatelier (1850–1936) and

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soon after elaborated for the Earth system by the Russian-French geochemist Vladimir Vernadsky (1863–1945). The "daisyworld" model of the UK chemist James Lovelock (Ph.D. 1948, London) beautifully illustrates the negative feedback concept. "Positive feedback," in contrast, increases any change and eventually leads to blowout if not checked. As a corollary, when positive ("non-Gaian," K.J. Hsü) feedback is at work, relatively modest forcing can result in enormously large changes. In a very general way, negative feedback supports Lyell and Gaia and traditional geologic thinking going back almost two centuries, while positive feedback does not: it produces unexpected results and abrupt change and tends to be rather hazardous, therefore.

1.1.2 Plate Tectonics and Endogenic Forcing

A little more than half a century ago, it was still possible to think that sediments on the deep seafloor offer information for the entire Phanerozoic, that is, the last half billion years. In fact, some geologists thought the deep-sea record might lead us back even into the Precambrian, into times a billion years ago, or more. Today, it is no longer possible to harbor such thoughts. The deep seafloor is young, geologically speaking. The most ancient sediments recovered are about 150 million years old, that is, less than 5% of the age accorded to ancient rocks on land. Deep-sea deposits older than Jurassic presumably existed at some point, but it is thought that they vanished, entering the mantle by subduction in trenches.

The main relevant activity providing for the forcing of plate tectonics is in the mantle of Earth (Fig. 1.1). It is not known just how the mantle operates in the context of plate tectonics.

"Seafloor spreading" is responsible for the world-encircling mid-ocean ridge and for the Atlantic. Volcanic activity associated with subduction (Fig. 1.2, lower panel) is rampant all around the Pacific Basin (hence, the label "Ring of Fire" for the Pacific margin). It is especially evident in South America (Fig. 1.3) but also on the northern West Coast of the USA, in Alaska, and in Japan. Earthquakes generated in the subduction belts can and do produce "tsunamis," that is, waves that travel on the surface of the ocean at the speed of jet aircraft and that grow to enormous size in shallow water.

1.1.3 Northern Ice-Age Cycles

Unlike his predecessors, the ice-age pioneer Milutin Milankovitch supposed that it is forced melting of ice in high northern latitudes that holds the answer to the cycles and not the pulsed making of ice (which is the question that was traditionally addressed). According to Milankovitch, what matters is whether the sun's warmth is strong in northern summer or not. The influx of solar heat is controlled by the changing tilt of the Earth's axis (the "obliquity") studied quantitatively by the French astronomer Pierre-Simon de Laplace (1749–1827) and by the changing Earth-Sun distance, a topic tackled by Johannes Kepler (1571–1630) some four centuries ago. Milankovitch's proposition, worked out mathematically, is now known as "Milankovitch Theory," reflecting its

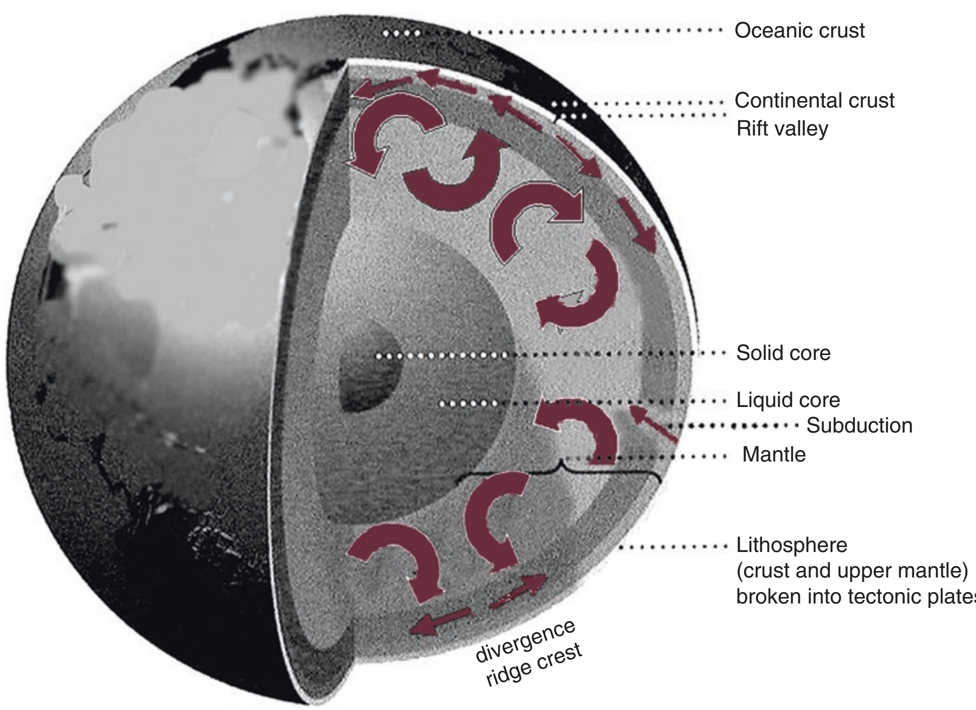

Fig. 1.1 The main elements of the planet's structure, not to scale and hypothetical regarding mantle motion. Roughly one half of the Earth is mantle; crust is negligible by volume, in comparison. The mantle convects, largely in unknown ways. Alternatively to the picture shown, fast convection may be restricted to an upper layer, while convection in the lower mantle is much slower. The lithosphere (the stiff uppermost part of the mantle) is roughly 100 km thick. Together with the crust, it defines the "plates" (Background graph courtesy of S.I.O. Aquarium Education Program, here modified)

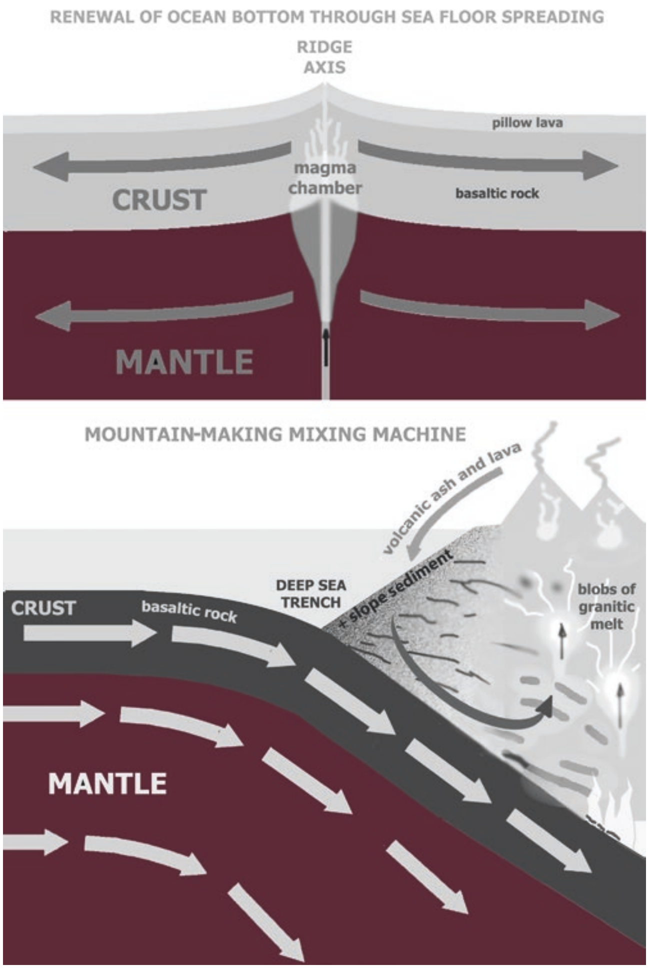

Fig. 1.2 The crux of modern global tectonics involving the seafloor. *Upper panel*: seafloor spreading; *lower panel*: seafloor subduction. The latter involves the making of mountain chains. The assumption is that creation and destruction of seafloor are balanced, so that Earth does not change its size. The downward-going lithospheric slab (about 100 km thick) eventually dissolves within the mantle. Drawing is not to scale (After W. H. Berger 2009, Ocean. U. of California Press, Berkeley, with some minor modifications)

Fig. 1.3 Volcanic eruption linked to subduction "Ring of Fire": Turungahua, Ecuador, in 2003 (Photo courtesy of Chung Luz, UCSD). The volcano ("the Black Giant") has frequently erupted in recent years, causing evacuations. On the positive side, volcanic ash acts as fertilizer

great power in explaining ice-age cycles (and earlier climate cycles on a multi-millennial time scale).

Milankovitch Theory employs seasonal effects as influenced by Earth's orbital variation. Right now Earth is closest to the sun early in January, so according to Milankovitch, ours would not be a good time to melt snow and ice with the help of astronomic forcing. Ten- or twelve-thousand years ago, the Earth was closest to Sol in high northern summer, and melting of the great northern ice sheets proceeded at a rapid clip. Sea level rose rapidly (as far as is known from marine sediments), the average exceeding a rate of 10 m per millennium. Apparently, after having been around for more than 50,000 years (among other things depressing the land it sat on, lowering elevations), large ice masses were ready to go away. The degree of eccentricity of Earth's orbit varies on a time scale of a hundred thousand years, so that a 100,000-year natural oscillation could conceivably lock on to an orbital energy cycle of similar length (such as eccentricity), even if not closely related. The tilt of the axis (that is, the deviation from an idealized vertical position) varies with a period of 41,000 years. As far as can be ascertained, overall, the Milankovitch mechanism matches observations quite well, although the nature of the match varies with geologic time (Fig. 1.4) for reasons poorly understood.

Correct dating of the cycles seen within a reliable sedimentary record was necessary to support Milankovitch's proposition. Once it is realized that the evidence simply was not available (e.g., no records from long deep-sea cores), it becomes much more understandable why Milankovitch Theory did not prevail earlier. It is not that geologists are uncommonly obtuse, it is that they are skeptical, as are almost all well-trained scientists when examining the hypotheses of colleagues. Verification of Milankovitch Theory required much effort and a challenging list of information. Even today there are obstacles to an unfettered acceptance of Milankovitch Theory. One of these arose toward the end of the twentieth century and became prominent. Milankovitch studied but the last one third of the ice ages, a time period when 100,000-year cycles dominated ice-age history. He did not, however, consider fluctuations of that length. Instead, Milankovitch focused on the effects of "precession" and of "obliquity" (changes in the season of perihel and changes in tilt of Earth's axis). The matter of the dominant 100,000-year cycle is still unresolved and is being discussed. Many scientists believe that Milankovitch-type forcing alone cannot resolve the issue.

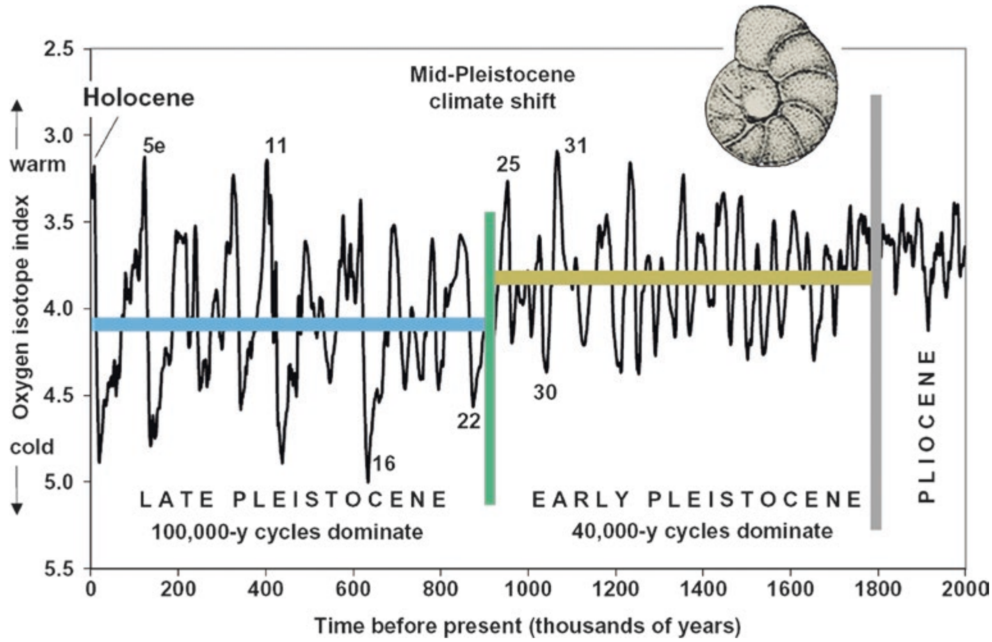

Fig. 1.4 Ice-age record in deep-sea sediments (oxygen isotopes in benthic foraminifers, a proxy for temperature and ice mass) (Data interpolated from Zachos et al. 2001, Science 292:686). The benthic foraminifer (closely related to the benthic signal carriers, genus *Cibicides*) is after a drawing by Jan Drooger. Numbers are MIS designations (marine isotope stages) following Cesare Emiliani – odd for warm stages. The last prominent MIS (25) in the warmish early Pleistocene occurred about a million years ago. Horizontal bars denote average conditions; vertical ones are climate boundaries. Note the changes in dominant cycles. Such changes are not explained by Milankovitch Theory

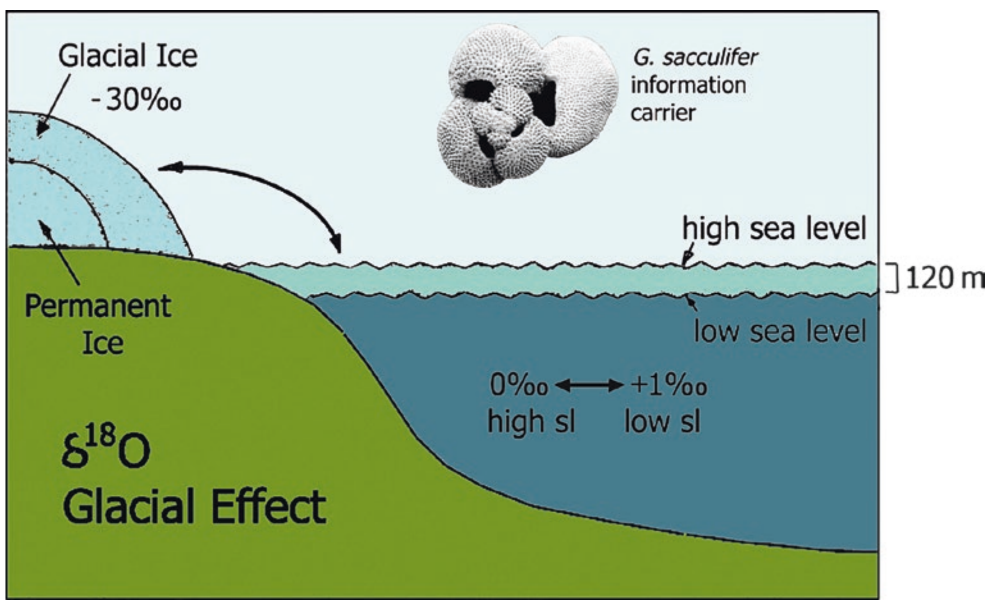

Fig. 1.5 Oxygen isotope ratios as a proxy for polar ice mass according to the Italian-American paleontologist and physical chemist Cesare Emiliani (1922–1995). The ice differs in composition from seawater. Thus, as the ice waxes and wanes, the ocean changes isotopic composition, which is reflected in the shells of foraminifers. Ratios are expressed as deviation from a standard, in permil (equal to ten times the value of percent). There also is a temperature effect, which was emphasized by Emiliani. The signal carrier shown (*Globigerinoides sacculifer*) is a warmwater planktonic species

1.1.4 Deep-Ocean Drilling: Discovering New Worlds

The rise of the new tectonic paradigm greatly encouraged the launch of a major venture in marine geology: the exploration of the deep seafloor by drilling. In fact, perhaps the most celebrated result of deep drilling was the support garnered for plate tectonics during Leg 3 of the Deep-Sea Drilling Project (DSDP). Drilling provided for long sequences. As a result, changes in the environment of the sea could be reconstructed for the last 100 million years in some detail. Thus, the other important result of drilling was the establishment of long-term deep-ocean history focused on environmental change.

Long-term ocean history had been addressed before, of course, based on shelf sediments. It was part of general geologic knowledge that there was a great cooling of the planet in the last 40 million years culminating in the northern ice-age cycles that started not so long ago, geologically speaking. However, the history of cooling was first clearly documented by drilling into Cenozoic sediments on the deep seafloor and the cycling of ice by using oxygen isotopes (Fig. 1.5). The drilling vessels were the *Glomar Challenger* and later the *JOIDES Resolution* (Fig. 1.6). The record on land is patchy

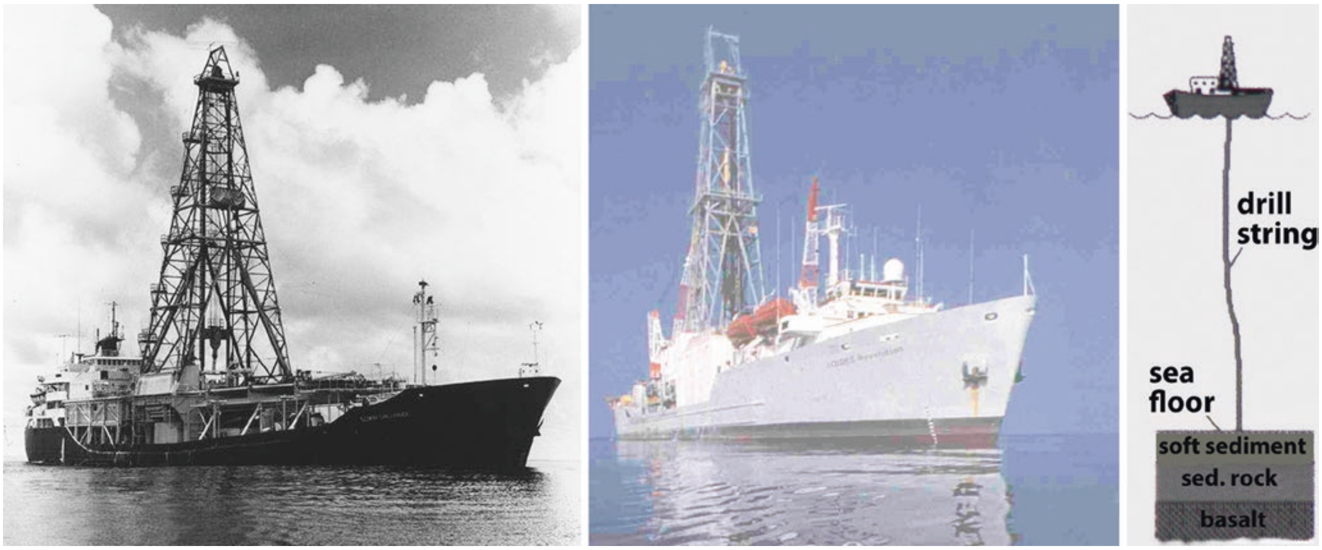

Fig. 1.6 The two drilling ships that explored the deep seafloor between 1968 and 2003. In essence, these were floating platforms for derricks such as used in drilling for oil. In each case, the drill string is lowered through a central hole in the bottom of the vessel (Photos courtesy DSDP, S.I.O., and ODP, Texas A&M; graph from M. Kastner et al., 1995, S.I.O. Ref. 95-15, simplified)

and incomplete by nature – land is eroded and delivers sediment to the sea. This renders suspect all land-based arguments about the pace of evolution, as already by Darwin pointed out. Only the record in the deep ocean can promise complete sequences, and even here gaps (called "hiatuses") proved to be quite common, frustrating much effort at reconstruction.

As a result of the drilling (and associated investigations), the stepwise cooling that characterizes the nature of sediment deposition since the Eocene (Fig. 1.7) was discovered by early DSDP participants in the 1970s. The Eocene itself, the longest of the three periods making up the early part of the Tertiary (called the "Paleogene"), is seen to be dominated by a general cooling. The step toward the end of the Eocene may be considered the culmination of the preceding trend. It is widely assumed that this step and the various others (marked as vertical-shaded rectangles in the graph) reflect positive feedback from albedo changes on ice and snow in high latitudes, including sea ice. Of course, such feedback becomes very strong rather suddenly, at critical temperature levels. The steps quite possibly indicate a sudden drop in carbon dioxide in the air as well. A trend toward lower CO_2 values in the Cenozoic is widely assumed.

The general trend of high-latitude and deep-water cooling has been ascribed to the rise of mountains (that is, to endogenic or mantle-based forcing involving plate tectonics). Mountain building changed the wind field (and thereby heat distribution) and the albedo of the planet. Also, it likely increased weathering rates and hence decreased carbon dioxide concentrations in the air. The same agents based on mountain building, by extension of the argument, were presumably responsible for starting the northern ice ages. Ways

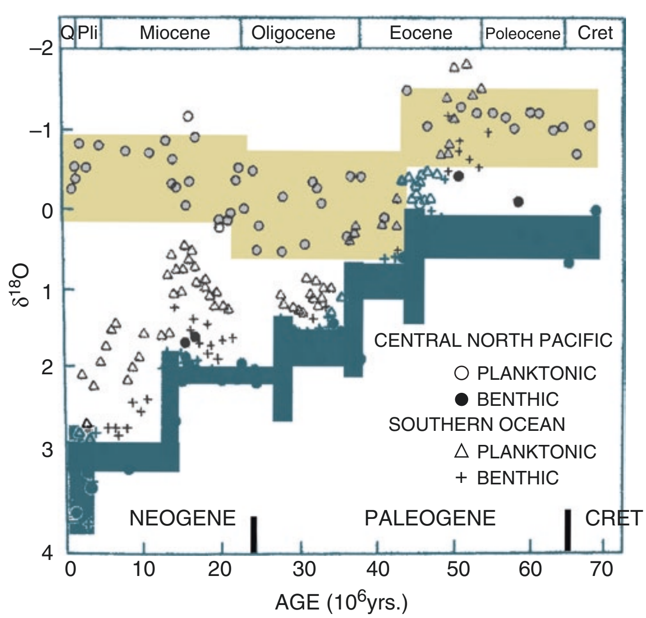

Fig. 1.7 Evidence for stepwise cooling in the Cenozoic (the proxy for temperature – and ice on land – is the δ-value for oxygen isotopes plotted as y-value). Note that the general cooling trend is most obvious in the high-latitude and in the benthic data, not in the low-latitude pelagic ones (Graph after H.R. Thierstein and W.H.B. 1978, Nature 276: 461, modified; one step added in the late Eocene; data from R. Douglas and S. Savin (1975; DSDP Leg 32) and from N. Shackleton and J. Kennett (1975; DSDP Leg 29), as well as from S. Savin (1977, Rev. Planet. Sci. 5, 319–355)). A reliable scale is necessary to put rates of evolution and other items of geologic history into perspective (Fig. 5)

to decrease atmospheric carbon dioxide and to increase albedo from causes other than uplift also are being discussed.

1.1.5 The Gubbio Event

The discovery of relatively high iridium concentrations within a layer of indurated clay initiated another major geologic revolution. The clay layer separates Cretaceous rocks from the following (somewhat similar) Cenozoic limestone layers in an exposed marine sediment section in Italy. Chemical analysis and interpretation were by Luis and Walter Alvarez and associates. Walter Alvarez sampled this section (near Gubbio) because it was well documented for foraminifer content by the paleontologists Hans Luterbacher (German-Swiss) and Isabella Premoli Silva (Italian). They had found (as had others earlier and elsewhere) that tropical planktonic forms (i.e., creatures of a warm and sunlit zone in the sea) became extinct at the end of the Cretaceous, followed by new forms, small, inconspicuous, and presumably initially largely nontropical.

The high iridium concentration at the level of the striking change in planktonic foraminifers is thought to have been caused by a "bolide" coming in from space, a large rock, perhaps around 10 km in diameter. Its impact on Earth (a huge crater with the right age was later found in Mexico) presumably created the havoc that produced the extinction of many creatures and thus ended the Cretaceous (and thereby the Mesozoic, the so-called Age of Reptiles). An even larger crater of the right age is said to have been identified near Mumbai, India. This Mumbai crater, suggested to have been made by a bolide 60 times bigger than the Mexican one, may be contemporaneous with the Deccan lava flows in India and thus presumably involved in releasing volcanic gases.

Victims of the end-of-Cretaceous mass extinction included the dinosaurs and in the marine realm the ammonites and the dolphin-like reptiles called ichthyosaurs. Other well-known victims included the reef-forming bivalves called "rudists" and a host of large reptiles such as the fearsome mosasaurs (lizards longer than large crocodiles but with equally awful teeth and with a snakelike expandable lower jaw). Mammals presumably benefited from the demise of the giant reptiles, which had dominated Earth for millions of years. In any case, mammals soon took over in the vertebrate world.

It has been known for decades that minute missives from space are hitting Earth all the time. Amounts arriving are measured in tens of thousands of tons. But only rarely is one of those missives several km in diameter, in which case it is called "bolide," and produces a crater on the surface of the planet. That impacts by bolides from outer space are not all that rare in geologic history was established by the NASA geologist Eugene Shoemaker (1928–1997), who had studied the great meteor crater in Arizona for his Ph.D. thesis. He and his associate (the chemist Edward Chao of Caltech in Pasadena, California) recognized the Miocene Ries crater in Bavaria as an impact feature, in 1960, providing an unusually large and impressive case study. Details of the story can be studied in the local museum at Shoemaker-Platz 1 in the City of Nördlingen in Bavaria. The crater is more than 20 km in diameter, that is, it is considerably larger than the meteor crater in Arizona. The impact is of middle Miocene age. It is said by many geologists to have had very little or hardly any effect on Earth history. Whatever the truth is, it is an interesting and instructive feature and well known to most geologists for the impact rock "suevite," the name a play on the local (German) dialect.

1.2 Great Revolutions in Geobiology and Geochemistry

1.2.1 Hot Vents, Cold Seeps, and Prokaryotic Microbes

The four major geologic revolutions identified in the previous sections concern mainly geophysics, although, of course, there are effects on organisms, as well. In addition to the geophysical phenomena, there are several very important revolutions related mainly to the biology and chemistry of Earth and its ocean. They include the discovery of hot vents and the nature of their fauna on ridge crests (Figs. 1.8 and 1.9), along with chemical reactions between hot basalt and seawater. Pioneers were the geoscientists Jack Corliss, Richard von Herzen, Clive Lister, Peter Lonsdale, John Edmond, and others. Marine biologists included Holger

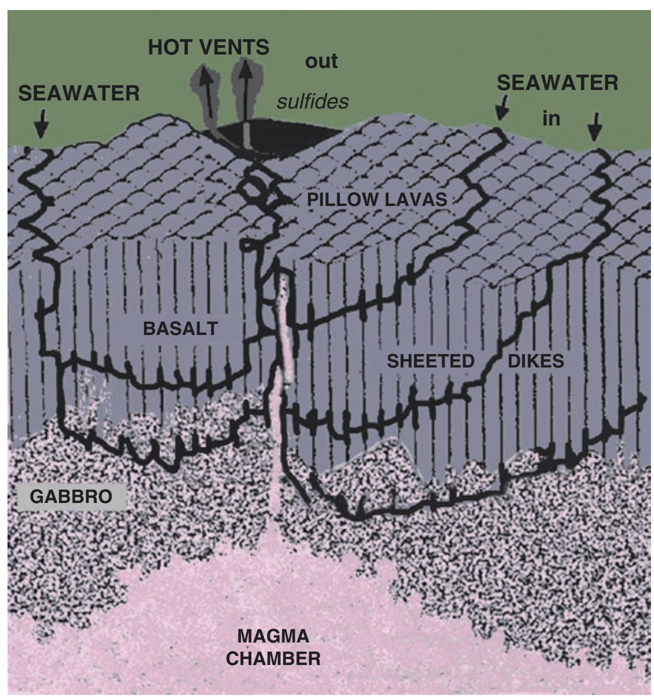

Fig. 1.8 Hot vents accompany the flow of hot seawater through freshly formed basalt in settings involving seafloor spreading. "Pillow lava": basalt with very small crystals and with pillow-shape structures; "gabbro" – basalt with large crystals, commonly "olivine" (*greenish*) and "pyroxene" (*dark gray*). Seawater percolating through cracks acquires heat from the basalt and changes its chemistry, thanks to reacting with the hot rock (Graph after schematic drawings by Enrico Bonatti and Peter Herzig)

Fig. 1.9 Tube worms at a hot vent in the Pacific, as seen from the Woods Hole scientific submarine ALVIN (Photo courtesy of Robert Hessler, S.I.O., here modified)

Jannasch, Robert Hessler, Cindy Lee van Dover, and many others, commonly using *ALVIN*, the scientific submarine of Woods Hole Oceanographic Institute. The discovery of cold seeps in certain continental slope areas also belongs into the category of fluids from the seafloor changing seawater chemistry (see next section). Pioneers were Erwin Suess, Gerhard Bohrmann, Miriam Kastner, Lisa Levin, and many others. A third geobiological revolution also concerns microbes, specifically bacteria and archaea. The discovery of such prokaryotic microbes, by drilling, deep within continental margins in marine sediments millions of years old, represents a major puzzle.

The vent setting (Fig. 1.8) in many places harbors the famous tube worms (Fig. 1.9) that rely on the oxidation of sulfide for sustenance, via the activity of microbes that live within the worms in symbiosis. Outstanding pioneers include Jack Corliss (Ph.D. 1967, La Jolla), who first saw the worms, and Scripps geophysicist Fred Spiess (1919–2006), who provided means to locate the first vent discovered. Woods Hole pioneer biologist Holger Jannasch (1927–1998) is among a long list of users of the Woods Hole scientific submarine *ALVIN*, platform of observations and for obtaining biological samples.

1.2.2 Cold Seeps

In addition to the "hot vents" and their unique fauna, we have "cold seeps" along many of the active margins, also with organisms relying for sustenance on microbe symbiosis (studied by Erwin Suess, Gerhard Bohrmann, Miriam Kastner, Lisa Levin, and many others). The cold seeps are commonly associated with methane-rich ice ("burning ice"; Fig. 1.10). The methane ice ("hydrate" or "clathrate") is being intensely studied for the associated fauna but also for possible retrieval as an energy source and as a potential hazard. Methane clathrates generate concern because of the potential for methane release during global warming. Methane is a powerful greenhouse gas. Also its release can facilitate large submarine landslides and thus generate large tsunamis.

Fig. 1.10 "Burning ice," that is, ice rich in methane ("clathrate"). It is commonly found in regions where interstitial waters are expelled from the seafloor, in continental slope areas bearing "cold seeps" (Photo courtesy G. Bohrmann, now MARUM Bremen, and E. Suess, GEOMAR Kiel)

1.2.3 Bacteria and Archaea

Ever since the end of the ninteenth century, evidence has been accumulating that prokaryotic microbes are intimately involved in the geochemistry of the planet (e.g., see the documentation of microbe activity in soils by S. Winogradsky in 1949). Recently, vast and impressive evidence for living microbes at great depths below the seafloor has emerged from drilling in continental slope sediments.

What, if anything, do these various geobiological revolutions have in common? They are intimately associated with the discovery of strange bacteria-like organisms, the archaea. These microbes define a new biological kingdom, proposed by the biologist Carl Woese (1928–2012) of the University of Illinois and his associates. Their proposal results in three domains of life, rather than just the traditional two (the "prokaryotes" lacking a well-defined nucleus and the "eukaryotes" which possess one). The traditional prokaryotes contain the familiar bacteria, while the eukaryotes cover everything else, including foraminifers, people, and plants. Presumably, the bacteria and archaea found by drilling well below the seafloor (Fig. 1.11) have a strangeness of their own – they must be able to survive for enormously long time spans. Their mass is said to be a substantial fraction of the mass of living things on the planet. The Ukrainian-French

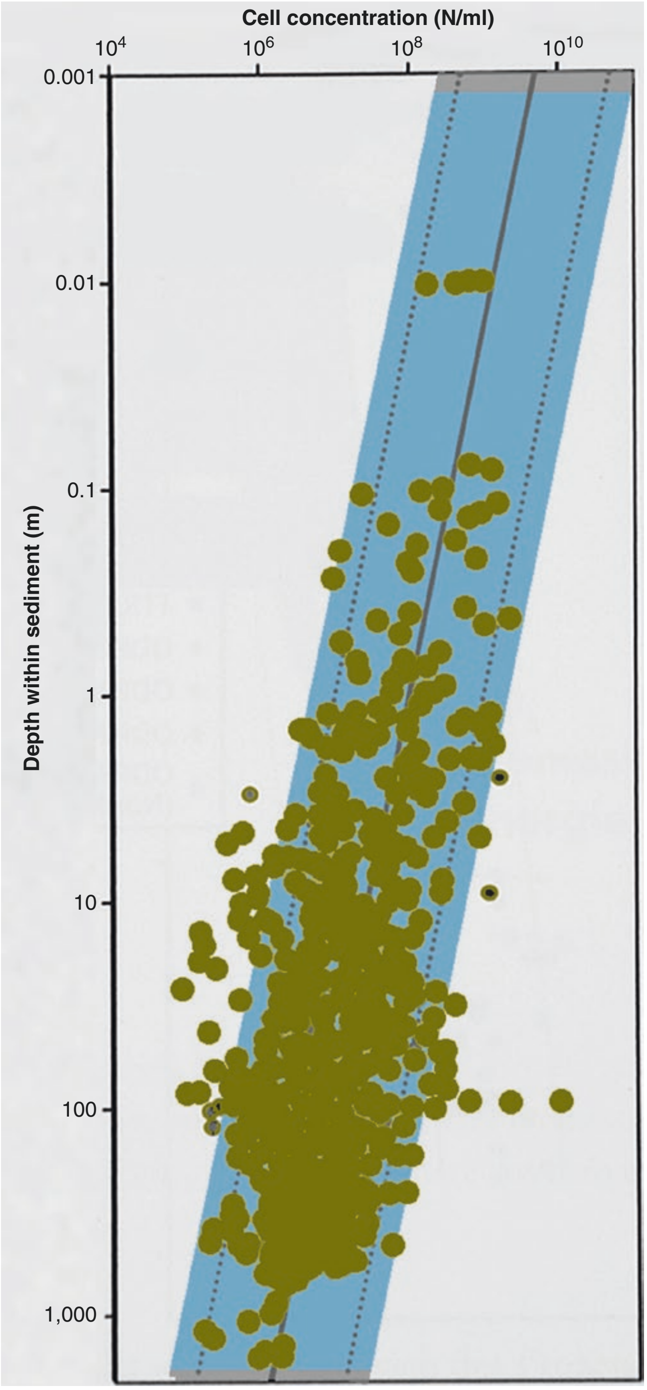

Fig. 1.11 Abundance of microbes within marine sediments (by volume of sediment) (After Hinrichs et al., 2010. In: Expedition Erde (3rd ed.), MARUM Bremen, ed. by G. Wefer and F. Schmieder, p. 144. Data (here simplified; leaving off area of origin): R.J. Parkes et al. (2000), S. d'Hondt et al. (2004), B. Engelen et al. (2008), E.G. Roussel et al. (2008), G. Webster et al. (2009))

microbiologist pioneer Sergei Winogradsky (1856–1953) had been right in surmising that microbial life is pervasive in its effects on the planet, being intimately involved in the cycling of carbon, of sulfur, of iron, and of nitrogen, among other things.

1.3 More on Pioneers Who Identified Basic Questions in Marine Geology

1.3.1 Johannes Walther and Persons in P.D. Trask's Book

Marine geology, as a discipline in geology, started well before the various revolutions listed, of course. It began with geologists noting the processes that helped produce the marine rocks which they saw on land. The German geologist Johannes Walther (1860–1937), known for "Walther's Law," was a pioneer in these types of studies. ("Walther's Law" asserts that vertical sequences of marine sediments reflect originally neighboring types of deposits.)

In his book, *Bionomy of the Sea* (Jena 1893, in German), Walther gave an account of marine environments and the associated ecology of shell-bearing organisms (and the pertinent literature of the time). Research on marine sedimentation in the four decades that followed is summarized in the volume *Recent Marine Sediments* edited by the USGS geologist P.D. Trask (Tulsa 1939). The symposium spans the range of sedimentary environments from beach to deep sea, and many of the articles were written by famous pioneers in marine geology of the time, geologists from various countries.

1.3.2 John Murray and the *Challenger* Expedition

As marine geologic studies progressed and moved farther out to sea, there was a gradual change of emphasis in the set of problems to be solved. The seafloor itself became the focus of attention, not for the sake of the clues it could yield for the purposes of geology on land but for the record it has for its own evolution and its role in the present processes relevant to the Earth's environment. The new emphasis is first obvious in the works of the Scotsman John Murray, naturalist of *HMS Challenger* on her world-encircling expedition (1872–1876). John Murray's chief opus, *Deep Sea Deposits* (written with the Belgian geologist A.F. Renard and published in 1891) laid the foundation for the sedimentology of the deep ocean floor, that is, of the sediments that cover most of the planet.

1.3.3 Philip H. Kuenen, B.C. Heezen, M. Natland, and Turbidites

Some of the great volumes of deep-sea deposits are nothing but shallow-water sediments that moved down to rest on the deep seafloor. This was proposed (in 1950) by the Dutch geologist Ph.D. Kuenen (1902–1976), who demonstrated that sediment clouds could rush downslope from shallow areas to great depth, due to the fact that muddy water is heavier than the clear water surrounding it. The Canadian-born geologist R. A. Daly (1871–1957) (Prof. at Harvard), the Lamont geologist D.B. Ericson (curator of the core collection), and the Californians Manley Natland (1906–1991), E.L. Winterer (Ph.D. 1954, U.C. Los Angeles), and Chicago-trained F.P. Shepard (1897–1985) came to similar conclusions at the time. The sediment thus transported settles out of suspension at the site of deposition, large grains first, fine grains last. The resulting layer (called a "*turbidite*") is *graded*. Graded layers are common in the geologic record (e.g., in the Alpine *flysch* deposits and in the Neogene sequences of sedimentary rocks in the Californian Transverse Ranges) suggesting that Kuenen's concepts are indeed important for all of geology. The Lamont geologist B.C. Heezen (1924–1977), together with Lamont's founding director M. Ewing (1906–1974), offered evidence (in 1952) that turbidite-producing flows race down the slope off the US East Coast occasionally. Similar events are now postulated for the origin of much of continental slope sediments everywhere, and to explain the origin of sea valleys within them, and for the origin of the flatness of *abyssal plains*, which are among the most extensive features on the planet.

1.3.4 *Meteor* and *Albatross* Expeditions and *Vema* Collections

Studies in marine sedimentation led into ocean history when geologists started to take cores. The pioneering expeditions were those of the German vessel *Meteor* (1925–1927), whose participant W. Schott (1905–1989) first proposed rates of deep-sea sedimentation at around 1–2 cm per thousand years, and the one of the Swedish vessel *Albatross* (1947–1948), led by the Swedish physicist and oceanographer Hans Pettersson (1888–1966). The several meters long Albatross cores established the presence, in all oceans, of cyclic sedimentation, which documented evidence for climate fluctuations during the last million years, including several ice ages (see Chap. 11). Enormous collection efforts followed, led by Lamont's research vessel *Vema (*also *R.V. Conrad).* Several Lamont scientists would have to be mentioned as pioneers, foremost the geophysicist M. Ewing (the founding director) and the geologist Bruce Heezen, already mentioned in connection with the turbidite discovery but also prominent in numerous others.

1.3.5 Deep-Ocean Drilling: *GLOMAR Challenger* and *JOIDES Resolution*

The last (and biggest) effort in the line of deep-sea expeditions elucidating matters of marine geology is the various programs for deep-ocean drilling, including the Deep-Sea Drilling Project (head office at S.I.O.), the Ocean Drilling Program (head office at Texas A&M), and the Integrated Ocean Drilling Program (core repositories in Texas [US Gulf Coast], in Bremen [Germany], and in Kochi [Japan]). Countless samples became available for the study of ocean history, thanks to the drilling expeditions. Together, the various drilling programs recovered hundreds of kilometers of core sections from the deep sea. The main results from drilling in the seafloor concern stepwise cooling within the Cenozoic (Chap. 12) and the shortage of oxygen in the Cretaceous (Chap. 13). Important pioneers are J.P. Kennett, N.J. Shackleton, S. Savin, and R. Douglas, as mentioned above in Sect. 1.1.1.

1.3.6 Morphology (Landscapes)

While the drilling ventures attracted a large amount of attention in the second half of the last century, traditional investigation of the seafloor proceeded unabated. Coastal landforms are the most accessible of the marine environments, and considerable information concerning them had been accumulated early in the twentieth century. Much additional work on these topics was subsequently done by the Scripps geologist Francis P. Shepard (1897–1985), who took his observational skills to sea. In France, it was the geologist Jacques Bourcart (1891–1965) who was able to test theoretical concepts against field data collected in shallow waters. Shepard's textbook *Submarine Geology* (first published in 1948) summarized the results of this and other early works. In the same year that Shepard's first marine geology text was published (1948), M.B. Klenova's textbook *Geology of the Sea* was printed, with ample coverage of Russian work. Among several outstanding Russian pioneers, there were many who made crucially important contributions to the understanding of deposition on the seafloor. For example, N.I. Andrusov (1861–1924) studied deposits in the Black Sea, and N.M. Strakhov (1900–1978) explored the origin of marine rocks in general.

The marine geomorphologist *par excellence* was B.C. Heezen (Fig. 1.12, inset). His physiographic diagram of the seafloor (produced together with his collaborator Marie Tharp, 1920–2006) now is in many or most textbooks. It was republished by the US Navy as a memorial to Heezen. Much of Heezen's work is summarized in the beautifully illustrated book *The Face of the Deep* (New York, 1971), written with C.D. Hollister (1936–1999).

1.3.7 Alfred Wegener (1880–1930)

It is perhaps no coincidence that a geophysicist first formulated a global hypothesis for the overall geography of the

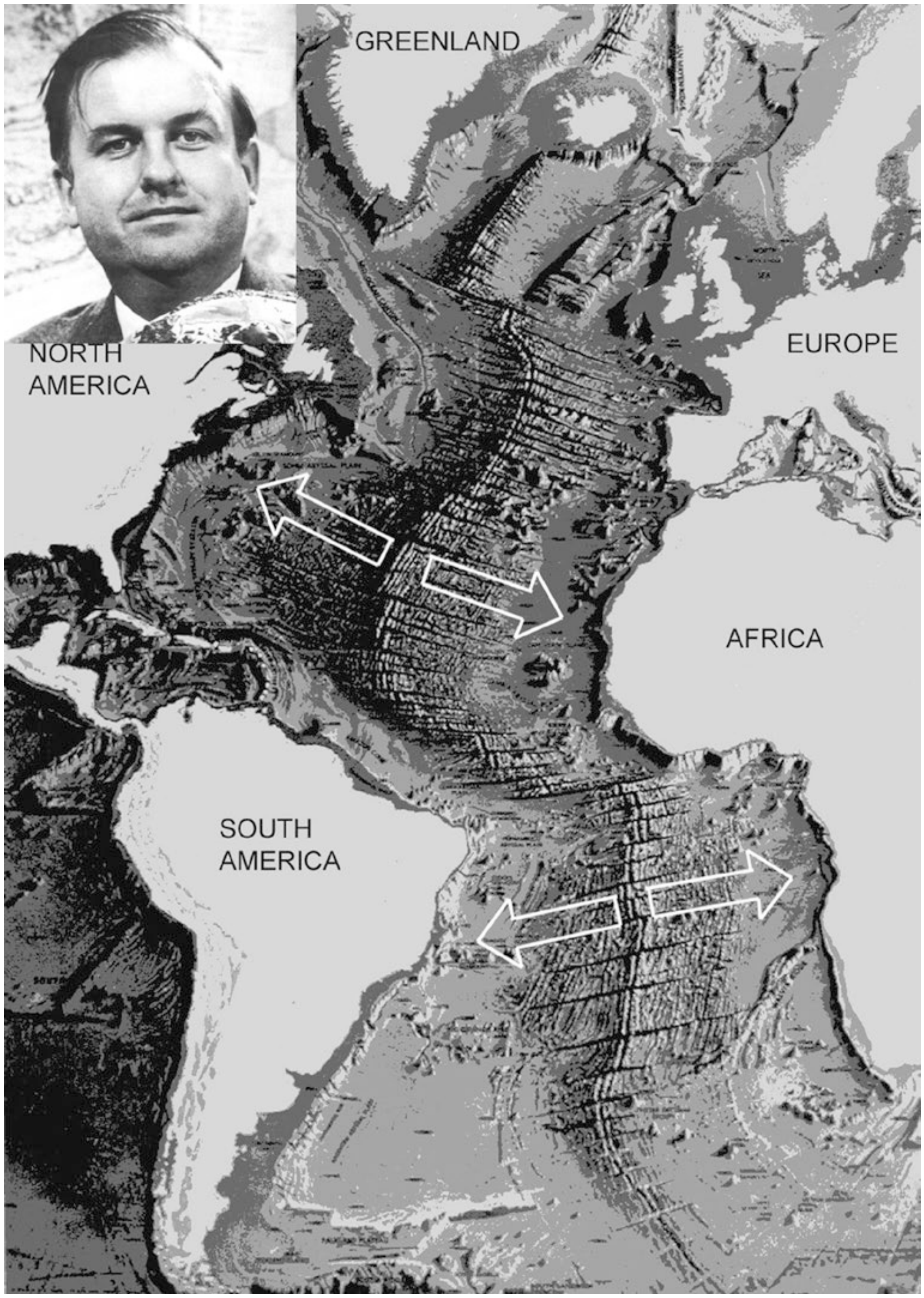

Fig. 1.12 Physiographic diagram of the Atlantic Ocean floor according to Bruce Heezen and Marie Tharp, Lamont, Columbia University, as repainted by H. Berann (National Geographic Society). Rendered here with omissions and additions with a portion of the memorial map of the US Navy. Insert portrait of Bruce Heezen: courtesy of the Lamont–Doherty Earth Observatory

planet that proved to be viable: the proposition regarding continental drift, by the meteorologist Alfred Wegener (1880–1930). Wegener was intrigued with the fact that the coast lines bordering the Atlantic Ocean are parallel. (The observation was widely known. It had been made decades earlier by the naturalist and explorer Alexander von Humboldt, 1769–1859.) Wegener rejected the land bridges proposed by leading paleontologists of his time to explain striking similarities of Paleozoic land fossils on both sides of the Atlantic. Instead, he postulated that the continents were once united. With this hypothesis, he started the "debate of the century" in geology. He proposed his hypothesis in an article in 1912 (in the journal *Geologische Rundschau*) and again in his book *The Origin of Continents and Oceans* (Braunschweig, 1915, in German; later translated into English). He envisioned granitic continents floating in basaltic mantle magma like icebergs in water (Fig. 1.13) and drifting about on the surface of the Earth in response to unknown forces linked to the rotation of the planet.

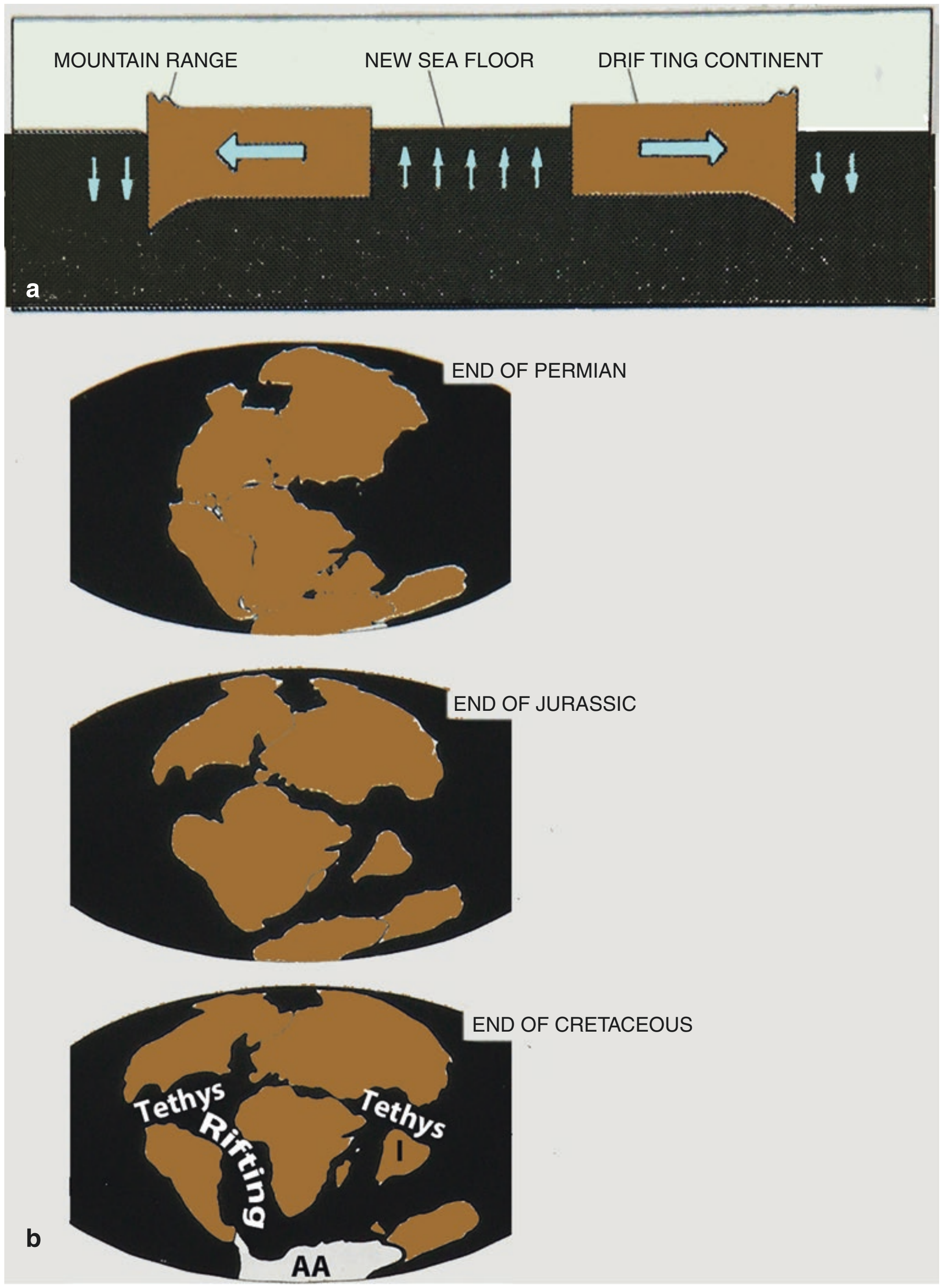

Fig. 1.13 Diagrammatic representation of the hypothesis of continental drift of Alfred Wegener. (**a**) Continental blocks made of rocks rich in silicon and aluminum float iceberg-like in the heavier mantle material rich in silicon and magnesium. As the continental blocks drift apart, mountain ranges form at the bows of the drifting blocks (the "active" margins), while new seafloor forms between the slowly sinking "passive" margins. (**b**) The breakup of the ancient continent of "Pangaea," as envisaged by A. Wegener, in a modern reconstruction by R.S. Dietz (US Navy) and J.C. Holden (1970; J. Geophys. Res. 75: 4939, simplified and with addition of labels). *AA* Antarctica, *I* Indian subcontinent, moving northward toward the ancient seaway called "*Tethys*" (erstwhile, part of a tropical connection between a growing Atlantic and the Pacific, now closed; color here added)

When mentioning Alfred Wegener in connection with plate tectonics, one might assume that "continental drift," the main brain child of Wegener, initiated the thinking that led to the final concept. This was actually not the case: Wegener's ideas were long rejected. It was the magnetism frozen into seafloor basalt solidifying from mantle melt during seafloor spreading that clinched the new concept, a concept that has many parents (see Chap. 2).

1.3.8 E.C. Bullard (1907–1980) and M. Ewing (1906–1974)

Wegener's hypothesis, in modified form, is now an integral part of *plate tectonics*, the ruling theory explaining the geomorphology and geophysics of the ocean floor. It was the work of sea-going geophysicists which eventually led to the acceptance of seafloor spreading and plate tectonics.

The pioneering efforts of the British geophysicist E.C. Bullard (1907–1980) on earth magnetism, heat flow, and seismic surveying must be mentioned, as well as those of the US geophysicist M. Ewing at Columbia University. Although M. Ewing may not have believed that plate tectonics describes reality correctly, he and his associates were of central importance in the development of the new paradigm. Both scientists caused enormous amounts of relevant observations to be made. Of course, there are many others who contributed as well (see Chap. 2). Ewing's and Bullard's studies were important to traditional geology in that they included the nature of continental margins, a region of prime interest in the study of marine layers on land (see the book edited by C.A. Burk and C.L. Drake).

1.3.9 H.H. Hess (1906–1969) and R.S. Dietz (1914–1995)

The turning point in the scientific revolution that shook the Earth sciences and which was culminated in *Plate Tectonics* is generally taken to have been the seminal paper by the former Navy officer and Princeton geologist Harry H. Hess (1906–1969), entitled *History of Ocean Basins* and published in 1962. It is probably fair to say that the paper was widely ignored when first published and for a few years afterward. Hess started his distinguished career working with the Dutch geophysicist F.A. Vening Meinesz (1887–1966) on the gravity anomalies of deep-sea trenches. The investigations resulted in the hypothesis that trenches may be surface expressions of the down-going limbs of mantle convection cells (which is similar to today's explanation). As a naval officer, Hess discovered and mapped a great number of flat-topped seamounts (which he called "guyots") during WWII, in the central Pacific.

Stimulated by the discoveries concerning the mid-ocean ridge (extent, morphology, heat flow, and other items), Hess proposed that seafloor is generated at the center of the mid-ocean ridge, moves away from the ridge and downward as it ages, and finally disappears at trenches. The term *seafloor spreading* was coined by the US Navy geologist R.S. Dietz (1914–1995) in San Diego and published in 1961. B.C. Heezen published on the nature of the Mid-Atlantic Ridge (that is, on rifting) in 1960, leaning on his work with the cartographer Marie Tharp (1920–2012). Years later some discussion ensued about priorities, documenting the importance of the new ideas (but not necessarily their first appearance). The fact that the Austrian geologist Otto Ampferer (1875–1947) and the British geologist Arthur Holmes (1890–1965) published ideas reminiscent of plate tectonic motions decades before others did may render the gist of such discussions moot.

1.4 Seafloor Spreading and Plate Tectonics: The New Paradigm

1.4.1 New Paradigm Accepted, Wegener's Hypothesis Triumphant

The new hypothesis of seafloor origin and mobility was generally accepted around 1970. Deep-sea drilling (Leg 3) probably had a role in it, by showing that micropaleontology agreed with geophysical assessment. This was an important demonstration for many geologists. Of course, many leading *geophysicists* (physicists doing geology) had already signed on to the new paradigm, relying on evidence from paleomagnetics. The plate tectonics theory recognizably starts with thoughts of A. Wegener, who argued from geography, paleontology, and crustal geophysics (Fig. 1.13). In other words, it took more than half a century for Wegener's ideas to prevail. A reluctance to accept new ideas, among geologists, also is obvious for the related propositions of Heezen, Hess, and Dietz. It took one decade for them to succeed (with insights largely based on a plethora of the US Navy data and confirmed by seismic and paleomagnetic information). The amount of information available had grown immeasurably, removing doubt more rapidly than before.

1.4.2 Euler's Theorem

Shortly after Hess published his provocative article on postulated seafloor spreading, E.C. Bullard and associates introduced "Euler's Theorem" to global tectonics (Fig. 1.14). The theorem, generated by the Swiss mathematician Leonhard Euler (1707–1783), states that uniform motion on a sphere is uniquely defined by the rotation about an appropriate point, called a "pole." The path of migration of any point on a plate in uniform motion appears as a portion of a circle about that pole. Bullard and collaborators used Euler's Theorem, in 1965, to produce a new "fit" of the continents bordering the Atlantic. Shortly after, the geophysicist and geologist James W. Morgan of Princeton University realized that fracture zones in the eastern North Pacific run parallel to small circles of Euler poles, thus cementing the connection between plate motions and Euler's Theorem.

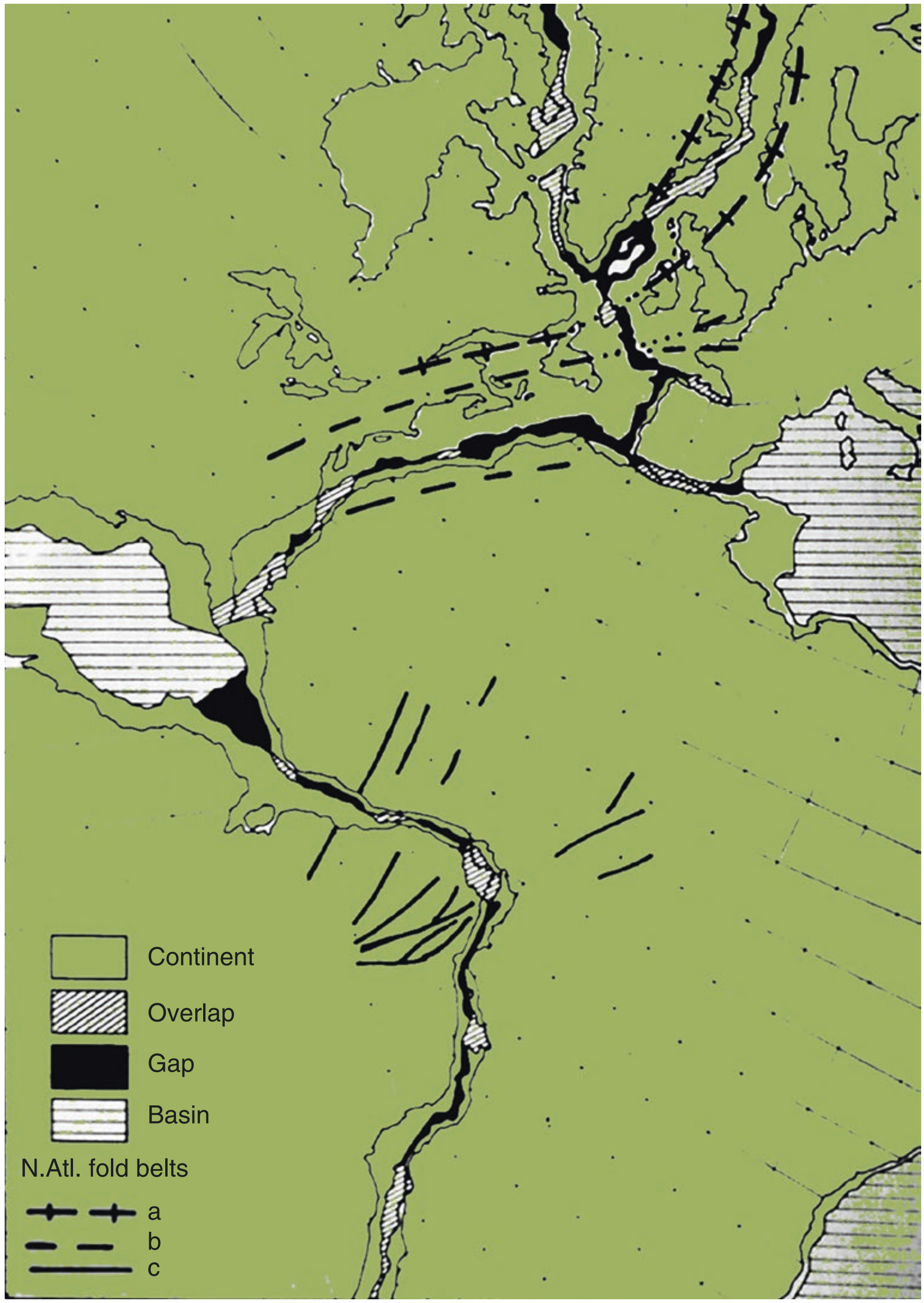

Fig. 1.14 The fit of the continents bordering the Atlantic, as proposed by E.C. Bullard et al., (in Blackett et al., eds., 1965. A Symposium on Continental Drift. Phil. Trans., Royal Soc. London, 258:41). Note the Niger Delta overlap, which is expected since the delta is geologically young, and the Bahamas overlap, which is unexplained. The alignment of the Paleozoic fold belts in North America and Europe and in South America and Africa is striking. (**a**) Caledonian fold belt (Norway-Scotland), (**b**) Hercynian fold belt, and (**c**) Pan-African fold belt (The latter is adapted from C.J. Archanjo and J.L. Bouchez, 1991. Bull. Soc. Géol. France 4: 638. Color here added)

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* A full reading of the references offered is not recommended. The prime intent is to provide a short list of tomes worth perusing for relevant information. Another is to provide access to books that preserve the historic context of the most remarkable revolutions in marine geology within the last century.

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2 Origin and Morphology of Ocean Basins

2.1 The Depth of the Sea

2.1.1 General Distribution of Elevations

The obvious question to ask about the seafloor is how deep it is and why. The overall depth distribution largely became known through the voyage of *HMS Challenger* in the nineteenth century, and it could then be combined with available information from the land surface (Fig. 2.1). We see that there are two most common elevations on the solid planet: an upper one just above sea level and a lower one centered near the average depth of the ocean. The higher of the two main levels presumably mainly represents the action of erosion to sea level (*base level*), as well as uplift of continental crust. The lower one, one assumes, reflects the volume of seawater along with the availability of space to put the water (i.e., the room that is left after making the continents). The seafloor between the two elevation modalities, that is, the part connecting shelves and deep ocean floor, is a transition consisting largely of the continental slopes and rises. A portion of this intermediate category of elevations also must be assigned to the worldwide mid-ocean ridge and its flanks. In addition, there is a special (small) portion of seafloor that is more than twice deeper than the average: such great depths only occur in narrow trenches, mainly in the subduction ring around the Pacific Ocean.

2.1.2 Notes on the Mean Continental Elevation

The substantial height of the mean continental elevation (several hundred meters; Fig. 2.1) deserves some comment. Apparently it results from uplift, since a general prevalence of erosion would presumably result in lowering the continental surface to sea level, that is, to the *base level of erosion*. For an overall imbalance of rates of uplift and erosion, the mean elevation of continents must vary on geologic time scales; that is, what is being measured geologically are transients, when considering elevations above sea level. A general cooling in the Cenozoic, strontium ratios in marine carbonates, and the onset of ice ages suggest a prominent role for uplift since the beginning of the Eocene and then all through the rest of the Cenozoic. Himalayan uplift in the Neogene supports the suggestion.

2.2 Endogenic and Exogenic Processes

2.2.1 Endogenic Forcing: Tectonics

As is true of all of the face of the Earth in general, the seafloor is shaped by two kinds of processes, those deriving their energy from inside the Earth, the *endogenic* ones (e.g., volcanism, mountain building, and uplift), and those driven by the sun, called "exogenic" (e.g., erosion). In addition, there is astronomic forcing, which is a type of exogenic forcing depending on variations of input of energy from the sun directly, à la Milankovitch.

The forces inside the Earth are responsible for volcanism and earthquakes; we meet them in the eruptions on Hawaii, in the geysirs of Yellowstone National Park, and in the quakes in California. We see their handiwork in the mountain ranges of the Alps, the Sierra Nevada, and the Himalayas and in the gigantic rift of the Rhine Graben and the even larger one of the Red Sea. We shall have to examine tectonic forcing in some detail in this chapter, since the geography of the seafloor largely depends on it.

2.2.2 Exogenic Products Conspicuous

When contemplating questions of biological production (Chaps. 7 and 8), we are obviously in a zone of overlap between ecology and geology. The overlap is prominent in marine geology notably in the portion dealing with

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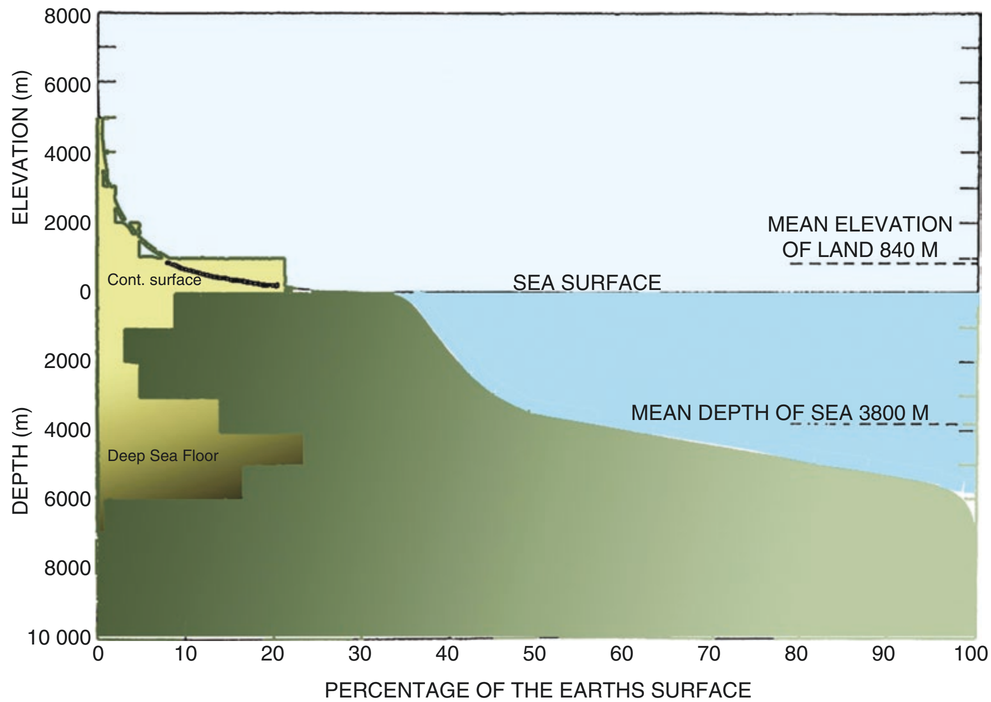

Fig. 2.1 Overall depth distribution of the ocean floor and of land elevations. Buff to *dark gray*: frequency distribution of elevations and depths on the planet. *Greenish gray*: seafloor, more greenish below coastal ocean (After H.U. Sverdrup et al. 1942, modified)

sedimentology and life on the seafloor (rather than with processes related to tectonics). Regarding non-biological sedimentation, there is the origin of turbidites. There is no question that exogenic forces are at work in generating the type of seafloor called *abyssal plain* – incredibly flat areas hundreds of miles in diameter, such as are seen off the MOR in the North Atlantic (Fig. 2.2). On land, the enormous playas that are dried-up lake bottoms in the west of the United States can convey an (somewhat diminished) idea of the extent and flatness of abyssal plains.

The abyssal plains are vast undersea playas collecting fine-grained debris from nearby continents, debris that is produced by the ever-present agents of weathering: rain, wind, and ice. The chippings made by these sculptors that wear down mountains are carried to the ocean by rivers and by winds. Here they build up sedimentary continental margins and abyssal plains beyond some of the continental slopes. The bulk of the mud and sand on the basin floors presumably is brought in by catastrophic events represented by the types of sporadic turbidity currents mentioned in the introduction when discussing Kuenen's contributions (Sect. 1.3).

Most of the sediment accumulating on the deep seafloor away from the continents, however, is not washed off from continents (i.e., it is not *terrigenous*) but arrives at the seafloor as a more or less continuous rain of particles: shells and skeletons of plankton organisms, wind-borne dust, and even cosmic spherules. Through geologic time (here: since the Jurassic), the slowly accumulating pelagic sediments built up a layer a few hundred meters thick on the seafloor of the deep sea, a layer that forms a thin exogenic veneer on the endogenic oceanic crust, which is basaltic. On land, analogous deposits of a pelagic nature are made by snow, albeit at a rate a thousand times greater than sedimentation rates typical for the deep seafloor. It is this thin veneer of pelagic sediments on the deep seafloor that contains a detailed history of the ocean for the last 100 million years or so.

Exogenic processes tend to *level* the Earth's surface by erosion and deposition. However, they also can build mountains, especially within the sea. The outstanding example is the *Great Barrier Reef* off northeast Australia. The reefs covering shelves (and ancient Cretaceous reefs on top of undersea mountains) consist of calcium carbonate secreted by stony corals, coralline algae, mollusks, and small unicellular organisms called foraminifers. The carbonate-secreting algae, as well as the minute algae symbiotic with the coral ("zooxanthellae") and with many of the foraminifers, depend on sunlight for energy, of course. Thus, typically coral reefs grow in shallow, sunlit waters. There are exceptions: some coral species (e.g., of the genus *Lophelia*) do grow at great depth in the dark (obviously they do without photosynthesizing symbionts).

With this briefest of introductions to the opposing effects of endogenic processes (which wrinkle Earth's surface) and

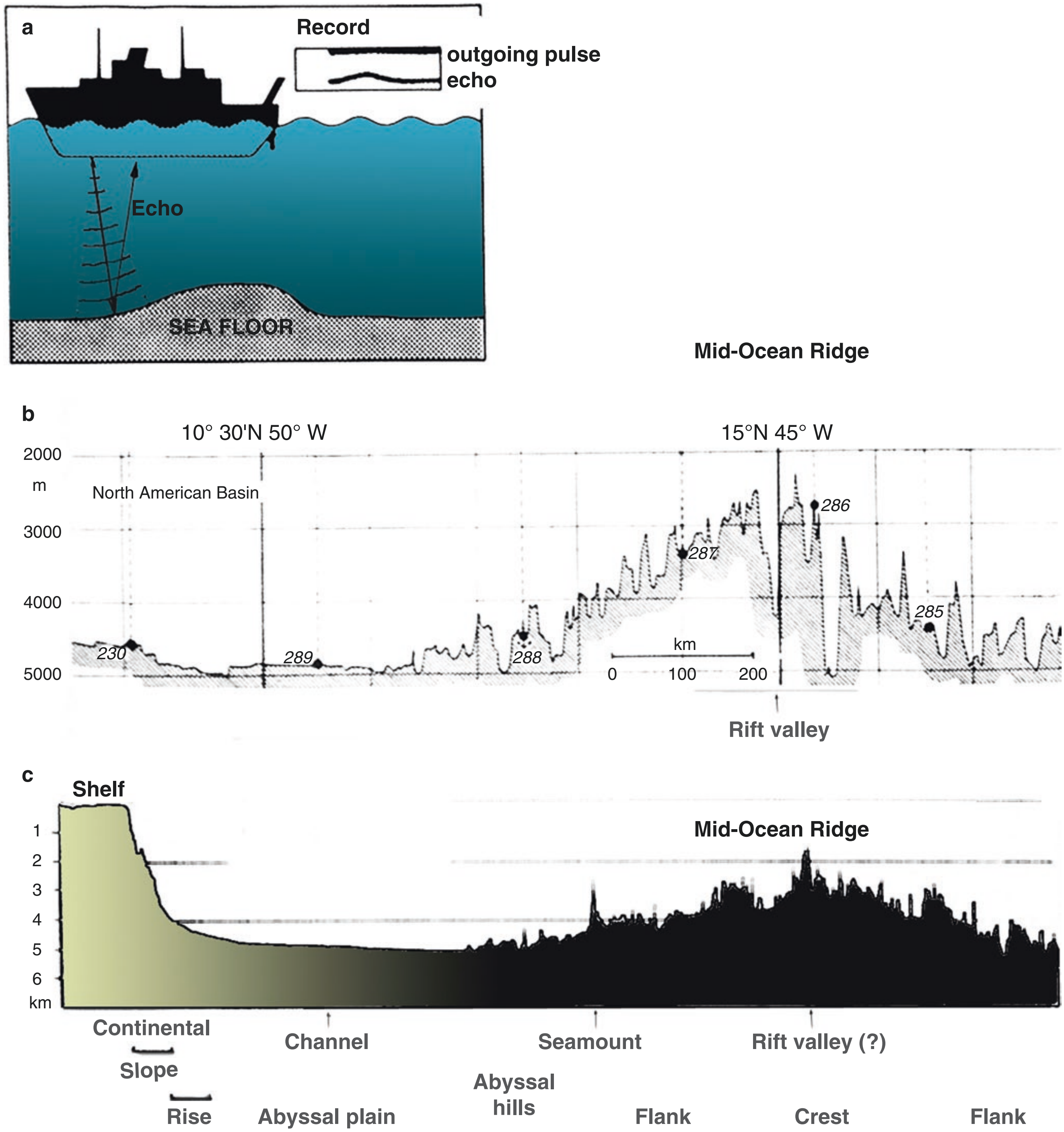

Fig. 2.2 Topography of the Mid-Atlantic Ridge as observed by echo sounding. (**a**) The depth emerges from the time it takes the echo to return and the speed of sound in seawater. The two-way travel time is recorded on the sounding vessel (inset: "Record"). (**b**) Results of echo sounding obtained by the German Meteor Expedition (1925–1927). Numbers refer to sampling stations. Note the prominent central rift. (**c**) Modern topographic profile (after B.C. Heezen et al. 1959; see Geol. Soc. Am. Sp. Paper 65). Labels for physiographic features here added. Note change of scale between (**b**) and (**c**), presumably emphasizing different latitudes

exogenic ones (which mainly smoothen it), let us now return to the nature of the grand morphology of the seafloor, that is, to the endogenic processes that express themselves in seafloor spreading and plate tectonics. Just how do the new concepts bear on the morphology of the mid-ocean ridge and that of the trenches, and how do volcanic island chains fit into the picture?

2.3 Morphology of the Mid-ocean Ridge

2.3.1 Background

On *HMS Challenger*, depth determinations were done by laboriously sending a weight to the ocean floor and measuring the length of the line paid out. When the scattered soundings were connected for drawing depth contours, the ocean floor looked smooth. Despite such difficulties, the fact that a central barrier separates western from eastern Atlantic basins did emerge during the expedition thanks to differences in the temperature of abyssal waters in the eastern and the western trough of the Atlantic. However, only when echo sounding was used routinely (in the twentieth century) did it become obvious that large parts of the ocean floor consist of immense mountain ranges whose cragginess rivals that of the Alps and the Sierra Nevada. Perhaps the most impressive of these ranges is the Mid-Atlantic Ridge, discovered by the famous *Meteor* Expedition (1925–1927) (Fig. 2.2) and subsequently recognized as a portion of the globe-encircling MOR (mid-ocean ridge) by the Lamont geologist Bruce Heezen and his associate Marie Tharp (Fig. 2.3).

Heezen's realization that the Mid-Atlantic Ridge is part of a worldwide phenomenon (the mid-ocean ridge) was a crucial piece of insight. The MOR is continuous and endless, much like the seam on a baseball. Thus, its existence shows a process of planetary dimensions at work.

2.3.2 The MOR a Product of Seafloor Spreading

The mid-ocean ridge is produced by spreading of the seafloor (Fig. 2.4). It is more than 60,000 km long and takes up roughly one third of the ocean floor (the precise value depends on where one puts the outer boundaries of the flanks of the ridge). One third of the seafloor translates a little less than one fourth of the Earth's surface. In the Atlantic and along certain other portions of the ridge, the crest is marked by a central rift, that is, a 30- to 50-km-wide steep-walled valley with a depth of 1 km or more.

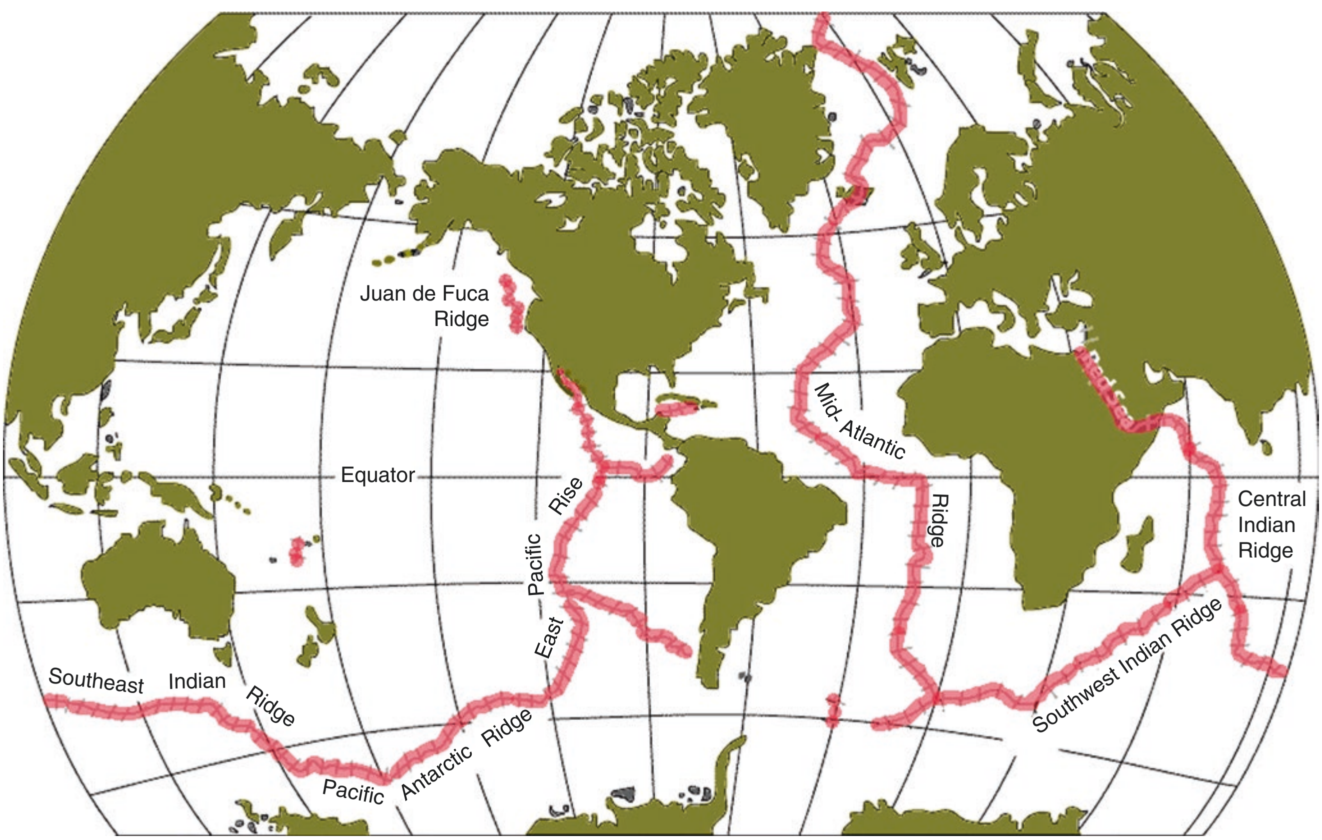

Fig. 2.3 The worldwide mid-ocean ridge, of which the Mid-Atlantic Ridge is a part (From the US Geological Survey)

2.3.3 Open Questions

What proof do we have that the theory of seafloor spreading is sound? The question was raised well into the 1970s. It is no longer asked – seafloor spreading no longer is a crazy idea possibly worth looking into. Instead, it is a fully accepted concept and has become part of textbook science, even at the grade school level. It is now fundamental, much like the insight that a fossil once was part of a living organism, that Earth has an age of 4.6 billion years, that large continental glaciers once covered vast areas of North America and northern Europe, or that biological classification reflects ancestry and evolution. None of these insights got into the textbooks immediately after they arose, being vehemently opposed by "experts" at the time they originated.

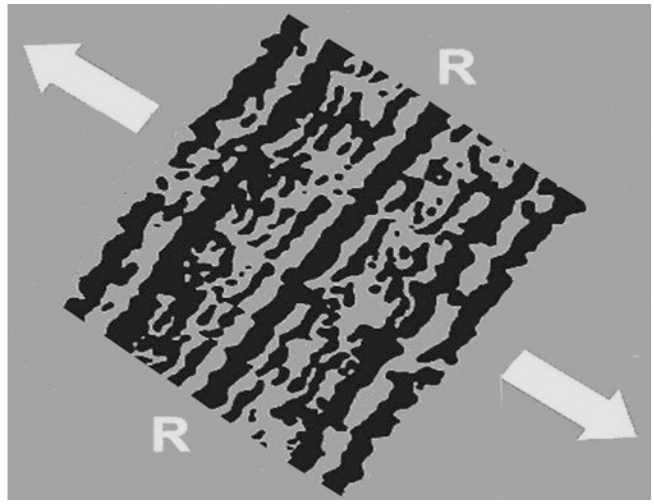

Fig. 2.4 Magnetic anomalies on Reykjanes Ridge according to J.R. Heirtzler et al. [see 1966 Deep-Sea Res. 13: 427] and interpretation in terms of seafloor spreading. The letter **R** (for "ridge", here added, along with arrows) denotes the center of the MOR, which also is a center of remarkable symmetry of the magnetic lineations or "stripes." An earlier (unpublished) report on the 1963 survey became available in 1965. The survey was cosponsored by the US Navy and Lamont Geol. Observatory

2.3.4 The Magnetic Stripes

Seafloor spreading is now treated as a fact. What makes us confident that this is justified?

Our confidence derives from the patterns of magnetic anomalies on the seafloor, notably patterns across spreading centers. As every compass-bearing school kid knows, the planet has a magnetic field, which is linked to the rotation around the Earth's axis (hence, we have "magnetic north" and "magnetic south"). Sporadically, on time scales typically tens or hundreds of thousands of years long (and for reasons unknown), the Earth's magnetic field reverses, so that magnetic north becomes magnetic south and vice versa. The reversals are recorded within the freezing basaltic rock of the new seafloor generated at spreading centers.

When mapping seafloor magnetism, one finds anomalies of magnetic intensity over the seafloor resulting from the combination of the current magnetic field with the one frozen into the basalt when it solidified (the *magnetic stripes*). The stripes were first discovered in the late 1950s. However, their origin remained a complete mystery for several years. One of the problems was that the area where magnetic anomalies were first mapped is geologically complicated, and the symmetry of the patterns about the center of the MOR, which holds the key to the explanation, is not obvious there. It is rather obvious elsewhere, fortunately, such as on the Reykjanes Ridge south and west of Iceland, where the expected pattern was impressively documented by J.R. Heirtzler (Ph.D., New York, 1953) and colleagues, in 1966 (see Fig. 2.4). Much of the original magnetic reversal stratigraphy (including dating) is the work of pioneers A.V. Cox (1926–1987) and his collaborators at the USGS, R.R. Doell (1923–2008) and G.B. Dalrymple (PhD 1963, UC Berkeley).

That there should be these types of magnetic anomalies at a spreading center was proposed by F.J. Vine (then a graduate student at Cambridge University) and D.H. Matthews (his advisor), in 1963. Their suggestion was strikingly simple: put together the ideas on seafloor spreading by R.S. Dietz and H.H. Hess, and combine them with the evidence for periodic reversals in the Earth's magnetic field, as then already in the literature, put there by the USGS geophysicists A. Cox and R.R. Doell, a team later including G.B. Dalrymple. The result is the generation of "blocks of alternately normal and reversely magnetized material 'drifting' away from the center of the ridge and parallel to the crest of it," exactly as documented in subsequent publications.

2.3.5 Fracture Zones and Magma Chambers

The MOR is segmented by fracture zones, as already recognized in the 1960s (Sect. 2.4.3). Thus, the MOR occurs more or less in straight segments that are offset from each other. The consequence of such offset is that a lateral fault must form at the ends of the straight crestal portions. As there is motion along this fault during active seafloor spreading, there are earthquakes on it. These quakes are shallow and define the *active* part of the *fracture zone*, that is, the *ridge-ridge transform fault*. Beyond this active portion, the fracture zone is the frozen trace of the inactive portion of the fault zone. Along this trace, any scarp becomes smaller as the seafloor ages and subsides on both sides of the zone but with the younger and warmer side subsiding faster. Fracture zones tend to have an irregular topography, leaks of magma produce seamounts, and friction and eruptions can result in ridges, troughs, or escarpments.

Each of the ridge segments (typically 300–500 km long) has its own morphology and geologic history. Quite generally, everywhere along the axis of the ridge, there are subtle changes, including elevation changes of a few hundred meters. The changes are in large part a result of the way the central magma chambers along the ridge axis are supplied with molten basalt, from 30 to 60 km below. The supply is discontinuous in time and space and very uneven, with some sections sated with magma, and others starved. Mushroom-shaped magma chambers (some with roofs only 1.5 to 2.5 km below the seafloor) cause local uplift and can generate narrow axial graben structures. Such valleys can then be filled by sporadic lava flows, whose products (pillow basalts and basalt tubes) are ubiquitous evidence of volcanic activity along the ridge. The fast-spreading ridge of the East Pacific Rise has no distinct graben, as mentioned, although in other respects, it may be rather typical of ridge structure (Fig. 2.5).

The molten rock that wells up from the magma chamber forms pillow basalts and lava sheets after contact with the seawater. Seismologists studying the seafloor know the *pillow lavas* and the underlying basaltic dikes as "Layer 2A" and "Layer 2B." The basaltic rocks called *gabbro* ("Layer 3") and the olivine- and pyroxene-rich upper mantle rocks called *peridotite* ("Layer 4") are separated by the discontinuity in sound velocity known as the *Moho* or "Mohorovičić discontinuity" (named after the Croatian seismologist Andrija Mohorovičić , 1857–1936). Below the deep seafloor, the Moho is near 10-km depth. In slow-spreading ridges, much more complicated sequences may obtain.

2.3.6 Crest, Elevation, and Sinking of the Flanks

Generally, the crest of the mid-ocean ridge has a depth of 2500–3000 m below sea level all around the world. From this similarity of elevations, we may conclude that the upwelling material and its temperature are likely rather uniform. In any case, differences are quite subtle.

The crest is characterized by shallow earthquakes (centers are at depths of less than 60 km), by active volcanism, and by high heat flow values. The rift valley is a result of a growing opening of marine crust, with opposite sides pulling apart. It is commonly associated with slow spreading and may be missing elsewhere. The rate of spreading tends to show values between 1 and 10 cm per year (low rates are found in the Atlantic, high ones in the Pacific). The new hot material filling the central gap is light relative to common mantle material, because of thermal expansion. As a result, the ridge stands high above the most common seafloor. Away from the crest, the flanks sink because the basalt becomes denser as it ages and seeks a lower elevation, therefore. The sinking slows exponentially; that is, the sinking is more or less predictable, being linked to 1/square root of time of cooling.

In numbers, the sinking is by about 1000 m during the first ten million years. The subsequent 1000 m of sinking takes about 26 million years. The relationship follows this rule (approximately):

$$\begin{aligned}&\text{Depth added to original depth (in m)}\\&= 320\,\text{times square root of age in millions of years.}\end{aligned} \quad (2.1)$$

Using this equation, we can calculate the average age of the deep seafloor from its average depth (after correcting for sediment cover). The result is 60 million years,

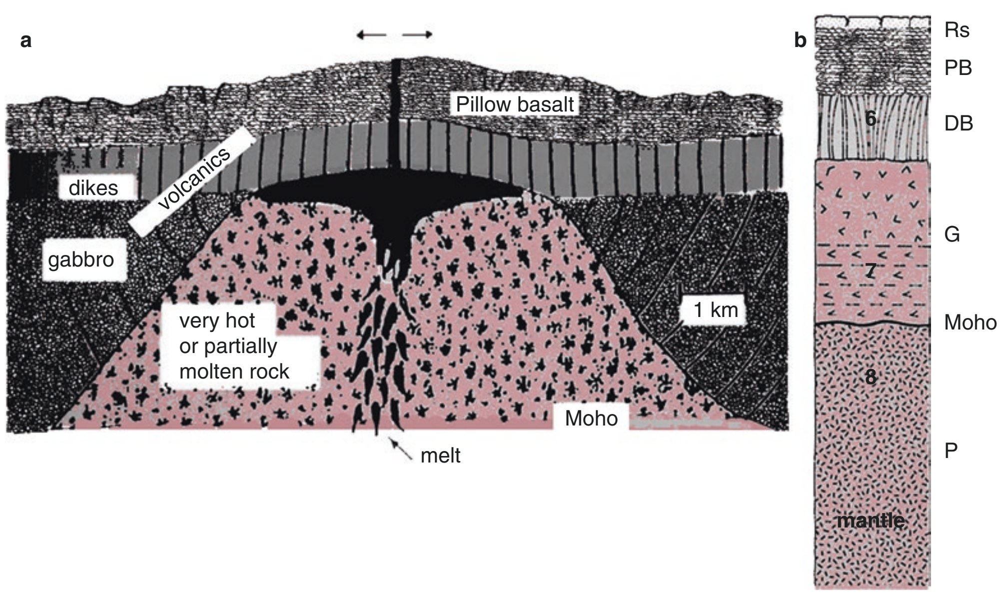

Fig. 2.5 (**a**) Comparison of the internal structure of the East Pacific Rise with (**b**) field observations on an ophiolite sequence on land. (Mainly after sketches by K.C. Macdonald et al., Nature 339, 178; and C. Allègre, 1988, modified) The numbers indicate rough estimates of typical speed of sound in km/s. The "Moho" (base of the crust) represents a jump in sound velocity. Letters in **b** refer to rock types: *RS* radiolarite (chert), *PB* pillow basalt, *DB* basaltic dikes, *G* gabbro, *P* peridotite (mantle rock)

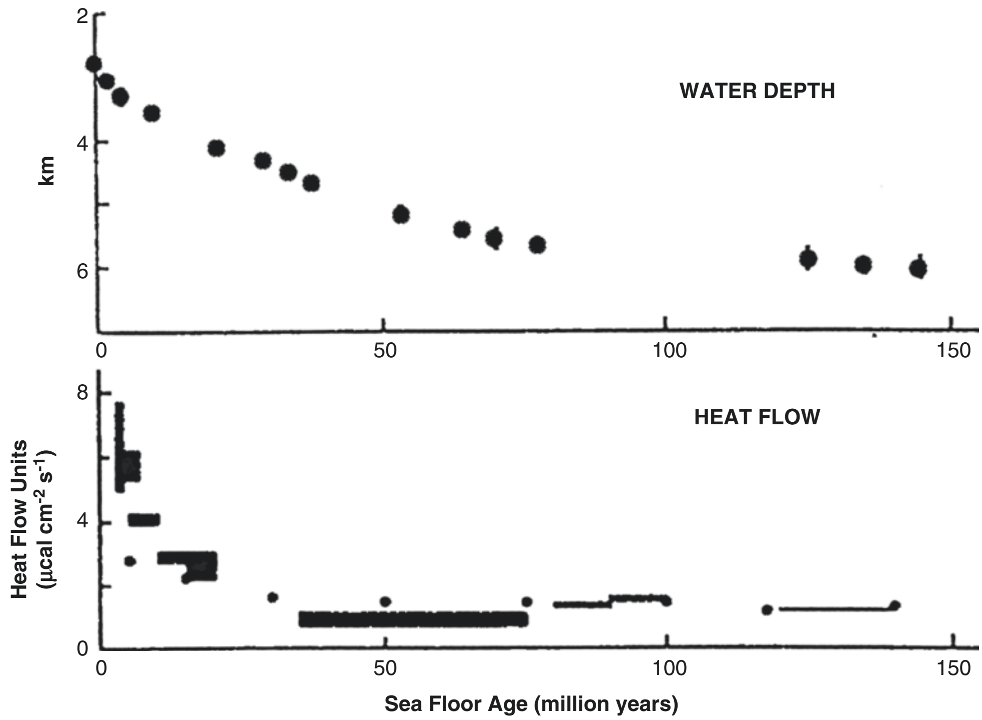

Fig. 2.6 Sinking of the flanks of the MOR (J.G. Sclater et al., 1971) and associated heat flow according to J.G. Sclater and J. Francheteau (1971; dots) and to Sclater et al. (1976; bars) (From E. Seibold et al., 1986. The Sea Floor, Japanese edition. Springer, Heidelberg & Berlin)

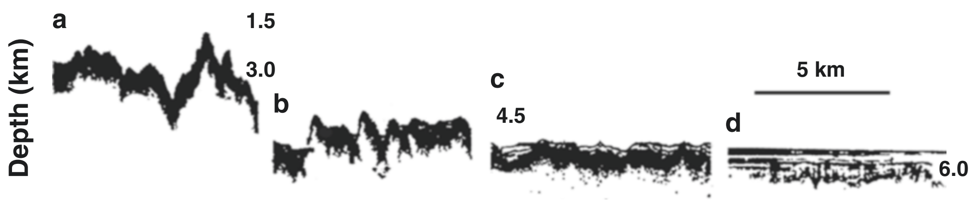

Fig. 2.7 Portions of continuous profiling records, from Mid-Atlantic Ridge (leftmost panel, **a**) to Hatteras Abyssal Plain (rightmost panel, **d**). Note the smoothing of the seafloor by sediment, with increasing distance from the crest. The sediment making the flat abyssal plain is likely to be largely terrigenous mud (After T.L. Holcombe 1977, GeoJournal 1: 25; simplified)

which is indeed close to the observed average age of the seafloor. The evidence linking sinking to cooling is excellent (Fig. 2.6).

The sinking starts commonly near 2500 m, at the average elevation of the crest. There are, however, exceptions to the general rule of ridge elevation. The ridge is much shallower than average in the North Atlantic, where it has, not so coincidentally, one of the most active "hot spots" associated with it, namely, Iceland. Iceland has volcanoes and "geysirs," that is, hot vents fed by locally available water. In fact, the word "geysir" is of Icelandic origin. Evidently, geysirs cool the rocks that make them.

2.3.7 Overall Morphology from Crest to Abyssal Plain

The morphology at the crest commonly is very rugged and complicated, while the flanks tend to be smoothed by sediment and (in the Atlantic) covered by turbidites at the low end (Fig. 2.7). During the sinking of the spreading seafloor, the rough topography produced by volcanism and faulting moves down the ridge flanks, being gradually smoothed by the sediment cover. However, abyssal hills with a relief in the range of 50–1000 m and with slopes between 1° and 15° remain as expressions of the underlying basement morphology over enormously large regions. *Abyssal hill* morphology provides for the most common type of landscape on the face of the Earth: in the Pacific Ocean, about 80% of the seafloor belongs into the category, that is, almost one third of the planet's surface.

2.4 Morphology of the Ring of Fire

2.4.1 The Shrinking Basin

For the Atlantic to grow while the planet retains its size, the Pacific basin has to be shrinking. The chief evidence that it does so is provided by the trenches rimming the basin, trenches being the manifestation of tectonic destruction of seafloor by subduction. The logic demanding a shrinking basin may be somewhat surprising, given the fact that seafloor spreading on the East Pacific Rise has values defining the fast end of the range, while spreading in the Atlantic is comparatively slow. But the demands of the balance notion are rather modest: the rate of subduction in the Pacific should just somewhat exceed the sum of seafloor production by Pacific spreading and the spreading in the Atlantic.

The length of the trench zone in the Pacific is comparable in magnitude to the MOR (Fig. 2.8). In fact, it is the only feature on the seafloor which achieves that particular distinction. Also, the *Ring of Fire* (which is, in essence, the trench zone with associated volcanoes) is a prominent source of deep earthquakes in contrast to the shallow-quake MOR. The MOR and the trenches are linked to the chief plate boundaries. There is no surprise in their great extent and in the occurrence of earthquakes along their boundaries. The boundaries mark zones of friction.

2.4.2 Morphology of Trenches

Trenches, of course, were described well before subduction became a textbook item (Fig. 2.9). The transition between oceanic and continental crust was a complete mystery to begin with. Problems are now regional and local, as the nature of the contact zone changes along and behind the trenches. Strong contrasts in mineralogy, fluid chemistry, and physical properties must be considered when dealing with this zone in any detail.

Trenches are abundant all around the margin of the Pacific basin, notably where continents meet ocean. It is not obvious why there are not more mid-ocean trenches. To answer this question, we would need to focus on processes within the mantle, that is, on largely unknown processes. What we do know is that trenches are narrow, being roughly 100-km wide (in their shallower part) and they are long (hundreds to thousands of kilometers). For example, the Alëutian Trench is 2900-km long. The cross-section is commonly V shaped (inset in Fig. 2.8), except that the deepest part may be flat due to sediment ponding. Sediment at the bottom of trenches commonly shows plain horizontal layering – an observation that

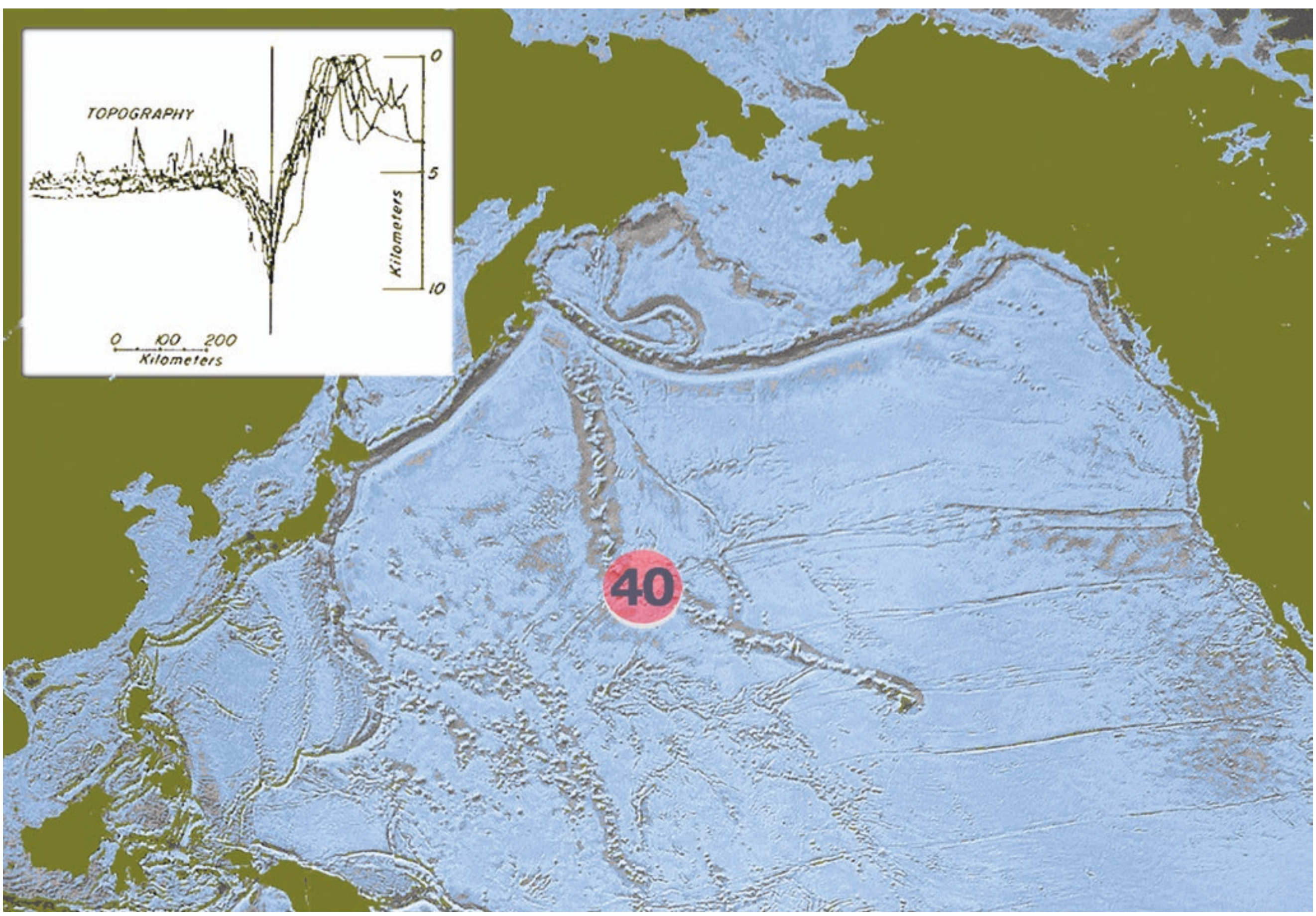

Fig. 2.8 Gravity-based morphology of the seafloor in the North Pacific. Trenches show minimum gravity (Map courtesy of D. Sandwell, S.I.O.; color here added) *Inset*: topographic profiles across trenches in various places in the ocean (M. Talwani 1970, in The Sea vol. 4 (1): 282). **40**, approximate age of the bend, in millions of years

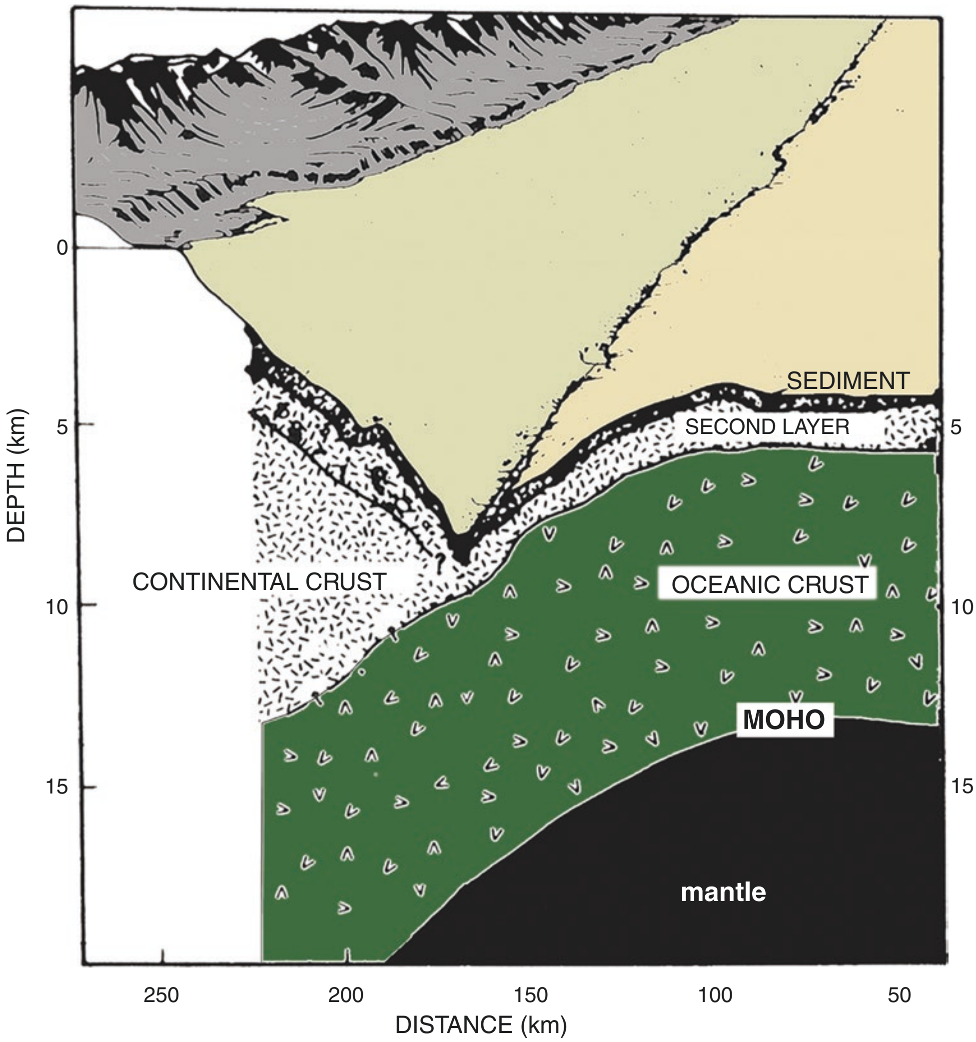

Fig. 2.9 An early version of a trench structure, based on pre-plate-tectonic interpretation of seismic refraction studies, and showing trench morphology off Chile (R.L. Fisher and R.W. Raitt, 1962. Deep-Sea Res. 9: 423. Shading and color here added). The down-going slab of the rigid upper mantle (the lithosphere) is roughly ten times thicker than the oceanic crust it carries. The "Moho" (here added) marks a sudden increase in sound velocity. The second layer actually has rocks that are different from the continental crust

was used on occasion as an argument against subduction when plate tectonic concepts were in their infancy. (However, as geologically young phenomena horizontally layered, turbidites in many trenches are not diagnostic of origins of regional morphology.) The trench walls usually have slopes between 8° and 15°. However, very steep sides (up to 45°) as well as steps also have been mapped. In cases, outcrops of basalt have been discovered by dredging and by deep-sea photography, notably on the inner wall (the side away from the deep ocean floor).

The greatest depths are in the western Pacific, in sediment-starved trenches off island arcs: Mariana Trench (maximally 10.915 m deep), Tonga Trench (10.800 m), Philippine Trench (10.055 m), Japan Trench (9.700 m), and Kermadec Trench (10.050 m). As locally measured features, depth values are not precise. Besides, they were determined from echo soundings affected by temperature and salinity of waters traversed by the sound. The resulting uncertainties, however, cannot mask the similarities in the maximum depths observed. As in the similarity of ridge-crest elevations noted above, the coherency points to the action of similar processes involving similar materials in each of the trenches of the western Pacific. Elsewhere, trenches tend to be shallower: Puerto Rico Trench about 8.600 m; South Sandwich, 8.260 m; and Sunda, 7.135 m. The eastern Pacific basin has trenches directly adjacent to continents, without intervening island arcs. Such trenches are being filled with continental debris.

2.4.3 The Zone of Deep Earthquakes

The ring of trenches girdling the Pacific is the site of deep earthquakes and volcanic activity (Fig. 2.10). Shallow quakes are abundant on the MOR, as mentioned. Within the subduction zones (landward of the trenches), when the depths of the quake centers are plotted below their *epicenters* (nearest point to the quake on the surface), they are seen to occur on dipping planes, presumably the contact planes between oceanic and continental crust. The planes intersect the surface near the trenches in typical arc fashion. They dip underneath the adjacent island arc or continent to a depth of about 700 km, within the mantle.

It seems reasonable to conclude that the earthquakes are largely caused by friction between the down-going slab of seafloor and the surrounding lithosphere and that the friction ceases when the rising temperature at depth becomes high enough to permit ductile deformation and flow. That the trenches are part and parcel of the new global tectonics was recognized by seismologists in the 1960s, based on the interpretation of earthquakes outlined (Fig. 2.11). In fact, the main features of plate tectonics all are well recognized in the seismic environment. Depending on the type of subduction at issue, volcanoes and mountain building proceeds on the landward side of the trench system. Much of such growth apparently also depends on the accretion of *terranes* (sometimes referred to as "exotic terranes"). These are pieces of real estate of various sizes coming from elsewhere with plate motion and refusing to partake in subduction, being less dense than oceanic lithosphere. Thus, they become part of the active margin. A trench clogged by a terrane must move seaward. Much of the West Coast of the USA is thought to consist of "terranes" that moved in with eastward plate motion in the Pacific.

2.4.4 Snake Rock

"Ophiolite sequences" are rocks of seafloor material that incorporate altered basaltic seafloor. In California, one sees "serpentinites" in certain coastal mountain belts, rocks that represent hydrated basalt (Fig. 2.12). "Ophiolite" and "serpentinite" both denote "snake rock" (the first in Greek, the second in Latin). The names are owing to the surface appearance of certain altered basaltic rocks; glossy and green and black and evoking the image of certain snake skins.

With few exceptions, the volcanic rocks forming the oceanic crust on the MOR (mid-ocean ridge basalt or MORB) are olivine-rich tholeiites (a basaltic rock, named after ancient basalt in Tholey near Saarbrücken in Germany). In addition to the minerals olivine (pronounced oly-veen) and pyroxene, MORB (mid-ocean ridge Basalt) has abundant plagioclase, that is, feldspar rich in calcium. (Basalts of island arcs may contain material from continental crust, which has relatively high sodium and aluminum content.)

Quite generally, the different types of basalts are characterized by differences in trace element content (e.g., rubidium, cesium, barium) depending on the history of the original melt. Interpretation of the chemical information is quite difficult and many problems arise when attempting to explain the (nonuniform) composition of the mantle material in terms of different types of basalt and their alteration products found at the surface of the seafloor and in continental margins.

Fig. 2.10 Ring of Fire. Deep earthquakes are common along the Ring of Fire. Very deep and powerful ones are associated with subduction (Fig. 1.2). The photo (Courtesy Taryn Lopez, University of Alaska) shows earthquake country near Fairbanks, Alaska (note the active volcano in the background to the right)

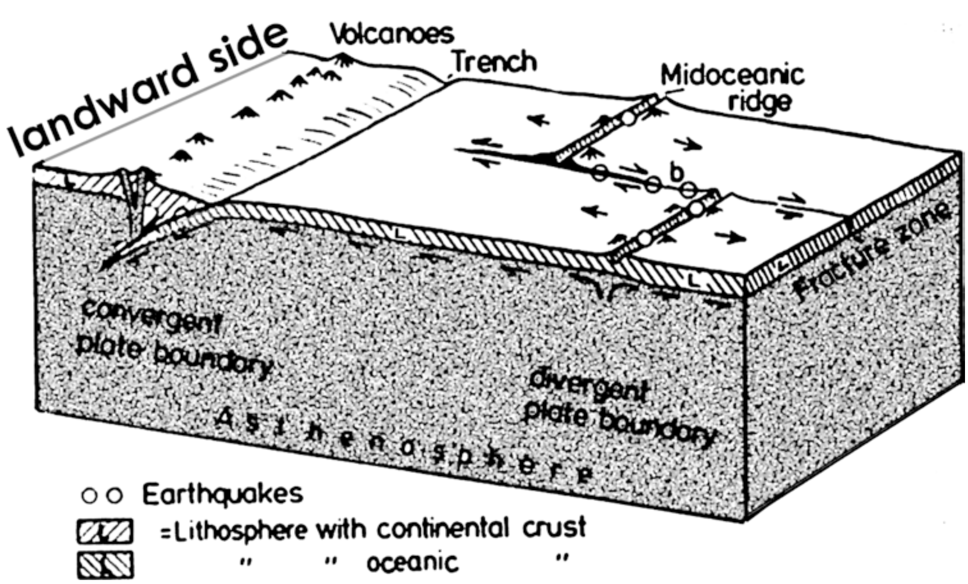

Fig. 2.11 Sketch of trench as an integral part of plate tectonics, based on a 1960s earthquake interpretation (After a famous diagram by B. Isacks, J. Oliver, and L.R. Sykes, 1968. J. Geophys. Res. 73: 5855; here modified for clarity) *L*, lithosphere (traveling toward trench, where it is subducted, accompanied by deep earthquakes)

2.5 Plate Tectonics, a Summary

2.5.1 Discovery

The features defining plates are the boundary-forming mid-ocean ridge (MOR), the trench zone, and *transform faults* connecting the MOR to a trench. These boundaries had to be explained in terms of spreading (Heezen, Hess, Dietz, Vine, Heirtzler, Bullard), subduction (Vening-Meinesz, Hess, Dietz, Isacks, Oliver, Sykes), and appropriate fault motions (Wilson, Vine, Morgan, Vacquier, Menard) to have "plate tectonics" emerge. While spreading and subduction are the central items, fracture zone activity is important. Fracture zones connect the end of a ridge crest portion to a trench. Such zones constitute the third type of plate boundary. Their importance was early recognized by the Canadian geophysicist J.T. Wilson (in the mid-1960s). First motion studies of earthquakes showed which way the slabs tend to move. Each of the plates, it turns out, has its own particular motion, whose long-term aspects can be read from the magnetism of the seafloor (A.V. Cox, J.R. Heirtzler, V. Vacquier). The quantitative development of these concepts was initiated in the late 1960s by pioneer geophysicists such as W.J. Morgan, D.P. McKenzie, R.L. Parker, X. LePichon, B. Isacks, I. Oliver, and L.R. Sykes (see A. Cox, 1973; listed in suggested readings).

Fig. 2.12 Serpentinite (hydrated basalt) found along the coast of California. (Here: Big Sur road cut) The rock glistens and commonly is *green* and *black* in color. A *bluish* tint is seen in places. Large rock lens (ca 1 inch diameter) for scale (Photo W.H. B)

2.5.2 Theory and Observations

"Plate tectonics" takes its name from great pieces of moving real estate on the planet, plates that are perched on uppermost mantle and carry crust and continents (Fig. 2.13). The thickness of the plates, which is typically near 100 km, varies significantly. A great variation in thickness also is typical for the crust that makes up continents (ca. 15 to more than 30 km) – in contrast, the ocean's crust is relatively uniform and is only some 10-km thick. The motions of the plates (lithosphere and crust) are parsimoniously described as rotations on a sphere, each uniform motion being linked to a pole of rotation as required by Euler's theorem. The fracture zones provide traces for the latitudinal circles around the respective pole of rotation (which can be close to, but need not coincide with, a pole of the daily rotation of Earth). Spherical geometry requires that spreading rates must increase away from a pole describing the separation of two plates, toward a maximum value. Remarkably, a given plate can contain both oceanic and continental lithosphere (but does not have to).

The kinds of rock dredged from deep clefts in the mid-ocean ridge and those recovered by drilling into the deep basaltic crust at sea presumably most closely resemble the material within the uppermost mantle and involved in plate motions. It may differ from the original rock, however. It is surely altered, with exposure on the surface of mantle and crust, especially through reaction with seawater.

Fig. 2.13 Sketch of the major plates (and a few minor ones). The subduction edge of the Pacific Plate makes up more than one half of the "Ring of Fire" around the Pacific (a zone of volcanoes and earthquakes). *Arrows* show the general sense of plate motion for two of the largest plates; the letter **R** denotes the mid-ocean ridge, the site of spreading. Direction of relative motion reverses across MORs (Graph courtesy of S.I.O. Aquarium, modified)

The heat that drives the motions in the mantle most probably comes from the decay of radioactive elements. Another possible source is the gravitational segregation of heavy and light material, which produced the onion structure of the Earth in the first place (heavy stuff in the core, light stuff in the crust). To decipher the *messages from the mantle*, that is, information about processes in the interior, geologists use seismic signals. Also, they study the patterns of magnetic and gravity fields. This activity is in the realm of geophysics. Mineralogists and petrologists investigate the behavior of rocks under high pressures and temperatures, and geochemists collect indirect evidence on the interior, in part by studying elemental abundances in the solar system. All this information is used in the reconstruction of the history of plate tectonics, a most difficult subject.

An especially important aspect of plate tectonics, to geologists, is the way mountain chains form in the trench zone. Continental accretion takes place here. It is one way in which the endogenic forces oppose the wearing down of continents by exogenic agents. Thus, the continued existence of continents that rise well above the sea is intimately linked to the processes associated with seafloor spreading.

2.6 Seamounts, Island Chains, and Hot Spots

2.6.1 Hot Spots and Plate Tectonics

Hot spot phenomena are routinely linked to processes deep in the mantle. What we see on the seafloor are edifices apparently resulting from volcanism, including hot spot volcanism. Hot spots on land have been invoked when contemplating certain major eruptions such as those of volcanoes in Wyoming and Idaho. Hot spots in the sea are thought responsible for some island chains, such as the one including the Hawaiian Islands. With very few exceptions, oceanic islands are made of volcanic rock, with or without a crown of reef carbonate. Coral reefs (or rather the algae thereof, symbiotic or coralline) depend on sunlight and grow only in shallow water, therefore. Thus, if a seamount is found with a top of reef carbonate and deeply submerged below the present sea level, it must have sunk from the sunlit zone. Such seamounts are common in the western Pacific. They usually have flat tops and they are roughly 100 million years old. It has been suggested that the mid-Cretaceous was a time of intense hot spot activity, possibly intense enough to elevate the seafloor elevation significantly.

No rocks older than Cretaceous have been found on flat-topped seamounts (or "guyots"). It is in fact unlikely that all the guyots identified have a similar history. There are thousands of seamounts in the Pacific basin, at various depths, so there is plenty opportunity for different signs of development.

2.6.2 The Hawaiian Islands

When discussing marine hot spot activity, one usually is faced with the Hawaiian Islands – large volcanoes along a chain suggesting motion of a plate across a hot spot. The volcanic origin is readily verified (Fig. 2.14). Everything else is not certain, even if very plausible.

The striking thing about the Hawaiian Islands is not, of course, that they are of volcanic origin. Practically all oceanic islands are. There are but a few that are not, being broken

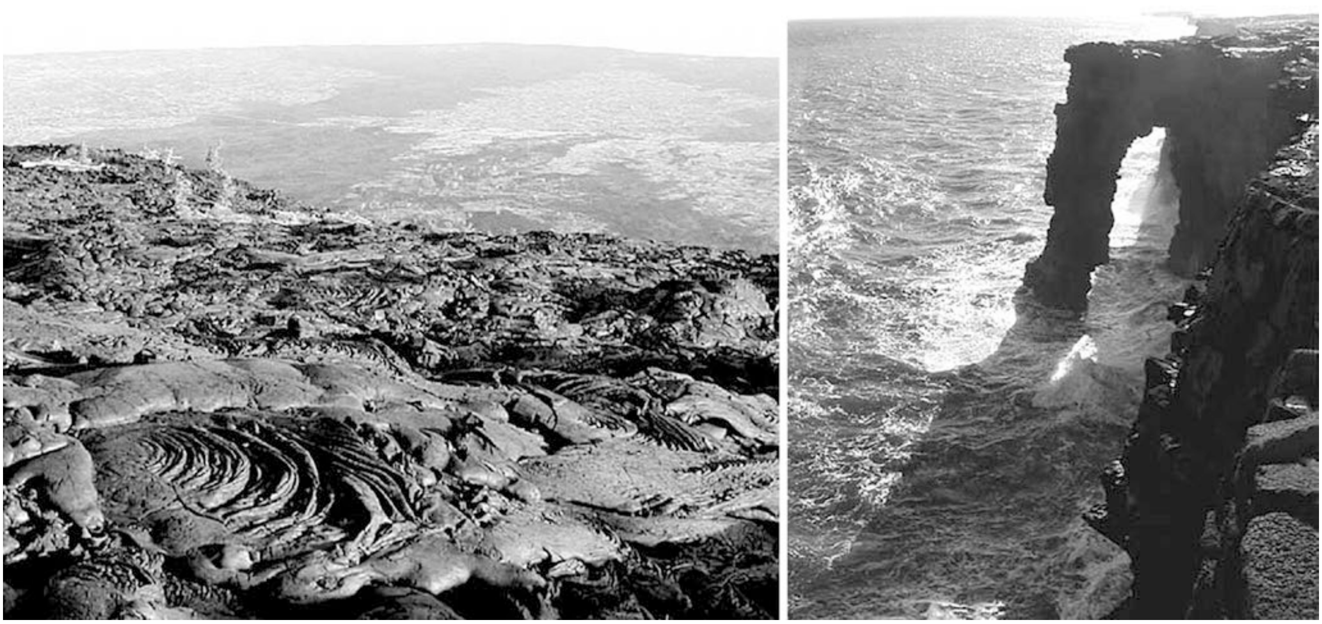

Fig. 2.14 The "Big Island" of Hawaii is an active volcano. Lava can be seen flowing into the sea, as well as frozen into rock on the flanks of Mauna Loa and at the shore (Photos W.H.B., the one to the right also is in the book "Ocean", UC Press, Berkeley, 2009)

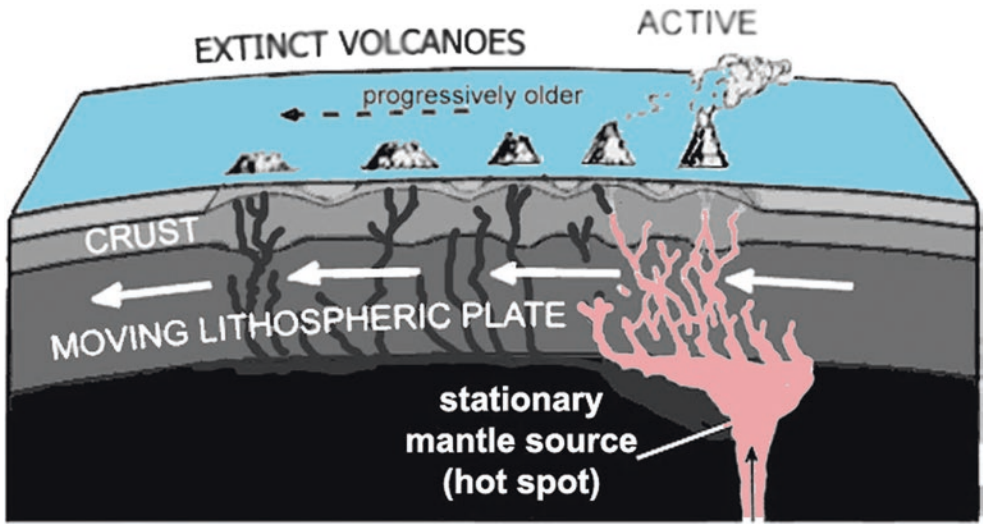

Fig. 2.15 Hypothesis concerning the origin of the Hawaiian island chain (After ideas of J.T. Wilson)

off from continents. What is interesting about the Hawaiian and some other island groups is that they occur in a chain and that commonly a chain in the middle of the Pacific has a kink connecting two linear segments (Fig. 2.8).

One possible explanation for the arrangement is to assume that the volcanoes making the chain lie on a long zone of weakness in the crust, a deep propagating fracture along which magma can rise to form volcanic structures. For the Hawaiian chain, where there is a progression from geologically young high and large active volcanoes in the southeast to sunken older islands with extinct volcanoes in the northwest, the movement of a plate over a hot spot has been postulated. The concept was put forward by J.T. Wilson (in 1965) and elaborated by J.W. Morgan (in 1968). In the view of these geologists, volcanoes built up on top of the oceanic crust over a "hot spot," a stationary source of melt deep in the mantle (also referred to as "plume"). As the plate moved over the hot spot, thus the hypothesis, a series of volcanoes formed, with the older ones cut off from the source and going extinct (Fig. 2.15). Along the Hawaiian chain, the knee between two distinct linear arrangements (circle in Fig. 2.8) has been dated near 40 million years. Within the framework of the Wilson-Morgan concept, a change in direction of plate motion is indicated on approaching the late Eocene. If this is correct, we must assume major reorganizations of upper mantle motions at that time. What we do know for sure is that major changes in sedimentation were going on toward the end of the Eocene.

2.6.3 Open Questions

Textbook assertions are not commonly ended on a question mark. Perhaps they should be in the case at hand. We are relatively ignorant of the mantle processes that drive plate tectonics and hot spot activity.

Regarding hot spots, what we can do is make some reasonable guesses. Large hot spot plumes are commonly thought to originate from the lower mantle and to contribute significantly to convection in the mantle. Basalts from this source are said to differ from basalts of the MOR in having a greater abundance of so-called *incompatible elements* (potassium, rubidium, cesium, strontium, uranium, thorium, and rare earth elements). Presumably, the upper mantle was stripped of these elements in its long history of making continents, so that high values are now only preserved in the lower mantle. However, the distribution of hot spots on the surface of the planet is as yet unexplained, emphasizing a lack of knowledge with regard to these phenomena.

Suggestions for Further Reading

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https://www.earth.ox.ac.uk/~tony/watts/downloads/Plate_Tectonics_2009_ABW_2.pdf

3 Origin and Morphology of Ocean Margins

3.1 General Features of Continental Margins

3.1.1 The Concept of "Continental Margin"

It is likely that the term was coined before its correctness (in terms of crustal rocks) was realized. The ocean floor, as we have seen, is made of geologically young basalt, unlike the continents. The basaltic rock that forms the ocean floor basement is rather close in composition to the mantle rock it came from (see the section on rocks, in the Appendix). Basalt is somewhat heavier than continental rock largely due to its high iron content. The lesser specific weight of the continental mass makes it protrude above the surrounding seafloor (Fig. 3.1). A thick wedge of sediment forms at the resulting boundary between ocean and continent. (In this view much of the shelf is continent.) The sediments may be well layered or strongly deformed, largely depending on the tectonic forces active at the particular margin considered.

The ocean margins, that is, the regions of transition between continent and deep ocean, differ greatly in their characteristics, depending on whether they occur on the continent's trailing edge or on the collision edge, or along a shear zone, and depending on the amount of volcanic rock in evidence. The one thing most of the ocean margins have in common is the occurrence of large masses of sediment (commonly in wedge form). All ocean margins are generally referred to as *continental margins* – a reflection of our original focus on coastal morphology and our landlubberly point of view.

The great difference in character of the margin (collision or trailing) is readily seen on the North American continent (Fig. 3.2). The western margin, where new continental crust is being made, is rising and displays cliffs with old sediments or with igneous rocks. The trailing eastern margin is sinking and bears large areas of wetlands and lagoons. Of course, photos taken at the surface can only show surficial differences along the coast. Most of the margins are in fact deeply submerged. The shelf (typically up to 200 m deep and with a slope of less than one half of a degree) has proportions of the seafloor of between less than 2% (Pacific, dominantly collision margin) and 8% (Atlantic, dominantly trailing edge) of its host ocean, the continental slope from 5% to 8% (the slope being typically very gentle, less than 1.5°), and the continental rise from 1.6% to 6.2% (with an inclination commonly well below 1°). Despite the relatively modest extent of the ocean margin, its sediment wedge tends to be very substantial, thanks to the input of materials from erosion on nearby land and (in places) of explosive volcanism. Also, here at the margin we find much of this input mixed with marine products from the coastal ocean, as was long ago recognized by John Murray of the *Challenger* Expedition and all marine geologists since.

3.1.2 The Coastal Zone

There is one type of seafloor that everyone is familiar with: the beach, at the shallow end of the shelf. In Southern California, it is typically a thin strip of sand on a narrow terrace cut into cliffs (Fig. 3.3). Elsewhere, on trailing margins, beaches can be much wider. Also, in between the Californian beaches, there are marshes and wetlands, commonly filled to create real estate for human use. Some are left in their natural state, sporting abundant marine benthic creatures living in and on the surface of dark mud. The mudflats are important sites of geochemical reactions (e.g., denitrification), and they are also important as sources of larval plankton to the ocean offshore. They occur at sea level in association with Holocene delta deposits piled into bays that formed as a result of the great melting at the end of the last ice age (Chap. 11).

The seafloor of the permanently submerged portion of the continental margin is much less well known than the exposed part, of course. Its morphology is accessible through sounding, notably using acoustics, and its cover is discerned by taking spot samples and reconstructing the resulting large-scale patterns. The continents, which serve as a prolific source of sediment from erosion of uplifted continental crust,

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_3

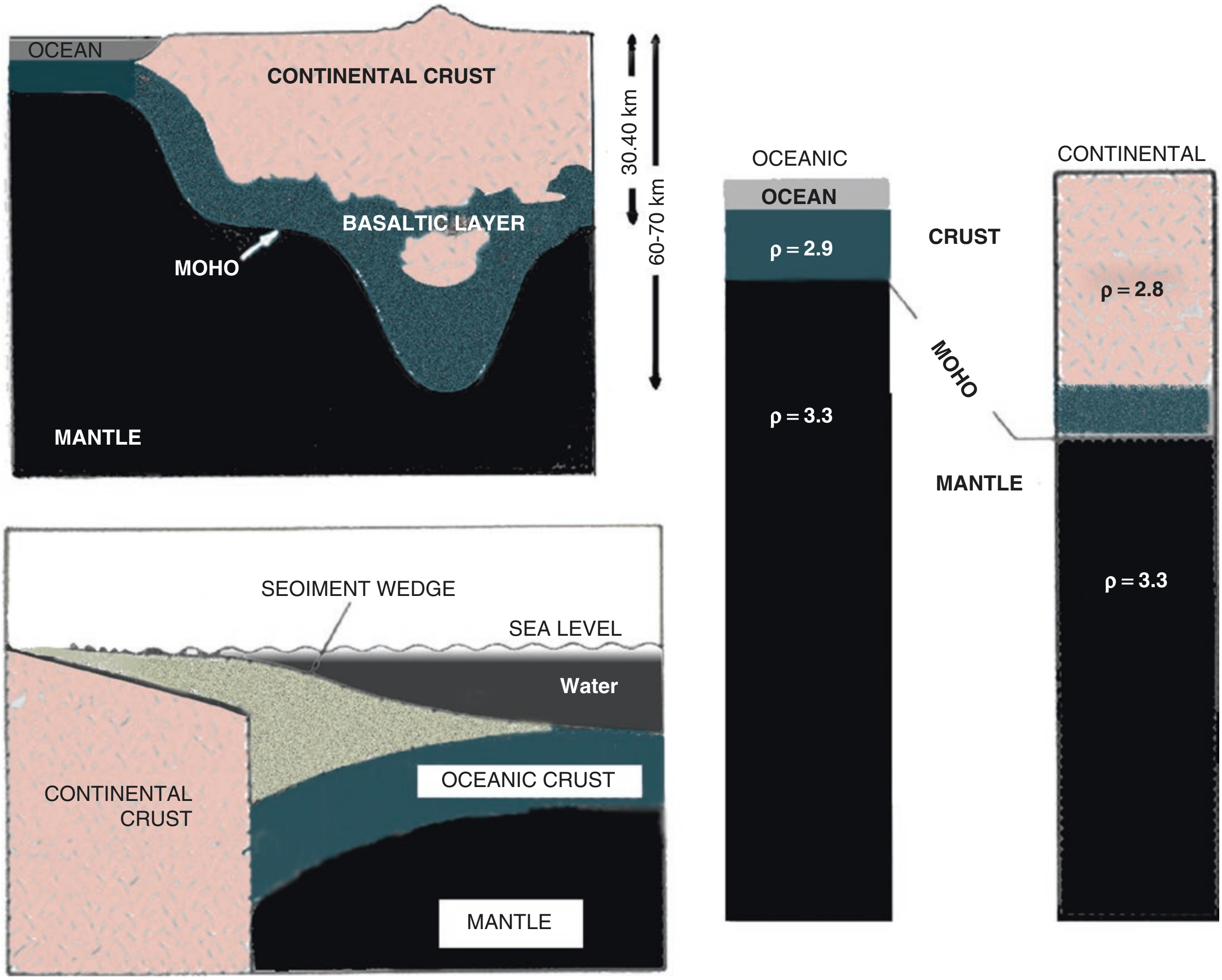

Fig. 3.1 Schematic isostatic block models for the continent-ocean transition and the significance of the "Moho" as a crust-mantle boundary (Mainly after S. Uyeda, 1978)

Fig. 3.2 Comparison of ocean margins at the western and eastern boundaries of the North American continent. *Left*: Big Sur, California (active margin). *Right*: Coastal New York from the air (passive margin; trailing edge) (Photos W.H. B)

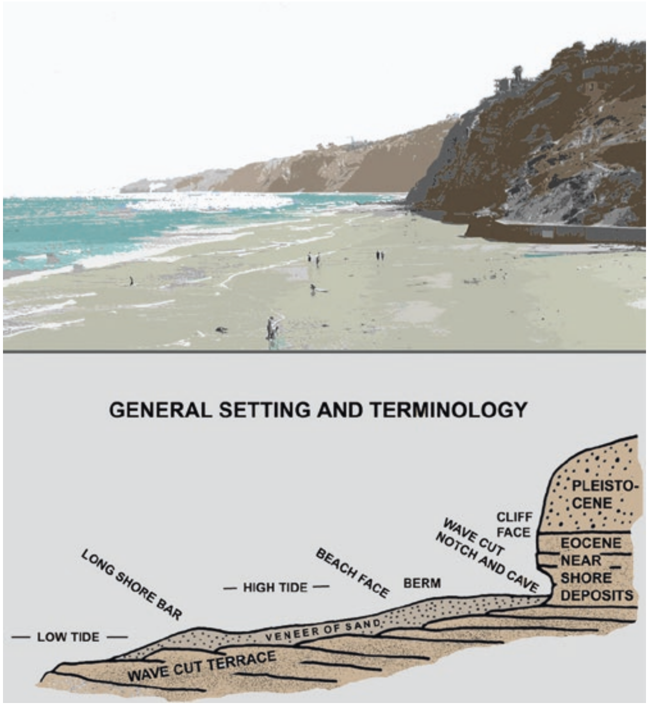

Fig. 3.3 Schematic of the beach north of the Scripps Pier, La Jolla, Southern California. The beach is the uppermost portion of the continental margin. It is occasionally covered by seawater (during high tide). The beach here consists of a narrow and thin band of sand on a terrace cut into the adjacent cliff, made of rising sediments of Tertiary age (*lower panel*) (W.H.B., 1976. Walk Along the Ocean. Mountain Press, Ocotillo; 'general setting' mainly after F.P. Shepard. Color here added)

Fig. 3.4 Mangrove covers on shallow seafloor in an embayment in Baja California. The height of the forest is typically around 6 to 7 m or 20 feet (Photo W.H. B)

tend to dump their debris first of all in the coastal zone from where it passes on to the great depths offshore. Unsurprisingly, the correct interpretation of well-traveled sediments proved to be a serious challenge.

Mangrove growth is commonly observed at tropical margins. In some subtropical bays of Baja California, the shallowest part of the seafloor bears dense mangrove vegetation (Fig. 3.4). Mangrove is a sea-level indicator. See Chap. 6.

3.1.3 Margins Trap Sediment from Land and from the Coastal Ocean

The continental margins (or ocean margins) are the dumping sites for the debris coming from the continents, that is, the *terrigenous* sediments. The margins also harbor the most fertile parts of the ocean. Thus, much organic matter becomes buried within the continental debris and the added pelagic material. In addition, such burial takes place in reef debris. If conditions are right (mainly a matter of history of burial and of heating), this organic material can eventually develop into petroleum, over millions of years. This happened in several places, one of which (conspicuously) is the Gulf Coast area, where oil is found buried under immense masses of sediment (see Chap. 14). Both oil and tar are found in the coastal zone of Southern California, where a thick wedge of sediments accumulated in the Neogene, with plenty of organic matter from upwelling. From experience most Californians know that drilling for offshore resources of oil and gas comes with considerable risks for underwater blowout.

Which margins are likely to have especially thick sediment wedges? Presumably, the largest and most active drainage areas are likely to emerge with the thickest wedges. One such super wedge, for example, is created by the deep-sea fan in the Bay of Bengal, a fan fed by the erosion of the Himalayas, the highest terrestrial mountains on the planet, and linked to a large and high plateau. Among the regular margins of ocean basins, however, it is those of the Atlantic that have the thickest wedges of sediment bordering it, up to 10 km thick and more. The Atlantic also has the largest proportions of slope and continental rise areas of the major ocean basins. The reason for this is not only sediment supply but also the fact that the Atlantic margins are "trailing edges" that have been sinking (and collecting sediments) for tens of millions of years.

3.2 Atlantic-Type (Passive) Margins

3.2.1 Background

Continental margins differ greatly depending on their origin. As early as 1883, the Austrian geologist Edward Suess (1831–1914) coined the terms *Atlantic margins* and *Pacific margins* to emphasize the major differences (Figs. 1.2 and 3.2). As mentioned, Atlantic-type margins are steadily sinking regions accumulating thick sequences of sediment in layer-cake fashion. In contrast, Pacific-type margins are rising on the whole and are associated with volcanism, folding, faulting, and various other mountain-building processes. Atlantic-type margins are now referred to as "passive" and Pacific-type margins as "active," because of the differences in tectonic style and the intensity of earthquakes and the absence or presence of active volcanoes.

It is now obvious that the origin of the margins must be understood in the context of seafloor spreading and plate tectonics. In the Atlantic the continental margins originated from a tearing apart of an ancient continent, along a line of weakness or great stress, with the torn edges sinking and accumulating sediment over large flat areas. A modern example of the rifting process associated with sedimentation can be studied in the Red Sea, where mantle material pushes up and tears the Arabian Peninsula from Africa (Fig. 3.5).

The initial phase of the process of passive margin formation can be studied in the East African Rift Valley, where the sea has not yet entered the rift. The pulling apart (and therefore the thinning) of the continental crust opens a window for mantle material (Fig. 3.5, left), melt that intrudes from the upper mantle. Heat flow increases correspondingly, and there is a bulging upward from the rising material, much as at the mid-ocean ridge. An increasing gap develops between the separating edges of continental crust, and pieces of continental crust break off at the edges at "*listric faults*." The process stretches the continental crust. Volcanism can be prominent in the early stages of rifting margins (as readily seen in thick *volcanogenic margins* of the Greenland-Iceland-Norway Sea, for instance). However, not all rifting margins start out in volcanic fashion. Non-volcanic types can be seen in the Northern Bay of Biscay and at Georges Bank, for example. Nevertheless, the outflow of lava is common at the time of initial rifting. If sufficiently extensive, such outflow can produce volcanic margins of impressive dimensions, as happened in East Greenland (DSDP Leg 38 and ODP Legs 104 and 152). Volcanogenic margins, of course, are likely to be especially prominent wherever hot spots contributed to the initial rifting.

At the Red Sea, submerged margins are sinking on the cooling lithosphere. Thick reef structures can grow in tropical and subtropical regions on such sinking blocks, building up a *carbonate shelf*, and further depressing the crust with their weight. If the Red Sea were only slightly less open, salt deposits would form – indeed there is evidence from thick evaporite deposits that this happened here in the geologic past.

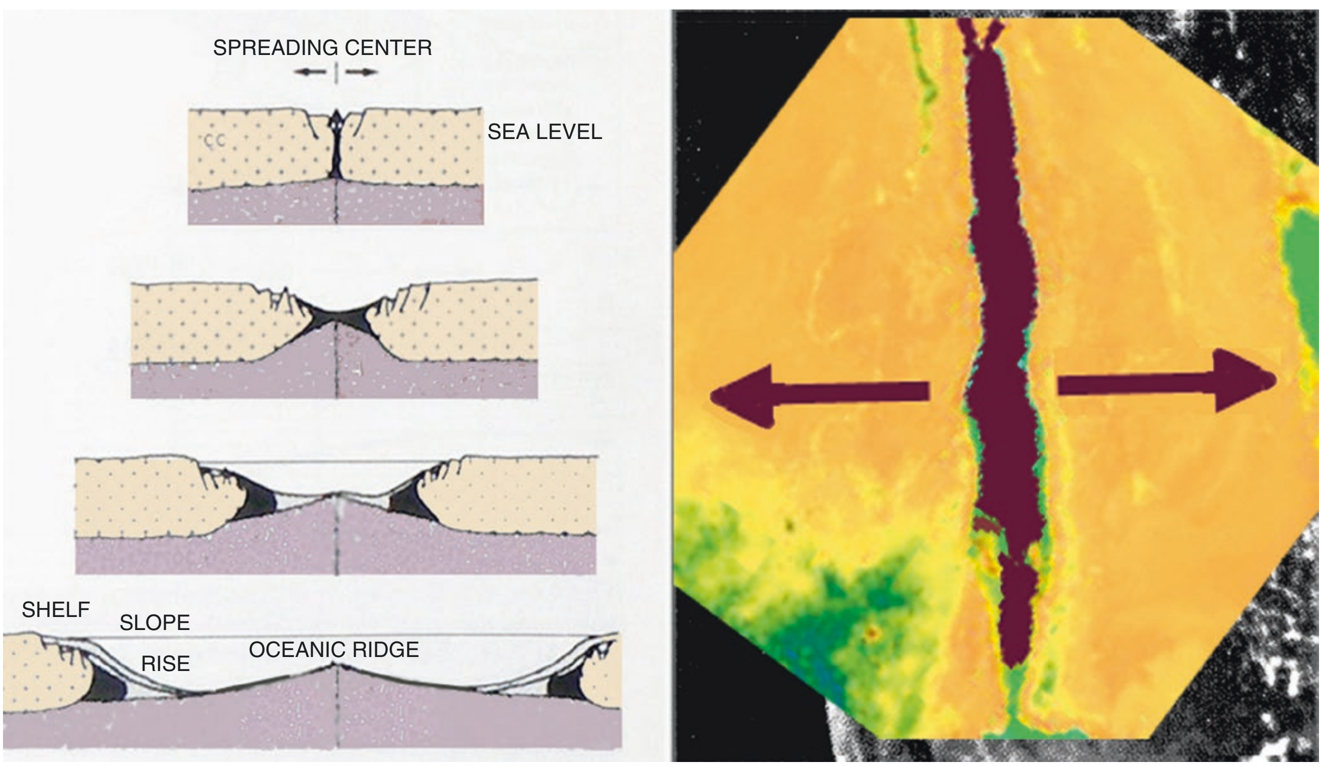

Fig. 3.5 Evolution of Atlantic-type continental margin. *Left*: sketch of sequence from uplift and expansion of the continental crust and volcanism to accumulation of sediment (stippled) and volcanic deposits (*black*) to regular seafloor spreading with sinking of the trailing edges. *Right*: space photo of Red Sea, where such rifting has started geologically recently and is now proceeding. (Space photo U.S. NASA; arrows and color here added)

3.2.2 A Plethora of Sinking Margins

In the Atlantic, coral reef margins are mainly found around the Caribbean Sea. The Atlantic margins off Africa bear excellent examples of trailing edges loaded with thick mixtures of terrestrial and pelagic sediments. The sediment stacks are commonly mapped using acoustics, that is, with a vessel pulling a sound-making device across the margin and recording the return, much as in echo sounding, but with the sound penetrating into the sediment (rather than being reflected from the seafloor only). In the example here shown (Fig. 3.6), the results were used in preparing for ODP Leg 175. (Prior to any ODP drilling, proposed drill sites had to be carefully surveyed, to prevent the choice of hazardous sites, that is, sites that could conceivably produce petroleum or large amounts of gas.)

Present and Neogene conditions, of course, are not necessarily typical for the geography prevailing during the evolution of the Atlantic. The early Atlantic saw the influence of restricted access to a newly rifted basin, as well as that of warm climate and a high sea level. Ancient *salt deposits* and *reef ramparts* are what we see along many of the Atlantic margins, although other sediments, of a more continental character, do exist also (Figs. 3.7 and 3.8). Salt deposits, of course, also are well known from the Gulf of Mexico, where they push up salt domes ("diapirs"), providing a path for petroleum migration (see Chap. 14). In the Atlantic proper, good evidence for salt deposits exist especially off Angola. The salt in the South Atlantic presumably accumulated when the ocean basin was narrow and closed to the north. Presumably it would have had restricted exchange with the world ocean due to the Walvis Ridge – Rio Grande Barrier to the south (now near 30°S). Large petroleum reserves may be associated with the salt deposits because the South Atlantic also was the site of deposition of organic-rich sediments during a lengthy period in the middle Cretaceous.

The end result of *rifting*, then, are continental margins consisting of thick sediment stacks piled both on the sinking blocks of a continental edge and on the adjacent oceanic crust. We can generalize these conclusions to all margins that originated by rifting and are now riding passively on a moving plate (that is, *passive* margins). Besides the bulk of the margins of the Atlantic there are the East African margin, the margins of India, much of the margin of Australia, and practically the entire Antarctic margin in this category of passive margin. In the Antarctic, of course, special conditions have prevailed with respect to erosion and deposition, at least since the formation of ice sheets there. Thick ice sheets have been present there for sure since the late Tertiary and quite probably much earlier, at least after the Eocene.

3.2.3 Unresolved Questions

There are a large number of unresolved questions in the study of passive margins. The commonly used analogies for the evolution of rifting – from East Africa and Red Sea to Gulf of California to Atlantic Basin – provide guidance as to the processes that might be at work. However, such analogies rarely answer geologic questions arising when studying actual rifting history. How much stretching (if any) was there

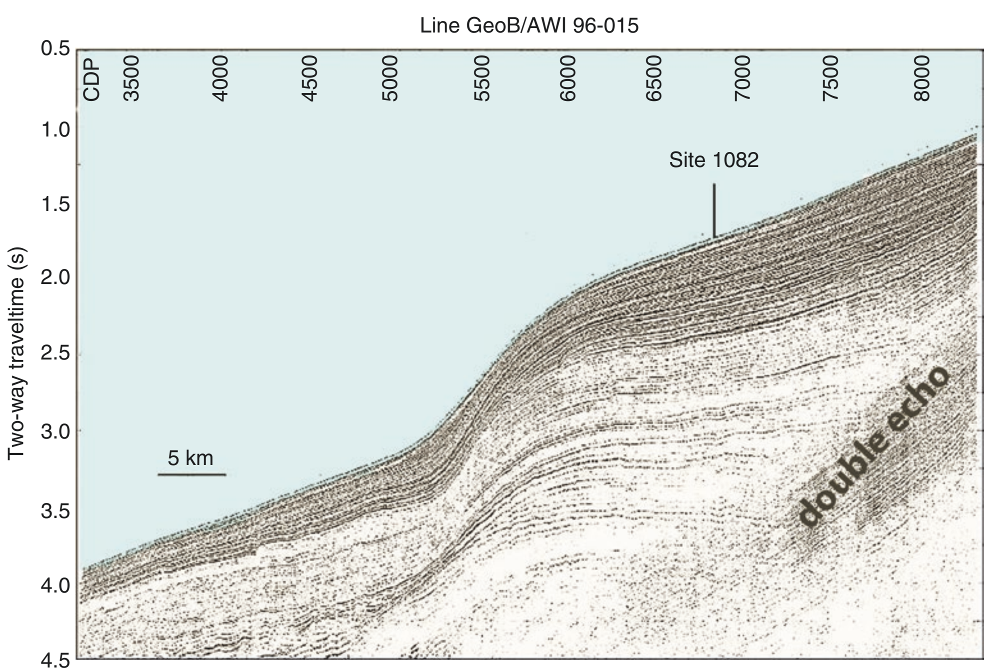

Fig. 3.6 Sediment stack off Namibia, southwestern Africa, explored with a small air gun in preparation of ODP Leg 175 (Seismic record courtesy of Volkhard Spiess, Bremen; color here added)

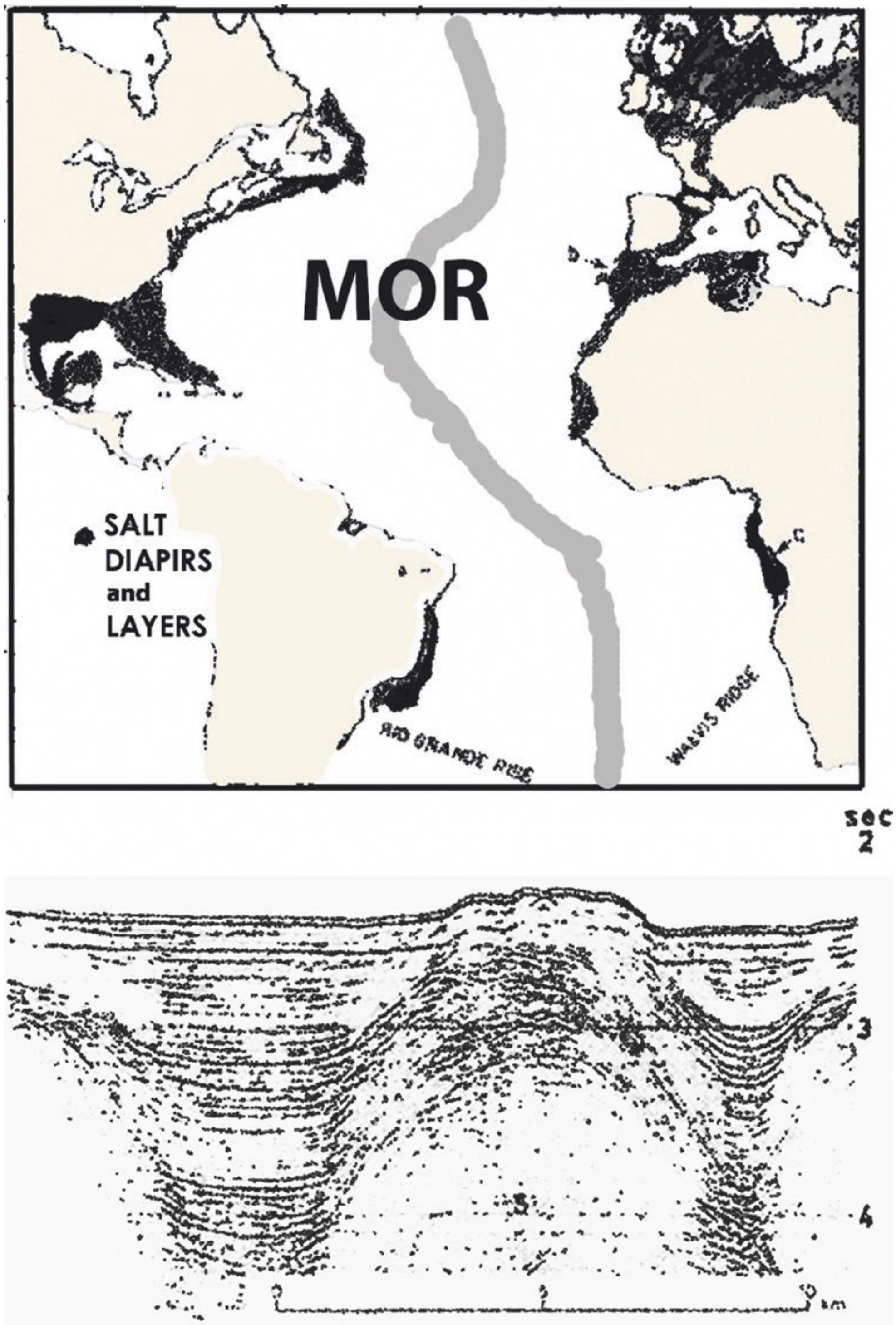

Fig. 3.7 Evaporite distribution in the Atlantic. *Upper panel*: geographic distribution of Mesozoic evaporites (After K.O. Emery; see AAPG Continuing Education Course Notes Ser 5: B-1 (1977); mid-ocean ridge (here added) for orientation) *Lower panel*: salt diapir structures (**S**) as seen on an air gun profile of Meteor Cruise 39, off Morocco (near 30°N). Depth of seafloor is approx. 1800 m (Graph after E. Seibold et al. 1976)

before rifting? How wide was the original rift valley before the sea entered? How does the sinking of the marine parts of the continental edge affect landward crustal blocks? What were the rates of uplift and subsidence through time? And what were the rates of associated erosion and sedimentation? With regard to the history of subsidence, what is the relative role of "floating in the mantle" (isostatic equilibrium) of the blocks, versus gravitational sliding? What are the forces responsible for the very long-lasting *uplift* of certain parts of the continental margin and for the formation of long deep-seated *barrier ridges* along some margins? What is the significance of the lack of sediments of a certain age, in many margins? Was the lack caused by erosion or by nondeposition or by huge landslides?

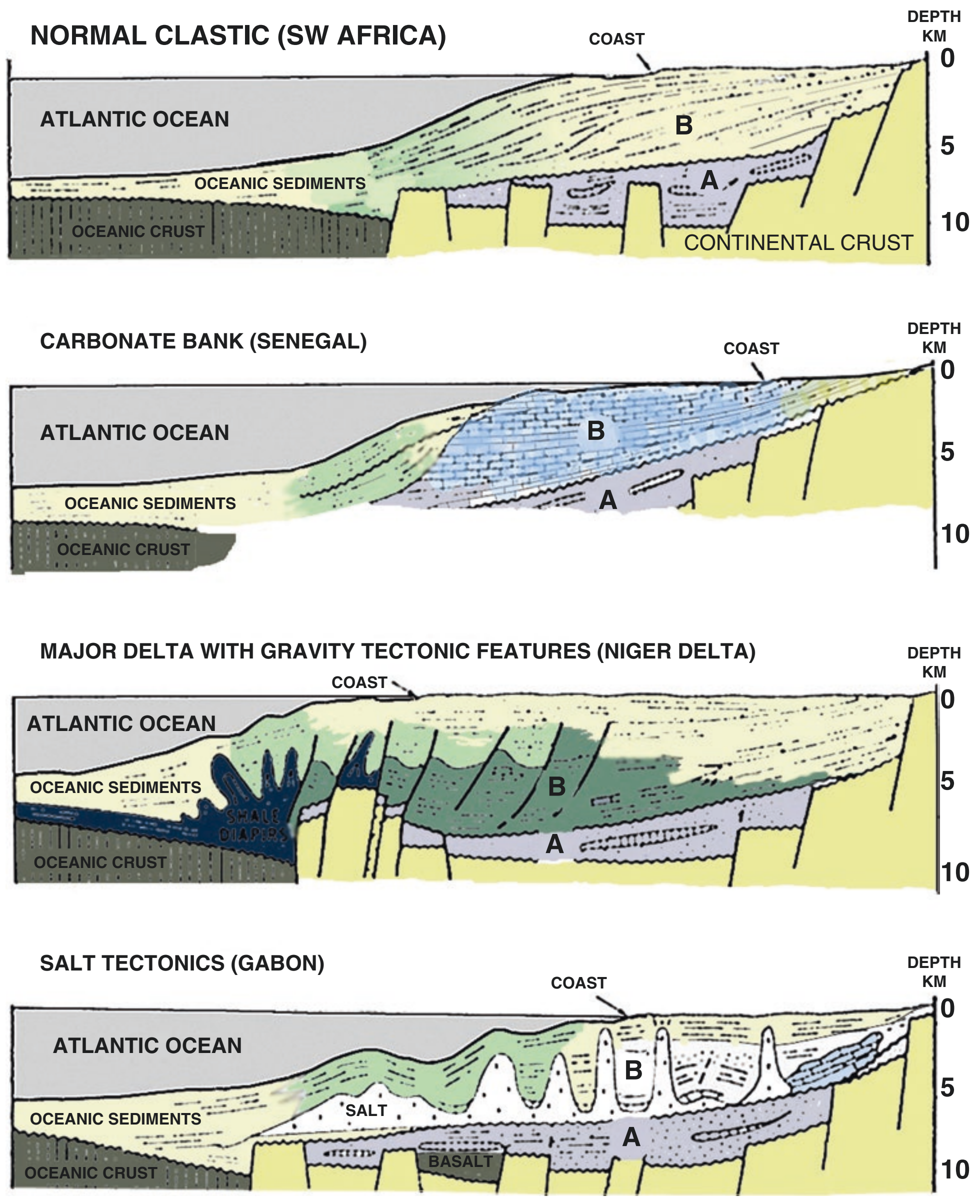

Fig. 3.8 Various types of passive continental margins. *A*, early deposits in the rifting zone, with non-marine deposits; *B*, deposits in a fully developed rift, with marine sediments. *Dark gray*: basaltic

In each particular case, a mixture of causal factors usually applies. "Silver bullet" situations with only one factor are rare.

One question of fundamental interest is whether and where the thick sediment stacks piling up on the passive margins will eventually be found in the geologic record on land. After all, the Atlantic margins cannot just go on moving apart – sooner or later a widening Atlantic runs out of space. One suggestion (by the Canadian geologist J.T. Wilson, 1908–1993) is that a proto-Atlantic once was formed by rifting, presumably in the early Paleozoic, *and then closed again*, running the previously passive margins into each other, thus turning a rift into a collision zone. A presumed product of this process (the "*Wilson cycle*") is the chain of (Paleozoic) mountains from Norway through Scotland and Newfoundland and on to the southern Appalachians. Checking the positions of the mountains on the Bullard fit (Fig. 1.14), we find that Wilson's suggestion makes sense as far as geography.

3.3 Atlantic-Type versus Pacific-Type Margins

3.3.1 Background

"Pacific-type" margins are the product of collision of plates. There are at least three types of collision margins that we need to consider: the ones produced by continent-continent collision as in the Himalayas, the ones reflecting continent-ocean collision as at the Peru-Chile Trench, and the ones

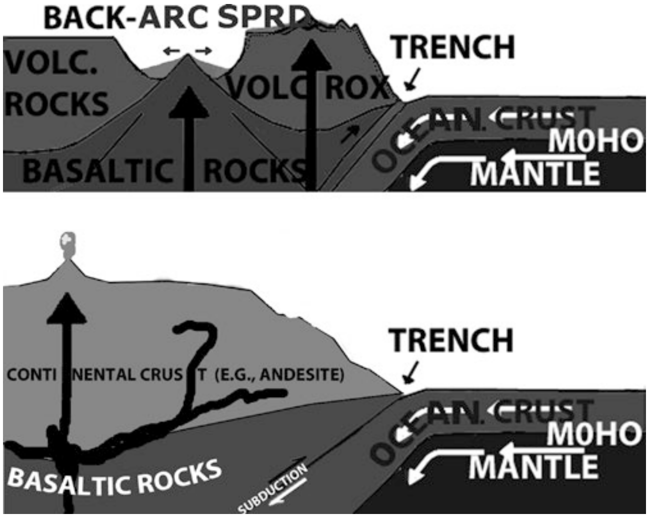

Fig. 3.9 Sketch of collision margins in profile (vertically exaggerated). *Upper panel*: back-arc spreading behind subduction zone. *Lower panel*: Peru-type collision (continent-ocean). Back-arc basin situation, with spreading center largely according to Dan Karig, with some info from D.R. Seely and W.R. Dickinson, 1977. AAPG Geol. Continuing Education Notes Ser. 5. Peru-type collision mainly after standard textbook models; see Uyeda, 1978

where subduction takes place along volcanic islands, as along the Marianas, for example (see Fig. 3.9). There are two fundamentally different types of the latter. As pointed out by the Scripps-trained marine geologist Dan Karig (then a graduate student) around 1970, one of these features is back-arc spreading (in Fig. 3.9 the two active margin types are compared schematically).

Among the most important characteristics of collision margins are folding and shearing of sediments along faults, as well as the addition of volcanic and plutonic material derived from mobilizing matter from the down-going lithosphere. The fractionation processes associated with partial melting on the descending slab and with hydrothermal reaction (seawater with hot basalt) can lead to local enrichment of deposits with heavy metals – and hence to the formation of ore deposits (see Chap. 14).

Based on deep-ocean drilling, we learned that as a rule the types of rocks that characterize the continental margins next to subduction zones are extremely varied. Among volcanogenic types we see solidified ashes, as well as altered basalts and gabbros. Among other igneous rocks, we see andesitic types along with highly deformed metamorphic rocks. Among sediments we can recognize both pelagic and shallow-water contributions. Certain rock types originating from the deep seafloor, when found on land, are referred to as ophiolites and are mapped in an effort to find ancient subduction zones. It is like hunting for lost oceans on land. Occasionally, the reward is the discovery of massive copper sulfides and other ores. The mechanism that allowed these observable ophiolites to escape subduction and to enter mountain building is a matter of research and discussion.

The steep slopes leading into trenches are favorable for large-scale submarine landslides. The jumbled masses (*mélange*) thus created are commonly sheared and metamorphosed (i.e., baked, cooked, and deformed) under great pressure and elevated temperatures. Temperatures may be lower, though, than is commonly found at the considered depths within the upper mantle). Blueschists and, following that, amphibole-dominated rocks (amphibolites) can form under such conditions.

The nature and history of active margins are the subjects of ongoing research. It still holds many surprises. An early simple concept of an origin from scraped-off material left behind by the down-going crust and lithosphere had to be substantially modified, for example. There is little transfer of material in places. In fact, there is *tectonic erosion*, whereby portions of the growing margin are swallowed by subduction. The swallowing trench may be fed by massive slumping (as is the case for portions of the Japan Trench), a process that delivers material that, after modification, can be used to build crust. The role of fluids in shaping margins has received increased attention. Tectonic motions (faulting, thrusting) and chemical reactions within the accretionary prism are both generally influenced by the presence of such fluids and their composition. The fluids are largely expelled from porous rocks by tectonic compaction and also stem from dehydration reactions involving mineralogy. Gases are important, too. In the Caribbean Barbados Ridge Complex, for example, the low-angle fault called "décollement" above the oceanic crust (Fig. 3.10) is greased by methane-bearing fluids within the fault zone. The presence of fluids aids in keeping the sedimentary wedge detached from the down-going slab.

3.3.2 Very Active Margins

Margins in zones of volcanic island chains are especially active. Spreading may co-occur with subduction where *back-arc spreading* is in evidence (Fig. 3.9, upper panel). Back-arc spreading occurs landward of volcanic arcs as, for example, in the Philippines or west of Guam. More than three fourths of such marginal basins are found in the western Pacific. That extension (by spreading, which lets magma rise) should be associated with collision is surely surprising. Complexity also arises from adding shear to active or to passive margins or from replacing trailing edges or collision margins with shear margins. Shear margins, like active ones, tend to have narrow shelves. Not all margins can be readily classified, of course. The processes characterizing them may be difficult to identify and categorize.

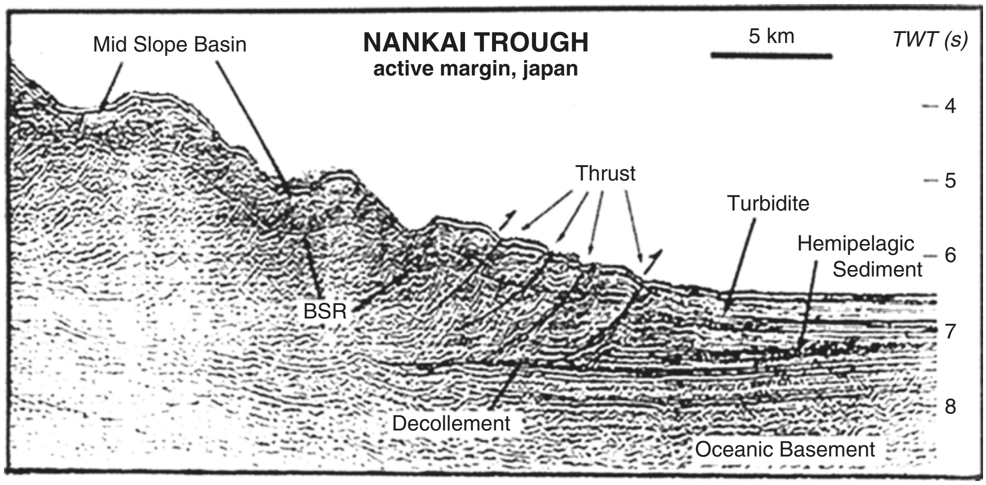

Fig. 3.10 Prominent features of active margins: reflection profile of Nankai Trough southeast of Japan. TWT, two-way travel time in seconds (1 is equivalent to 750 m). BSR, bottom-simulating reflector (commonly associated with methane ice). For "décollement," see the text (After A. Taira and Y. Ogawa, 1991; see Episodes 14, 3: 209)

Active margins also are sediment traps, although perhaps less obviously so compared with passive margins. In active margins sediments are piled up into chaotic mixtures of various types of rock, unlike in the passive margins, where well-ordered layering is the rule. One must keep in mind that in active margins enormous masses of material simply disappear deep within the mantle. The scale of subduction activity is difficult to imagine – the lithospheric slab now entering the Japan Trench is more than 10,000 km long! At present rates it will vanish in about 100 million years. Fluids leaving the landward wedge piling up in the subduction zone make "cold seeps." They may come from sediments, rather than from volcanogenic sources, depending on local conditions. Fluids travel along faults, mainly.

3.4 Shelves and the Shelf Break

3.4.1 Wide Atlantic Shelves

The *shelf* is the submerged part of a continent. Its typical maximum depth is between 100 and 150 m. It is rather flat, connecting the nearshore zone with the shelf break over some distance. Some shelves are quite wide, especially on passive margins (such as those in the Atlantic) (Fig. 3.11).

On the whole, the wide Atlantic shelves are depositional features, that is, they reflect sediment buildup on a sinking site of deposition. Narrow and rocky shelves, on the other hand, are common on active margins (as off California). Here erosional processes play an important role in shaping the shelf. Uplifted shelves (largely displaying alternation of erosion and deposition) and erosion-made nearshore areas once at sea level can make "marine terraces," features that are familiar along the coastal zone of California. Here they commonly serve as platform for roads or as sites for housing.

3.4.2 Shelf Seas

Some shelves extend deeply into continents and harbor *shelf seas* such as Hudson Bay, the Baltic Sea, or the Persian Gulf. Most of the marine sediments found on land were originally deposited in shelf seas. Thus, to understand these sediments – which cover a large portion of the continents – one must study sedimentary processes in modern shelf seas (see Chaps. 4, 5, and 6). Hudson Bay and the Baltic Sea, in their morphology and their sediments, have a memory of the ice ages, though, a rather unusual condition within Earth history. Low latitude shelf seas, for example, on the rim of the Mediterranean Sea or in Indonesia, might be more useful in delivering analogs for ancient shelf sediments. Given the ice buildup in the Neogene and a concomitant drop of sea level, useful shelf seas might not be as abundant as they were before the Neogene. Analogs for ancient shelf sediments are sometimes difficult to find.

Quite generally, present shelf environments and sediment types show considerable variety even over short distances. In large part, the variety owes to the fact that the sea level stood much lower than now only 15,000 years ago. Conditions were entirely different then from those of today, and many portions of the shelf still reflect those historic conditions in topography and sediment cover.

3.4.3 Ice Age Shelves and Fjords

The general nature of modern shelves, then, reflects tectonic factors (active versus passive) along with the recent rise of sea level by some 100 m, as ice on land melted during the transition from the last glacial maximum to the present. Thus, changes in sea level join tectonics as factors of prime importance in shaping shelves. To these sculptor processes must be added climate conditions (e.g., wave climate, storms, and various other local effects). In low latitudes, buildup by

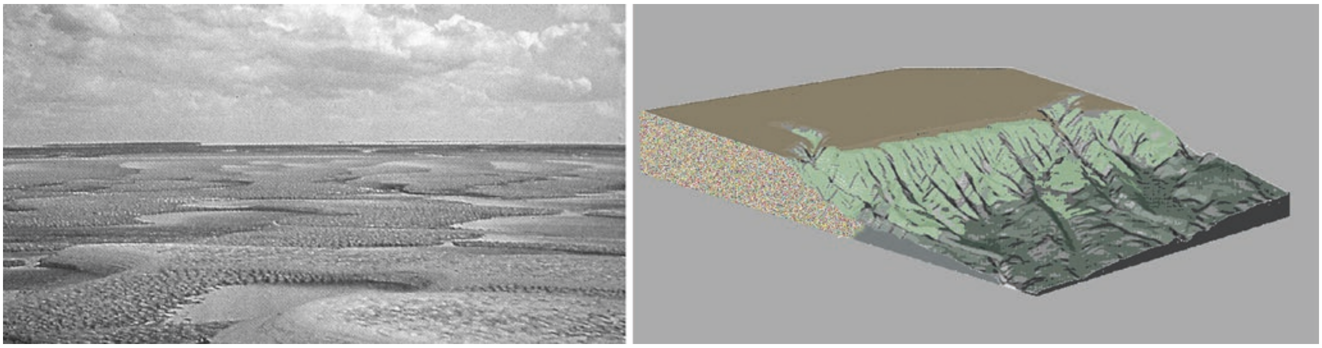

Fig. 3.11 Aspects of wide Atlantic shelves. *Left*: view of tidally flooded shallow portion (off the River Weser in northern Germany (Photo E. Seibold); *right*: passive continental margin with a wide shelf off Norfolk, USA (Drawing from NOAA; colors here added). Note the canyons beyond the shelf break break

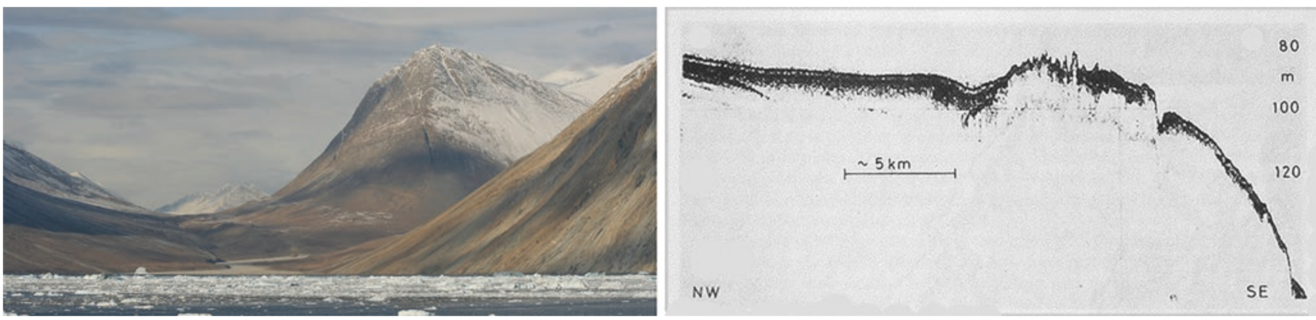

Fig. 3.12 Factors active in making shelves. *Left*: fjord in E. Greenland, attesting to erosive power of polar ice tongues (Photo W.H.B.). *Right*: shelf at the entrance to the Persian Gulf; subsurface echo profile by the research vessel Meteor (1965). The reef is dead. The shelf break in this case is somewhat above 100 m depth (From E. Seibold 1974. Der Meeresboden. Springer, Berlin)

reef-forming organisms is (and was for millions of years) important for shelf morphology in many places (Fig. 3.12, right panel). Some of the low-latitude shelves may be legacies of a distant past, even a pre-ice past. In high latitudes ice has been an important agent of shaping shelves for the last several million years (Fig. 3.12, left panel). The growth of ice not only led to exposure of the shelves to erosion but it also dumped enormous amounts of debris in places. Such debris includes large amounts of *moraines*. Much of this type of material still sits on the shelves off Newfoundland and in the North Sea. Ice moving out onto shelves also actively carved deep ravines and depressions that have not yet been filled with sediment. Runoff from melting ice can result in valleys carved into shelves. The fjords of Norway, Greenland, and Western Canada are witnesses to the powerful direct carving action of glacier ice. At the end of fjords, one finds sills made of moraines. Uplift (from ice unloading) can make such sills into barriers.

3.4.4 Delta Shelves

Shelves formed by large deltas off river mouths (e.g., Amazon, Mississippi) can be very flat and monotonous, in striking contrast to both rugged ice-carved shelves and irregular coral reef shelves. A rich supply of fine sediment brought to the delta environment allows for redistribution and much smoothing by waves and coastal currents. Of course, waves and currents can also build dunes, barriers, beach berms, and sand waves, depending on circumstances (see Chap. 9). Examples of conditions where the supply of terrigenous sediment (from rivers) is high and the shelf is flat include the North Sea, the shelf off the northern shore of Siberia, and the shelf of the Yellow Sea. Another prime example is the Senegal delta (West Africa), where the shelf has less than a foot of relief over several miles!

In any study of present shelves, we are chiefly faced with the question of how much of the observed morphology and sediment cover is of recent origin, and how much is inherited from the past. Time scale problems, again! Finding answers is complicated by the fact that "recent" includes at least the last several centuries. Within such time spans, the sea can produce effects whose causes may be hard to study, especially if rare but powerful hurricanes are included in the causal factors, as well as large earthquake-generated waves (*tsunamis*).

Tsunamis chiefly are produced close to the trenches rimming the Pacific; they travel over thousands of miles within a few hours. The waves are extremely long and quite low in the open sea, so that they are not noticed on board a ship riding them. But when such a wave reaches a shelf, it gains height on slowing down when "feeling bottom" in shallow water. Heights of tens of meters can be reached, and correspondingly severe damage can be produced along the coast suffering the onslaught.

3.4.5 Shelf Break

The shelf break, where the shelf ends and where the continental slope begins, is a prominent morphological feature of most continental margins. In principle, the break marks the depth below which the influence of sea level on erosion and deposition wanes rapidly. However, many details governing this feature remain to be discovered, especially since the origin of a break may involve a number of different factors. In many places off Antarctica, the break is uncommonly deep (near 400 m) compared with the usual depth near 130 m. Presumably, thick ice played a role in making the shelf break so deep. However, the break is equally deep off southwest Africa, where an explanation linked to ice action apparently would not work. Most commonly the shelf break is readily identified, especially where coral reefs are involved (Fig. 3.12, right panel). However, the transition between shelf and slope can be quite gradual in places.

Geologists have assumed for some time that the shelf break between 100 m and 150 m depth marks the low stand of sea level during maximum ice buildup for the last million years. Apparently the relevant low stand was reached repeatedly during maximum glaciation in the late Quaternary, thus exerting considerable control on shelf evolution. In any discussion of the depths of shelf and shelf break in a particular geographic area, the regional *isostatic response* of a crustal surface to loading and unloading with water owing to the changing sea level in the last several million years must be considered in addition to long-term regional uplift or subsidence.

3.5 Continental Slope and Continental Rise

3.5.1 Background

The classic profile of Atlantic-type (passive) continental margins shows a steepening of the slope beyond the shelf break and a gradually diminishing descent toward the deep seafloor (Figs. 3.13 and 3.14). The relatively steep part beyond the shelf break is the upper *continental slope*. The slope transitions into the extremely gentle (virtually flat) part of the margin that leads into the deep sea, the *continental rise*. The boundary between the lower slope and the upper rise is actually ill defined. Perhaps, it is helpful to think of the slope as definitely part of the continental margin, whereas the *continental rise* is built on oceanic crust and is essentially part of the deep-sea environment.

Not all slopes and rises, of course, fit the textbook outline of the ideal Atlantic-type margin – not even in the Atlantic. Deep marginal ridges, as off Brazil, sheer walls of outcropping ancient sediments and deep-lying plateaus, such as the Blake Plateau off Florida, can interrupt the ideal sequence.

The collision margins off Peru and Chile (Fig. 3.9b) are characterized by a steep slope and are without a rise – the trench swallows the material that would normally build the rise. A descent in a series of steps is typical for the collision slope, marking the complicated tectonics associated with the collision between opposing plates. Rather complicated conditions, in fact, prevail off much of the western US coast also. Earthquakes are common both at the rim of California and at that of Chile.

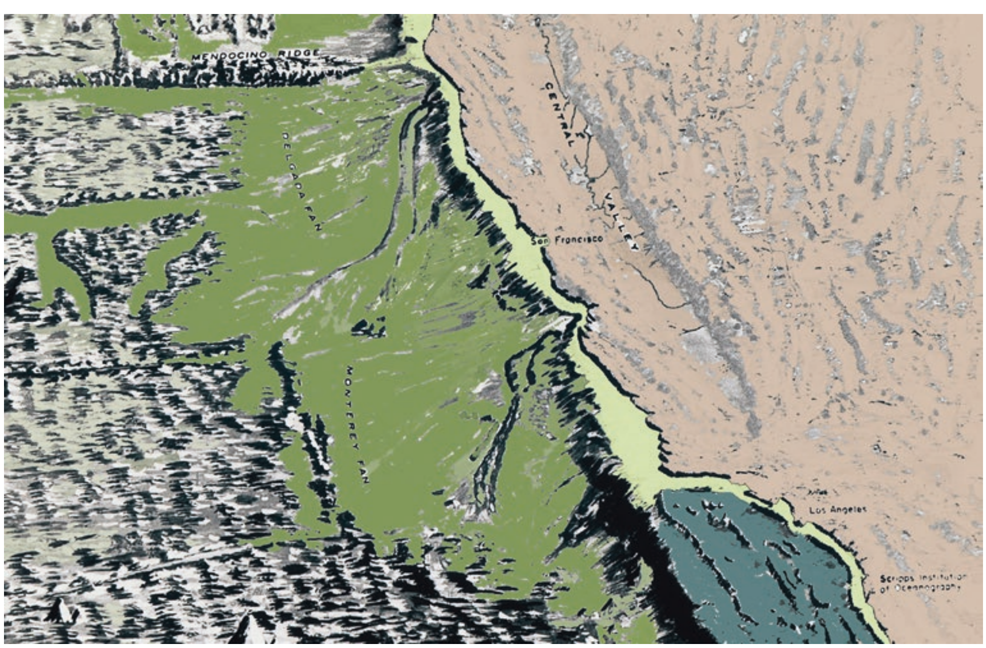

Fig. 3.13 Physiographic sketch of the continental margin off California. A portion of the California Borderland is seen in the southern landward portion of the diagram (Sketch based on a line drawing by H.W. Menard, 1964. Marine Geology of the Pacific. McGraw-Hill, New York). The narrow shelf is drawn *yellow*; continental slope mud is shown in *olive-green* and *grey*. *Bluish*: Borderland. (The origin is very complicated)

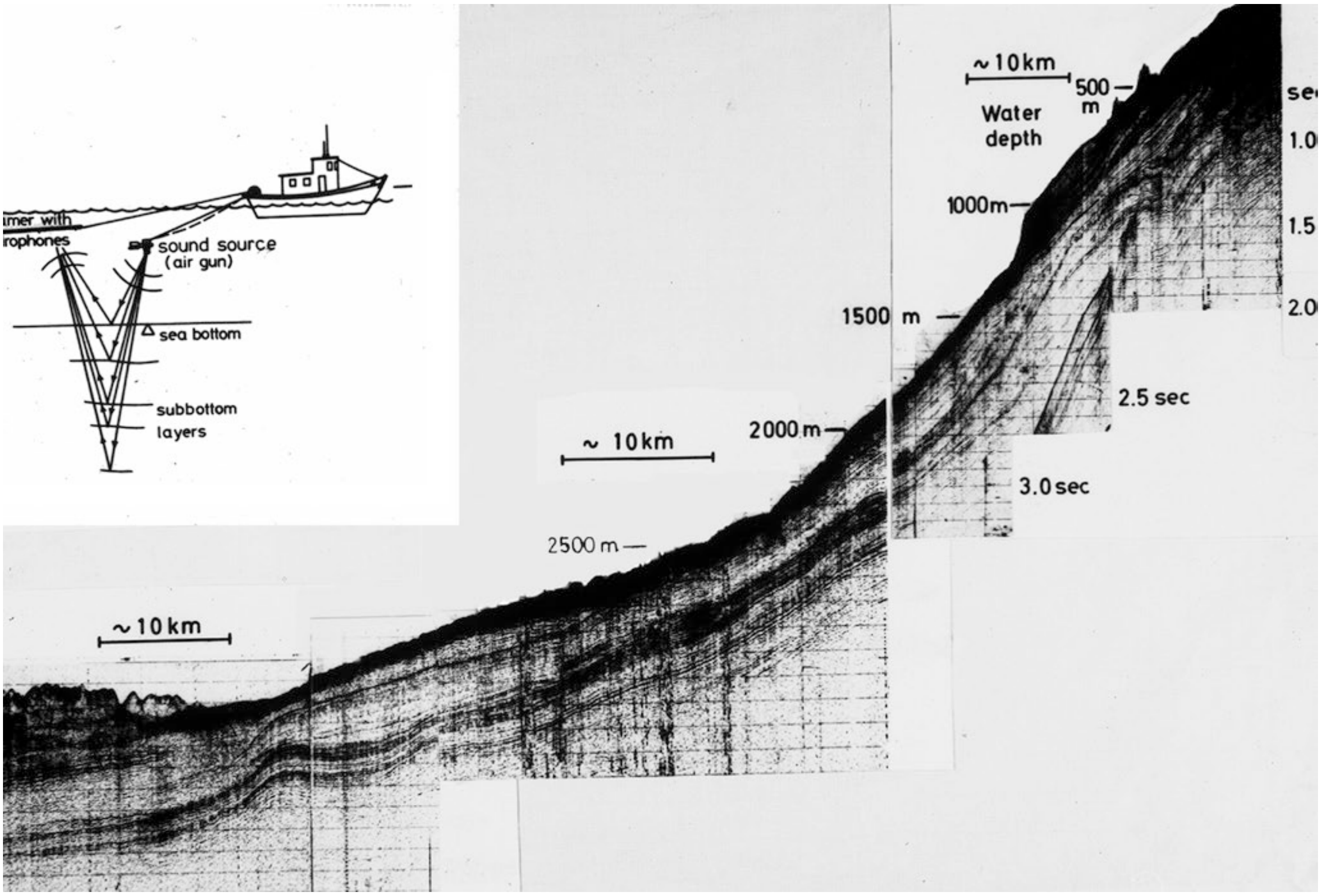

Fig. 3.14 Submarine mass movements off Dakar (Senegal, NW Africa). Air gun record of Meteor Cruise 25/1971. Thickness of slide is ca. 200 m. (Note vertical exaggeration.) *Insert*: air gun system with air guns as sound sources. Echoes from within sediment stack are picked up by the hydrophones in the trailing streamer (From E. Seibold, 1974. Der Meeresboden. Springer, Berlin)

3.5.2 Slope and Rise Off California

While the present slope and rise off northern and central California can be described simply enough in terms of coalescing deep-sea fan deposits (Fig. 3.13), no such description is possible for the margin off Southern California. The Southern California Borderland looks much like an extension into the sea of the basin and range topography familiar from the Mojave Desert, even though much complicated horizontal motion is involved within the Borderland.

The various physiographic diagrams commonly depicting the rapid descent from shelf edge to the deep seafloor (whether across passive or active margin) are somewhat misleading in that the slopes are, in reality, quite gentle (one to six degrees). A one-degree slope, of course, would appear as a plain on land to anyone standing on it.

3.5.3 A Variety of Slopes

The variety of types of slopes encountered suggests that a number of different forces are at work in shaping them. We have emphasized the differences between active and passive margins along with the endogenic forces that produce these differences. Changes in the thickness of continental crust, through melting and assimilation into the mantle (*subcrustal erosion*), have been proposed as one variety-producing mechanism active in collision margins. Such processes also need to be considered in trying to understand the Borderland morphology.

Basically, most slopes are the surfaces of thick accumulations of sediments washed off the continents and mixed with marine materials produced by organisms. Close to volcanoes, of course, there is a strong component of volcanic debris. The high rates of accumulation on many slopes result in a precarious situation in places. When there is not enough time to de-water and solidify slope sediments, immense submarine landslides can result from rather small disturbances (e.g., smallish earthquakes), even on very gentle slopes. The same is true if gas pressures within the sediment rise sufficiently. Such gases are commonly created by decay reactions of organic matter and may be unable to escape rapidly enough into bottom waters. They then can destabilize their host sediment.

3.5.4 Slides on Slopes

Slides on slopes tend to move on surfaces defined by clayey layers with high water content. Pressures may be unusually high in sloping horizons, thus decreasing friction between the overlying sediment stack and the base it rests on. Slides are not necessarily coherent. They can be chaotic (then called s*lumps*) and produce a jumbled mess whose origins are difficult to reconstruct in any detail.

Examples of slides on slopes are found readily along almost all continental margins. Off Cape Hatteras, for example, a tongue-shaped mass of displaced sediment on the upper continental rise is 60 km wide and over 190 km long, with a hummocky relief of up to 300 m in places. Another example of a large-scale slide is seen off Senegal (NW Africa) and off Dakar (Fig. 3.14). The jumbled slide masses at the foot of the slope now form part of the continental rise. The seismic profile shown in the figure was obtained by echo sounding *into the seafloor* with powerful "booms" of sound, rather than the "pings" used to define the surface of the floor below the moving ship. The method using booming sound is referred to as *continuous seismic profiling*. It is sketched as the inset in Fig. 3.14.

3.5.5 Erosion on Margins

As mentioned, margins are not just places of deposition but also of erosion. Erosion on slopes and rises can be produced by submarine landslides involving earthquakes and sediment instability and by the action of strong deep currents flowing horizontally along the slope. Commonly such currents have been dubbed *contour currents.* They are distinct from another type of important current: *turbidity currents*, mud-laden bodies of water flowing downhill. It is now thought that much of the sediment on continental rises and in adjacent abyssal plains is carried there by turbidity currents following slides and slumps that started somewhere below the shelf break, with contour currents doing the redistribution and shaping of the material. Thus, slides and turbidity currents apparently are largely responsible for shaping sediment-covered margins (Fig. 3.15).

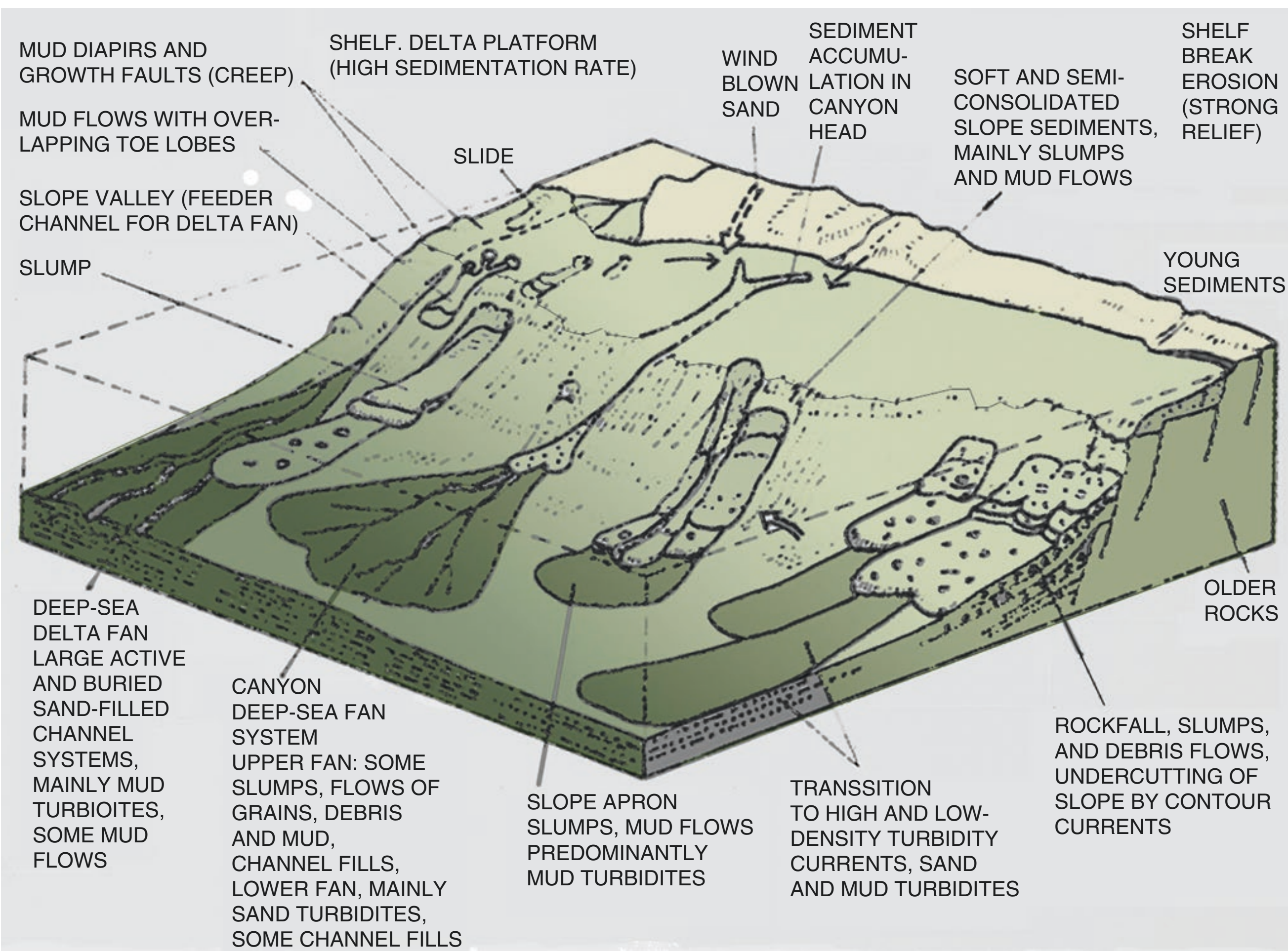

Fig. 3.15 Sketch of exogenic processes shaping continental margins (G. Einsele in Einsele et al. (eds.) 1991. Cycles and Events in Stratigraphy. Springer, Heidelberg) Note the olive-green color of the unstable mud off the coast (colors here added) and the prevalence of slumping and similar gravitation-enhanced processes

3.6 Submarine Canyons

3.6.1 Background

The continental slope is commonly cut by various types of incisions, ravines and valleys; the most spectacular of which are called *submarine canyons*. The origin of these impressive features has long puzzled marine geologists. Much has been learned concerning the matter, but the topic is still a matter of debate. The largest of such canyons off California is "Monterey Canyon" of Monterey Bay (Fig. 3.16, left). Its cross section has dimensions like those of the Grand Canyon of the Colorado.

One of the earliest of detailed studies of submarine canyons was off S.I.O., by Francis P. Shepard and his students and colleagues. By chance, S.I.O. has two of these features off its shores, named, appropriately enough, *La Jolla Canyon* and *Scripps Canyon*. The latter, a tributary to La Jolla Canyon, is somewhat smaller than the former. They join at a depth near 300 m, below the shelf break. The resulting La Jolla Canyon subsequently turns in a southerly offshore direction to end up in the San Diego Trough off the Point Loma Peninsula. The late geophysicist D. Inman, an expert on coastal processes, found out that beach sand sporadically moves down the canyons (Fig. 3.16, right). Indeed, such sand is found in the San Diego Trough, along with the remains of kelp, presumably delivered there by muddy gravity-driven flows down the canyons (that is, by turbidity currents). Note that the canyon heads shown (the largest nearby and the closest to S.I.O.) are on the shelf (which was exposed during the last ice age).

3.6.2 Origin

Many large submarine canyons are much like their counterparts on land: with tributary systems in the upper parts, and meandering *thalwegs along much of their course* (similar to river beds), and with steep sides in places (20° to 25°, even 45°). Overhanging walls also occur (Fig. 3.17). Walls may consist of hard rocks, including granitic rock in places. As in river canyons, there is a continuous descent of the valley axis, with maximum slopes of 15% near the coast and gentle slopes (about 1%) farther out to sea.

A large number of hypotheses have been put forward to account for the origin of submarine canyons. In fact, we must look for different origins for different types of canyons. For example, thanks to deep-sea drilling, we know that the Mediterranean became isolated from the world ocean at the very end of the Miocene, some six million years ago. It then dried up repeatedly during regionally dry periods or when sea level was lowered. At those times deep canyons could be carved along the edges of the basin. Indeed, the true floor of the Nile Valley (below geologically young sediments) is very deep, supporting the idea of canyon cutting during desiccation. However, we can hardly invoke such drastic falls in sea level everywhere in the world ocean. Thus, there must be a way to make submarine canyons by cutting them underwater. One common answer is that muddy waters heavy enough to flow downhill do the cutting, that is, *turbidity currents*.

3.7 Turbidity Currents

Turbidity currents derive from muddy downhill flow. We are probably quite familiar with muddy downhill flows: large amounts of mud are brought down a river during floods. In the sea (and in lakes), hurricanes and wave action may produce muddy water by the stirring up of sediments. Earthquakes presumably can be agents for starting submarine mudslides that turn into muddy downhill flow. During glacial periods, when sea level was lowered (commonly by

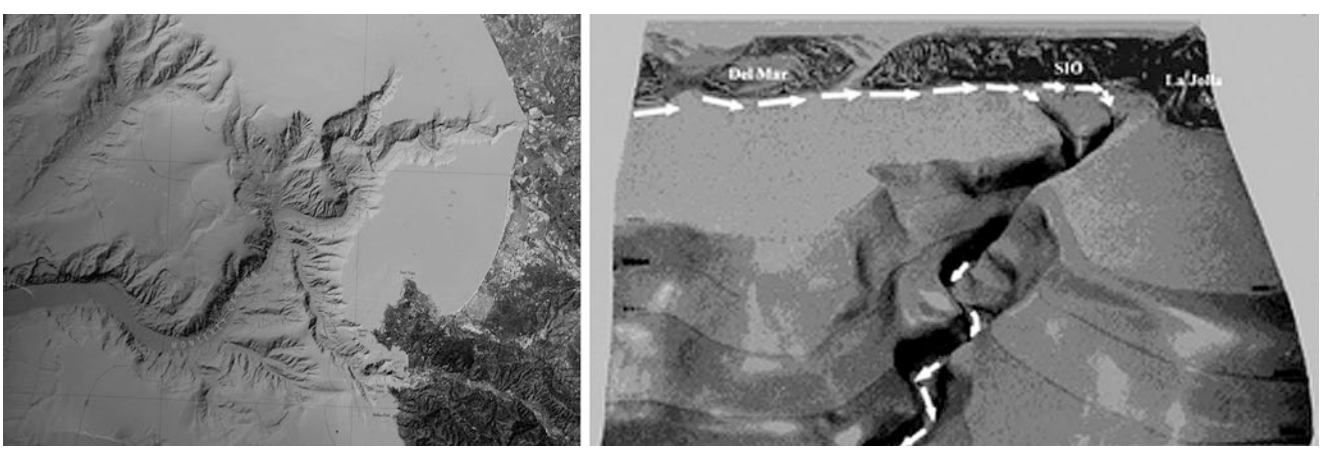

Fig. 3.16 Submarine canyons off California. *Left*: NOAA map of Monterey Canyon, in part based on space surveys. *Right*: beach sand moving down La Jolla Canyon, according to concepts of pioneer marine geologists D. Inman, F.P. Shepard, and G. Einsele. Canyon model (right) courtesy S.I.O. Aquarium. *Arrows* indicating sand flow added. Note large differences in size between *left* and *right panel*

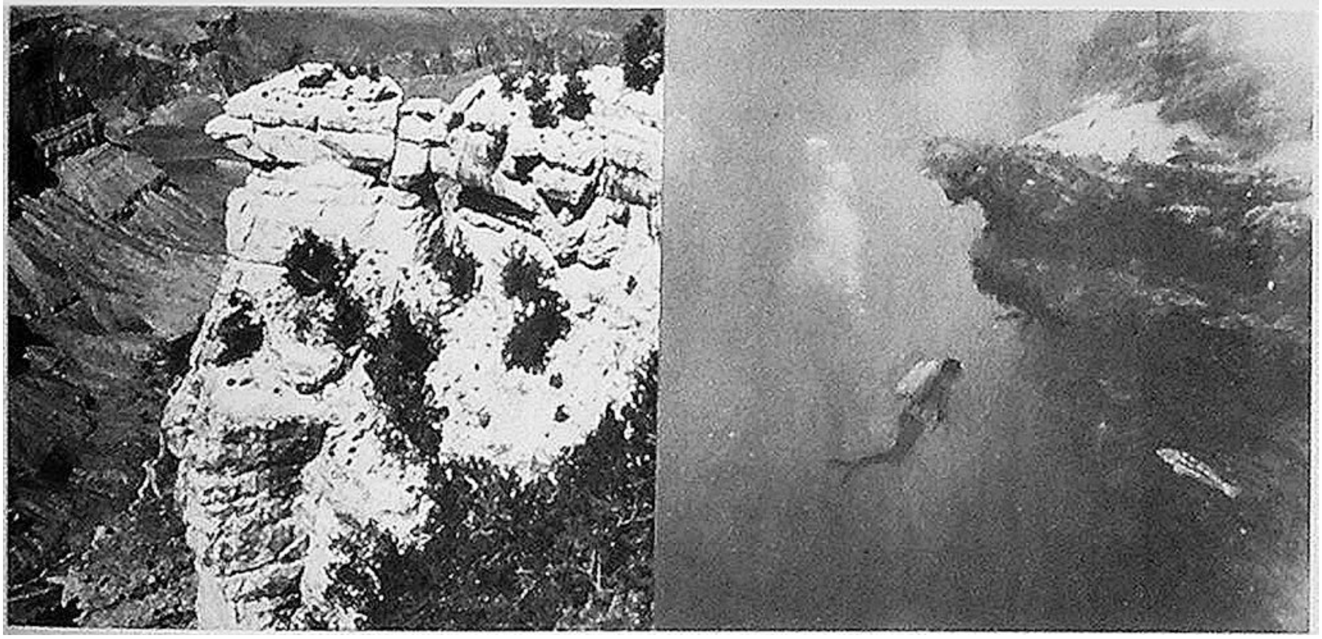

Fig. 3.17 Illustration of morphological similarities of a subaerial canyon (Grand Canyon, *left*) and a submarine canyon (La Jolla Canyon, *right*). Note steepness and overhang in both cases (Photo on left E. S; under-water photo courtesy R.F. Dill)

around 400 ft or some 120 m) by the buildup of high-latitude ice caps, shelves were largely bared and could not act as mud traps for the rich sediment load coming to the sea from the backcountry. Storms and storm waves may have been more frequent or more powerful (or both), making shallow sediments more vulnerable to periodic resuspension with the result of initiating turbidity currents. Even "normal" variations in storm activity presumably would have resulted in such currents.

The realization that turbidity currents play an extremely important role in present-day marine processes as well as in the geologic record, only came in the 1950s, largely thanks to the work of the Dutch geologist Ph. H. Kuenen (1902–1976) (see Sect. 1.3). Kuenen established, by experiment, that muddy downhill flowing currents can exist in nature and that their deposits would exhibit the *grading* familiar from previously unexplained sediment series in the Alps (called "flysch" by the Swiss locals, a term that was adopted as a technical one by geologists).

A standard sequence for graded beds seen in the field was proposed by the Dutch-American geologist A.H. Bouma in 1962. It is referred to as the "*Bouma sequence*." The entire series (from a massive, graded sandy layer at the bottom to the fine-grained shale at the top) is in fact rarely encountered. The common explanation for the sequence calls on a waning turbidity current, with turbidite deposition visible in the lower part of the sequence, and "normal," non-turbidite deposition in the uppermost part of the sequence, the portion that represents almost all of the geologic time captured.

3.8 Deep-Sea Fans and Abyssal Plains

A large part or perhaps most of the sediments in deep-sea fans consist of *turbidites*; that is, deposits of turbidity currents. Turbidites are not only common in continental slopes (the slopes being essentially coalescing fans, Fig. 3.13) but also in *abyssal plains*, perfectly flat portions of the seafloor bearing much material derived from continents. Such plains are largely extensions of fans into low-relief regions dominated by abyssal hills (buried below fine-grained fan material mixed with pelagic sediments). In places, high hills may pierce the cover of flat deposits, betraying original conditions before burial.

Channels are abundant on submarine fans. They are filled or refilled when abandoned. Details of the fan landscape with its distributary system (including channels, levees formed by spillover of muddy flows, slumps, slides, and other features) have become available through side-scan acoustic surveys.

Turbidites are generally recognized in deep-sea seismic profiles taken across abyssal plains as perfectly horizontal reflectors (Fig. 3.18).

Many turbidite layers are thin and are easily destroyed by bottom-living organisms and by bottom currents. Thick layers, of course, can escape such destruction and are then recognizable in the sedimentary record. Thus, thick layers tend to be overrepresented in the geologic records, and the preservation of thin layers says something about the rarity of strong storms and of large burrowers.

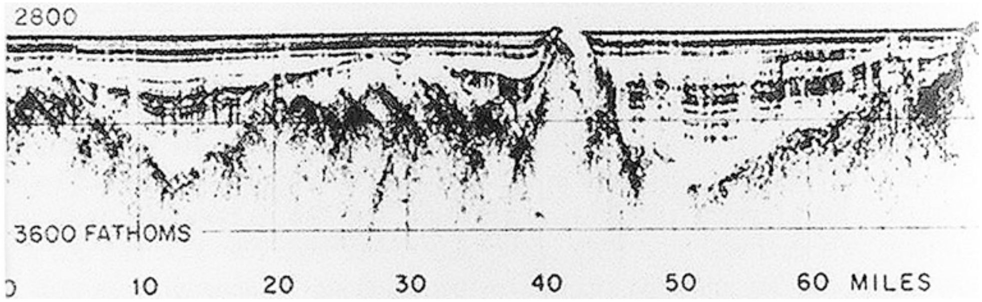

Fig. 3.18 Abyssal plains in a seismic profile. 2800 fathoms = 5100 m; 3600 fathoms = 6600 m (Graph of seismic profile courtesy C.D. Hollister)

The very distribution of abyssal plains supports the idea that turbidity currents run across the seafloor over very long distances. Direct evidence is available from a number of sites, including the turbidite-rich regions off the St. Lawrence River, as well as those off the Hudson, the Mississippi, and the Amazon. Some 3000 km of travel are indicated across the Bengal Fan south of the Ganges-Brahmaputra Delta.

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[PDF]Chapter 4 Continental Margins and Ocean Basins

4 Sources and Composition of Marine Sediments

4.1 The Sediment Cycle

4.1.1 The Concept

Marine sediments consist of deposits accumulating below the sea. They show great variety. There is the debris from the wearing down of continents and volcanic mountains, the shells derived from organisms, organic matter, minerals precipitated from seawater, and there are volcanic products such as ash and pumice. All such matter is transported in various ways, is deposited, and subsequently suffers diagenesis and redeposition to various degrees (Fig. 4.1). Terrigenous muds (with their tremendous fraction of continental erosion products) make up well over one half of the total volume of marine deposits. By area it is deep-sea sediments that are dominant, with roughly one half of the deep seafloor covered by calcareous ooze – largely shell material from coccolithophores and foraminifers, along with some reef debris along many tropical margins. In contrast, extraterrestrial matter, while interesting for its content of information, is negligible when considering abundance.

4.1.2 Notes on Geochemistry

Concerning seafloor geochemistry, reactions of seawater with the hot basalt of the oceanic crust are fundamental. Many of the reactions are conspicuous at the Ridge Crest (*hot vents*). Such reactions may contribute considerable amounts of matter to seawater. Also, they may largely relieve seawater of certain elements, such as magnesium. The types and amounts of elements added or subtracted are the subject of intensive studies, with the origin and evolution of seawater at focus.

As concerns the input to the terrigenous sediment reservoir by erosion, the sea itself takes a toll from the continents: waves and tides can eat into the land, taking debris offshore. Such material, along with the sediment delivered by rivers, by ice, or by winds, may stay on the shelf, or it may bypass the shelf to accumulate on the slope or below that on the rise or the abyssal plain. Sediments that stay on the shelf, of course, essentially stay on the continent. The sediment deposited beyond the shelf eventually is carried back to a continent (e.g., by mountain building) or disappears into the mantle by subduction, as discussed earlier.

4.2 Sources of Sediment

4.2.1 River Input

What rivers carry (the dissolved and particulate load) constitutes the main source of marine sediment making the coalescing fans that build the continental slope off California and elsewhere. The fans are largely riverine mud with an admixture of shell and organic matter. The admixed shell material (largely made of calcium carbonate and opal) also derives to a great extent from river input, that is, from the dissolved load brought by rivers to the sea. We can make a rough estimate of how much material is involved. Slope sediments typically accumulate at a rate of some 100 mm per millennium and deep-sea oozes at rates at one tenth of that, with one half of the deep seafloor ooze-free and accumulating clay at 1 mm/1000 years.

The sediment supply from rivers is reported to be near 12 cubic km per year. If this amount is distributed on the area of seafloor on the planet, we obtain an average sedimentation rate of about 30 mm/1000 years. The erosion rate for continents, at twice that value, would be 60 mm per millennium. The rates of input from reactions of seawater with the basalt of the oceanic crust and from leaching of soils and rocks on land are neglected in this calculation, an input that if considered would decrease the estimate of continental erosion rates somewhat.

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_4

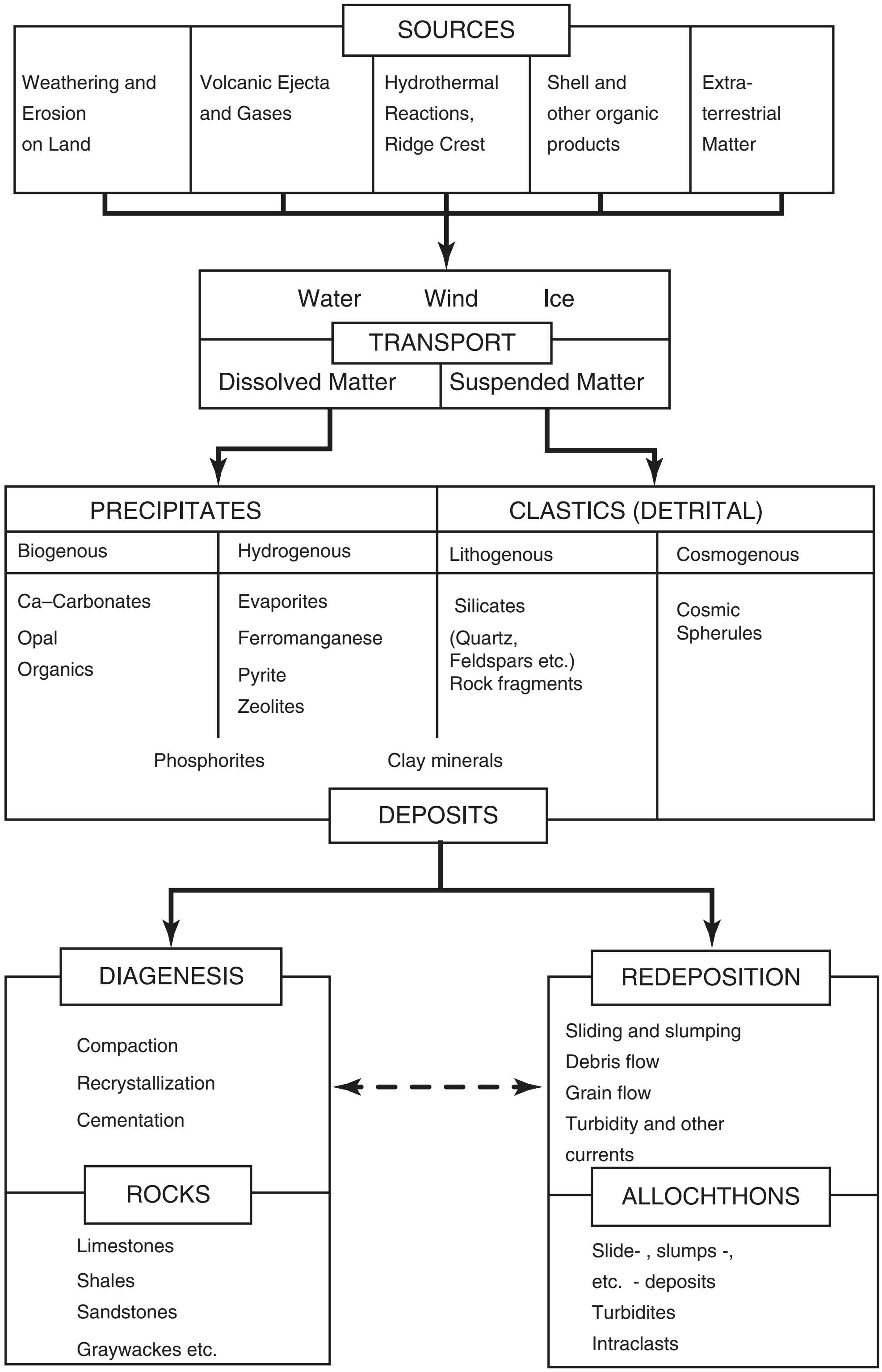

Fig. 4.1 Schematic summary of sources, transport, and destination of marine sediments

In general, mechanical erosion dominates in high latitudes, where much of the water is ice, and in deserts, where water transport is linked to flash floods and other sporadic action unimpeded by vegetation. Chemical weathering (leaching) is favored by rainfall and high temperatures and dominates in tropical areas. In extrapolating these and other present-day patterns into the past, we must remember that we live in a highly unusual period. The growth of mountain ranges and the powerful abrasive action of glaciers have greatly increased mechanical erosion for the last several million years.

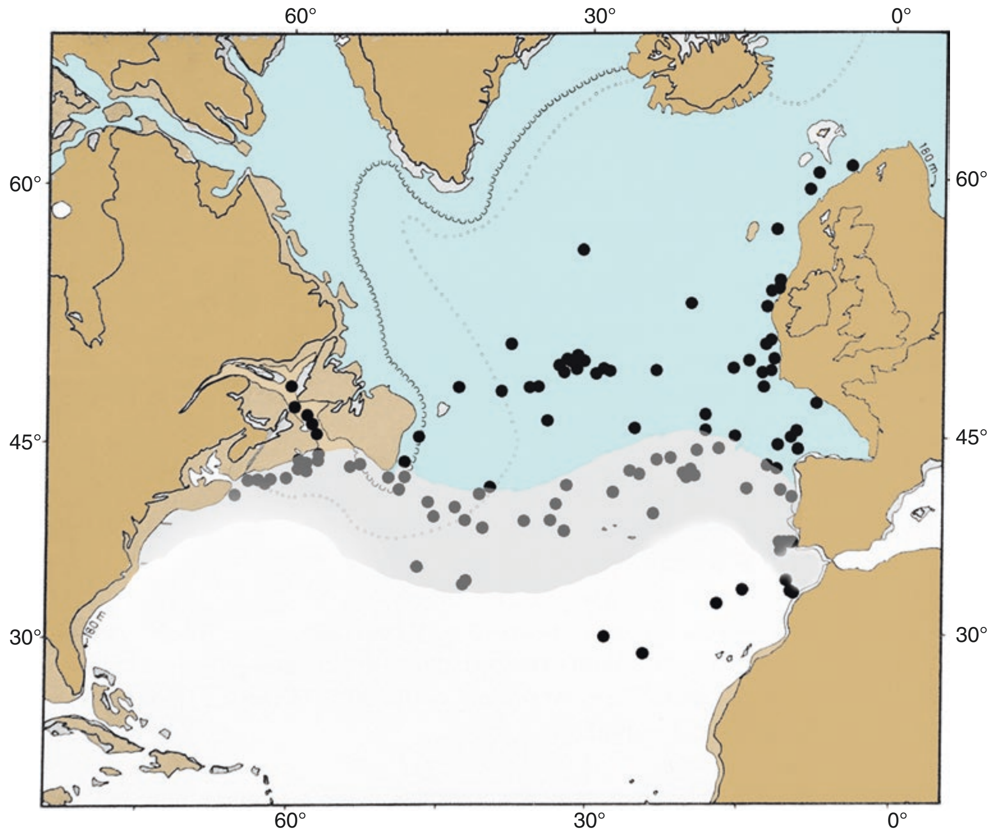

Fig. 4.2 IRD (ice-rafted debris) on the seafloor of the North Atlantic, as recorded by H. R. Kudrass, Bundesanstalt, Hannover. Present limits of drift ice shown schematically (broken lines). Samples of IRD (*black dots*) well toward the south presumably indicate conditions of cold phases of the ice age (suggested southern limit here added: shaded zone) (after H.R. Kudrass, 1973. Meteor Forschungs-ergebnisse Reihe C 13:1, modified). Note the assumed great extent of winter icebergs in the glacial North Atlantic, based on IRD (*blue*). Also note the addition of land as sea level dropped from ice buildup on land

4.2.2 Input from Ice

The great importance of ice in delivering sediment to the sea in high latitudes is readily appreciated when contemplating the immense masses of outwash material glaciers bring to the shores, material that is reworked on the shelves by wave action and by nearshore currents.

Less important as concerns sediment mass, but very useful in the reconstruction of paleoclimate, is the fact that icebergs from calving glaciers can transport both fine and very coarse materials far out to the sea. Upon melting, an iceberg drops its load (as *ice-rafted debris* or *IRD*). Around Antarctica iceberg transport reaches to about 40°S, well off the continent. At present, about 20% of the seafloor receives at least some ice-transported sediment. During the *last glacial maximum* (*LGM*), as expected, the dropstone limit extended much farther toward the equator than today (Fig. 4.2).

4.2.3 Input from Wind

In striking contrast to ice, wind only moves fine material. Medieval Arabian scientists already noted the dust coming out of the Sahara into the “dark sea” of the Atlantic (Fig. 4.3). In the nineteenth century, Charles Darwin

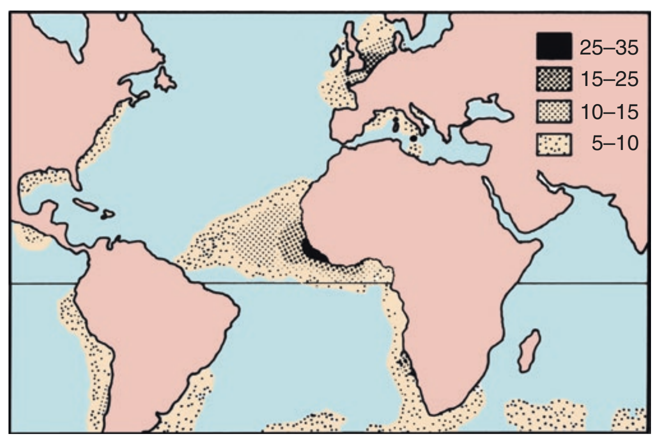

Fig. 4.3 Abundance of haze from dust over the Atlantic (G.O.S. Arrhenius, 1963. In M.N. Hill (ed.) The Sea vol. 3: 695; color here added)

assumed (correctly) that the dust blown offshore off NW Africa must end up on the seafloor. Particles in a large Saharan dust storm in 1901 had an average size of about 0.012 mm in Palermo. Sizes of particles were about half of that in Hamburg (i.e., Hamburg was reached by extremely fine silt, the rest having settled out on the way). During this same storm up to 11 g of dust per cubic meter were measured over the Mediterranean. Some Saharan dust ends up in the Caribbean, as noted in satellite images. In the Pacific,

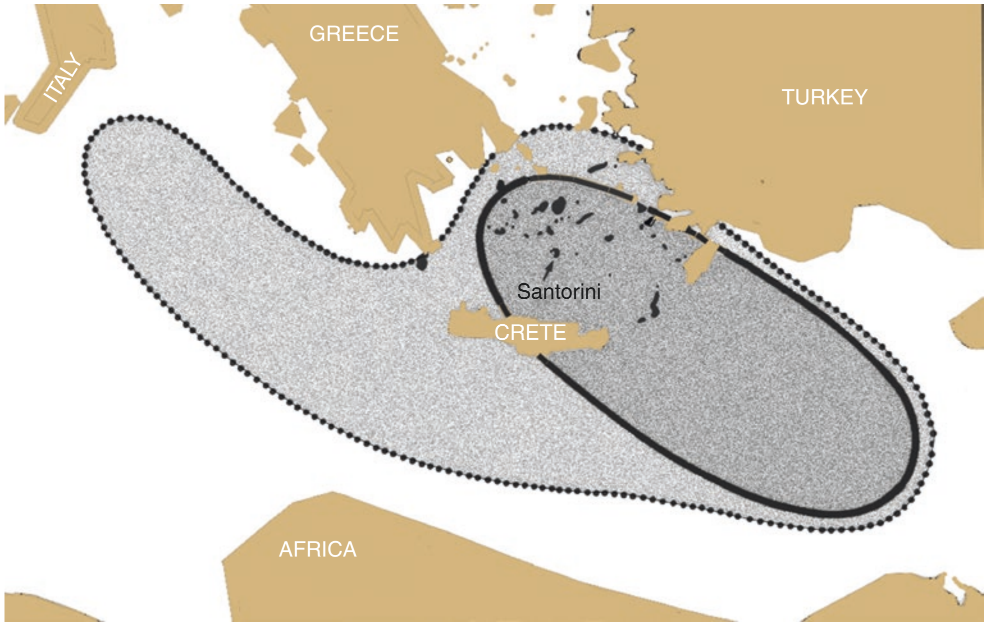

Fig. 4.4 Distribution of volcanic ash on the seafloor of the Mediterranean near Crete; ash is produced by two large explosions in the Aegean Sea (presumably Thera and Santorini). The lower ash layer marks a prehistoric event (>25,000 years). The upper layer is less than 5000 years old and possibly reflects an eruption that ended the Minoan culture (After D. Ninkovich and B.C. Heezen, from K.K. Turekian, 1968. Oceans. Prentice-Hall, New Jersey. Simplified topography; color here added)

we have the example of quartz on and around Hawaii, fine grains apparently blown in from sources in China, thousands of km away.

4.2.4 Input from Volcanism

Volcanoes supply a substantial amount of marine sediment, especially in the vicinity of active margins, such as the ones around much of the Pacific basin. While much of the "*volcanic ash" c*arried by winds is widely dispersed and makes thin layers, layers several cm thick are not uncommon in deep-sea sediments. Some of these, marking periods of major eruptions, *correlat*e over large distances (i.e., they originate from the same event and are of equal age, therefore). A well-known example is the Toba super-eruption in Sumatra (Indonesia) 73,500 years ago. Volcanic products that reach the stratosphere result in strong cooling. In the Toba case, the effect presumably was severe and lasted for several years. It has been suggested that such cooling may have contributed to the fast re-glaciation at the end of the last interglacial. While we may indeed be looking at tectonic feedback in climate change (re-glaciation may have changed the distribution of gravitational forces on the planet and thereby helped trigger eruptions), suggestions of a link between volcanism and re-glaciation are somewhat ad hoc and superfluous in a Milankovitch setting.

Near island arcs, volcanic ash layers (*tephras*) can build sediment aprons several thousand meters thick from tens of thousands of eruptions. The composition of deep-sea clay (Chap. 10) suggests that in the time before ten million years ago, before vigorous mountain building and glaciation changed the planetary environment in drastic fashion, the main source of deep-sea clay in the Pacific was the decomposition of volcanic ash. Not all volcanogenic sediment is brought by wind. *Pumice*, highly porous volcanogenic rock with plenty of air within, can float and move with ocean currents for long distances, even carrying gooseneck barnacles and other organisms. Of course, volcanic material also is eroded on land and is then brought into the sea as terrigenous sediment. Differences in color, mineral composition, glass properties, as well as subtle differences in chemistry hold the clues as to the source of a given volcanogenic deposit.

Geologically, volcanoes have a short life, with single eruptions marking flash-like events. Ash layers, therefore, are very useful in much of regional stratigraphy (as *tephrochronology*), for example, in the Mediterranean realm (Fig. 4.4) or in the vicinity of Iceland. Volcanic activity also adds gases and hydrothermal solutions to the sea. These fluxes have an important bearing on the evolution of the chemistry of seawater and of the atmosphere. As mentioned earlier, they are the subject of intense study.

4.3 Sediments and Seawater Chemistry

4.3.1 Acid-Base Titration

To a first approximation, seawater is a solution of sodium chloride (with a bit of Epsom salt thrown in). Sodium and chloride make up 86% of the ions present by weight. The other major ions are magnesium, calcium, and potassium (alkaline and earth alkaline metals) and the acid radicals sulfate and bicarbonate. The major cations form strong bases that are balanced, on the whole, by the acid radicals. However, bicarbonate forms a weak acid, and seawater is slightly alkaline, therefore, with a "pH" near 8, slightly basic (neutral: pH = 7). On the whole, the salty ocean may be understood as the end product of emission of acid gases from volcanoes (hydrochloric, sulfuric, and carbonic acid) and the leaching of common silicate rocks of oceanic and continental crust, the rocks having minerals of the form [Me Si_a Al_b O_c], where Me stands for the metals Na, K, Mg, and Ca and the remainder makes insoluble silica-aluminum oxides, that is, clay minerals.

How stable was the composition of seawater through geologic time? If we use the above concept of acid-base titration and assume that seawater is a solution in equilibrium with the sediments on the seafloor, the result is that the composition was rather stable. Paleontologic evidence certainly agrees with this assessment. Already in the early Paleozoic, there are organisms in various groups whose closest modern relatives have rather narrow salt tolerances: radiolarians, corals, brachiopods, cephalopods, and echinoderms. Naturally, we cannot rule out adaptation of the organisms in question to a changing salt content of the sea. There is no guarantee that the ratios of leached rocks did not change considerably through time. Thus, it is quite likely that the composition of seawater did change, contrary to the equilibrium argument, which implicitly assumes that present conditions are a permanent feature of the sea.

Comparison of solutes in river water and in seawater suggests that siliceous acid is removed from river water as it enters the sea. Diatom production is especially high at river mouths. This is one aspect of the observation that the ratios in the river influx of dissolved matter to the sea tend to be irrelevant to the makeup of sea salt. In the simplest terms, very soluble salts are abundant in seawater and the others are not. For example, iron is removed rapidly and efficiently, keeping concentrations very low. The implication is that the common iron compounds (hydroxides and sulfides) are not very soluble.

4.3.2 Interstitial Water and Diagenesis

Fine-grained sediments (clays and silts) have porosities (i.e., water content) of 70–90% by volume when first deposited on the seafloor, while sands have around 50%. As the sediments are buried, pore space is reduced by compression. The water expelled does not necessarily have the same composition as the water trapped originally. Instead, the expelled water has contents reflecting reactions within the sediment, reactions that correspondingly change the chemistry of the solids accumulating. Compaction and chemical reactions involving pore fluids (or solids only, as in recrystallization) constitute *diagenesis*, the process that ultimately transforms sediment into rocks.

Typically, diagenesis is most active in the uppermost meter or so of freshly deposited sediments, and the chemical reactions within the sediment are commonly driven by the reactions of organic matter. In fact, *redox* reactions usually dominate the process (called "early diagenesis"). They depend greatly on the organic carbon present. The oxidation of organic carbon leads to removal of dissolved oxygen from pore waters. Additional oxygen demand, after such removal, is satisfied by stripping oxygen from dissolved nitrate and from solid iron oxides and hydroxides (e.g., coatings on grains). If the demand is strong enough, sulfate also is stripped of its oxygen, resulting in an abundance of hydrogen sulfide (which produces the foul-smelling hydrogen sulfide, as well as the ubiquitous gold-colored iron sulfide or "fool's gold"). The redox reactions within sediments are mediated by bacteria and archaea, which are in evidence even deeply below the surface (Fig. 1.11), especially in the organic-rich deposits below the coastal ocean.

Once the sediment is buried, concentrations within the interstitial waters can increase to the point where re-precipitation must occur. This leads to *cementation* of remaining grains by carbonate and silica cements (most commonly calcium carbonate and also quartz with microscopic grains). These various processes, including recrystallization (crystal growth within preexisting solids), can be reconstructed by studying the distribution of the elements and compounds involved, within the solids and the interstitial waters. The distribution of certain isotopes (oxygen, carbon, strontium, and others) is of special interest in this context, because dissolution, migration, and re-precipitation under various conditions can result in altered ratios of the isotopes. Diagenetic processes are extremely important for making hydrocarbon source rocks, of course, and for migration and for reservoir rock porosity and permeability (see Chap. 14).

4.3.3 Methane

Escaping pore waters carry information from the redox reactions: they are enriched in gases such as methane (from fermentation of organic compounds), carbon dioxide (from plain oxidation of organic matter), ammonia (from reaction of water with oxygen-stripped nitrate). *Methane* (*Earth gas*) can react with water (given low temperature and high pressure) to make *methane clathrates* (Fig. 4.5).

The stability of methane clathrate depends on temperature and pressure; hence the temperature below the seafloor is of enormous importance in defining the depth at which the hydrate becomes unstable. Correspondingly, the relevant graph (Fig. 4.5, right panel) emphasizes that hydrates will not be found at great depth within certain sediment stacks, owing to the rising temperature below the seafloor. Since methane implies the presence of large amounts of organic matter for fermentation, the marine methane is likely to be found fairly close to continents (in the coastal ocean); that is,

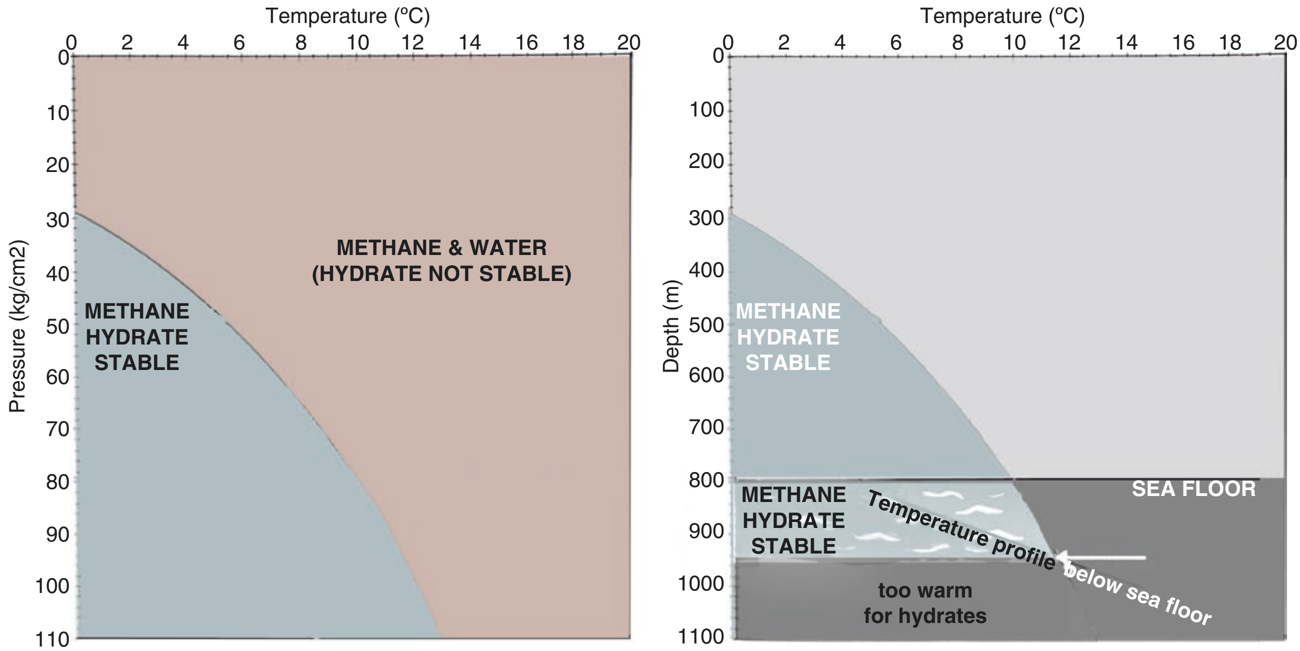

Fig. 4.5 Stability field of methane ice (*left panel*, **a**) and application to sedimentary deposits (*right panel*, **b**). A bottom-simulating reflector (BSR) can occur at the bottom of the layer bearing methane hydrate. At the BSR sound is strongly reflected. The distribution of clathrate is restricted; its presence requires appropriate temperature and pressure to ensure survival if clathrate is present, in addition to a source of methane (stability field and example of occurrence after Erwin Suess (Kiel and Corvallis) and Gerhard Bohrmann (erstwhile Kiel, now Bremen) in G. Wefer and F. Schmieder (eds.) 2010. Expedition Erde (3rd ed.), MARUM, Bremen Univ. Here modified for clarity). A high supply of organic matter is necessary to make methane. For clarity, the clathrate presumably forms within slope sediments, not in water

much of the seafloor likely does not have it. Nevertheless, the amounts of carbon fixed in methane ice are thought to be enormous, exceeding the carbon in the world's estimated coal reserves.

The evidence for methane storage within sediments includes pieces of *methane ice* (burning when lit, in spite of its icy nature; Fig. 1.10), *cold seeps* with its unusual fauna and microbes, *bottom-simulating reflectors* (*BSRs*), and chlorine-poor water from melting, occasionally obtained during drilling. In addition, escaping methane can produce *mud volcanoes* on the seafloor (Fig. 4.6), features that are identified in seismic profiles or by acoustic side scanning. Sizes of the gas-produced mud volcanoes vary; smaller ones are known as *pock marks*.

Besides redox reactions, the dissolution (and also the re-precipitation) of carbonate and of opal is of prime importance in diagenesis. During *early diagenesis* much of the dissolved matter can leave with the escaping pore water or depart simply by diffusion out of the sediment into the overlying water.

4.3.4 Residence Time

For steady state conditions, output must equal input of a geochemical system. Thus, if seawater composition is to stay constant, seawater has to rid itself of all new salts coming in, in a "sink." Where then are the *sinks* for this material? We might first look to the sediments for maintaining the balance, but this is not the whole story. In fact, the quantitative assessment of sinks is a major geochemical problem. For calcium carbonate the sink indeed is largely calcareous shells and skeletons built by organisms, and for silica it apparently is opaline skeletons. Metals presumably leave the ocean largely in newly formed minerals such as authigenic clay (produced in place), oxides, sulfides, and zeolites, as well as in products resulting from reactions between hot basalt and seawater near the crest of the mid-ocean ridge. Sulfur is precipitated in heavy metal sulfides in anaerobic sediments near the land and as gypsum in restricted basins. Some salt leaves with pore waters in sediment. Under the assumption that seawater maintains its composition, we can calculate the average time a seawater component remains in the sea before going out. This time is called *residence time.*

Calculating this time is analogous to figuring out how long people are staying in a museum from counting the people present and the number of people entering per unit time. The ratio is the average viewing time (commonly between half an hour and somewhat greater than an hour, within a museum):

$$t = A / r, \tag{4.1}$$

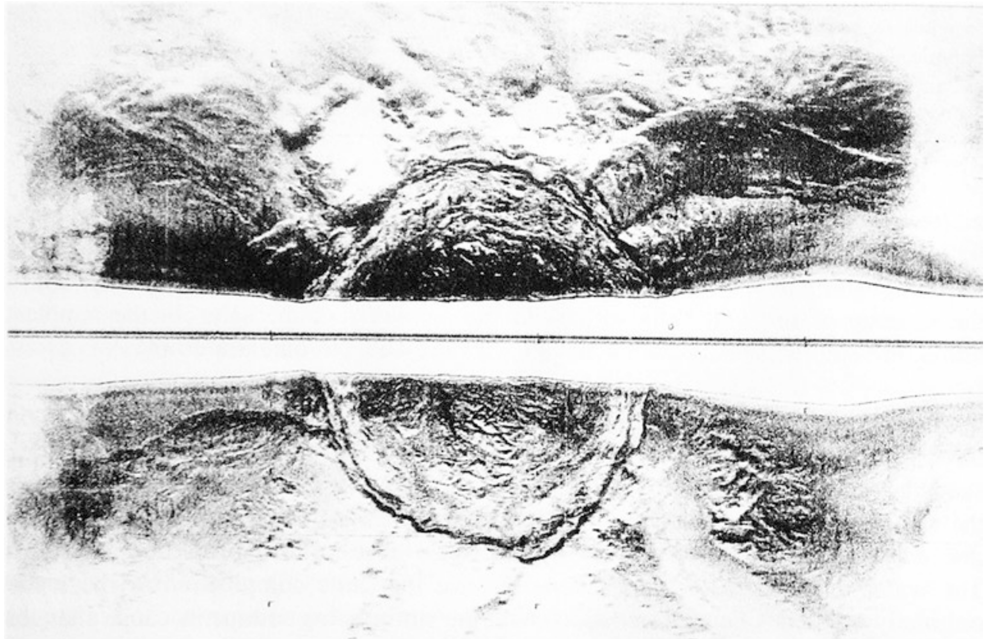

Fig. 4.6 Mud volcano in an acoustic side-scan image. Side-scan record is from the bottom of the Black Sea (Courtesy Dr. Glunow, Moscow; see UNESCO-IMS Newsletter 61, Paris)

where A is the number present and r is the rate of input. A successful exhibit has a large t. In the sea, the residence time is, in essence, a measure of geochemical reactivity. Sodium and chloride have a very long residence time, while silica has a short one. Sodium, having found its partner chloride, tends to stay in solution, while silica is readily precipitated and becomes close to inert as a consequence of diagenesis. Equation (4.1) was once used to calculate a *salt age* for the ocean, assuming that the ocean started out as a freshwater body and retained the sodium added. It was then a useful concept as an estimate for the scale of geologic time. The result came out near 100 million years, much longer than many other estimates and closer to the truth, therefore. However, it was still very short of reality: the calculation was flawed. (Earth is more than 40 times older than 100 million years; the age of the ocean remains unknown, but presumably exceeds a factor of 30 over the salt-age guess, based on various clues.)

4.4 Major Sediment Types

4.4.1 General

There are essentially three types of marine sediments: those that come into the ocean as particles, are dispersed, and settle to the seafloor, those that are precipitated out of solution inorganically, and those that are precipitated by organisms. We call the first type *lithogenous*, the second hydrogenous, and the third biogenous (see Box 4.1: Classification of Marine Sediments).

Around the ocean margins, lithogenous sediments are predominant, their source being largely mechanical weathering of continental rocks. Also, there are salt deposits here (hydrogenous) and biogenous sediments, but in relatively low abundance. On the deep seafloor, on the other hand, biogenous sediments ("oozes") dominate (Fig. 4.7), with *calcareous ooze* being much more common than *siliceous ooze.* In fact, calcareous ooze covers roughly one half of the seafloor. Defining components are (silt-sized) coccoliths and (sand-sized) foraminifers and very small mollusks for the calcareous ooze and (silt- to sand-sized) diatoms and radiolarians for the siliceous ooze. The organisms involved are eukaryotic microbes, with coccoliths and diatoms photosynthesizing (i.e., these are microbial algae).

4.4.2 Sand

Marine sediment particles mainly come in three sizes: sand, silt, and clay (Box 4.1, Fig. 5.4; Appendix A5). To geologists "sand" denotes solid particles between 0.064 and 2 mm in diameter, no matter what it is. Under the microscope, we can see that beach sand in Southern California (Fig. 4.8) and in most places elsewhere mostly consists of mineral grains, small pieces of rock, and shell fragments. We have known for more than half a century that sand grains are readily transported far offshore into the realm of continental fans and continental rises by turbidity currents running down submarine canyons, as discussed in the previous chapter. The same process works with sediments of the deep sea, of course, as long as the seafloor is sloped, as along the flanks of the MOR, except that we largely deal with oozes here that are being displaced. Foraminifers are largely sand (the smallest ones are silt size).

Beach sediments are mainly mixed lithogenous and biogenous.

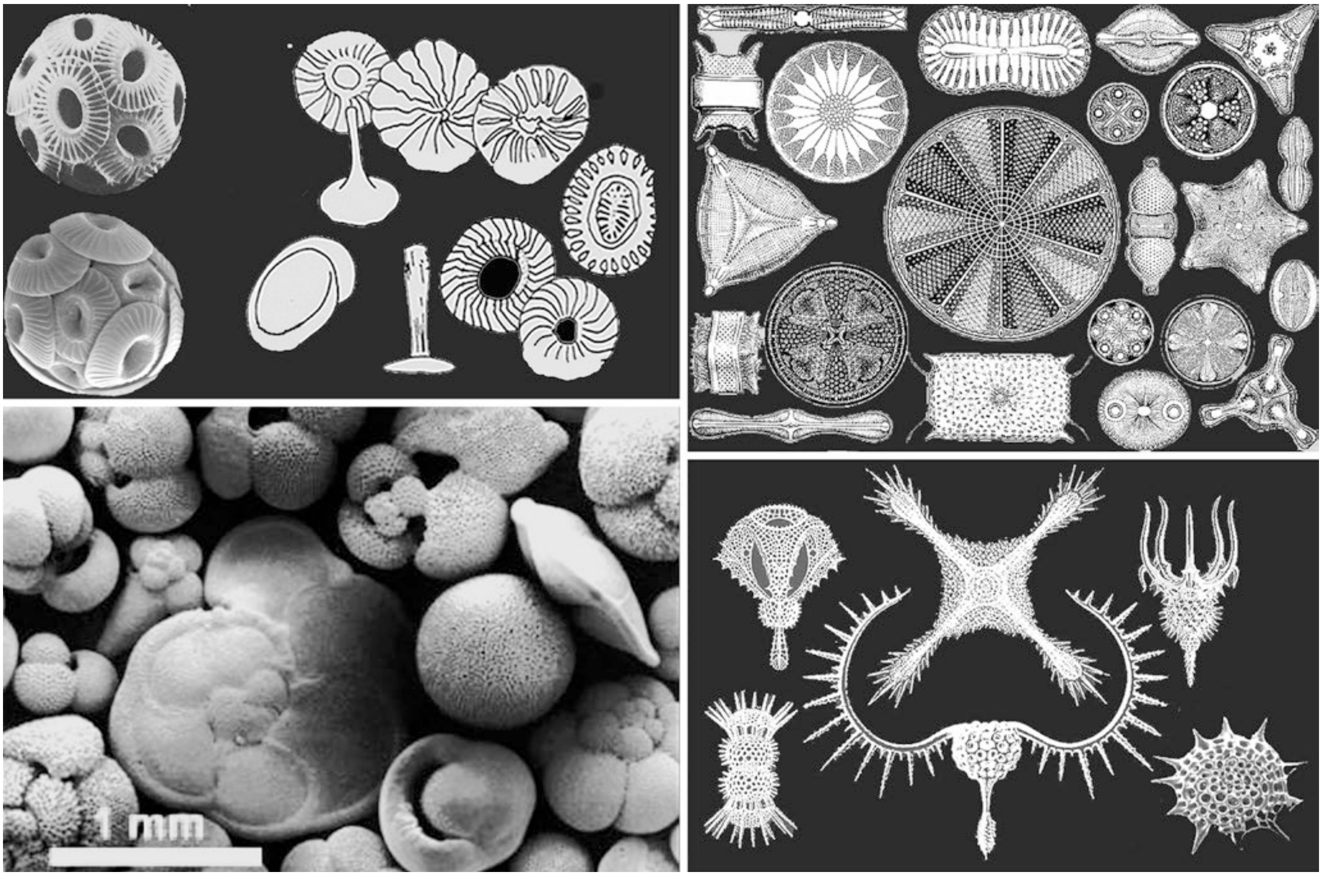

Fig. 4.7 Components of "ooze." (*Left side*; calcareous ooze; *upper panel*: coccolithophores and coccoliths (The former courtesy G. Wefer, and R. Norris; the latter mainly after A. McIntyre, Lamont); *lower panel*: foraminifers (SEM by M. Yasuda, S.I.O.). *Right*: components of siliceous ooze (diatoms and radiolarians; from E. Haeckel, 1904). Colors: calc. particles buff to very *light gray*; diatoms *greenish brown*; radiolarians glassy. Sizes: silt, except the foraminifers, many of which are sand. Coccoliths commonly are extremely fine silt

Fig. 4.8 Beach off S.I.O., in La Jolla. *Left*: pebbles. *Right*: sand grains. The beach sand has an abundance of (glassy) quartz grains (proportionally many more than produced by erosion). Quartz is resistant to chemical attack and to abrasion, other common minerals less so. The pebbles are igneous rocks from the cliffs (Photo of pebbles W.H.B.; microphoto courtesy of P.A. Anderson, S.I. O)

4.4.3 Classification

Box 4.1 Classification of Marine Sediment Types *Lithogenous sediments*. Particles derived from preexisting rocks and volcanic ejecta. Nomenclature based on grain size and various properties including composition, structure, and color. Typical examples, with most common environment in parentheses:

- Organic-rich *clayey silt* with root fragments (marsh)
- Finely laminated *sandy silt* with small shells (delta-top)
- Laminated quartzose *sand*, well sorted (beach)
- Olive-green homogeneous *mud* rich in diatom debris (upper continental slope)

(*Mud* is the same as *terrigenous clayey silt* or *silty clay*, commonly with some sand.)

Fine-grained lithogenous sediments are the most abundant by volume of all marine sediment types (about two thirds), largely because of the great thickness of sediment in continental margins.

Biogenous sediments. Remains of organisms, mainly skeletal parts (calcium carbonate from mollusks, calcareous algae, coral, foraminifers, coccolithophorids, etc.), hydrated silica from diatoms and radiolarians, and calcium phosphate from arthropods and vertebrates. *Organic* sediments, while strictly speaking biogenous, are commonly considered separately. Arrival is as particles (some precipitated in situ) or in aggregates, dispersal by waves and currents. Redissolution is common, both on the seafloor and within the sediment. Appellation is by organism source and by chemical composition and by other properties. Examples:

- Oyster bank (lagoon or embayment)
- Shell sand (tropical beach)
- Coral reef breccia (fragments, reefal debris)
- Oolite sand, well sorted (strand zone, Bahamas)
- Light gray to buff calcareous ooze, bioturbated (deep seafloor)
- Greenish-gray siliceous ooze (deep seafloor)

Biogenous sediments are widespread on the seafloor, covering about one half of the shelves and more than one half of the deep ocean bottom, for a total of around 55%. About 30% of the volume of marine sediments being deposited at the present time may be labeled biogenous, although there may be considerable admixture of lithogenous material, in part volcanogenic.

Hydrogenous sediments. Precipitates from seawater or from interstitial water, especially within freshly deposited sediments (*early diagenesis*). Redissolution is common. Nomenclature is based on origin and chemical composition, as well as some other obvious properties. Examples:

- Laminated translucent halite (salt flat)
- Finely bedded anhydrite {i.e., calcium sulfate} (Mediterranean basin, subsurface)
- Nodular grayish white anhydrite (ditto)
- Manganese nodule, black, mammilated, 5-cm diameter (deep seafloor, Pacific)
- Phosphatic concretion, irregular slab, 5-cm thick, 15-cm diam., light brown to greenish, and granular (upwelling area, upper continental slope)

Hydrogenous sediments are widespread but not important by volume at present.

4.5 Lithogenous Sediments

4.5.1 Grain Size

The bulk of sediment around the continents consists of debris washed off the elevated areas of the continents. The debris results from mechanical breakup with or without chemical attack on continental igneous and sedimentary rocks. Transportation of the resulting rock fragments and minerals is by rivers. Thus, the process is part of the hydrological cycle (Fig. 4.9).

Grain size is extremely important for assessing source and transport processes. One distinguishes *gravel, sand, silt,* and *clay* (see Hjulström Diagram; Fig. 4.3). *Sand*, as mentioned, is all solid material of a size between 0.063 and 2 mm, regardless of composition or origin. Silt is the next smaller size category, at sizes between 0.063 and 0.004 mm (see Appendix A5). The next smaller category, clay, has particles smaller than 0.004 mm (or 4 µm). In some scales 2 µm is taken as the upper clay limit. The term "clay" is somewhat confusing, *clay minerals* being sheetlike minerals of a certain type regardless of size (albeit being common within the size category of "clay"). For clarity, "pebble," "sand," "silt," and "clay" are *size* categories to a geologist. The material needs naming if it is to be specified.

Overall there is a gradation of grain size from source to place of deposition, with coarser particles (including boulders) closer to the source and clay-sized material far away, commonly carried off by water and wind. *Gravel-size* (2–256 mm) and *boulder*-size (>256 mm) material does not commonly travel far, except when taken along by ice (which is hardly sensitive to the size of the load and thus gives rise to *erratics*, that is, boulders without an obvious source in

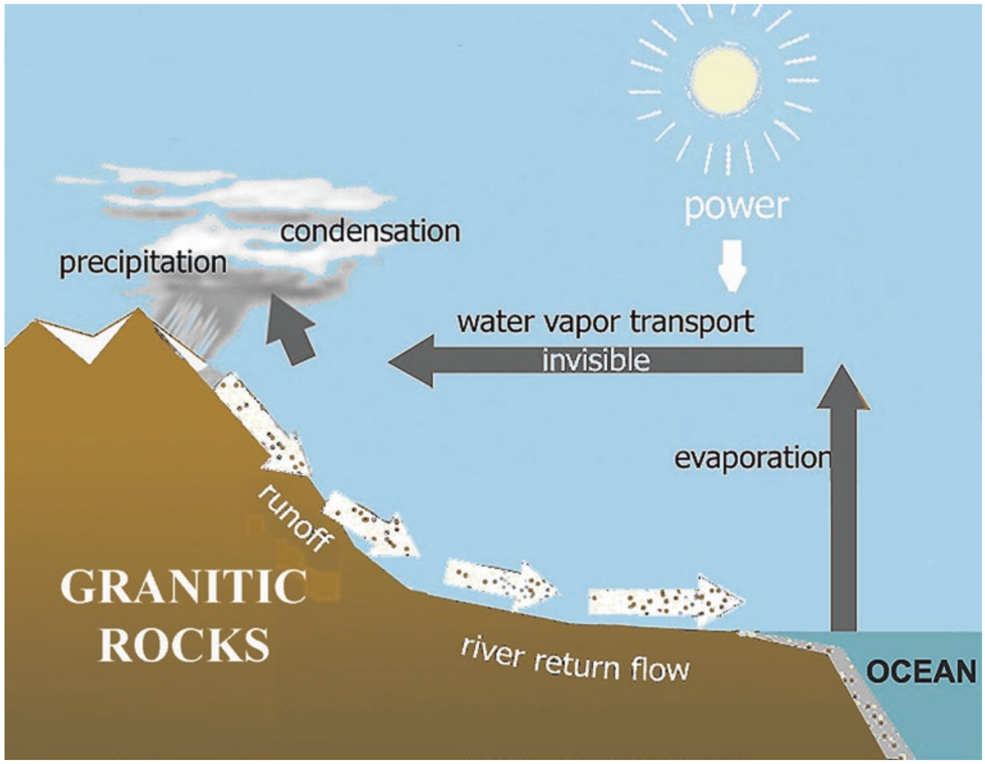

Fig. 4.9 Origin of lithogenous sediments. Weathering of source rock and river transport as part of the hydrologic cycle. The situation schematically depicted is typical for the western coast in North America, as well as for many other places where erosion of mountains delivers sediments to the sea (W.H.B., 2013. San Elijo Lagoon, UCSD, modified)

strange surroundings. Except around reefs and in high latitudes, boulders and gravel are not common in marine sediments.

4.5.2 Lithic Sand

Lithogenous sands are typical of certain beach and shelf deposits, as in Southern California, for example (Fig. 4.8). The sand may consist of minerals (commonly dominated by the resistant quartz) or of rock fragments, as is the case for volcanic material. A striking example of the latter is provided by black beach sands in the Hawaiian Islands (Fig. 4.10). Also, green mineral sand of volcanic derivation can be found on occasion. Other volcanic islands such as Iceland and the Galapagos Islands also have dark gray lithogenous beaches – light brown beaches are largely of continental origin, while white beaches are typical for carbonate environments.

The source areas and dispersal history of sands can be explored by noting the compositional types of *heavy minerals* (densities >2.8 g/cm^3; examples are the silicates hornblende, pyroxene, and olivine; also the iron-rich compounds magnetite and hematite; and the titanium-bearing ilmenite and rutile (see Appendix A4)). The heavy mineral association allows the mapping of depositional provinces. In turn, such mapping provides clues to the action of shelf current wave climate and other factors.

The *shape* of sand grains has been used to obtain clues about their origins. For example, sharp edges have been linked to recent mechanical action, while rounding has been ascribed to reworking. Problems arise with recycled sand grains (i.e., *polycyclic sand*) and with post-depositional etching of grains within sediments.

Fig. 4.10 Lithogenic volcanic dark gray beach sand on Hawaii. Note volcanogenic gravel and boulders in background. The young coconut tree presumably brought in by waves as seedling (foreground) is unlikely to survive on the shifting sand (Photo W.H. B)

4.5.3 Lithic Silt

Lithogenous silt is common on the continental slope and rise, although largely mixed with coccoliths and other biogenous particles (Fig. 4.11). The composition of the silt is eas-

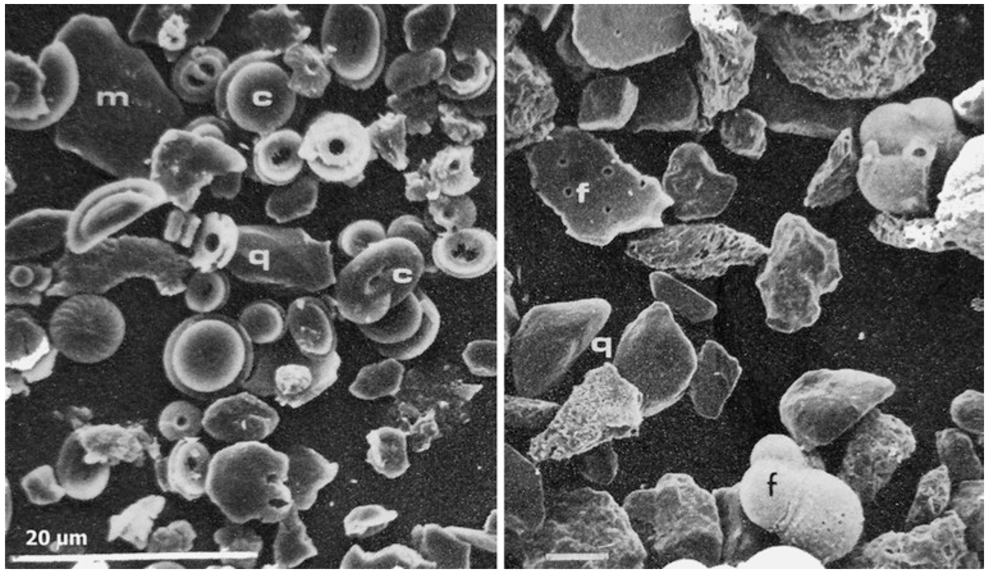

Fig. 4.11 Mixed terrigenous and biogenous sediment makes up the fine silt fraction (2–6 µm) on the continental rise off Cape Verde (NW Africa). The medium-size silt (*right panel*) likewise consists of a mixture. *c* coccoliths, *f* foram shell, *m* mica, *q* quartz (SEM photos courtesy D. Fütterer, then Kiel)

ily rationalized by comparing with adjacent sand deposits and with associated clay. Mica, a platy terrestrial mineral delivered both by igneous and ancient sedimentary rocks, is commonly especially well represented.

Sand is commonly studied with a binocular microscope, while the composition of clay is investigated by X-ray diffraction and other sophisticated methods. The study of silt used to fall into the crack between the methods. It has had a rather low popularity rating. For the last several decades, the scanning electron microscope (SEM) has made it possible and attractive to investigate this size fraction in some detail. Not surprisingly, it turned out that the composition of the silts is commonly closely related to that of the associated fine sand fraction. Biogenous contributions (plankton remains of the coastal ocean) can be dominant in places (including diatoms and radiolarians), rather than continental debris (Fig. 4.11).

4.5.4 Clay-Sized Sediment

Clay-sized particles are ubiquitous both on the continental margins and on the deep seafloor. Much like fine silt, the presence of clay, where abundant, indicates *low-energy environments*. Clay is easily transported, although, in places, pickup is hindered by a bacterial mat covering the sediment. In well-oxygenated environments, however, such mats are commonly disturbed by the churning of sediment (*bioturbation*) by larger organisms, if they form at all.

Common *clay minerals* are montmorillonite (or smectite), illite, chlorite, and kaolinite. When mapping their distributions on the deep seafloor, patterns emerge that are readily interpreted in terms of origin and paleoclimate, with links to volcanic activity, metamorphic processes, and physical erosion as well as deep chemical weathering on land (Chap. 10).

The high surface areas of clay particles that come with their minute size give clayey sediments special chemical properties. For example, clays readily absorb a large variety of substances, and many react easily with the ions in seawater and in interstitial waters, establishing chemical equilibrium with their surroundings. During diagenesis within buried sediments, new clay minerals can form at elevated temperature and pressure. Ultimately such reactions have important implications for the chemistry of seawater and for geochemical processes in general.

Clayey sediments in regions of high sedimentation rate are commonly rich in organic matter, partly because organics cling to the clay during deposition and partly because where conditions are quiet enough for the deposition of clay, they also are favorable for the deposition of fluffy organic particles. Much of the clayey matter actually may be brought to the seafloor within fecal pellets of organisms filtering the water. Such clay is associated with food particles. Experience with sediment-trapping equipment (in the Baltic and off California, among other places in the sea) suggests that the *fecal pellet transport* mechanism is a significant agent of sedimentation.

4.6 Biogenous Sediments

4.6.1 Types of Components

Organisms produce sediments in the form of shells and other skeletal materials and organic matter. The label "biogenous" is generally applied only to the hard parts, that is, to calcareous, siliceous, and phosphatic matter. The organisms involved actually include bacteria and archaea (e.g., in producing metal oxides, hydroxides, and sulfides). Easily recognized as fossils are diatoms, radiolarians, silicoflagellates, and primitive multicellular forms such as sponges (hydrated silicon oxide). Coccolithophores, foraminifers, various kinds of algae, mollusks, corals, bryozoans, certain

Fig. 4.12 Biogenous sediments on the beaches in the Hawaiian Islands. *Left*, pebble-size material; *middle*, coarse sand; *right*, beach, dark gray: volcanogenic pebbles among coral-derived pebbles thrown up by storm waves (Photos W.H. B)

brachiopods, arthropods, echinoderms, annelid worms, calcareous sponges, and vertebrate remains (i.e., solids made of calcite, aragonite, Mg-calcite, and those made of calcium phosphate minerals) are especially common.

Perhaps the most abundant and certainly the most conspicuous biogenous types of sediment are delivered by calcareous skeletal parts of organisms. In shallow tropical waters, there are abundant remains of coral and associated materials, depending on location (Fig. 4.12). On the deep seafloor, there are the remains of coccolithophores and of pelagic foraminifers (Fig. 4.7). The shells of benthic foraminifers and coccoliths are ubiquitous and are especially abundant in shallow water. The benthic "forams" show great variety (Fig. 4.13) and are useful therefore in stratigraphy and in environmental studies. We have already pointed out the strong biogenic aspect of low-latitude silt in slope sediments and on the continental rise. In places, the biogenic component can be truly striking, as in certain shell pavement deposits in the "wadden" of the North Sea and in similar tidal flats.

One source of carbonate that is ubiquitous, involving both shelves and the deep seafloor, is a tremendous variety of foraminifers, with hundreds of common species in the benthic types (Fig. 4.13). Pelagic forms, with relatively modest diversity (tens of species), may have arisen several times during geologic history from benthic stock.

For siliceous (opaline) hard parts, the gradient pattern from shallow to deep is similar to the carbonate pattern: remains of benthic organisms dominate in shallow water (sponges, benthic diatoms) and pelagic materials dominate in sediments of deep water (diatoms, radiolarians). Pelagic diatoms are typical for the coastal ocean (overlying shelf and upper slope) and especially for upwelling areas. Likewise, river mouths are prime areas of diatom deposition, especially benthic ones.

In addition to carbonate and silica, phosphatic particles are produced by organisms (including phosphatic fecal pellets). The sedimentation of such particles plays an important role in the phosphorus budget of the biosphere and is of great interest geochemically, therefore. The various other types of hard parts – strontium sulfate, manganese, iron, and aluminum compounds – are intriguing in the context of evolutionary studies but do not materially contribute to marine sediments. Organogenic sediment is important in the context of the productivity of the sea (Chap. 7) and of hydrocarbons (Chap. 14).

4.6.2 Sediment Contributions by Benthic Organisms

The great importance of carbonate-secreting benthos in producing sediment is apparent all through the Phanerozoic record, that is, more than the last half billion years. There is some indication that the less stable carbonate minerals aragonite and Mg-calcite are more abundant during cool and cold conditions (latest Precambrian, latest Paleozoic, Neogene) than during warm time periods such as much of the Cretaceous. At present, the most conspicuous biogenic edifice in the sea is the Great Barrier Reef. It is largely composed of the remains of benthic organisms: corals, calcareous algae, mollusks, and benthic foraminifers. Among the calcareous matter, both calcite and aragonite (same chemical composition, different arrangement of elements) are found. Aragonite is much more soluble than calcite. Also, there is a tendency for higher Mg content in the calcite precipitated in warm water, compared with pure calcite.

Mesozoic and later Cenozoic platform carbonates once were deposited all around the *Tethys,* the ancient tropical seaway that once linked the western Pacific with the central Atlantic. The shallow-water limestone rocks of the geologic record, which originated in the shelf environments of ancient oceans, commonly contain an admixture of siliceous rocks, as

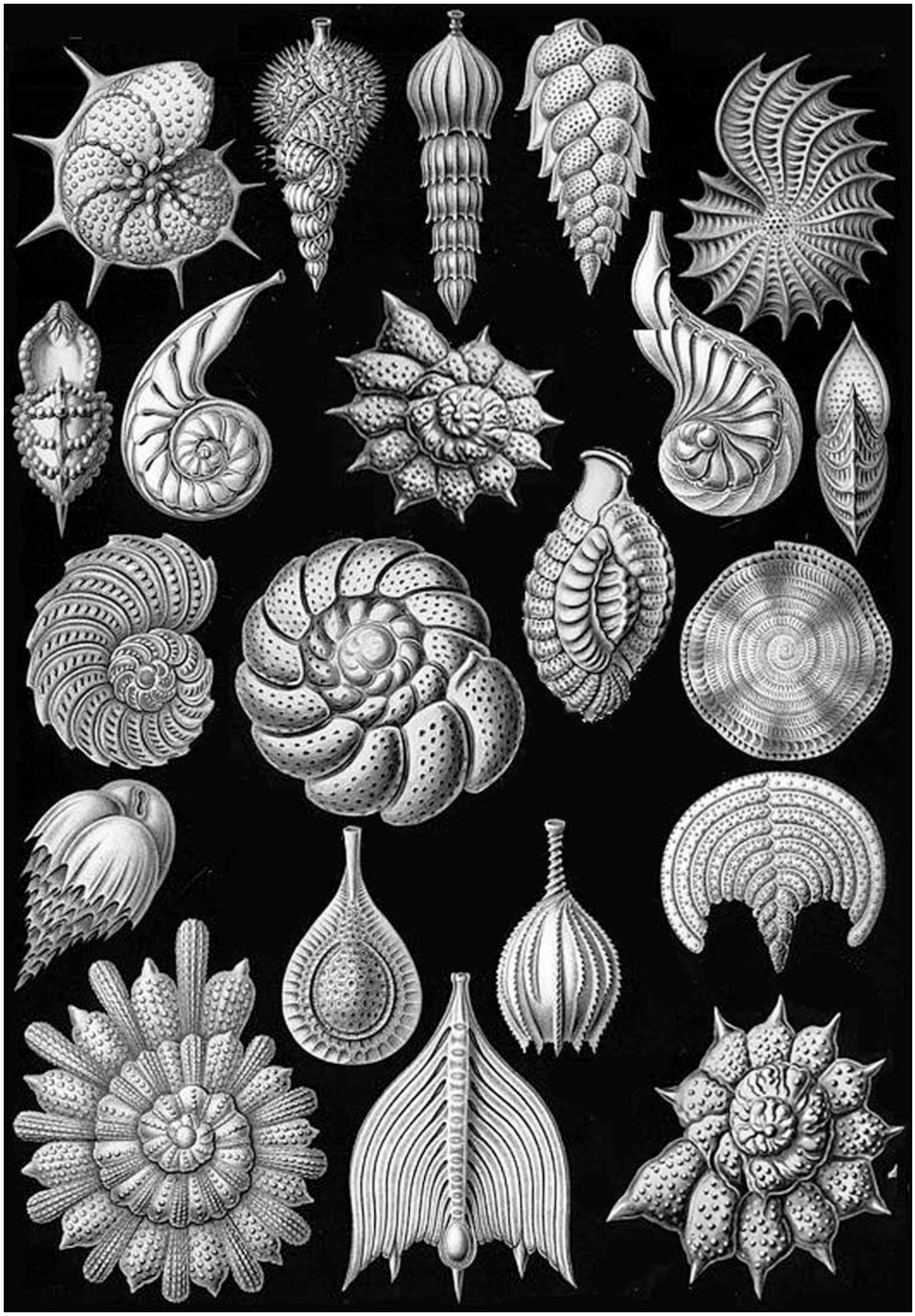

Fig. 4.13 Shells from benthic foraminifers, illustrating the great variety in these forms (From E. Haeckel, 1904. Kunstformen der Natur. Leipzig)

layers of *chert* arranged along horizons parallel to the bedding. The "chert" usually appears as microcrystalline quartz lumps and beds and originated from recrystallization after expulsion of the water from the opal. Any of the silica-producing organisms can be responsible for the origin of a given piece of chert, even though some fossils (e.g., sponge remains, radiolarians) may be more conspicuous within it than other fossils.

Why is there little or no evidence for incipient chert formation in modern shelf carbonates?

The answer is not clear. The reason may be that silicate concentrations are comparatively low in present tropical waters. Seawater (especially shallow seawater making shelf carbonates) commonly is highly undersaturated, presumably stripped of its silicate by diatoms in upwelling regions. In fact, a general rarity of chert (as is evident after the Eocene) may be an inevitable consequence of planetary cooling and associated upwelling. In any case, the once popular idea that waning volcanism is to blame for the post-Eocene rarity of chert has been largely abandoned. Eocene chert in deep-sea sediments attracted much attention by deep-sea geologists when deep-sea drilling started, its presence being rather inimical to any drilling.

4.6.3 The Remains of Planktonic Organisms

Remains of pelagic species are much in evidence in slope sediments, as we pointed out earlier when discussing silt and Fig. 4.11. The bulk of deep-sea sediments, to be sure, is rich in planktonic skeletal matter. Shelf sediments as well can contain considerable amounts of planktonic remains. Thus, for example, the English chalk of Cretaceous age is extremely rich in *coccoliths,* that is, particles produced by *coccolithophores* (more commonly addressed as *nannofossils* by geologists*).* Siliceous remains of diatoms, radiolarians, and silicoflagellates (all planktonic and opaline) are characteristic of offshore high-production areas, as mentioned. This general pattern must be added as a clue to distance from the shore to the gradient in the benthic-planktonic ratio in foraminifers that parallels distance from the shore. In deep-sea sediments, planktonic remains (specimens, not species) tend to outnumber benthic ones by ratios of 10:1 to 100:1, the higher values being typical in areas of good preservation.

4.7 Nonskeletal Carbonates

4.7.1 Carbonate Saturation and Precipitation

In the present ocean, carbonate precipitation occurs either within organisms (shells, skeletons, "internal precipitation") or in association with their metabolic activity (e.g., algal crusts, "bacterially mediated precipitation," "external precipitation"). This need not always have been the case: at very high levels of saturation, there might have been inorganic precipitation, which may have produced certain types of limestone seen in the geologic record. Where would one look for possible inorganic precipitation today? One would need to look in regions of unusually high carbonate saturation. Seawater that spontaneously precipitates a mineral is said to be *supersaturated* with this mineral phase. In contrast, seawater that dissolves the mineral is *undersaturated. Saturation* obtains at the point of balance between precipitation and dissolution in a solution. For two-ion compounds (such as calcium carbonate), the degree of saturation is expressed as the ratio between the products of observed concentration of the two ions involved and the product of concentrations at saturation. This criterion defines much shallow tropical seawater as being supersaturated with calcium carbonate. Yet spontaneous inorganic precipitation is not observed. Presumably the expected reaction does not occur because of interfering factors, that is, the presence of a blocking agent or several such agents.

In the presence of interference, if inorganic precipitation is to be observed, especially high levels of supersaturation need to be extant (and any blocking agent is to be removed), assuming that appropriate nuclei for crystal formation are sufficiently abundant. Naturally, when saturation is lowered substantially, for example, by the addition of carbon dioxide (in modern times from human energy use), it becomes more difficult to make carbonate sediment or shell. This is, in a nutshell, the effect from the *acidification* of seawater. When carbonate is dissolved, carbon dioxide is used up in the reaction. Conversely, when precipitating carbonate, carbon dioxide is released and any undissolved excess is at least partially expelled from the sea to the atmosphere. Evidently, then, the reactions involving carbonate are of prime importance when discussing changes in the atmospheric concentration of carbon dioxide and associated greenhouse effects on various time scales.

4.7.2 The Bahamas

One place that seemed ideally suited for finding out whether inorganic precipitation is now occurring or not is the archipelago known as "the Bahamas."

The islands are surrounded by some of the warmest and most alkaline waters in the ocean. There is hardly any terrigenous input, so that pure carbonates can accumulate. The Great Bahama Bank is very flat and very shallow over large areas (Fig. 4.14). Evaporation and warming by the sun combine to increase saturation. Ubiquitous benthic algae remove carbon dioxide when growing. Thus, conditions would seem very favorable indeed for the precipitation of carbonate directly from seawater.

Fig. 4.14 Carbonate oolite deposits on the Bahama Banks, seen from the air (Air photo courtesy D.L. Eicher, Boulder, Colorado)

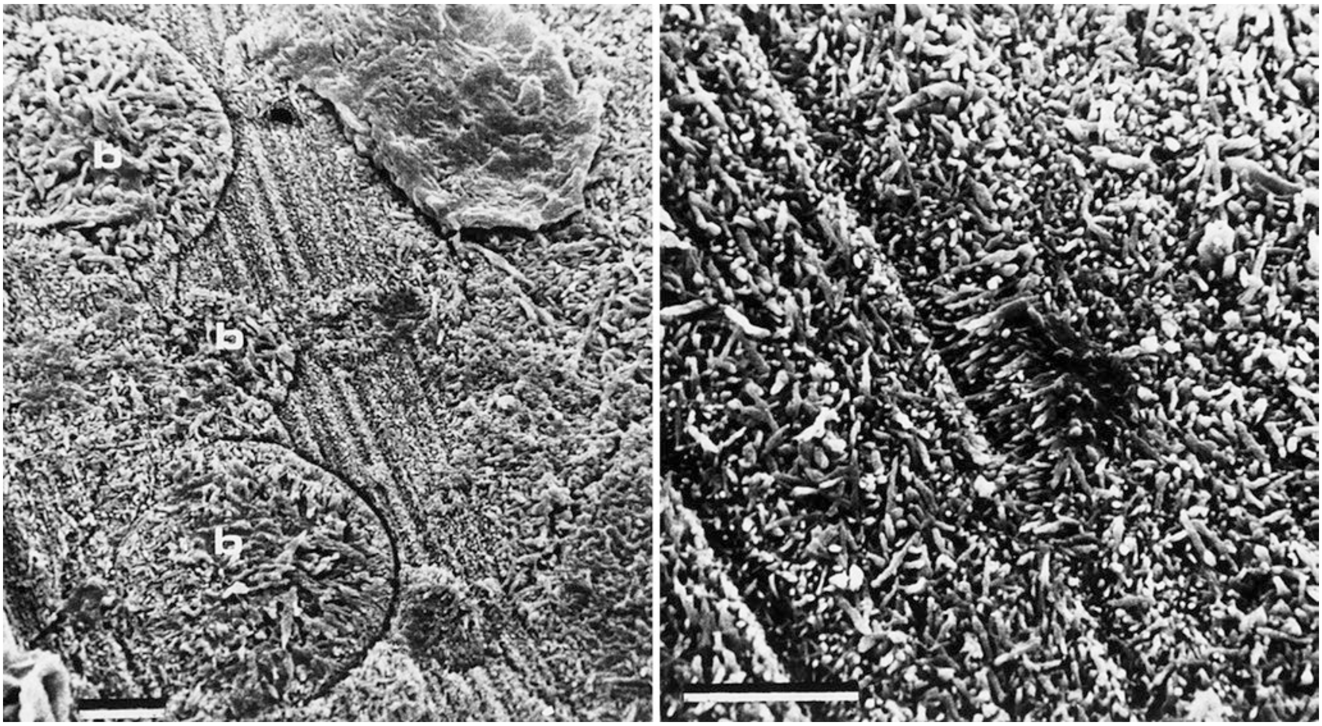

Fig. 4.15 SEM images of ooids, scale bar 5 µm. *Left*, slightly etched section. Three secondarily filled borings (marked "b") intersect the concentric laminae of primary oolitic coating. *Right*, close-up of oolite laminae showing acicular aragonite needles (Images courtesy of D. Fütterer, Kiel)

Two types of calcareous particles on the seafloor of the Great Bank have been considered as possible products of inorganic precipitation*: aragonite needles* (length of a few micrometers) and *oolites* (consisting of spherically layered particles (*ooids)* with a diameter near one third of a millimeter) (Fig. 4.15). The origin of the aragonite needles has puzzled geologists for some time; apparently they form within certain algae. Oolites are a favorite object of geologic discussion, famous for being mentioned by Mark Twain (in an ironic remark on the quality of geologic reasoning). Oolites are abundant in the geologic record, especially in subtropical shelf sediments. In the Bahamas they occur especially on the outer rim of the Great Bank in the shallowest water, suggesting a strong influence of tidal currents and breaking waves in their origin. Apparently the ooids only form in the presence of sufficient organic matter. This is not good as support for the action of inorganic precipitation. In fact, observations now suggest that precipitation is largely through biocalcification by unicellular algae. It appears, then, that even here in the Bahamas precipitation directly from seawater is negligible in present-day conditions.

4.7.3 Dolomite

Calcite may contain magnesium in different concentrations (*magnesium calcite*). In the present shelf environment, as mentioned, magnesium-rich calcites are found in greater abundance than pure calcite. *Dolomite,* a carbonate mineral with equal amounts (by atom numbers) of magnesium and calcium, is a different story. It does not precipitate in shells or other biogenic products. Its presence in the geologic record poses many unsolved questions. Thus one might expect that dolomite would precipitate from seawater because it is much less soluble than aragonite. There is, after all, plenty of magnesium in seawater. However, such precipitation has not been observed. Instead, dolomite apparently forms *within* sediments, either through partial replacement of calcium with magnesium in preformed carbonate or possibly also by precipitation from pore waters. Either process would be referred to as "diagenesis." Much of "diagenesis" is about postdepositional geochemical problems. Thus, geochemists have done much fundamental work on dolomitization (Scripps examples: M. Kastner, P. Baker). In recent years, petroleum and methane geologists and chemists have shown much interest in the topic.

A classic area for the study of incipient dolomite formation is the southern margin of the Persian/Arabian Gulf. Here the intertidal flats lie behind barrier islands within lagoons with warm and very saline water. In the pore waters of sediments above low tide, magnesium is greatly enriched with respect to calcium, while sulfate (which apparently hinders dolomite formation in many surroundings) is reduced in abundance owing to the precipitation of gypsum (calcium sulfate) in adjoining evaporite pans (which are common in the *sabkha* environment). Microbial sulfate reduction in the uppermost few tens of meters of continental margin sediments similarly seems to favor "dolomitization." One problem interfering with solving the dolomitization question is that unknown conditions of the geologic past and in a warm ocean may have played a large role in some of the dolomite formation.

4.8 Hydrogenous Sediments

4.8.1 Marine Evaporites

Evaporites form when seawater evaporates. Desert belts – high ratios of evaporation rates over those of precipitation – are located near 25° of latitude, on both hemispheres. Unsurprisingly, these also are the latitudes of the highest salinity in the surface waters of the open ocean. For precipitation of salt, however, concentrations have to exceed saturation values, and this happens only upon restriction of exchange of shelf-water bodies with the open ocean, commonly in a semi-enclosed basin on the shelf (the one widely quoted exception being the Mediterranean basin at the end of the Miocene, as documented by deep-sea drilling during DSDP Leg 13.

How much salt can be produced by evaporating a 1000-m-high 1-m^2 column of seawater? Salt is 3.5% (or 35 per mil) of the weight of the column; thus the answer is 35 tons or 14 m at the common density of 2.5 tons per cubic meter. Most of the readily recognized salt obtained would be kitchen salt ("halite"). The salts precipitating first would not include halite, though, but have abundant carbonates and sulfates.

To precipitate halite the brine would need to be concentrated about tenfold. Many evaporites only contain carbonate and gypsum (or "anhydrite"), whereas others have thick deposits of halite or (even more rarely) final layers of the valuable (and very soluble) potassium salts. In marine sediments, to get any one salt without the others, "fractionation" in linked basins is necessary or periodic removal of certain salts by dissolution, while preserving others.

4.8.2 Phosphorites

Phosphatic deposits presumably are largely of biogenic origin. Phosphorites are of prime importance in the cycling of phosphorus and hence in the productivity of the ocean (Chap. 7). Also, they constitute an important marine resource, used predominantly in the fertilizer industry. Thus, phosphorites are encountered again in the chapter on resources (Chap. 14).

4.8.3 Iron Compounds

Microbes are notably involved in the origin of both iron sulfides and iron hydroxides, as has been known for more than a century. Iron apparently is of prime importance in the productivity of the ocean, with at least some of it mobilized from the seafloor upon loss of oxygen. Sulfides and hydroxides are important items in the geochemical cycling of temporarily free oxygen, the abundance of which changes in Earth history, with consequences for marine sedimentation (black mud, clay, and shale and greenish deposits versus brown, reddish, and yellowish deposits).

An iron-bearing mineral that has attracted much attention by marine geologists is *glauconite.* It is a greenish silicate common in shallow marine areas and is commonly found in association with phosphatic sediments in high-productivity regions along certain continental margins, as, for example, off Angola.

The origin of iron oolites, common in Mesozoic marine sediments (e.g., "minette" ores of Alsace-Lorraine), is not known. Apparently minette ores are not forming in the present ocean in noticeable abundance.

4.9 Sedimentation Rates

The idea of *geologic time,* so fundamental in all of geology, is in fact quite young, compared with the age of various branches of science. It was early discussed by James Hutton (1726–1797), and its chief protagonists were Charles Lyell (1797–1875) and Charles Darwin (1809–1882). Lyell, like Hutton, invoked geologic time to create Earth's morphology and the geologic record. Darwin made use of geologic time in his explanation of evolution. However, before the discovery of radioactive decay at the end of the nineteenth century and the application of this discovery to the geologic record, there was no reliable way of telling just how much the geologic time scale differs from the chronology derived from multimillennial human time, that is, the account in Genesis, summarized by Bishop Sam Wilberforce, FRS. Wilberforce's clever but derogatory pronouncements on Darwin's ancestry were forcefully attacked by T.H. Huxley, in 1860. The disputations reflected different belief systems, not certain knowledge.

We now know that the guesses proposing millions of years of Earth history were closer to the truth than those postulating thousands of years. Modern determinations of sedimentation rates span the gamut between 1 m per million years and many km per million years, depending on the environment (see Fig. 4.16). A million years is a useful time interval to work with, for geologists contemplating pre-ice age processes. To apply the rates listed to ancient sediments, one has to consider compaction and loss of porosity – about 40% in sands and about 70% in muds. On the whole, high rates of sedimentation (say 100 m per million years) occur at the edges of the continents. Exceptionally high rates are found off many glaciers. The lowest rates occur on the deep seafloor far away from continents.

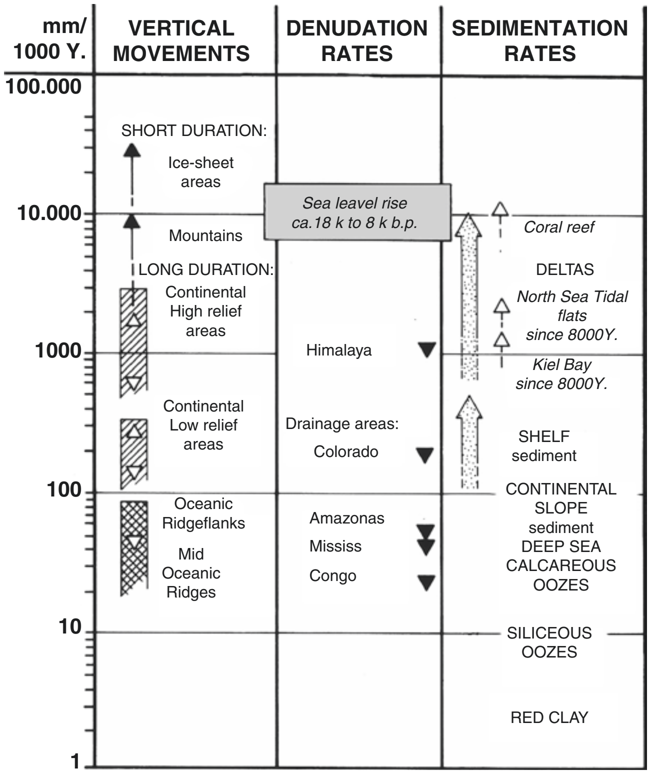

Fig. 4.16 Typical rates of vertical crustal motion, of denudation, and of sedimentation rates. Scale in mm per thousand years (meters per million years). Postglacial sea-level rise shaded (average of some 100 m in 10 thousand years; *k* kilo years). Logarithmic scale. The values are approximate and for comparison (E. Seibold, 1975. Naturwissenschaften 62: 62; modified)

Characteristic values of sedimentation rates on continental slopes are 40–200 mm per thousand years, with a typical value near 100 mm per millennium or 100 m per million years. Theoretically, coral reefs can build up at rates near 1 cm per year, that is, 10 m per millennium or 10 km per million years! More commonly, one finds values somewhat less than half of that in reef growth. In any case, such high rates make the sinking of the seafloor irrelevant to the growth of stony reefs. Such findings throw doubt therefore on the commonly quoted Darwinian origin of atolls.

Reliable estimates of high sedimentation rates are possible in the case of *annual layers* or *varves*. Counting varves in the Black Sea have yielded rates near 400 mm per millennium. In a bay off the Adriatic Island Mljet, 250 mm per millennium was found. In Santa Barbara basin off Southern California, counts of varves yielded values near 1 mm per year, that is, 1 m per millennium. Such high rates of sedimentation are favorable for the detailed reconstruction of environmental changes over the last few millennia and centuries (Chap. 15).

In concluding this chapter about the composition of marine sediments, some of the more esoteric components may be mentioned. So-called cosmic spherules were first described by John Murray of the *Challenger E*xpedition. Black magnetic spherical objects up to 0.2 mm in diameter, commonly rich in Fe and Ni, can be found in pelagic sediments, typically several spherules per gram of pelagic clay. Because it was thought that they represent a steady influx of matter from space, their abundance was used on occasion as an indicator for accumulation rates.

Glassy objects, normally up to 1 mm in diameter, may be produced by the impact of meteorites on rocks. They are called *microtectites* or simply *tectites*. Tectites were found in deep-sea sediments in strewn fields around Australia (dated at 0.7 million years), off the Ivory Coast (1.1 million years), and in the Caribbean (33–35 million years), for example. Where abundant, they can be used as a time marker and serve for correlation of sedimentary rocks by ev*ent stratigraphy*. Tectites commonly serve as evidence for impact events.

Suggestions for Further Reading

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5 Effects of Waves and Currents

5.1 Sediment Transport and Redistribution

5.1.1 General Remarks

We have seen that the morphology of ocean margins is largely determined by tectonics and sediment supply (Chap. 3), and we have reviewed the various types of sediments involved in building margins in the previous chapter (Chap. 4). We now turn to the all-important role of water motion in determining the distribution of sediments on the seafloor. A first impression of this role can be gained from contemplating Fig. 5.1, which illustrates the redistribution of sediment supplied to the sea by a river in a setting typical for Southern California. To be sure, the water motions indicated in the graph are not the only ones that need to be dealt with, as we shall see.

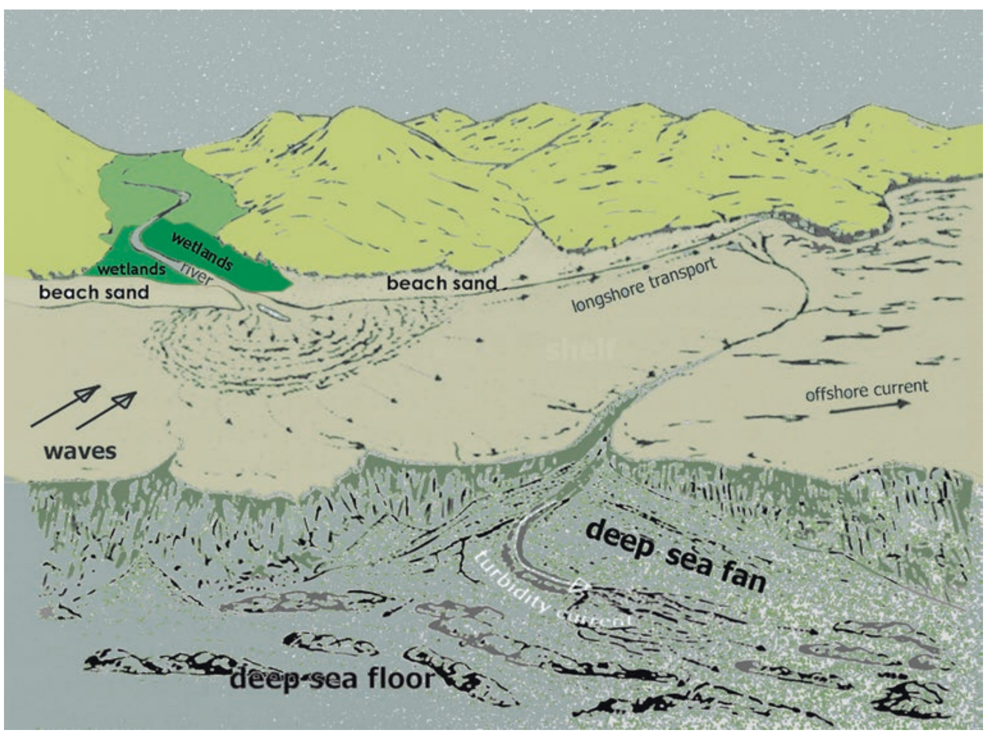

Fig. 5.1 Redistribution of sediment on the continental margin, by water motion. The drawing reflects a La Jolla (West Coast) setting. Note the river bringing material from nearby mountains (*green*), the longshore transport powered by waves, and the interception of the "river of sand" (D. Inman) by a submarine canyon that delivers the sediment to a submarine fan on the continental slope. Also note starved beaches and rocky shelf beyond the canyon. Vertical scale greatly exaggerated (Based on a drawing by D.G. Moore, US Navy Electronics Laboratory, modified; see Geol. Soc. Amer. Spec. Pap. 107: 142 (1969))

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_5

Fig. 5.2 Evidence for the action of waves and currents. *Left*: ripple marks in the intertidal flats off the mouth of the Weser, northern Germany. *Middle*: ripple marks on the beach of Southern California. *Right*: Large submarine sediment dunes on the continental slope off Baja California, as found by F.P. Shepard, using seismic profiling (Photos **a** and **b**, W.H.B., graph to the *right* (**c**) courtesy of F.P. Shepard, S.I.O., excerpted)

Waves and currents leave their imprint on the seafloor in many ways, as depositional and erosional features. Familiar examples are waveforms on the sediment surface, from the smallest ripple marks to large submarine dune fields (Fig. 5.2). Others are bedding structures within the sediment, from beach laminations to thick graded layers, and also the grain of the sediment, from muddy lagoon deposits to the highly sorted sands on wave-washed beaches. Scour marks, channels, and clean-swept banks and submarine plateaus also are conspicuous evidence for erosion by currents in the sea.

Just how effective are waves and currents as sculptors on the ocean floor? To this question we turn next. Also, we need to examine the clues to be studied if we wish to reconstruct the wave and current regimes of the past from the geologic record. One way to do this is to study ripple marks visible at low tide or in echo-sounding profiles or documented by diving (Figs. 5.2 and 5.3).

Fig. 5.3 Measuring ripples. Diver is assessing the inclination of a lee slope of a giant ripple north of Fehmarn Island, Baltic Sea (Photo Diving Group, Kiel)

5.1.2 Role of Grain Size

The details of the relationships between water motion and sediment response are by no means clear, despite of considerable study of the matter. One problem in trying to capture relationships precisely is the complicated feedback between modified water motion at the interface of water and sediment and the changing character of the seafloor surface. It is difficult to predict what will happen in a given situation based on studies of another one because grain-sized distributions, porosity, and cohesiveness of the sediment may not be precisely the same for any two situations. With a different mix of properties come different interactions.

5.1.3 The Hjulström Diagram

One of the most basic questions one can ask in the context is "how strong does a current have to be to move sediment?" Answers are still being found in the "Hjulström Diagram" (Fig. 5.4).

It is reasonable to expect that coarser grains need more of a push to move than finer ones, and this is indeed what is found by experiment. Also, there is no surprise in consolidated sediment being more difficult to erode than unconsolidated deposits, with consolidated clay being rather resistant. The principles are readily established; details remain somewhat obscure, however. For example, how far above the bottom the velocity should be measured poses a problem, when contemplating the "Hjulström Diagram" (Fig. 5.4).

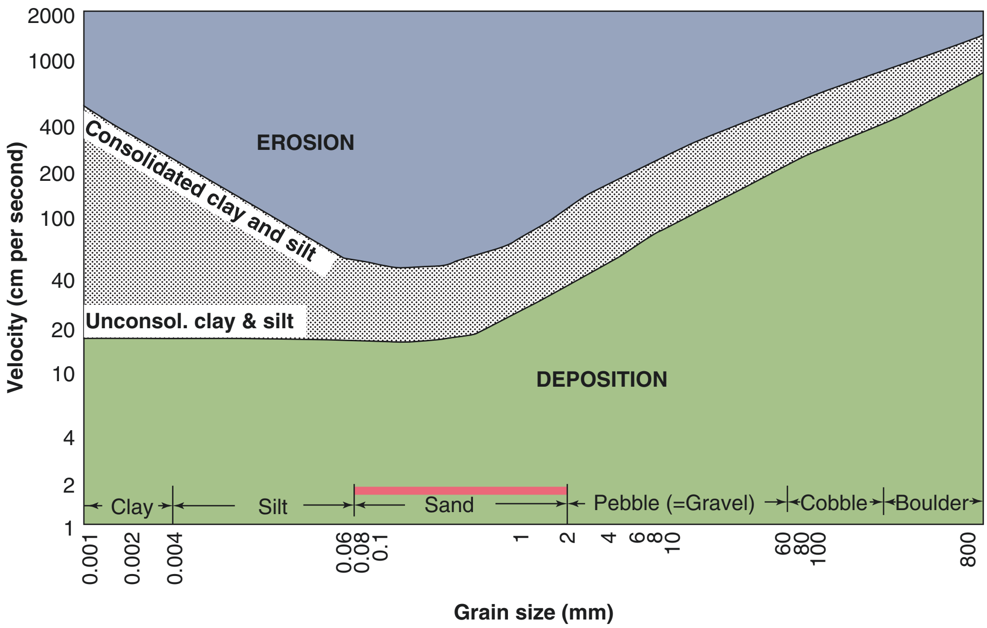

Fig. 5.4 In the Hjulstroem diagram, one might explain the cloth pattern (difference between relating current velocity somewhat above the seafloor to transport of grain-sized particles (well sorted)) (After A. Sundborg, 1956, cited in J. Gilluly et al., 1968. Principles of Geology. W. H. Freeman, San Francisco, modified). Note that 1 mm describes sand size, with 0.063 mm being the lower boundary for the category, and 2 mm the upper one (*solid red line*)

Regarding the principles, the graph of Fig. 5.4, modified from one introduced by the Swedish geographer Filip Hjulström (1902–1982) to illustrate transport by rivers, is easy enough to read. For example, we can see that sand grains of 1 mm in diameter move in a current of around 0.4 m per second (almost one *knot* in oceanographer's lingo) and that boulders are starting to move only at velocities of several meters per second. Note that the diagram is a log-log plot (a plotting method that unfailingly greatly reduces the visual impact of experimental scatter).

According to the diagram, very fine material remains in suspension more easily than material coarser than fine sand. Thus, the fines will be carried farther from the source than the coarser stuff. As a result, grain size will normally decrease in the downstream (or down-current) direction of transport, a clue to motions that can be applied along the coast, on the shelf, or even in the deep sea in places (where shape and fill are very important). However, this simple and obvious relationship between grain size and water velocity is valid only down to sizes of 0.1–0.2 mm. For grain sizes smaller than that (very fine sand and finer), water velocities may have to *increase* again to initiate erosion. Why should this be so?

The chief reason for the aberrant behavior of clay and silt is that when particles settle on the floor, the very fine sediments tend to produce a smooth surface, which reduces turbulence at the interface and thus the opportunity for pulsed pickup of sediment particles by fast water particles. The large surface areas of fine particles also provide for increased cohesion between them. Another reason for resistance to erosion is that *bacterial mats* form on fine sediment layers, mats that likewise reduce the impact of water on sediment particles by smoothening the surface. Also, bacteria increase cohesion between sediment particles.

5.1.4 Role of Water Velocity

How exactly does water move the grains? The average velocity of the current decreases on approaching the bottom and goes toward zero at the interface itself. Hence the values of current velocity given in the Hjulström diagram are applicable for some distance *above* this interface. Because of the difficulties in defining the proper distance for different conditions, modern investigations use *bottom shear stress* produced by the flow regime. Exactly how much effect a current has at the interface depends on the roughness of the seafloor and the turbulence created on the bottom. Turbulence produces sudden changes in the impact of water on a grain sticking out at the surface. Roughness is not part of the diagram, which applies to grain sets of equal size. In the real world, as current velocity increases, the frequency and force of impact pulses increase and some of the smaller grains start to move. This leads to the impacting of grains by other grains and also to a small-scale increase in roughness of the floor. Soon more and more grains start to roll and jump over the floor. The rolling and jumping grains are addressed as *bed load* of a current in contradistinction to the *suspension load,* a term applied to sediment within the water (Fig. 5.5).

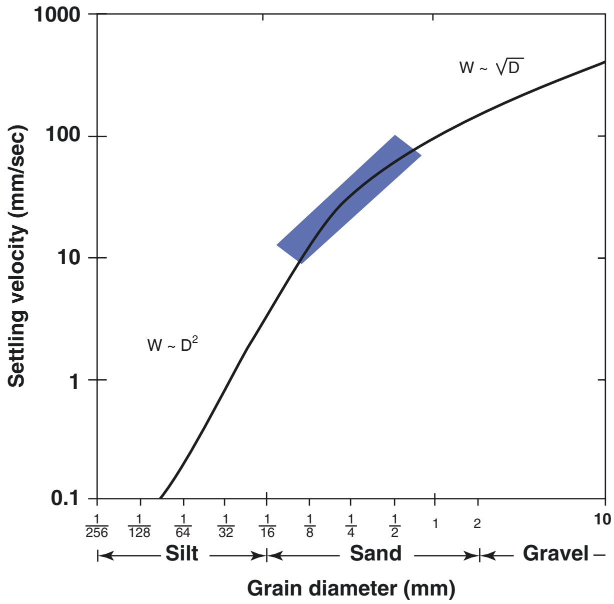

Fig. 5.5 Settling velocities of quartz grains in water. Silt and fine sand settle according to one equation (velocity a function of square of diameter; Stokes' Law) and coarse sand and gravel according to another (velocity a function of the square root of diameter). The transition zone i(*blue*) is shaded (After W.W. Rubey, 1933, as quoted in C.O. Dunbar and J. Rodgers, 1957. Principles of Stratigraphy. John Wiley, New York, modified). Note that sizes are plotted in steps of factor of two

A grain, once moving, greatly decreases contact with the seafloor. Contact vanishes entirely for the *suspended load*. Since fine grains settle rather slowly and turbulence moves many grains upward, they stay in the water for some time, once brought into suspension. Bed load and suspension load tend to go parallel, therefore. Correlations depend on grain-sized distributions and turbulence. The bed load particles, which impact each other, experience abrasion, which therefore increases with the length of transport especially for pebbles, which are thought to be 300– 400 times more sensitive to abrasion than sand grains. Below a size of 0.25 mm rounding and length of transport have no simple relationship. In fact, rounding may *decrease* down-current as the sands become finer and hence more irregular.

As is evident from Hjulström Diagram, the values for current velocities at which the sediment comes to rest are considerably lower than the values for erosional velocities. Clearly it is easier to keep sediment moving than to set it in motion from rest. One implication of the behavior of sediment captured in the Hjulström Diagram is that any current activity will tend to separate clay-sized from sand-sized material. The clay-sized particles stay in suspension long after most of the sand has settled out, as is evident in the top layer of graded beds. For the suspended load, there is no minimum velocity for transport. *Settling* dominates the scene (Fig. 5.5) and not conditions at the interface.

We have presented the processes involved in erosion, transport, and deposition of detrital sediments in a very simplified and mostly qualitative manner. The information given in the Hjulström Diagram, for example, which might be taken to be of a quantitative nature is, in fact, an illustration of a *concept* rather than a nomogram describing observable relationships with some precision.

Quite complicated experiments and theories are necessary when attempting the quantification of the processes involved, processes that are of considerable interest to geologists studying sediments and to coastal engineers who deal with changing beaches and with shore erosion. One difficulty arising is that in many circumstances the condition of the sediment being studied is inherited from the recent past and does not reflect present wave and current dynamics. *Relict* sediments in this category show distribution patterns appropriate for the past conditions. This confusing situation can be very misleading when trying to interpret present sediment patterns in terms of current conditions.

5.1.5 Role of Unusual Events

Events that result in suspension and thereby allow considerable transportation can be rather short. Also, once sediment particles are brought into suspension by currents, even a relatively weak current can keep the material moving. In providing for suspension by waves, mid- and high-latitude events are commonly linked to winter storms, while in low latitudes, hurricanes or typhoons provide common sources of forcing for various events. Effects are widespread and include the action of both waves and currents. Seasonal changes may be very important in providing for the kind of forcing that turns out to be effective. In the Yellow Sea, for example, sediment distribution is entirely dominated by the combined action of strong winter waves and associated currents. All around North America, beach sands move southward, driven by winter waves.

The occurrence of sediment-relevant events may be separated by long time intervals – long, anyway, with respect to the duration of the events. Storms do much of the moving of sediments along the shores and much of the damage observed from cliff erosion and other factors, even though the time intervals of the processes involved are brief. In the record, a geologist dealing with millions of years may see the results of events unheard of in the normal course of history. Especially large storms can rework a relatively thick layer of sediment on the seafloor, a layer subsequently undisturbed. Such layers are identified as *tempestites*, that is, as a record of unusual storm activity.

5.2 Effects of Waves on Sediments

5.2.1 Beach and Shelf

One notes chiefly three kinds of waves (Fig. 5.6) in a nonscientific way: (1) the everyday waves, commonly originating

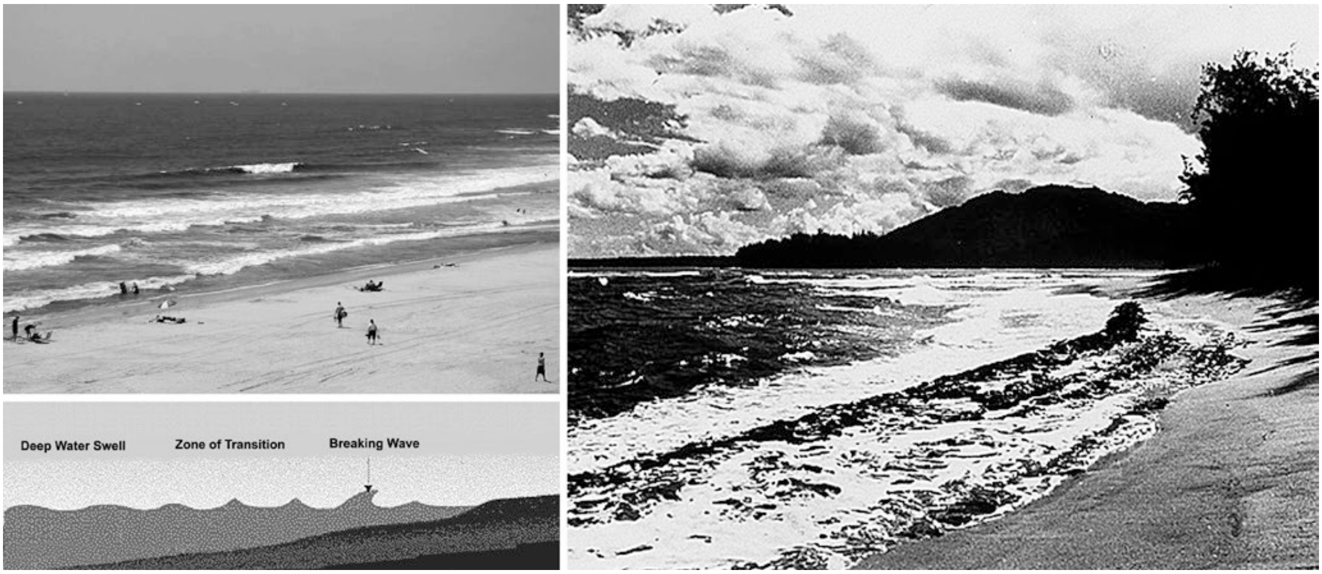

Fig. 5.6 Regular wave climate. *Left*: on a beach in San Diego County. *Right*: in Hawaii. Waves slow in shallow water and steepen. They break when the slowdown of the wave base attains a critical value so that the top of the wave becomes faster than the base (Beach photos W.H.B., graph of breaking wave: NOAA)

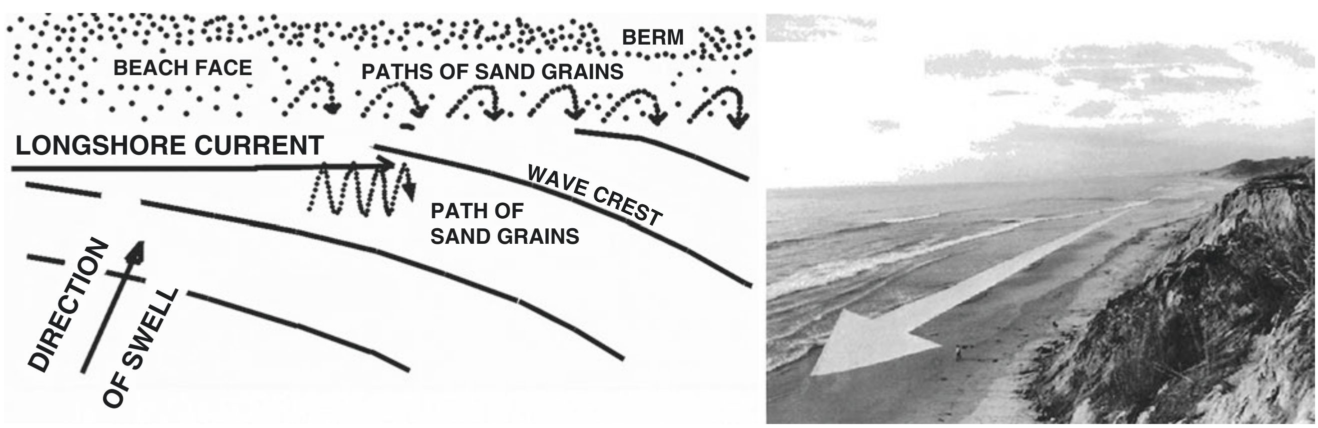

Fig. 5.7 Sand motions on the beach. Overall, sand moves southward on both West and East Coast of North America, driven by waves from northern storms (Graph and image inspired by W. Bascom (*graph on left*) and D. Inman (*image on right*)) see SW. H.B., 1976, Walk Along the Ocean. S.I.O., La Jolla

from storms and traveling many hundreds of miles before arriving at the shore, (2) the long waves that traveled thousands of miles and originated in large southern storms, even storms from off Antarctica, and (3) the winter storm waves, which in California come from northerly regions and make rather choppy surf upon breaking, a surf poorly suited for surfing. Scientifically, the various types of waves are characterized by length and period.

Choppy waves made by local winds are well distinguished from the swells that carry energy from distant storms and typically have a longer period than the chops. All types of waves move sediment after breaking along the shore, as seen in comet marks and other sand patterns on the beach. But the *winter storm waves*, as the most energetic ones, are chiefly responsible for transport of beach material (Fig. 5.7), a transport that ends with filling canyon heads and creates the conditions for turbidity currents running down those canyons (Fig. 5.1).

The energetic winter waves of Southern California, generated by northern storms, move sand out, exposing a rocky beach terrace on occasion (Fig. 5.8). Also, powerful waves can result in the coastal erosion that produces so many problems along many shores (Fig. 5.9).

Not all beaches are chiefly made of sand, of course. In latitudes with morainal debris, one finds pebble beaches and other beaches with coarse material (e.g., in East Anglia and around the Baltic Sea and in the Svalbard archipelago and in Alaska). Also in low latitudes, storms can add plenty of coarse material to the beaches (such as reef debris, e.g., in the Caribbean and in Hawaii). In some places there is pebble-sized volcanic debris.

Fig. 5.8 Seasonal change in sand cover, beaches of La Jolla, California. Summer condition: *far left*. *Middle*: winter condition. *Far right*: wave-cut terrace at sea level near SIO (Photos W.H. B)

Fig. 5.9 Cliff erosion in Oceanside, San Diego, endangering roads on a terrace next to the shore (Air photo W.H. B)

Wave motion quickly decays with depth in the water. At depths greater than one half the wavelength, there is virtually no motion from surface waves. Surface waves involve only the uppermost part of the water column. We can be fairly certain that strong winds were responsible for making the waves in the first place and that wave-driven motion of sediment must be largely confined to the region bordering the beach. *Internal waves* also exist, at the thermocline and at other density discontinuities in the water column. Such waves have been shown to induce currents in submarine canyons, and they may be rather important in pumping nutrients into surface waters above canyon heads. On the whole their origins may vary (the tides apparently are commonly involved) and their effects are not well known. Waves called *tsunamis* can be generated in deep water by earthquakes and associated motions, by submarine landslides, and by volcanic eruptions. Such waves, traveling at the ocean surface, have lengths measured in kilometers and they move at the speed of a jet plane.

In general, the maximum depth to which sand is being moved below a surface wave, the *wave base*, is near 10–20 m. In exceptionally strong storms, wave motions can reach considerably deeper. Evidence is hard to come by, however. Thus, for example, while ripple marks are seen on shelves even out to the shelf break, they presumably formed during glacial times, aided by a drop in sea level. In any case, the relative importance of surface waves, past waves, internal waves, tides, and currents in producing such ripples is not necessarily clear. Some years ago symmetrical ripples were discovered in fine sands of the outer shelf off Oregon at depths to 200 m, with crest-to-crest lengths of 10–20 cm and with the crests parallel to the coast. These oscillation ripples are thought to have been produced by winter storms on a shelf then only thinly covered.

Despite of numerous problems in interpreting extant ripple marks on the shelf, "wave base" is a useful concept. The concept is important not just for wave motions per se but also for the distribution patterns of sediment types and for benthic organisms. The composition of benthic organisms changes greatly across the wave base boundary, presumably in response to changes in sediment motion.

5.2.2 Storm Action and Storm Damage

We already mentioned the ability of strong waves and associated currents to clear a terrace of its sediment and to vigorously erode cliffs (Figs. 5.8, 5.9, and 5.10). At Boomer Beach in La Jolla, the beach forms in summer next to a slightly uplifted (geologically young) terrace made of marine rocks of Cretaceous age (rocks that were made of sediment once deposited well below the shelf edge). The place has become a prime tourist attraction, in part because of the spectacular breaking waves on the rocky point (hence the name "boomer") and in part for the fact that the beach lately is a favorite hangout for harbor seals and their pups at the southern end of its extent, where a man-made barrier calms the waves and sand is present all year. In winter, in response to wave attack, much of the sand along the northern and the middle portions of the (unprotected) beach moves offshore, leaving underlying rocks and gravel exposed. The sand that collects offshore forms bars whose position can be recognized from the shore by observing the breakers that form on top of them. (Very shallow water slows down waves and makes breakers.)

Where beaches are narrow or missing altogether, storm waves can hit the coast with force. Armed with gravel or sand, breakers can dig a deep notch into cliffs, especially in places where the rocks at the base of the cliff are somewhat softer than normal. When the notches and sea caves become large and deep enough, the overlying material collapses and the cliff retreats. This type of cliff erosion (*notch cutting*) makes for wave-cut terraces (Fig. 5.8) as well as for steep cliffs (Fig. 5.9). Also, by grinding up the fallen rocks, storm waves produce additional beach sand while cutting into the land.

An obvious response to wave attack is the construction of defensive walls to protect the shore. However, walls are quite vulnerable to wave attack. Attacking waves remove the support at the base, so that a wall eventually topples over toward the sea. A more effective way to tame a furious sea is to pile up "rip rap" (i.e., large boulders) or "tetrapods" (i.e., man-made concrete structures). Or else one can construct dams with gentle slopes, where breakers can spend their energy

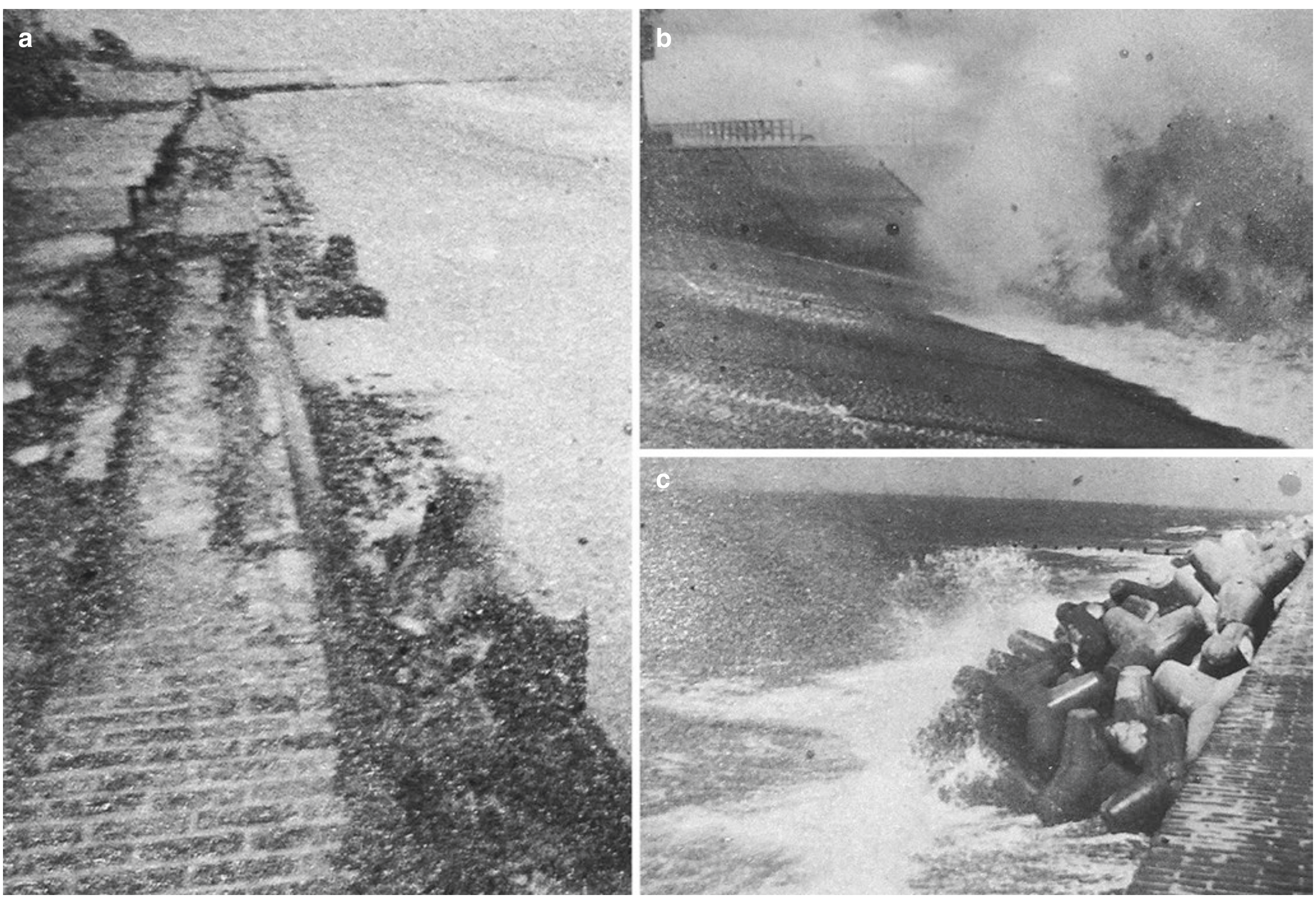

Fig. 5.10 Defense against violent surf action in Westerland, Island of Sylt, German North Sea. (**a**) Damaged beach wall and displaced tetrapods; (**b**) gently sloping wall for breaking the power of storm waves; (**c**) interlocking tetrapods dissipating energy before damage is caused (Photos E.S. (**a**) and courtesy J. Newig (**b** and **c**))

(Fig. 5.10). Also, energy-consuming friction and turbulence can be enhanced by providing for rough surfaces on such slopes.

Whenever dealing with shore defense, or in fact whenever dealing with the dangerous side of nature, it is well to recall the advice of the British Renaissance philosopher Francis Bacon (1561–1626) who believed in scientific observation and experiment: “Who would rule Nature must first obey her.”

Bacon thought it is worthwhile to study what actually happens in nature, before trying to steer events.

5.2.3 Tidal Waves

Tidal currents are ubiquitous on continental shelves. They are produced by tidal waves, which owe their existence to the Earth’s moon and the interaction of its gravitational field with that of the sun and the tidal forcing of both experienced by Earth. In Southern California, including La Jolla, we have mixed diurnal tides, with one of the two crests per day being somewhat higher than the other. Elsewhere the smaller crest may disappear entirely (*diurnal tide*), or it may reach the elevation of the larger tide (*semidiurnal tide*).

The puzzle for the uninitiated is that some simplified textbook schemes with a water-covered Earth show two similar tidal crests associated with each Earth rotation in an explanation of diurnal tides. Given the sense of gravity, why is there a bulge opposite to the moon? The solution to this puzzle is that the moon does pull a bulge where it is closest to Earth, but it also generates a second one at the opposite side, which reflects the difference between the lunar gravity and the opposing centrifugal force between the moon and the Earth. The forces keeping the moon up there are in balance, but only on average, not locally.

The normal tidal wave has a period of 12 h and 25 min, one half of the length of the lunar day, which is 24 h 50 min long – slightly longer than the rotation period of 24 h. The lengthening of the lunar day relative to the Earth day is due to the fact that the moon moves eastward. Thus it takes slightly more than one Earth rotation (a day) to get the moon to reappear at the same meridian as the day before. Tidal action does depend on the sun as well. When in line (sun-Earth-moon or Earth-moon-sun), the forces add up straight and we have a spring tide, with tidal action at a maximum. But when the celestial bodies are at right angles (with Earth at the point of intersection of the defining lines), the sum of forces is much weaker and we get a neap tide.

Tidal currents are set up especially along the continental margins – typically on the shelves. Whenever there are no shelves during glacial times, most of the tidal energy must be lost elsewhere, presumably in the deep sea. Even at present, we think we see evidence for tidal action on the deep seafloor. However, the most conspicuous evidence at present is on shelves, in shallow water. Effects include the uncovering of burrowing mollusks, as well as ripple marks (Fig. 5.11).

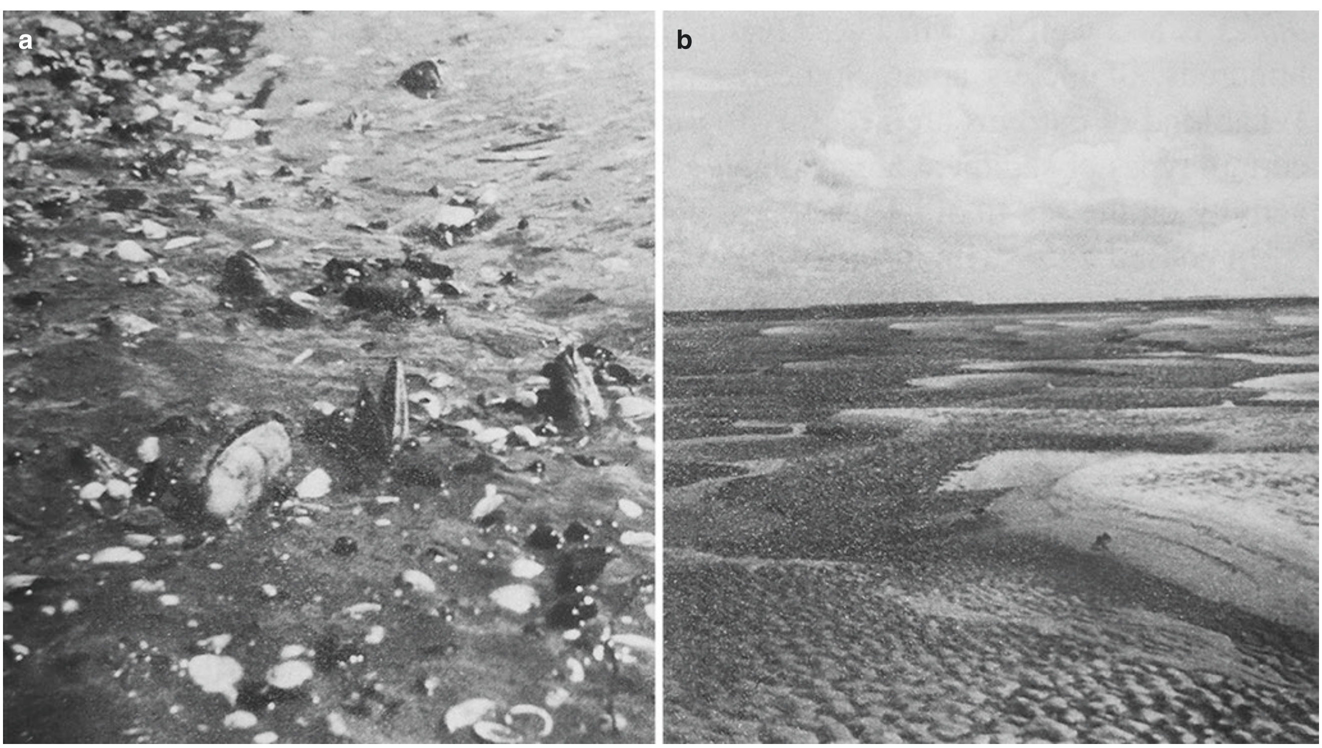

Fig. 5.11 Current action in tidal flats, German Bight, North Sea. (**a**) *Mya arenaria* in life position uncovered by erosion by migrating tidal channel; (**b**) ripples off the Weser, produced by currents of the outgoing tide. Note the presence of both large and small ripples (Photos E. S)

Current ripples tend to be at right angles to the current creating them. Their shapes are controlled by the flow of water over the seafloor that results from the presence of the ripples themselves. Thus, we deal with a complicated feedback system, whereby currents make ripples, and ripples stabilize ripples and make more of them, extending them sideways (hence the washboard aspect of ripple marks).

5.3 Effects of Currents on Sediments

5.3.1 Shelf Currents in General

There are many clues to the direction and strength of past currents at the seafloor, as mentioned when discussing size distributions. Winnowing (the separation of fine from coarse material) is one type of clue; scour marks behind obstacles is another. Responding to density gradients within an almost unconstrained environment, flow in the sea is turbulent. Also, it is influenced by the Earth's rotation. The use of side-scan sonar, both on the shelf and in the deep sea, has greatly increased our knowledge about the nature and distribution of bottom-near currents (Fig. 5.12). The currents are not seen directly on the sonar graphs but are inferred from their effects on sediment patterns on the seafloor. Streaks of coarse material in fine sediments, a result of current action, were discovered at the entrance to the Baltic Sea by such acoustic sensing (Fig. 5.13). In the deep sea, also, current-generated sediment streaks, ripple marks, and other markers have been discovered.

5.3.2 Currents in the Open Sea

Of all ocean currents, the *Gulf Stream* js perhaps the most familiar (Fig. 5.14). It is the largest and most important current of the northern hemisphere, transporting nearly 100 million cubic meters per second of warm water northward into colder regions. Some of this water even reaches the shores of Norway, keeping coastal fjords free of ice all winter. For comparison regarding the volume of transport, extraordinary peak floods of the Mississippi might carry as much water as one half of a thousandth of the usual Gulf Stream transport.

The Gulf Stream is best understood as the western limb of the great North Atlantic gyre, driven by the trade winds in the south and the west winds in the north. The eastern limb is provided by the Canary Current and is associated with upwelling along the coast, a phenomenon of central importance when discussing the productivity of the sea (Chap. 7). The center of the gyre (which marks a desert area of the ocean in terms of production) is near 30°N, in the Sargasso Sea. There are five such great gyres on the globe: North and South Atlantic, North and South Pacific, and southern Indian Ocean. In each case, the west winds and the trades provide the driving forces, and the western and eastern boundary

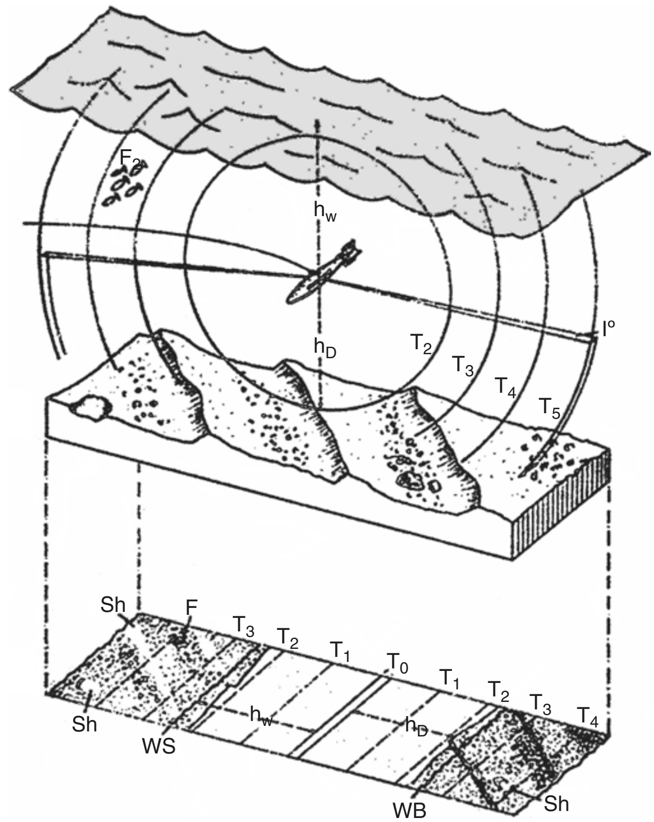

Fig. 5.12 Principle of side-scanning echo sounder. (**a**) Water surface; (**b**) seafloor with ripples and rocks; (**c**) acoustic recording of the area (sonograph). T0, outgoing sound pulse (see in *panel c*); other Ts with subscripts: sound waves emanating from the sound source, the "fish"; Sh, acoustic shadow; F1 and F2, fish schools, and their acoustic image; hw, distance of the sound source to the water surface; hb distance of sound source to seafloor (R.S. Newton et al., Meteor Forschungsergeb. Reihe C 15:55; simplified)

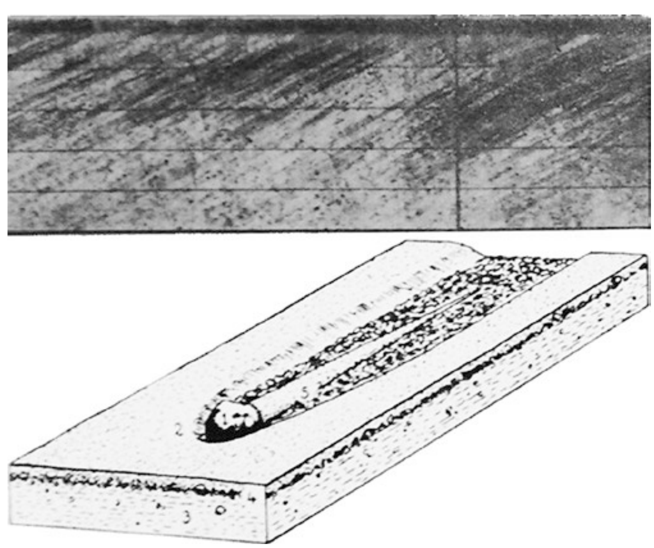

Fig. 5.13 Bottom current indicators in the Big Belt Channel in the Baltic Sea ("Store Belt") between the Danish islands of Langeland and Lolland. *Upper panel*: excerpt of sonograph (one side only). Length of the record is 2 km. Water depth is approximately 12 m (Source: F. Werner, Kiel). A "comet mark" (details by diving) may result from turbulence behind an obstruction interfering with the current. Sediments consist mainly of moraine material (Courtesy F. Werner, Kiel; See F. Werner et al., 1980. Sediment. Geol. 26:233)

currents complete the "gyres" (a term introduced by the physical oceanographer Walter Munk, S.I.O.).

Normally, the surface currents of the open ocean reach down to about 100 or 200 m, and their velocities are low: a fraction of a knot. The Gulf Stream, however, and other fast narrow boundary currents like that (e.g., its sibling off Japan, the Kuroshio) reach to a depth of some 1000 m and have velocities of a couple of knots or so (~100 cm/s). According to the Hjulström Diagram (Fig. 5.4), such fast currents can carry grain sizes beyond the coarsest sand. The Florida Current, which brings Caribbean waters into the Gulf Stream through the Florida Straits, reaches velocities of six knots (300 cm/s).

5.3.3 Northern Heat Piracy

The Gulf Stream is a major means of heat import into high northern latitudes in the Atlantic (*North Atlantic heat piracy*). The North Atlantic heat piracy ultimately results in a large warmwater anomaly in and around the Norwegian Sea, a fact that is reflected in the planktonic shells accumulating on the seafloor in the area. The warm current that starts as the Gulf Stream is intimately linked to the ocean's deep circulation in the Atlantic. Surface waters moving north have a large component of salt-enriched water thanks to enhanced evaporation in the central North Atlantic. Upon cooling in high latitudes, this salt-rich water becomes heavy enough to sink to great depth. In doing so, it starts deep-water and bottom-water southward flow in the region (Fig. 5.15). The resulting stratification pattern in the deep Atlantic reflects anti-estuarine circulation and is an important control on sediment patterns.

Present circulation patterns changed markedly over geologic time, as suggested by the sediment sequences obtained by deep-ocean drilling (Chap. 12). The evolution of an anti-estuarine North Atlantic starting in the middle Miocene

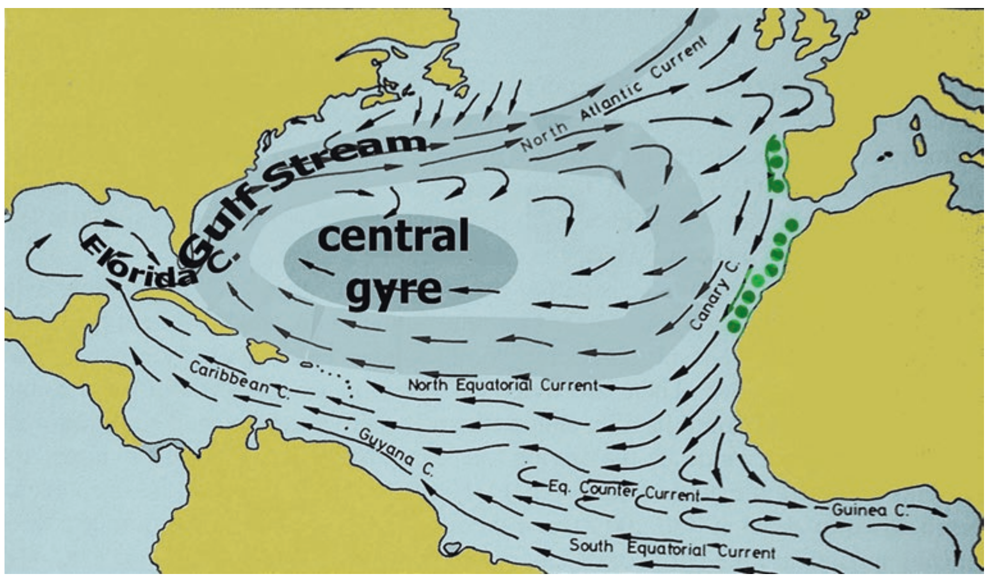

Fig. 5.14 The Gulf Stream off the Atlantic US coast, carrying warm water to the offshore areas of NW Europe. It is fed warm and salt-rich water by the Florida Current (and also by the central gyre). Upon cooling, such water can sink to the bottom. *Green dots*: coastal upwelling (Background chiefly after G. Neumann and W.J. Pierson, 1966. Principles of Physical Oceanography. Prentice-Hall, Englewood Cliffs)

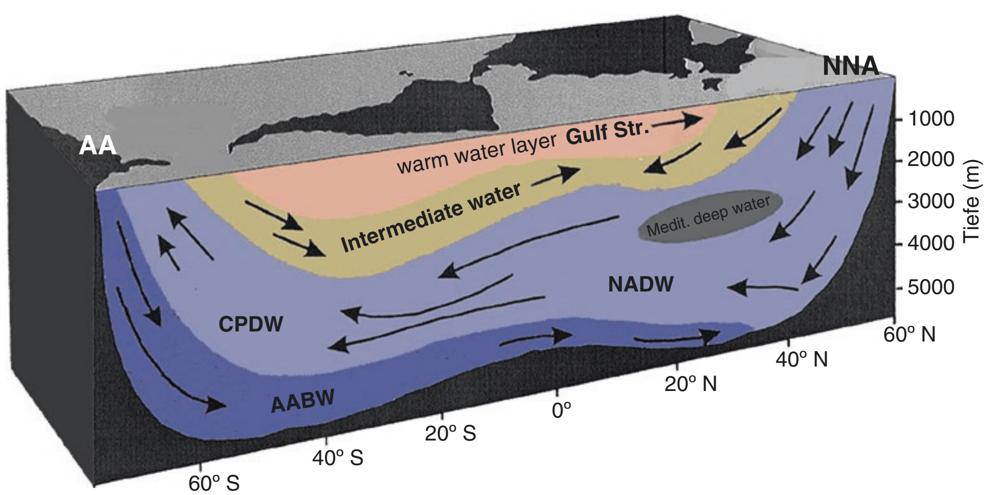

Fig. 5.15 Anti-estuarine nature of deep circulation in the Atlantic. *NNA* northern North Atlantic, *NADW* North Atlantic Deep Water, *AA* Antarctic, *CPDW* circumpolar deep water in the Antarctic ocean, *AABW* Antarctic bottom water (After G. Wefer et al., in G. Wefer and F. Schmieder (eds.) 2015. Expedition Erde. Marum, U. Bremen, p. 329; modified and color added)

implies the development of a balancing estuarine circulation in the deep North Pacific with implications for deep basin-to-basin exchange patterns.

Before the serious cooling of the planet in the Oligocene, heavy bottom waters presumably were not made in high latitudes, but elsewhere closer to the equator. Shelves in the ancient seaway called “Tethys” are attractive candidates, owing to present-day observations in the Adriatic Sea east of the Italian peninsula, where Mediterranean deep water is made by winter cooling of saline shelf water. Salinity effects, of course, are important everywhere when making heavy deep waters. In today’s polar regions, the effects from the rejection of salt from freezing sea ice commonly dominate. Evaporation is important in the subtropics, and always was, one assumes.

The North Atlantic deep circulation with its strong link to heat transport represents the most conspicuous portion of the “global conveyor,” the inferred worldwide circulation system reproduced in many oceanography textbooks and commonly credited to Lamont geologist and chemical oceanographer W. S. Broecker. The *conveyor* links the motion of surface waters and deep waters and reflects well the heat transport between ocean basins and between hemispheres. The sedimentary record suggests a beginning in this circulation roughly 15 million years ago, in an event that was arguably linked to the buildup of ice masses in Antarctica (the middle Miocene “Monterey Event” in Chap. 12).

5.3.4 Bottom Water Circulation

There was a time, a century ago or even less, when the deep ocean was thought to be a quiet and calm environment with weak currents, currents that were considered rather unimportant as far as shaping the seafloor. The first indication that this concept could be quite wrong came from the calculations of the physical oceanographers Georg Wüst (1890–1977) and Albert Defant (1884–1974) in the 1930s, in the wake of the famous German *Meteor* Expedition (1925–1927) (which also delivered initial data on the stratification of the Atlantic shown in Fig. 5.15). These two scientists showed that the distributions of temperature and salinity seen in the closely spaced profiles of the *Meteor* in the deep central Atlantic implied strong bottom-near flows driven by density differences. Such currents hug the slopes of ocean margins and the flanks of mid-ocean ridges following density surfaces.

The currents at issue are commonly referred to as *contour currents* (by marine geologists) or a*s deep geostrophic currents* (by oceanographers*)*. Their effects are clearly seen on deep-sea photographs as streaks or ripples reminiscent of the features in shallow water. Deep currents are strong enough to erode the seafloor in places, especially where confined to passages. Erosion occurs, for example, where abyssal bottom water passes through the Vema Channel off Argentina or through the Samoan Passage in the deep central Pacific.

The NADW (North Atlantic Deep Water) that sinks in the Norwegian Sea falls over shallow passages in the Greenland-Faroe Ridge northwest of Scotland in enormous undersea cataracts. As the NADW moves south in the western trough of the Atlantic toward the Circumpolar Current, it makes the *Deep Western Boundary Current,* which affects the seafloor along the deep eastern margin of North America. Coarsening of silts, alignment of magnetic grains, and the making of mud waves, ripples, and scour marks indicate its presence, forming *contourites*. Deep *contour currents* were investigated in detail during the *HEBBLE* (High Energy Benthic Bottom Layer Experiment) project at 4800 m depth off the US East Coast near Woods Hole. One surprising finding of the study was the discovery of strongly pulsed *abyssal storms* lasting between 2 and 20 days and occurring several times a year. Currents of such “storms” can approach velocities of one knot and carry sand over enormous distances. The origin of the storms is commonly sought in eddy formation within shallow water, eddies whose energy is partially transferred to great depth.

In general, the frictional interaction between bottom currents and the seafloor results in a *benthic boundary layer,* which is on the order of a few hundred meters thick. Compared with the overlying deep water, it has a high content of suspended matter and is characterized by turbulent motion. Also, its chemistry is distinct. This water layer, in combination with the bioturbated layer of sediment on the seafloor, forms a system within which seawater/seafloor interaction is a dominant process controlling benthic life, sedimentation, and geochemistry.

At present the *Antarctic bottom water* (*AABW)* covers vast regions of the deep seafloor. Also, it causes dissolution of calcareous sediment – a process that profoundly affects all of deep-sea sedimentation (Chap. 10). The AABW originates on the shelf of the Antarctic mainly in the Weddell Sea. The processes involved consist of cooling of locally available seawater loaded with salt from the making of sea ice. Being heavier than other water in the deep ocean, AABW sinks and fills abyssal basins. For the filling to proceed in any one basin, of course, there must be access to the basin.

Besides small-scale scour features and ripple marks on the deep seafloor, there are some large-scale features in places, features that likewise indicate the activity of bottom currents. Among these are giant submarine dunes (Fig. 5.16). The deep-side-looking sonar developed in the late twentieth century by the marine geophysicist and engineer Fred Spiess and his colleagues at S.I.O. has revealed an abundance of dunes, sediment ridges, and erosional ravines in many places at abyssal depths, even where deep currents were not suspected. (Of course, one must always keep in mind that any attempts

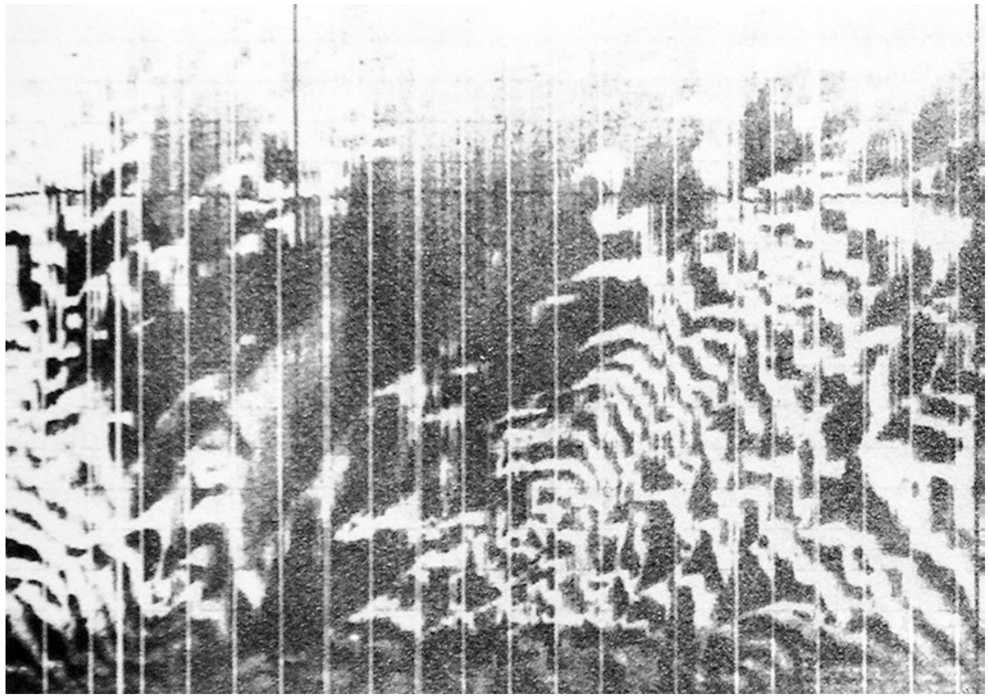

Fig. 5.16 Giant submarine dunes in the Carnegie Ridge area, eastern tropical Pacific. Side-scan sonograph taken at 2.4 km depth. Long edge of the area surveyed roughly 1 km sediment is calcareous ooze (Recording courtesy P. Lonsdale and B.T. Malfait, S.I. O)

to link sediment features to measured currents could be futile if the features seen have nothing to do with present water motions, being inherited from conditions of a distant past. We point out, again, that when shelves are drained during maximum glaciation, tidal action is confined to the deep sea.)

5.3.5 Exchange Currents

The currents that provide for the exchange of waters between marginal semi-enclosed seas and the open ocean are geologically highly significant. The *nature of the exchange* (commonly known as "estuarine" or "anti-estuarine") entirely dominates the chemistry and productivity of a marginal basin through regulation of oxygen and nutrient abundance, with corresponding control on sedimentation.

In arid zones excess evaporation over precipitation produces relatively heavy surface waters in marginal basins, water masses that flow out to the open ocean at depth and are replaced by import of nutrient-poor surface waters of the open ocean. Prime examples for this type of circulation are the Mediterranean Sea and the Persian Gulf. As a result of this *anti-estuarine circulation*, the Mediterranean seafloor has very little diatomaceous sediment and is poor in organic carbon and phosphatic materials, but has abundant calcareous deposits, as does the Persian Gulf. The inverse situation – shallow current out, deep current in – is typical for estuaries and occurs on a large scale in the Black Sea and the Baltic Sea. Excess precipitation over evaporation is necessary to develop this *estuarine circulation* or freshwater inflow, as in estuaries.

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6 Sea-Level Processes and Effects of Sea-Level Change

6.1 Importance of Sea Level

6.1.1 Sea-Level Position as Geologic Calendar

When studying sedimentary rocks on land, the first question a geologist will ask is whether the sediment was laid down above or below sea level, that is, whether or not it is of marine origin. For marine sediments, the next question usually is about the depth of deposition, that is, about the *position of sea level* relative to the sedimentary environment. At present, seafloor depth of deposition is one of the most important factors determining the major facies patterns of the material accumulating on it. Quite generally, *sea-level fluctuations* on scales between millennia and millions of years dominate the calendar of geologic history (Fig. 6.1).

Where the sea level intersects the continental margin, physical, chemical, and biological processes are of high intensity, as elaborated in the preceding chapter. Mainly, in geology sea level is the baseline of erosion and deposition. Exposed areas are subject to erosion, while submerged areas commonly receive sediment. The erosional and depositional processes at and near sea level are *sea-level indicators*. Also they help determine the *coastal morphology* (Fig. 6.2). However, since the type of sediment present has more information than an absence of material, we may safely assume that in general depositional processes are more suited as indicators than is erosion.

6.1.2 Types of Sea-Level Change

Sea-level fluctuations are of two kinds: *global* and *regional*. Global fluctuations produce contemporaneous transgressions and regressions on the shelves of all continents. Such changes in sea level are called *eustatic*; they originate from changes in the volume of ocean water (most commonly by buildup or destruction of ice masses on land) or changes in the average depth of the ocean. Regional fluctuations result from *transgressions* and *regressions* on a regional scale due to uplift or sinking of continental or oceanic crust.

Sea-level fluctuations result in changes of environmental conditions. Flooded shelves are darker than dry ones; they absorb sunlight and thus add energy to the global budget. Also, flooded shelves are less likely to experience chemical weathering than exposed shelves. Flooded shelves build up carbonate sediment (a process releasing carbon dioxide dissolved in the sea). Thus, a rise in sea level is assumed to go parallel with long-term warming of the planet and a drop with cooling. At present, we are experiencing a rise. It may result in further rise owing to positive feedback (albedo change, lifting ice at icy margins).

As far as the carbon cycle is concerned, global changes in sea level bear on the production of hydrocarbon and of coal. Marine sediments rich in organic matter, which have the potential to deliver petroleum after being heated under pressure, accumulated especially during periods when there was extensive flooding of continents. The hydrocarbon-rich black shales of the lower Jurassic and the middle Cretaceous developed during such widespread transgressions. Transgression may provide for enhanced trapping of organic matter on the shelf. Thus, sea-level fluctuations are of great interest in *economic geology*, the science that focuses on obtaining resources by mining (Chap. 14).

Sediment mass distributions in general are thought to reflect sea-level position. For example, when sea level is high, the transport of sediments by turbidity currents will presumably be reduced. Such currents depend on a high supply of mud to the outer edge of the shelf, supply that is reduced when sea level is high, creating traps on the shelf.

In summary, changes of sea-level position through geologic time are of crucial importance in the interpretation of geologic history. However, sea level also is of central interest to people living at and near the coast. Their homes and livelihoods are at risk, whether there is a rapid drop of sea level

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_6

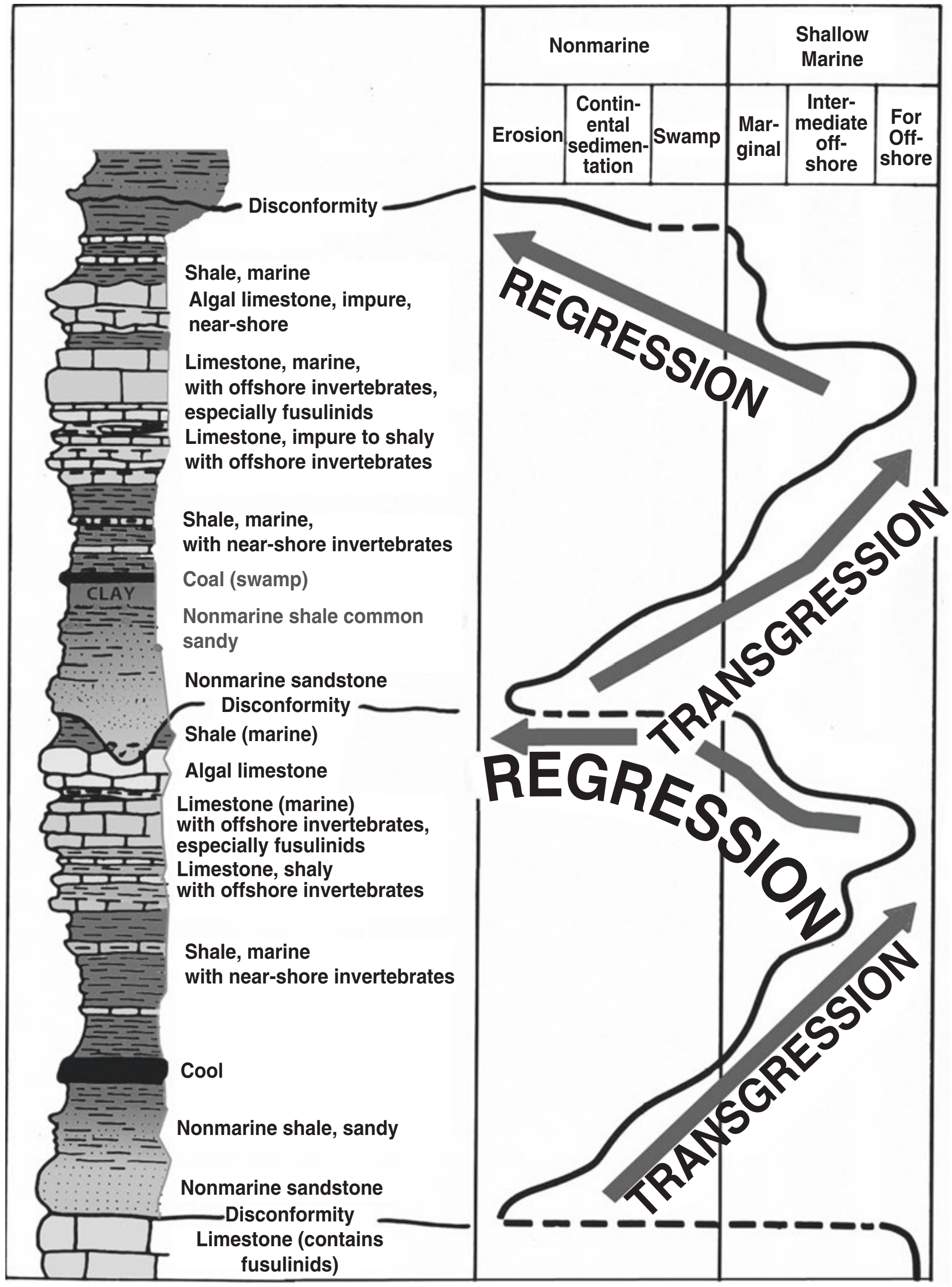

Fig. 6.1 Sea-level fluctuations as calendar of geologic history (*arrow* of time is upward, as in nature): alternating marine and nonmarine sediments in a late Paleozoic succession in Kansas. Coal layers interpreted as remains of swamp deposits. Probable habitat of marine fossils mentioned on the illustration and the curve of rise and fall of sea level drawn suggest a depth range of middle shelf to at least just above sea level for fluctuations. ("Invertebrates" denote macrofossils such as mollusks and echinoderms.) Estimates of elevation changes implied are between ca. -50 m to terrace height (not known for this time). "Fusulinids" are extinct grain-shaped foraminifers that lived on the shelf seafloor, presumably in need of sunlight for the benefit of microbial symbionts (Drawing after R.C. Moore as redrawn by J.C. Crowell, 1978. See Am. J. Sci. 278, here strongly modified)

(as in formerly glaciated regions, rising after shedding their heavy load) or a rapid rise (as observed in many places today). There is much concern that a human-caused rise will accelerate with continued and enhanced global warming from the input of carbon dioxide (Chap. 15). The rise has been short of an inch per decade for some years, but has been accelerating in recent years, according to the best estimates available. The concern thus raised has compelled the question just how fast sea level did rise, given the present outlook regarding planetary warming. The correct answer is that we do not know. However, we can study the last deglaciation period (roughly 10,000 years long) in some detail to find out about the maximum rise rates in the geologic record. Answers are given in terms of meters per century for thousands of years. The time scale remains problematic, though, as far as applications of geological findings to environmental changes of interest to humans: the best resolution achievable in non-varve marine sediments (commonly ice-age studies) is near 1000 years.

Fig. 6.2 A wave-cut terrace as an example of sea-level indicator. Enoshima, Pacific coast of central Japan (Photo E.S.)

6.2 Sea-Level Processes and Indicators

6.2.1 The Intertidal Zone

Processes in the intertidal zone include waves and currents at sea level, in narrow zones, as well as broad intertidal flats bearing tidal currents and displaying tidal variation of water cover. The tides that we witness at the coast are large rotational waves radiating from "points of no motion" situated in the central areas of the ocean basins, with the highest amplitudes on the basin edge.

Presumably, with the tides linked to both astronomy and to basin morphology, we can potentially obtain important clues about changes in the Earth-Moon system and about the evolution of basin morphology from fossils. However, convolution of possible causes normally greatly complicates interpretation of the observed record: it is most difficult to separate the causes.

Tidal flats are widely associated with river deltas as in the Bay of Bengal, at the mouth of the Amazon and the Mississippi, or as in the Gulf of Korea and off the Nile in Egypt or at many regions in the Wadden of the North Sea. Such tidal flats typically are receptacles for enormous masses of sediment, deposited as extensive beds with flat surfaces (as reflected in the name "tidal flat"). Typically, sediments in such areas display small-scale flaser- or lenticular bedding, bimodal cross-bedding ("herringbone pattern"), a variety of erosional and bioturbational markings, channel lag deposits, and graded laminae. Rain imprints and desiccation cracks (Fig. 6.3) also are seen occasionally, the latter especially in desert zones. On a larger scale, a landward decrease of grain sizes owing to decreasing current activities and time of water cover marks a contrast to normal offshore conditions.

Intertidal flats are subject to rapid changes in the conditions of sedimentation. Episodically peak winter storms wreak havoc along the coastal lowlands of the North Sea, for example. In 1362 an enormous storm caused the "Great Man Drowning" during which sea level was locally raised by about 6 m along the western Friesian Coast, where Germany borders with Denmark. The storm floods created new access for the tides into the salt marshes and moors, landward of the intertidal flats. The tides brought mud that soon covered the marshes and the peat as a *storm layer*. To the people living there, this geologically common event meant death by drowning and destruction of homes and pastures. In 1634 another fierce storm resulted in widespread flooding, destruction of villages, and the death of thousands of people. The present coastal geography in the area, in essence, is a product of those two storms, almost three centuries apart.

Typically, the geologic record preserved on slowly sinking ground in the region consists of an intercalation of peat, salt marsh deposits, marine muds, and the shells and sands of the beach. Storm deposits are common within this intercalation. Similar sequences are quite familiar from the geologic record. Groundwater below tidal flats presumably is strongly affected by changes in sea level. The same is true for the landward freshwater body, which floats on the heavier seawater. A rise in sea level is immediately felt in the landward groundwater level, together with an increase in salinity of the groundwater from mixing, one assumes.

6.2.2 Photosynthesis

Photosynthesis can only proceed when sufficient light is available. Hence, photosynthetic organisms, including the

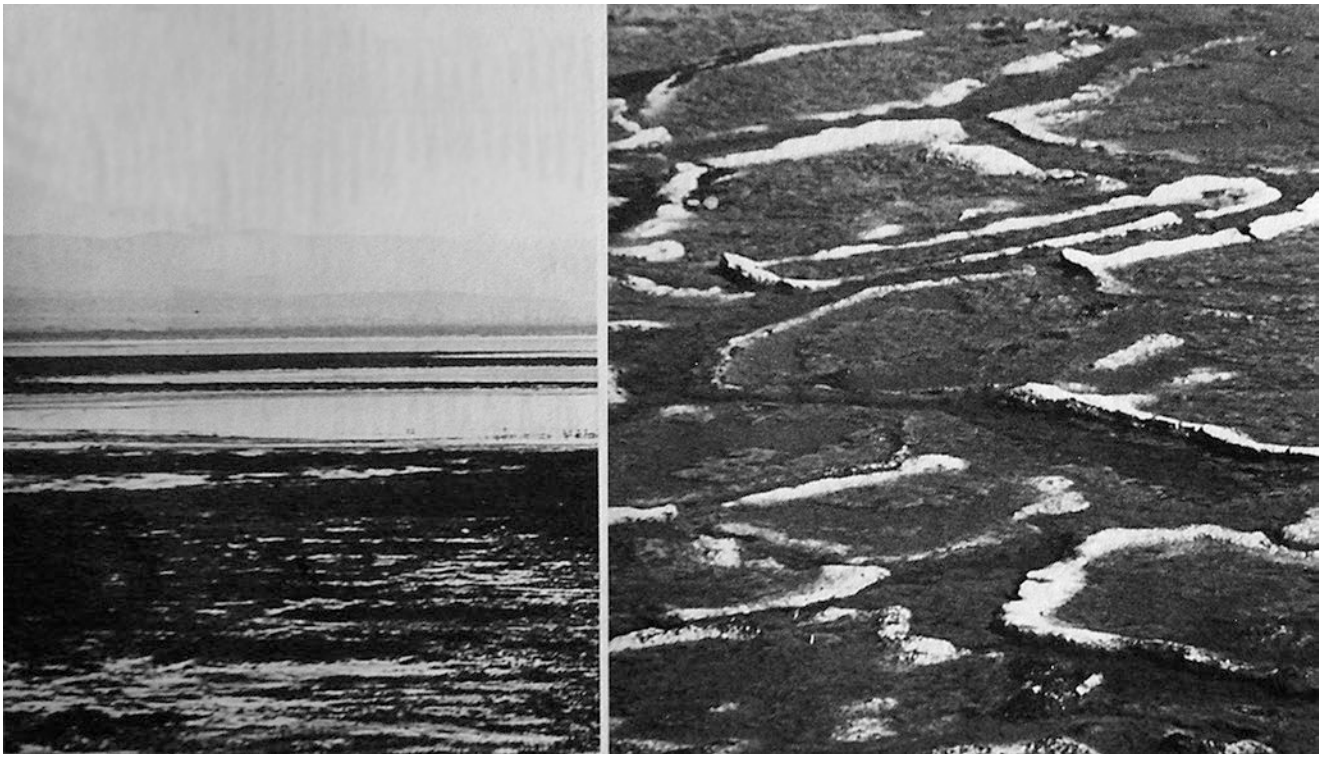

Fig. 6.3 Tidal flat at the mouth of the Colorado. Baja California (*left*) and desiccation cracks in algal mats there (*right*, 2-m-wide close-up). Algal mat fragments curl up when drying. Salt collects at upturned rims (Photos E.S.)

Fig. 6.4 Stromatolites (=mounds involving cyanobacteria and other photosynthesizing microbes) have a venerable history. The photo shows a fossil stromatolite mound in late Precambrian rocks (in this case almost a billion years old) at a roadcut in Nevada (Photo W.H.B.; penny for scale)

algal mats just mentioned, are indicators of shallow water, that is, of conditions near sea level. Stacked laminated sequences of algal mats largely constructed by *cyanobacteria (blue-green algae)* are known from the record back into the Precambrian, as *stromatolites* (Fig. 6.4). Modern examples of stromatolites were discovered in the 1950s, in Shark Bay, Australia, in a lagoon where extremely saline water excludes the chitons and snails that normally graze on the algae. Apparently the algae (or several dominant species among them) are not vulnerable to the high salinity. Stromatolites also were found in the Bahamas and in the South Pacific. Thus, they are not restricted to one particular environment, and their survival may have aspects differing from those in Shark Bay. The observed layering in stromatolites has been linked to tidal action, with implications for reconstruction of the slowing of *Earth's rotation* and acceleration of lunar orbital velocity over geologic time.

The high algal production in tidal flats presumably has implications for the abundance of benthic organism there, such as algae-eating mollusks. In Neogene rocks, by analogy with present conditions, a high abundance of alga-fed mollusks such as certain species of the snail *Turritella* and certain bivalves including oysters presumably therefore indicates a tidal flat or lagoonal conditions close to sea level. Likewise, sessile plants such as the geologically important calcareous algae and animals living in symbiosis with algae (e.g., many corals and certain foraminifers) indicate growth in shallow water near sea level.

Some caution is indicated, however, when using fossils to determine sea-level position. For example, modern stony corals of the genus *Lophelia* occur well below the photic zone on much of the seafloor in the North Atlantic. Thus, their presence does not indicate nearshore or sunlit conditions, even though very similar forms do live close to sea level.

6.3 Coastal Morphology and the Postglacial Rise in Sea Level

6.3.1 General Effects of the Recent Sea-Level Rise

In order to understand the coastal landscape, we must constantly be aware of one central fact: the recent rapid rise of sea level that accompanied the melting of polar ice masses after the last glacial maximum (LGM) about 20,000 years in the past to the end of deglaciation melting some 6000 years ago. Radiocarbon ages are slightly off (too young), commonly by more than 10% and with a varying error size (Fig. 6.5), which affects assigned timing of sea-level change and apparent rates.

The range of the motion exceeded 120 m (400 feet); that is, sea level typically rose from the average depth of the shelf edge to the present position. The melting of ice that produced the rise took place mainly on the Canadian Shield (labeled *Laurentia* for ancient times and rocks) and in Scandinavia. Antarctica apparently contributed also; the proportion is a matter of research and discussion. There is some indication in the data (especially information from ice cores) that deglaciation proceeded in pulses. Most researchers assert the presence of two major pulses (one before and one after the no-melt millennium referred to as the "Younger Dryas" (see Chap. 11)). There may have been more pulses, of course, on a shorter time scale.

In southern California, uplift of the margins is reflected in cliffs, and the rise of sea level is recorded mainly in the wetlands between the cliffs (Fig. 6.6). Thus, the coastal morphology here is a product of both tectonic forces and processes associated with sea-level position. Raised marine terraces are a feature caused by this interplay. In contrast, the coastal morphology of sinking margins is dominated by the processes associated with the sea level itself, the tectonic factors being entirely submerged and out of view. In North America in general, the Gulf of Mexico and the East Coast provide prime examples for this type of morphologic development; in Europe the effects of large-scale sinking are seen in the "Wadden Sea" mentioned earlier and in fact in all of the northwestern flatlands, emerged or not.

North America's East Coast has spectacular examples of *drowned river valleys*, valleys once carved when sea level was low due to ice buildup in high latitudes and later flooded when the ice melted. Prominent legacies of the deglacial sea-level rise include *Chesapeake Bay,* the mouth of the *Saint Lawrence*,

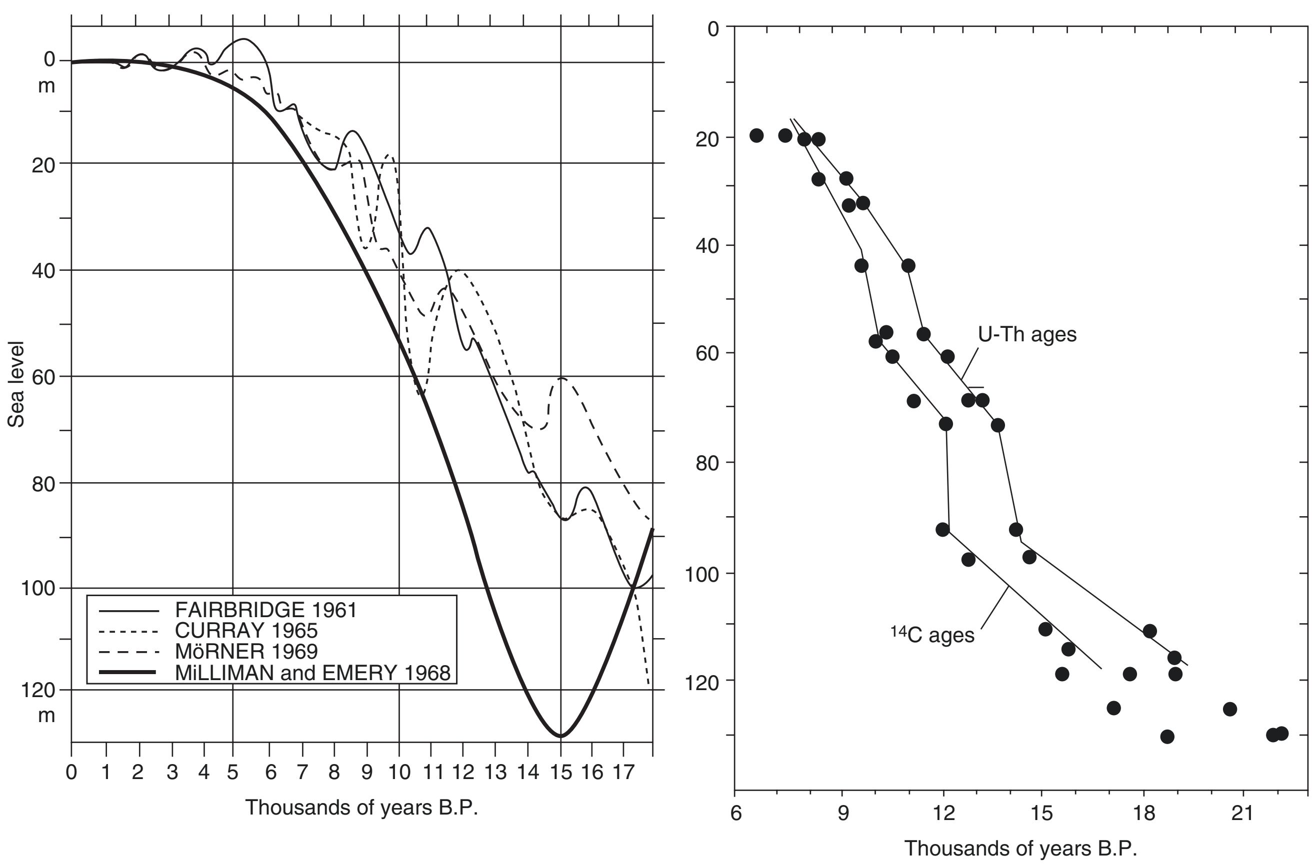

Fig. 6.5 Rise of sea level during deglaciation by various authors. *Left*: early estimates (Compiled by E.S. in R. Brinkmann,1974. Lehrbuch der Allgemeinen Geologie, Enke, Stuttgart). *Right*: age and position of corals on Barbados (*Acropora palmata*) during deglaciation (Courtesy Eduard Bard. See Bard et al., 1990. Nature 345:405; radiocarbon ages by R. G. Fairbanks, 1989. Nature 342:637) Note the 3000-year gap centered near 13,500 radiocarbon years

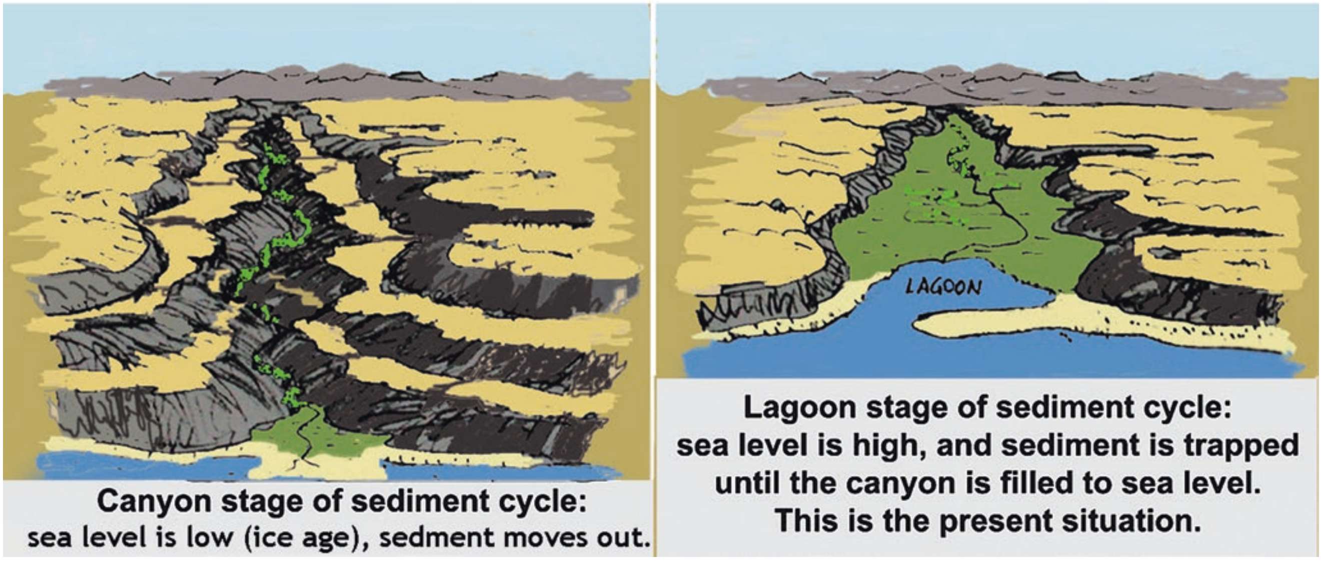

Fig. 6.6 Cliff and beach morphology of southern California and its origin (W.H.B., 2013. Coast to Crest and Beyond. F. L. Parker Education Program, S.I.O., UCSD)

and the lagoon landward of *Long Island* in New York. In addition to providing excellent harbors, drowned rivers trap sediment and starve the shelves offshore. Much of what we find on the shelves is relict material from the last ice age and not indicative of processes active there now.

Sea level is rising right now, after a long period of stability during the Holocene (i.e., the last 10,000 years). During the last few decades, it has risen by some 1–2 mm per year, with indications for acceleration with time. Presumably, the rise is forced by thermal expansion of the upper ocean layers, as a result of global warming, by melting of mountain glaciers, and by some input from polar ice (ready to contribute at the rate of meters per century, based on the record of deglaciation (see below)).

For any specific (i.e., localized and actual) rise of sea level, there is always the question of the proportion of the rise, if any, that must be assigned to local sinking of the ground. Clearly, it is also of interest, for northern latitudes, just how local uplift was influenced by the removal of the enormous ice masses that once depressed the crust, and how the crust responded to unloading by the melting of ice and to loading by addition of water on the shelf. Also, the rotation of the Earth is affected by such motions and by melting. Observed changes in rotation can be used to check on certain ideas about causes of ongoing sea-level change.

6.3.2 River Mouths

A glance at a map of the world shows that river mouths are not necessarily indentations, reflecting a rise of sea level. Many of them are represented by protrusions of the coastline – that is, by delta deposits. Evidently, the amount of sediment supplied by the river or rivers entering an estuary is crucial for the creation of the associated landscape. Tidal motions can reach far into river mouths, indented or not. Intrusion by marine saltwater bringing sediment is readily recognized by the fossil content of the sediment such as echinoderm remains, foraminifers, and marine ostracods. In southern California, the presence of kelp plants (and thus marine organic matter and fossils including bryozoans growing on kelp) may indicate the occasional influx of marine materials into normally continental water bodies, forced by tidal action.

Why a delta forms at a river mouth and not an estuary depends on many factors, besides the ratio of sediment supply to the energy available for redistributing the sediment. Young estuaries may reflect the fact that there has been insufficient time to fill up the drowned mouth of a river. Alternatively or in addition, much of the sediment supply may have been lost across the adjacent shelf, by moving down an ice-age channel continuing the river bed toward the uppermost slope. Examples of such processes are seen at the Congo and Hudson rivers. The best harbors, therefore, are those adjacent to a submarine canyon, where sediment is moved out, as off New York, or those where strong tidal action keeps the outer river channel open, as in London, Bordeaux, or Hamburg.

For a geologist who must learn to read the record in ancient deposits, the delta structure is of special interest (Fig. 6.7). Where coast and sea level move neither up nor down very much and a delta builds out to sea, the marine conditions retreat as the terrigenous material builds up. As a consequence, the record shows *regression*; that is, shallow deposits overlie deeper ones in accordance with *Walther's*

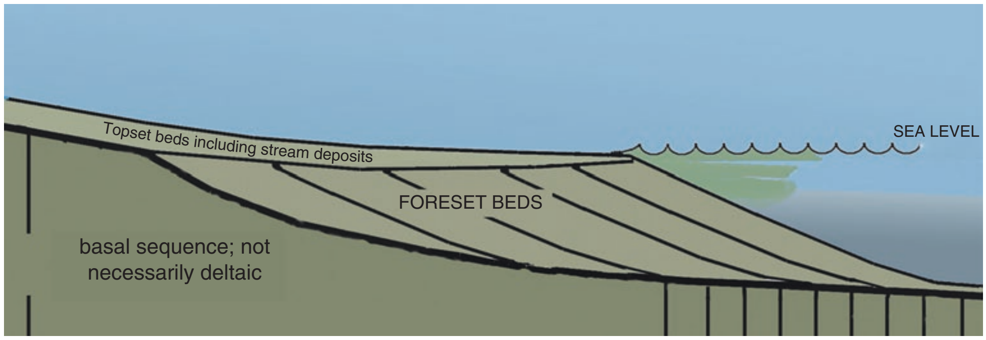

Fig. 6.7 Schematic structural profile of a delta deposit, showing positions of topset, foreset, and bottom-set units (After a graph of US NEL geologist R. Dietz, as reproduced in A. Lombard, 1956. Géologie Sédimentaire, Vaillant- Carmanne, Liège. Modified and with color here added)

Law. In accord with that same rule, a sufficiently rapid rise of sea level or sinking of the coast will result in a *transgressive sequence,* with marine sediments deposited at greater depth overlying the more continental delta deposits.

Deltas are economically interesting because of the *oil and gas* potential of delta deposits. High organic productivity and enhanced preservation of organic matter by high accumulation rates may form source rocks. Sands in distributary channel fills or in offshore bars at the delta front can become reservoir rocks (see Chap. 14).

6.3.3 Lagoons and Barrier Beaches

Next to deltas, we commonly find low-lying coast bearing lagoons or barrier beaches or both. Different types of sediment (e.g., lagoonal *sedimentary "facies"*) tend to parallel the coastline, with offshore bars or barrier islands commonly being accompanied by a sand beach. The beach may be backed by dunes consisting of beach sand gathered by the wind, sometimes strong glacial-age wind. The beach-dune complex in turn may form a seaward barrier for landward lagoons, as seen along much of the coast of the Gulf of Mexico and along the East Coast. The barriers are narrow and can be tens of miles long. Rivers emptying into the lagoons may cut one or several channels through the barriers creating a series of barrier islands. From the seaward side, storm waves may break through a barrier, forming *overwash fans* (Fig. 6.8). Tidal action tends to keep the channels open, building deltas on both their sides, conspicuously on the quiet lagoon side where regular wave action is negligible and storm-induced wave height is modest.

How does barrier-and-lagoon morphology reflect changes in sea level? Again, the rate of supply of sediments to the coast is a crucial factor. When sea level stabilized globally, some 6000 or 7000 years ago, sediment supply became very important indeed in determining landscapes. Off Galveston Bay, for example, a large supply of sand allowed a rapid building out of the offshore barrier (Fig. 6.9). In general, however, barriers and lagoons must have migrated landward

Fig. 6.8 Overwash fans on St. Joseph's Island, Texas, produced by storm waves. Open Gulf to the left, lagoon to the right (Air photo courtesy D.L. Eicher, Boulder, Colorado)

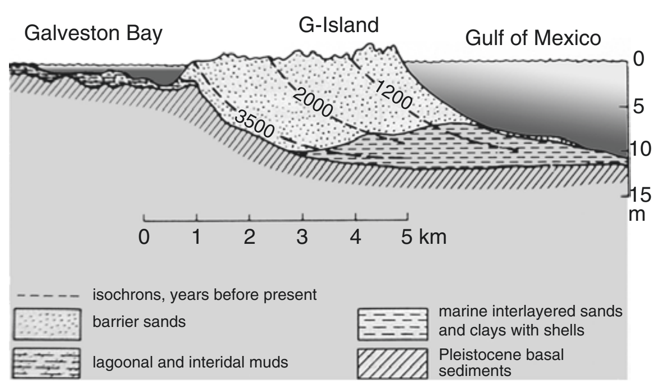

Fig. 6.9 Structure of Galveston Island. Regressive sequence below the barrier beach near Galveston, due to high supply of sediment. Age: late Holocene (From J. R. Curray, 1969. In: D.J. Stanley (ed.) The new concept of continental margin sedimentation. Am. Geol. Inst., Washington, D.C. Shading here added)

during the postglacial rise of sea level in the past 20,000 years, including in the early Holocene.

The barrier-type coast is very abundant: of 244,000 km coastline, its share is about 32,000 km, that is, 13%. North

America and Africa each have about 18% and Europe about 5%. A stable broad shelf and a high supply of sediment apparently are favorable for the development of barrier coasts, as illustrated by the barrier island coast of the southern North Sea. Coastal uplift is unfavorable for building a barrier island coast, as are narrow, dissected, and starved shelves.

6.3.4 Mangrove Swamps

During the postglacial rise of sea level in the tropics and subtropics, one of the most impressive migrations must have been that of the mangrove swamps landward. Mangrove growth dominates many intertidal zones in the tropics (Fig. 3.4). We see mangroves on Hawaii, in Florida, on Australian shores, on the Galapagos Islands, on tidal flats of Baja California, and indeed at the shores of tropics and subtropics all around the world. Several species are involved, species that are not all closely related. Some of the plants are high enough to represent trees; most are large shrubs.

Mangroves need year-round temperatures of 20 °C (ca.70 °F) or higher. In equatorial regions with high and variable rainfall, mangroves thrive, thanks to the high temperatures and a tolerance to salinity variation over a wide range. Off Cameroon and Guyana, for example, mangrove forests form a broad belt offshore and blend into the tropical rainforests at high tide level. The landward expansion and subsequent burial of mangrove swamps during deglaciation must have produced an organic-rich layer on many tropical shelves, much as peat growth did in higher latitudes.

Mangrove swamps, like coral reefs and other types of coastal bio-zones, are quite sensitive to disturbance by human activities. Changes in coastal habitats, whether by nature or from human impact, are being monitored using *remote sensing* from satellites. This type of information can greatly increase our appreciation of the rates of change in tidal flats, mangrove swamps, deltas, lagoons, barrier islands, and coral reefs. Also, it provides clues regarding the processes at work, including human impact through land use, pollution, and rise of sea level. Whether or not the increase in knowledge results in remedial action is a question concerning politics; that is, the question goes well beyond documentation and scientific analysis. The subject is correspondingly difficult (Chap. 15).

6.4 Ice-Driven Sea-Level Fluctuations

6.4.1 The Würm Low Stand

Sea-level change in response to the waxing and waning of large continental ice masses has been part of geologic history for hundreds of thousands of years (Chap. 11). As far as is known, the rise of sea level during the last deglaciation, witnessed by our ancestors and commented upon in ancient myths, was as big and as fast as any of such events ever (Figs. 6.5 and 6.6).

Rapid deglaciation (10,000-y scale) was the result of a maximum change of climate: from a peak cold period to a peak warm period. About 20,000 years ago, during the last ice age (the *Würm* in Europe, the *Wisconsin* in North America), enough ice was tied up on land (especially in Canada, but also in northern North America in general and in Scandinavia) to depress sea level (in low latitudes) by somewhat more than 120 m (or 400 feet). The sea-level drop being the size of average maximum shelf depth (according to F.P. Shepard), the effect on shelves was most notable. Large shelf areas fell dry as a result of the ice-caused regression of the sea. Rivers then crossed the shelves, entering the sea at the shelf edge and cutting backward into the shelf, making canyon heads. Their sediment load, dropped at the point of entry, apparently became unstable in many places and slid, starting turbidity currents rushing down the undersea canyons. In front of glaciers on the shelves outwash plains developed (or else moraine ridges) and dune fields could then evolve, which in turn could become sources for beach sand and other marine sediments.

Land animals spread over the emerged shelves (even in moderately high latitudes) and in places used them as land bridges to move from one continent to the next (as mammoths and other large mammals apparently did in Beringia, the shelf connecting northeastern Siberia with Alaska). In places (e.g., the North Sea), the remains of mammoths and of other mammals are found on the sea bottom now, together with tools of prehistoric hunters. It was the famous French naturalist and paleontologist Georges Cuvier (1769–1832) who first proposed that floods were a cause of extinction, a notion presumably based on this type of evidence from the seafloor.

6.4.2 Ice-Age Fluctuations

Ice-age fluctuations are the result of cyclic ice buildup and destruction, linked to solar input variations in high northern latitudes, from orbital variations. Throughout the late Pleistocene, major transgressions (caused by deglaciation) apparently were much more rapid than any of the regressions (caused by ice buildup). Building up large ice masses took much longer, one suspects, than melting them. The evidence that this is so provided by oxygen isotope series, which were introduced to the arguments about rates of sea-level change by the Italian-American Cesare Emiliani, as mentioned in the introductory chapter (Fig. 1.5) and have become very prominent in the reconstruction of ice-age fluctuations since (Fig. 6.10).

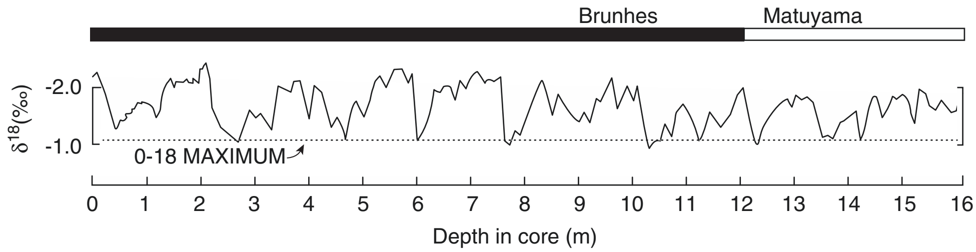

Fig. 6.10 Fluctuation in the oxygen isotope index of the planktonic foraminifer *Globigerinoides sacculifer* in a long core from the western equatorial Pacific. "Brunhes" denotes the present "normal" magnetic epoch. Its onset was later dated at 780 or 790 millennia ago. "Matuyama" is the name of the previous magnetic epoch, with a "reversed" field (Shackleton and Opdyke, 1973, Quat Res. 3:39) Note the evidence for maximum ice buildup (here added as dotted line), suggesting that controlling factors limit ice buildup

The determination of rates, of course, implies the knowledge of time differences within a record. Thus, for detailed analysis, it is absolutely necessary to correctly date the oxygen isotopic series. Three methods are widely used: firstly, a tie-in to the paleomagnetic time scale (notably the age of the Brunhes-Matuyama event). At the B/M event and other similar paleomagnetic events, the direction of the Earth magnetic field changed from "reversed" (magnetic North Pole in the south, Matuyama Chron) to the present "normal" situation (magnetic North in the north, Brunhes Chron). The event occurred some 780,000 years ago, as determined by dating using long-lived radioactive isotopes.

The age thus determined agrees with dating by the second method, that is, tuning to Milankovitch forcing. The *tuning* involves using astronomically defined cycles within the solar system as a yardstick. The dates have been calculated to great precision for millions of years into the past (by the Belgian astronomer André Berger and his associate Marie-France Loutre and subsequently by his colleagues extending the time span). By this second method, the position of the Brunhes-Matuyama boundary on the isotopic cycles emerges as about 780,000 years old, providing evidence that Milankovitch was right in principle and thus that tuning the marine isotope record to astronomically determined fluctuations works for dating. Other evidence also exists, notably the parallelism of Milanovitch forcing and changing elements of the deep-sea record discovered by J. Hays and colleagues.

The third method – the oldest among the three that work – is the dating of the last high stand of sea level, using coral from that time and dating it by means of long-lived radioisotopes (isotopes in the uranium series, not radiocarbon, which becomes entirely untrustworthy before 50,000 years ago because of its high decay rate and resulting low abundance). The various ages obtained for the time of high sea level (the *Eemian* in middle Europe, the *Ipswichian* in Britain, and the *Sangamonian* in North America) peak in the vicinity of 125,000 years ago, as expected for a Milankovitch time scale.

6.4.3 Effects on Reef Growth

The sea-level fluctuations of the Pleistocene left their imprint in various depositional systems including shallow-water carbonates, notably in tropical reefs. Each high-stand of the sea level resulted in the buildup of reef carbonates, while the low-stands resulted in erosion. In fact, the origin of atolls, those ring-shaped islands dotting the tropical Pacific, has been contemplated under this aspect in the early twentieth century by the Harvard geologist Reginald A. Daly (1871–1957). Thus, while Darwin's hypothesis of atoll origin invoking submergence of islands remains interesting for many aspects of atoll development, the influence of the fluctuating sea level in the quaternary must not be forgotten. Fluctuations of sea level within the ice ages were large and fast; they *must* have affected coral growth. There is a reason that there are so many fast-growing species among modern stony corals. Presumably corals were selected for fast growth during deglaciation – high growth rates of certain (abundant) species of coral may well be a legacy of the ice ages.

The depths of the lagoons, and their mud-free areas, may owe much to karst processes from tropical rainfall during glacial-time exposure, as suggested by the carbonate expert E. G. Purdy (Houston, Texas) and the S.I.O. geologist E.L. Winterer. In defense of Darwin and his neglect of ice-age cycles, at the time he introduced his widely cited ideas about the origin of atolls, the discovery of multiple ice ages was still to come (see Chap. 11).

6.5 Reconstruction of Ancient Sea-Level Fluctuations

6.5.1 Mesozoic Sea-Level Variation

Sea-level fluctuated considerably all through the Phanerozoic, including periods when no large ice masses were present, as far as known. Fluctuations since the Jurassic are of special interest: essentially they determined the sequences of sedimentary layers seen today within continental margins, including layers of economic interest. For a full understanding of the nature of these sediment bodies and their environment of deposition, we would like to know how sea level fluctuated over the last 150 million years. However, inasmuch as some Mesozoic variations apparently were not driven by waxing and waning of ice masses, they were not recorded in the isotopic composition of seawater in the shells of organisms then growing. Thus, we have to find ways to reconstruct the fluctuations from other types of information.

6.5.2 Sediment Bodies as Indicators

An intensive worldwide seismic exploration of continental margins after WW II has resulted in the proposition that the various reflector series seen show striking similarities between different locations. Sea-level changes have been suggested as the common cause responsible. In support of this hypothesis, the Texan geologists P. Vail, R.M. Mitchum, B.U. Haq, and their associates developed concepts and methods to derive sea-level fluctuations from the geometry of sediment layers as seen on seismic reflection records. For the time since the beginning of the Triassic (some 250 million years ago), they have had to postulate more than 100 major global sea-level changes, about one for every 2 million years, on average, to take account of certain changes in sedimentation pattern. The basic idea is rather simple (Figs. 6.11 and 6.12): During a relative rise of sea level (i.e., a transgression), sediment layers expand into shallower water and they become wider in the process. Conversely, during a fall of sea-level (regression) erosion sets in on the exposed shelf.

The erosion produces a *hiatus,* that is, a surface that joins older and younger sediments in a discontinuous way. The course of regression is poorly documented, therefore. The record looks as though regression happened on a short time scale, there being little to show for an erosional event.

Employing the concepts illustrated in Fig. 6.11, P.R. Vail, B.U. Haq and their associates reconstructed the geologic history of sea level for the entire Cenozoic (Fig. 6.12) and beyond. For the Cenozoic, there emerges a sudden large drop in sea level in the middle of the Oligocene (near 30 M years ago in their time scale), rather than near the end of the Eocene, as expected from other indicators. Whatever the proposed explanation for the proposed drop in sea level, the Oligocene apparently was cool enough so that a large regression is likely to involve large polar ice masses.

For the middle Cretaceous, the sediment piles of the continental margins and of shelves in general show the widespread global transgression that has been widely recognized for more than a century (Chap. 13). The enormous transgression in the middle Cretaceous presumably implies tectonic effects, such as an appropriately large change in the shape of the mid-ocean ridge, as proposed by the Lamont geologist Walt Pitman. In essence, Pitman's suggestion involves changing the average depth of the ocean by changing the temperature of the lithosphere to achieve general transgres-

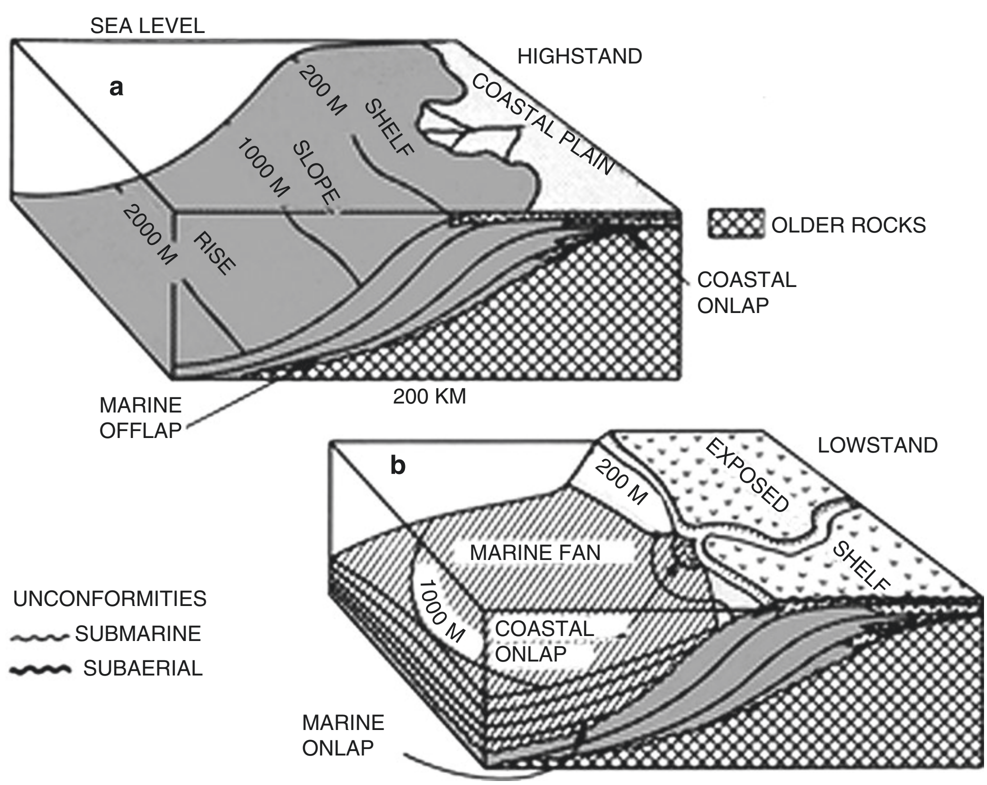

Fig. 6.11 Effect of sea-level position on depositional pattern at the continental margin (Courtesy Peter Vail; see P.R. Vail et al., 1977, Am. Assoc. Petrol. Geol. Mem. 26:49)

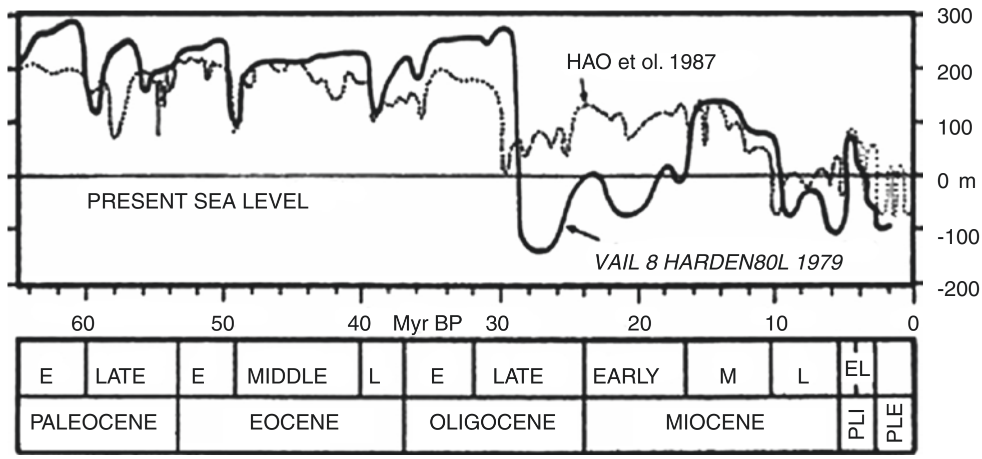

Fig. 6.12. Relative changes of sea level in the Cenozoic as deduced from the geometry of sediment bodies in the margins (Data used in P.R. Vail and J. Hardenbol, 1979, Oceanus 22:71; Haq et al. 1987, Science 235:1156; combined and redrawn by W.H.B. and L.A. Mayer, 1987, Paleocngr.2: 620)

sion or regression. The process is slow but presumably can be very effective on long time scales.

6.5.3 Problems Arising

The reconstruction of past sea-level positions is of prime importance in the study of geologic history, and the *Vail sea-level curve* has become a much-used tool for the correlation of seismic sequences of continental margin sediments and for doing historical geology of the last several hundred million years. Overall one chief problem persists: the sea-level changes deduced from sediment bodies call for a steady supply of sediment. With mountain building and a general cooling of the planet came a buildup of ice and an increased supply of sediment during much of the Cenozoic. Thus, conditions changed with global cooling in a way suggesting overall regression. In addition, assigning ages to the remotely found sediment bodies (mostly found by seismic profiling) can pose serious problems unless there is information from drilling.

While the Pitman mechanism for large-scale transgression (acceleration of seafloor spreading) is highly plausible, it is not always readily documented. For example, there is indeed evidence that spreading rates of the MOR did change and that they were higher than now in the Late Cretaceous, supporting Pitman's hypothesis in principle. However, large outpourings of basalt on top of oceanic crust must be considered as well. We do not know the necessary details about volcanogenic edifices on the seafloor.

Sea-level position is certainly expected to respond to a gradual change in the elevation of the global seafloor, the question remains how sea level can be changed in other ways and faster than in tens of millions of years. Mountain-building involving shallow crust stacked up within mountain ranges is such an alternative way to change sea level. For example, if one assumes a doubling of the continental crust for making the Tibetan Plateau, the corresponding drop in sea level would be ca. 40 m. However, it takes a long time to double the thickness of the crust (just like changing spreading rates does, so the improvement in rates may be marginal).

The quickest way to change sea level not using ice is to empty or to fill an isolated ocean basin, such as happened at the end of the Miocene to the Mediterranean Sea. The water that filled the basin before the great drying had to go elsewhere, presumably raising sea level of the global ocean by some 10 m. On filling up the basin, conversely, sea level dropped by the same amount (roughly ten meters). These are, however, small amounts in view of the hundreds of meters of sea-level rise postulated for the middle Cretaceous transgression in North America.

The examples here given illustrate the complexity of sea-level processes, complexities faced by geologists when attempting to interpret the geologic record of sedimentary rocks on land. Likewise, unfortunately, in our task to learn from the present for understanding the past, we face important obstacles. Perhaps the most vexing consists in the fact that the present ice-dominated period is not typical for much of geologic history. Another is represented by the problem of time scale mentioned in the preface – our short-term observations and measurements may not reflect properly the processes linked to the enormous time spans available in geologic history.

The time scale problem is well illustrated by the sinking of Venice, which reflects both geologic processes and human impact.

6.6 Sea Level and the Fate of Venice

6.6.1 Venice Is Sinking

The ancient city of Venice, with its San Marcus Cathedral, its palaces, and its canals, is slowly sinking below sea level, causing great concern in Italy and elsewhere. The stairs of entrances to mansions next to canals, once used

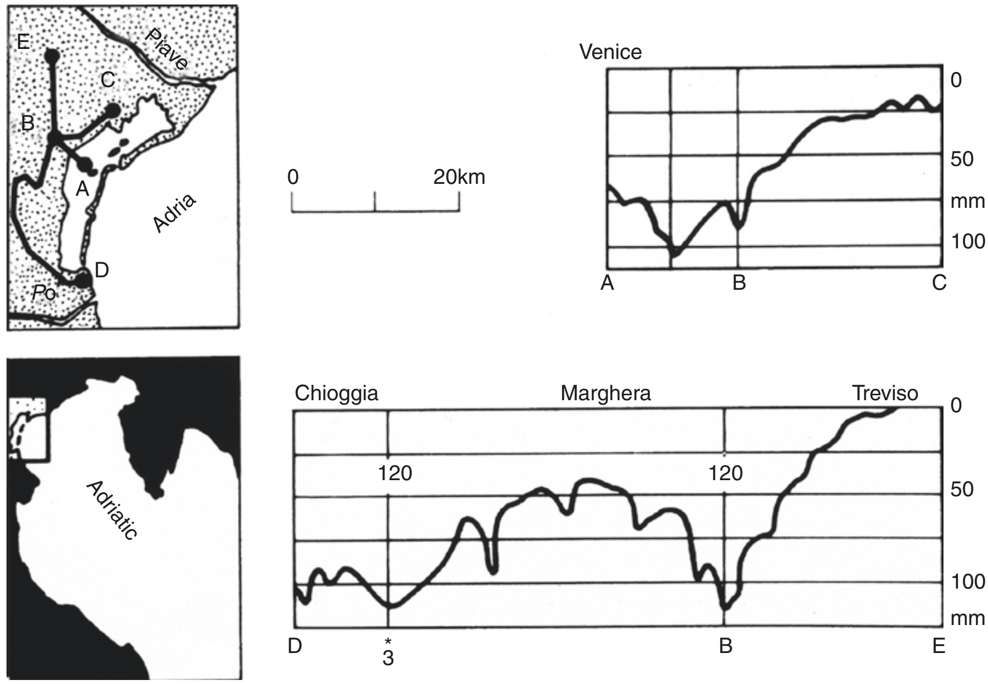

Fig. 6.13 Rates of subsidence in and around Venice. *Left*: location maps. *Right*: Total subsidence in the period between 1952 and 1968 at the profiles indicated in the map of profiles (*upper left*, *A-B-C* and *D-B-E*). The zero point is defined as such. The center of the city of Venice subsided by about 80 mm, that is, at a rate of 5 mm per year (Data from R. Frasetto, 1972. CNR-Lab. Stud. Din. Masse Tech. Rep. 4)

by people debarking from water taxis, are covered with seawater. As is obvious to every visitor, they are no longer functional but record conditions of a time long gone, when sea level was lower relative to San Marcos Square. Can Venice be saved? Let us take a look at the formidable scale of problems associated with the task of understanding the rise of the sea within the city. The necessary information has been gathered by the *Laboratorio per lo Studio della Dinamica delle Grandi Masse* of the *Consiglio Nazionale delle Ricerche*.

Venice was built surrounded by water many centuries ago. The city lies at the rim of the Po delta, in a lagoon protected by a barrier island. The setting presumably discouraged laying siege to the city. The Lido is breached by tidal inlets. That the city is sinking is rather evident: docks must be built up, entrances are bricked in, steps are under water, and the city is easily flooded during high water. The extent of sinking was determined by careful geodetic measurements starting with Treviso as a fixed point (Fig. 6.13). Between 1908 and 1925, Mestre-Marghera sank by around 0.15 mm each year. Between 1925 and 1952, the rate was 0.7 mm per year; the rate of sinking increased dramatically during the late 1950s and 1960s. Between 1952 and 1968, it rose to 3.8 mm per year (0.15 inches) in many parts of the city. Results are highly unwelcome, of course.

More recently (in ca. 2010), optical echo measurements by satellites suggest subsidence values up to 10 mm per year in places. A natural background sinking of around 1 mm per year has been proposed, with the remainder attributed to human activities. To evaluate such findings, we have to consider the causes for the sinking in some detail.

6.6.2 On the Causes of Sinking and the Rising of Sea Level

One cause for the sinking of Venice is the general sinking of the Po delta. It was found by drilling that there are more than 3 km thick Quaternary sediments below the central delta. If we set the "Quaternary" equal to around two million years, we obtain a subsidence rate of about 1.5 mm per year. Venice being at the rim of the delta, its share of this natural subsidence would be somewhat less. Thus, 1 mm per year apparently is a defensible guess for long-term geological subsidence.

The Quaternary sediments underlying the region are partly marine and partly continental, reflecting fluctuations in sea level resulting from both tectonic movements and changes in the sediment supply by the river Po. This finding warns us that natural fluctuations can mask the regional background sinking. As it happens, the fluctuations also are of the order of 1 mm/year or somewhat larger. Another component of uncertainty is added by compaction of sediments. The uncertainty can be locally reinforced by building and by infilling of lagoon areas. Compaction may be one reason for increased sinking in the center of the city. Although the effects from compaction cannot exceed the rate of growth of sediment, of course (i.e., about 1 mm per year), an additional sinking of 0.5 mm would seem conceivable from this source.

The sand layers making up much of the underground of the city are groundwater reservoirs (*aquifers*). When groundwater is removed faster than it is replenished, the ground sags. This type of subsidence can reach over considerable distances away from the pumping stations

Fig. 6.14 The flooding of San Marcus Square in Venice by more than 1 m in November 1966 (Photo courtesy A. Stefanon, Venice)

(1 km and more) if the pumping is done from a depth of several hundred meters. Pumping apparently was an important contributor to the sinking in the twentieth century. It has been stopped, but the sinking did not, since other causes persist (and new ones arise).

The normal tidal range is about 1 m at Venice. However, storms can considerably increase the rise of sea level produced by a high tide, even doubling it. It is no longer unusual, therefore, for San Marcus Square to be flooded (Fig. 6.14). It now happens several times each year, and the frequency may be increasing. The access of storm tides to the city presumably was facilitated by filling the lagoon (i.e., reducing its effectiveness as a buffer system) and by deepening shipping channels. Unintended consequences when applying a fix for unwelcome phenomena are quite common, of course.

The results obtained from the studies summarized above have a message: slow processes can have effects on a short time scale. Also, they suggest that the curtailing of groundwater pumping by government regulation was the correct decision, even though it did not remedy the entire problem. An effort to regulate tidal access by adjustable mechanical barriers holds promise. However, the problem of removing waste persists.

It is difficult to envisage allaying both the problems generated by geologic sinking and by the larger man-made rise of sea level. Sea level indeed is rising from general warming. Thus, local remedies may well be eventually overridden by a global effect.

Incidentally, not only Venice is threatened by sinking and by a rise of sea level, but so are many other large coastal cities including Bangkok in Thailand, Manila in the Philippines, New Orleans in Louisiana, Jakarta in Indonesia, Dhaka in Bangladesh, and Shanghai in Eastern China. Tens of millions of people are affected directly by the rise of sea level and many more indirectly. While groundwater pumping can be stopped and tides can be dealt with, nothing can be done about a general regional sinking. We cannot control geologic factors; we must live with them. Whether we can reduce or eliminate the rise of sea level from a general warming, ongoing and projected for the future, remains to be seen. Although the problem is identified, the political will to find remedies is not yet documented beyond doubt.

6.7 How Fast Did Global Sea Level Rise?

As is obvious from our discussion, the observed general current rise of sea level adds materially to all problems generated by regional subsidence. The current global rise of sea level has become a controversial subject for several reasons (see Chap. 15): Even though it may no longer be difficult to measure, its course remains extremely difficult to predict. To predict is to estimate. Estimates are bound to differ, depending on the person or agency doing the guessing. Here we use estimates published by the IPCC (Intergovernmental Panel on Climate Change). Its guesses presumably are relatively free of personal bias. We note from the relevant reports that the reported and projected rise of sea level are commonly of the same general magnitude (1 to several mm per year) as is the regional sinking rate for Venice and that there is an upward trend of measured rise values suggesting acceleration.

In recent years, relatively reliable information comes from satellite observation. Guesses for future development are obtained by extrapolation. Clearly, results will differ depending on assumptions concerning any acceleration of the trends seen. In 2007 the IPCC estimate of a rise till the end of the century (4th assessment) was one fifth of a meter. Taking then discussed acceleration into account, one might get one third of a meter. A rise of more than 1 m (well in excess of the IPCC estimates) has been proposed by some US government scientists. In fact, one distinguished climate researcher has issued warnings in a popular book about the possibility of a rise of several meters within the twenty-first century. He thinks we might be risking a Venus-like destiny, while betting on the future of the planet.

What, if anything, can the science of marine geology contribute to this discussion? Some insights are obvious, actually. Mainly, the time scales involved in geologic and human-impact research commonly differ by very large factors. What can be estimated in the deep-sea record is the maximum rate at which sea level has risen in the past on a millennium scale, once the problem of dating has been solved satisfactorily. The variation in sea level over the last million years is well known, compared with all the rest of geologic time (for which all age estimates are comparatively uncertain). The reason for our remarkably high confidence regarding the last million years is the application of the oxygen isotope method (with Milankovitch theory for precise dating) to a great number of cores from the deep seafloor. Available data were compiled by James Zachos at the University of California, Santa Cruz, and his colleagues (Fig. 1.4).

By making the Emiliani assumption that the oxygen isotope changes due to temperature changes went parallel to the changes in ice mass, we can interpret the isotope record purely in terms of sea level. We simply take the range in sea level for the last big change (say, 125 m) and set it equal to the range of the oxygen isotope index at the time of deglaciation. Next, we smoothen the binned distribution of calculated rates of change (which reduces bin contents with extreme values). The result emerges in the shape of a skewed bell-shaped curve, with a likely general distribution of past rise values and a few outliers (Fig. 6.15). For our purposes, we assume that only values for relatively warm conditions are relevant. This assumption reduces our number of cases to 47 (i.e., about 5% of what is available in principle). We preferentially employ data from postglacial periods for our analysis, as seems proper for our warmish (interglacial) time.

About one half of the cases (50%) show very little change; that is, sea level stayed roughly the same for 1000 years. Two cases show the average deglaciation rise of 1.3 m per century, the value that we get when considering the last deglaciation only, on a 10,000-year scale. It is remarkable that such a high rate is still possible even on a millennial scale after much of the vulnerable ice (the ice that fluctuates readily) has melted and disappeared. For 2 m per century, however, we seem to be close to a very low likelihood for the conditions we have chosen as relevant (we find that less than 1% of cases fit the 2-m-per century criterion). To what degree the distributions found in the present analysis are relevant to the current rise in sea level, of course, is another question. The answer is by no means clear.

What we *can* conclude from this exercise is this: In the not-so-distant geologic past, sea level rose by 1 m/century without much provocation about 4% of the time under conditions somewhat similar to the present ones (except for much slower change then). With the ongoing fast warming, therefore, the chance for a similar rise rate or a faster one might be a serious possibility. The other item of interest in Fig. 6.15 is the asymmetry of the distributions, suggesting the presence of positive feedback on the side of the rise of sea level (i.e., the destruction of ice produces more destruction). This finding is robust. Also, it is notable that sea level was several meters higher than present in the last warm period preceding the last major glaciation (several meters above the present position is commonly suggested for the peak Eemian (Europe) or Sangamonian (N. America)). The implication is not a happy one: it seems that our current sea level is on the low side of our type of environment, and a substantial (natural) rise would not be surprising. Encouragement of melting of polar ice definitively seems not advisable given this situation.

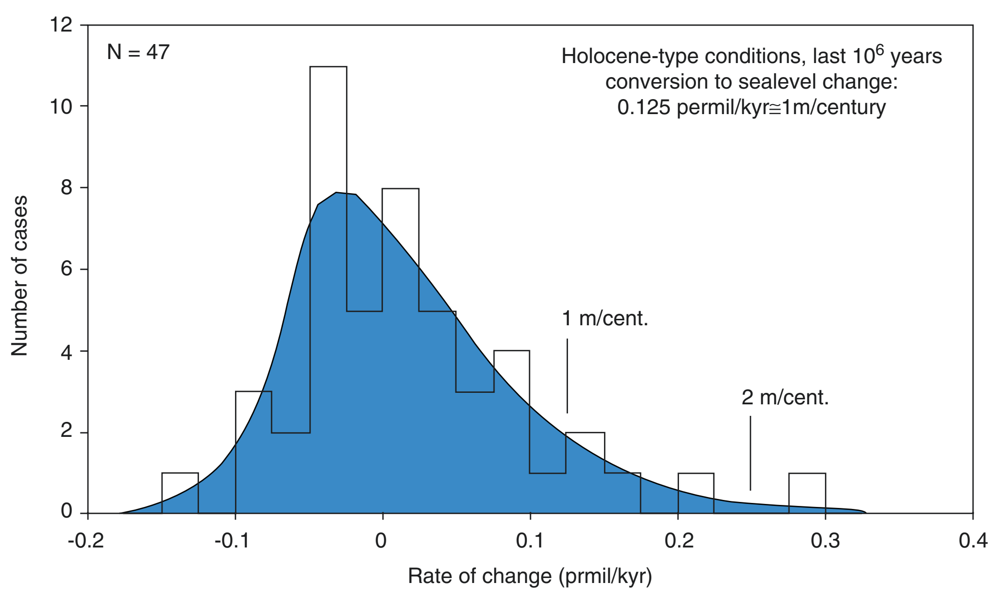

Fig. 6.15 Histogram of differences between sea-level positions of the last million years, for conditions of a warm climate (top 5% of cases). Note that rates are calculated for millennia (not centuries) and are given in per mil/kyr (not m/kyr, with only two such values given, as labels (1 and 2 m per century)) (From W.H.B., 2008. Int. J. Earth Sciences 97, 1143; color here added)

7 Productivity of the Ocean and Implications

7.1 General Comments

7.1.1 Limiting Nutrients and Recycling

The productivity of the world ocean (that is, its biological output) is of interest to marine geologists mainly for two reasons: (1) the distribution of biogenic particles (calcareous and opaline skeletal material, shells) and of organic matter depends on the patterns of productivity of the sea and (2) the history of productivity drives the evolution (that is, the geologic history) of organisms that eat food, which means all of the marine animals, including those of the zooplankton. Since the green primary producers need nutrients, they must evolve as well to adjust to changing nutrient supply. There are (at least) two more reasons to study productivity: (3) the evolution of marine microbes. (4) In economic geology, of course, the history of marine productivity is important in the origin of oil and gas.

The distribution of high-production regions is linked to the availability of nutrients in sunlit (surface-near) waters where photosynthesis can proceed. The bulk of the production is recycled; that is, it is based on nutrients released by microbial action in the water and on the seafloor through the decay of organic particles (Fig. 7.1). Production being limited by nutrients (a concept first introduced into terrestrial agriculture by the chemist Justus von Liebig, 1803–1873), the recycling of nutrients is of crucial importance. The nutrients traditionally contemplated by marine biologists and other oceanographers are nitrate, phosphate, and silicate. More recently, dissolved iron has received much attention. Distributions of the traditional nutrient compounds are highly correlated in the modern ocean – wherever one nutrient is abundant, there are the others, as well. The similarities in the distribution of nitrate, phosphate, and silicate make it difficult to identify any one of them as the "leader" in controlling production. As far as iron, it is exceedingly rare as a dissolved (and usable) nutrient and has its very own distributional patterns, strongly linked to oxygen. Wherever oxygen is high, dissolved iron is low, for the obvious reason that oxidation of iron in the modern (cold) ocean makes it highly likely that it precipitates. Within sediments, much iron is precipitated as sulfide and is thus taken out of circulation. In the marine realm, it seems, both an abundance of oxygen and its absence lead to a denial of iron to diatoms, which demonstrably use it.

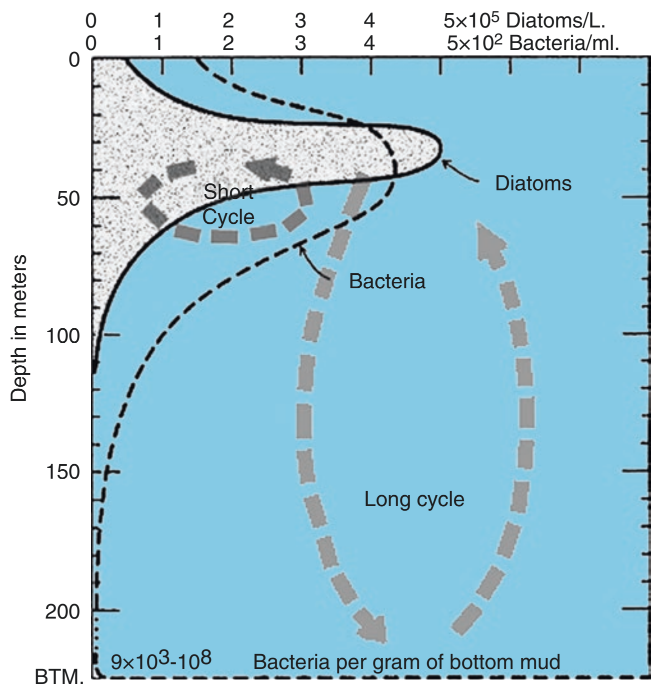

Fig. 7.1 Recycling of nutrients by microbial action (ascribed to "bacteria" in this graph). In the open ocean, recycling involves the distribution of primary producers (in the present example, diatoms) and of bacteria in the water column. Two types of cycling of nutrients are illustrated in this graph; one involves the seafloor (note the great potential abundance of microbes there). [After Sverdrup et al. (1942; Prentice-Hall); bacterial abundance presumably underestimated; from W.H.B., 2009. Ocean. UC Press, Berkeley]

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_7

7.1.2 Nitrate, Phosphate, Silicate, and Iron

From the relevant discussions among experts, the impression arises that a focus on short time scales favors nitrate as the leading nutrient, while a geologic time scale results in an emphasis on the role of phosphate. This is in keeping with the fact that the nitrogen cycle (which involves the atmosphere) is characterized by rapid cycling, while phosphorus involves long-term processes and sedimentation on the seafloor. Silicate is of special interest to geologists as its distribution in sediments reflects abundance of diatoms and other siliceous fossils, which are indicators of high production in the modern (Neogene) ocean. The presence of dissolved Iron has been shown experimentally to stimulate diatom growth, thus providing a possible explanation for low production in nutrient-bearing areas where high production is expected, by calling on a shortage of iron. An emphasis on iron implies that productivity patterns must be greatly influenced by oxygen distributions, that is, there is then built-in positive feedback on organic distributions in the marine production system, since high supply of reactive organic matter releases iron in affected localities by using up oxygen during decay. Since decay below the seafloor results in oxygen-free diagenesis and iron sulfide precipitation, diagenesis represents (long-term) negative feedback in the same system. The proxy use of nitrogen (for oxygen) may also result in employing negative feedback. It may be difficult to decide which sign on feedback (positive or negative) is relevant at any one time.

7.1.3 On Dominant Factors in Geological Production History

The history of production is strongly influenced by the changing availability of shelf regions and the overall cooling of the planet, which apparently started measurably since the early Eocene about 50 million years ago. Cooling and a drop in sea level affected ocean stratification and wind field, both of which are important controlling factors of productivity. True, wind stirs the uppermost waters, bringing nutrients up from thermocline depths. But also, wind takes green plankton down into the dark, in a poorly stratified sea. The end result of cooling and wind increase, therefore, is not immediately obvious.

Over shelves, it should be noted, nutrient-rich particles settling out from the photic zone cannot go very far away from that zone. Thus, nutrients are likely to stay in the area upon being released during decay, and shelf seas thus can attain rather high productivity from the seafloor release of nutrients (unless bottom waters are expelled in anti-estuarine circulation). As a general rule, shelf seas show large variation in production in large part because of the contrast between estuarine and anti-estuarine circulation, patterns that are created by regional conditions and can change rapidly. Thus, the various insights found for open ocean productivity may be of somewhat limited relevance to the (regionally strongly influenced) interpretation of marine sedimentary rocks seen on land and to the evolution of organisms in shelf seas, even when ignoring the role of vitamins, emphasized in recent studies.

7.2 The Grand Patterns of Production

7.2.1 A High Coastal Productivity

To beachgoers in Southern California, a high-production sea is the normal state of affairs. The water offshore of San Diego is relatively cold for the latitude – during much of the year, surfers wear a neoprene suit to keep warm. Some of the coldness owes to advection of surface water from high latitudes (by the "California Current"). But mainly the low temperature of the water near the shore attests to the fact that deep water is coming to the surface here: the water at depth is cold (it went down in the first place because of being cold and therefore heavier than the surface waters). Deep, cold water is rich in nutrients, and its ascent (commonly from no more than 300 m in coastal upwelling) provides for algal growth in the sunlit zone, including the famous large kelp of Southern California. In turn, algae feed the familiar nearshore animals, many of which filter the water for edible plankton. Mussels and gooseneck barnacles are conspicuous among filtering animals (Fig. 7.2).

The mussel-bed assemblages that characterize the rocky shores all along the western coast, from Alaska to Baja California, together with the kelp forests farther offshore, are testimony to the abundance of food and nutrients in the water. (Lately there are drops in abundance of native organisms and additions of certain newcomers from Europe and from Asia. Also, there are warmwater visitors from off Baja California in the coastal community. Both types of change are cause for concern and comment.)

There are good reasons why the coastal ocean in general is commonly referred to as the "green" ocean, while the sea far offshore is called the "blue" ocean. The "green" indicates photosynthesis by microscopic algae in the plankton, mainly, while the "blue" refers to the reflection of the sky in an ocean where primary producers do not much color the sea. Fishing birds everywhere will be seen at and over the green ocean but not commonly far out to sea. Coastal upwelling, where it occurs, is a crucial factor in generating high production next to the shore. But there are other reasons for high coastal production as well. An important cause for a relatively high nutrient content in the coastal zone is that longshore currents make eddies through friction with the continental margin, and the counterclockwise turning eddies (on the northern hemisphere) pump nutrients from dark, deep waters. The fact

Fig. 7.2 Life at the coast of California. *Left*: mussel-bed assemblage and other conspicuous shore animals on rocks (California mussels, gooseneck barnacles, and greenish sea anemone). *Right*: black cormorant flying along the shore, presumably looking for fish (Photos W.H. B.)

that nearshore waters are shallow and recycled nutrients are accessible contributes to high production. Iron is unlikely to be limiting at the shore thanks to mobilization in low-oxygen conditions. The ocean's oxygen minimum layer below the sunlit zone, which is the source of upwelling nutrients, intersects the seafloor and provides a sink for the oxygen that immobilizes iron in rust and related compounds.

7.2.2 The General Distribution of Productivity

The reasons for the observed global productivity patterns, including the blue-and-green dichotomy, are quite simple, in principle. They have to do with the fact that everywhere the deep ocean's temperatures are quite low and that almost everywhere warm water forms a relatively thin layer on top of the cold. As a rule, the top layer, the warm water, is impoverished in nutrients because in the sunlit zone minute floating photosynthesizing organisms constantly extract the nutrients, and the associated debris (including the fecal matter from animal plankton feeding on those organisms) moves the nutrients downward into the cold and dark water layers, within solid particles.

Because of the impoverishment of the warm surface layer, there is a desert at the center of each major basin, except right along the equator, where cold water is upwelling thanks to the action of trade winds (Fig. 7.3). High production means high delivery of organic matter and of diatom shells and radiolarian skeletons to the seafloor. This correspondence provides a means for the reconstruction of production patterns in the past, notably in the ice ages but also for time periods before that. In the Neogene both organic matter within the sediment and silica (opaline shells and skeletons) are useful in the task, along with many other indicators, including barium sulfate, phosphatic compounds, and certain fossils that are strongly linked to various production levels, especially the ones that are uncommonly abundant where there is a rich supply of food. In the modern ocean, these are often regions where oxygen levels are relatively low, especially near continental margins. In ancient settings (Paleogene and Cretaceous), the rules found for the modern ocean are of questionable relevancy, of course, increasingly so with the age of rocks targeted for interpretation.

Since the desert areas in the sea (centers of gyres in Fig. 7.3) are extremely large, we should expect the most abundant photosynthesizing organisms to be denizens of the ocean's desert. In fact, there are prokaryotic photosynthesizing microbes so small that they were largely overlooked not so long ago, when biologists mapped phytoplankton to define productivity patterns. In any case, for geologists the shell-making (hence sediment-making) coccolithophores and diatoms are of central interest, that is, eukaryotic microbes with hard parts rather than prokaryotic ones without such parts. Nevertheless, sediment structures involving prokaryotes, in this context, may be considered fossils on occasion (e.g., Figs. 4.15, 6.3, and 6.4).

7.2.3 The Record on the Seafloor

Within marine sediments on the seafloor beyond the shelves, we see relatively high concentrations of organic matter and of siliceous shells in areas with high productivity (Fig. 7.4). Surprisingly, the patterns of distribution of organic matter and of siliceous shells are strikingly different, despite all

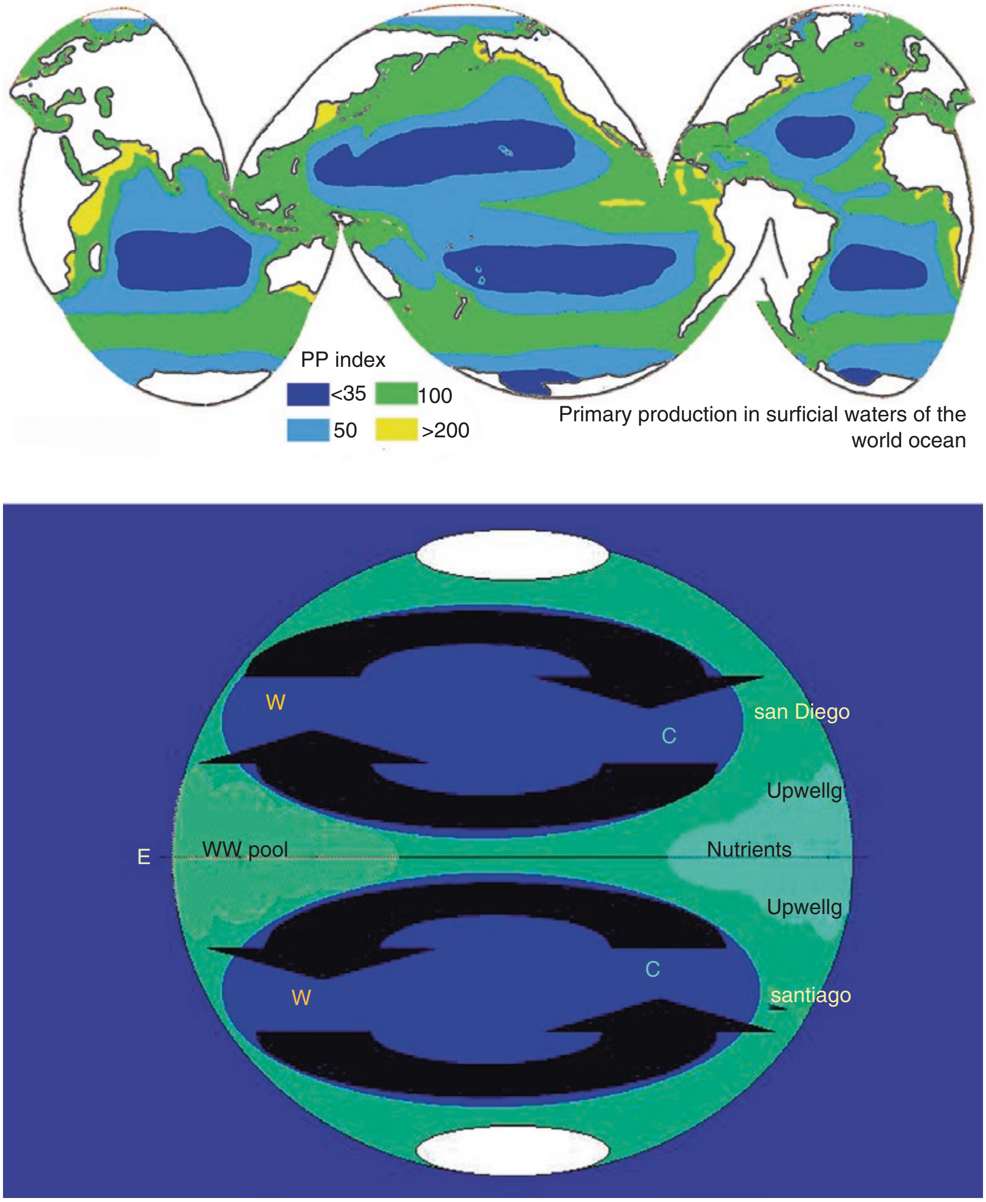

Fig. 7.3 Global patterns of productivity, simplified. *Top panel*: measured and estimated production values (After W.H.B., 1989. In W.H. Berger, V. Smetacek, G. Wefer (eds.) Productivity of the Ocean, Present and Past. John Wiley, Chichester, U. K). *Lower*: generalized pattern of productivity, relative to central gyres (driven by winds). The centers of the gyres (*blue*) are deserts

similarities. The reason for the differences, presumably, is that the abundance of organic matter is strongly linked to the oxygen content of the water in contact with the seafloor. In contrast, the preservation of opaline sediment (i.e.,, siliceous shells) follows different rules: it is best where silicate content is high in waters bathing the seafloor.

Wherever rapid burial enhances the preservation of either organic matter or of siliceous shells, the relative abundance of these high-production indicators is likely to suffer, since rapid burial implies high sedimentation rates, which in turn implies increased dilution of the component of interest. Thus, the relative abundance of these sought-after sedimentary proxies for productivity is quite a complicated matter, with both water chemistry (esp. oxygen and dissolved silicate) and burial playing important roles. Also, hiatus formation may falsify apparent sedimentation rates. Because of the many complications involved, the patterns of productivity proxies observed can be difficult to interpret.

7.2.4 The Biological Pump

The process of nutrient removal, which is responsible for the low nutrient content in central waters, is a result of "the biological pump" (Fig. 7.5). It is of great interest in the context of the sequestering of carbon by the ocean (and hence the removal of excess carbon dioxide introduced by human activities into surface waters). (The problems created by excess carbon dioxide gas are the reasons why carbon chemistry has become a central issue when discussing ocean processes. However, there are other reasons as well, especially in marine geology applied to the rock record.) Below the

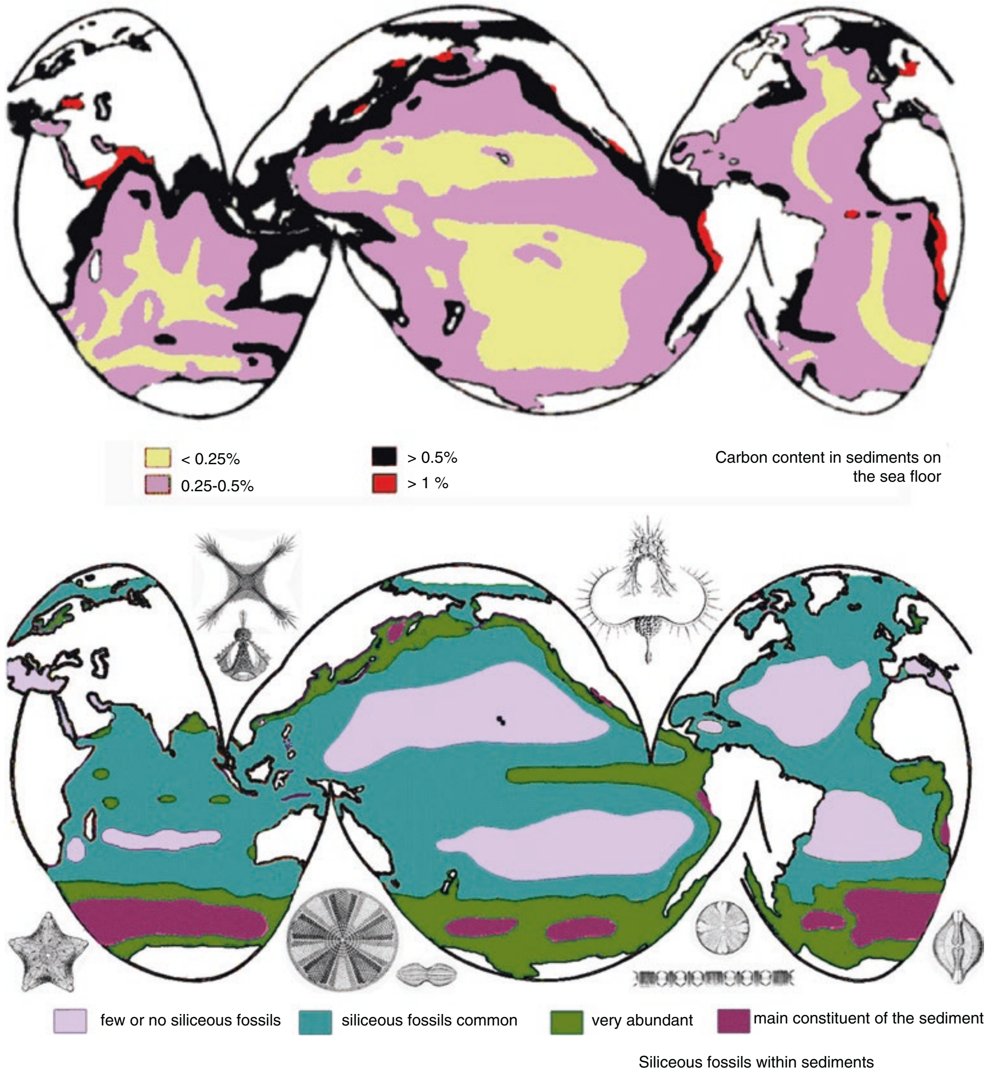

Fig. 7.4 Abundance distribution of organic matter and of siliceous fossils (radiolarians and diatoms) on the seafloor. Note the similarities and the differences of these two commonly used sedimentary indicators of high productivity. [W.H.B. and J. Herguera, 1992. In: P.G. Falkowski and A.D. Woodhead (eds.) Primary Productivity and Biochemical Cycles in the Sea. Plenum Press, New York.] The organic matter distribution is largely based on Romankevich, (1983). Opaline microfossils illustrating siliceous particles (radiolarians, diatoms) are taken from E. Haeckel (1904)

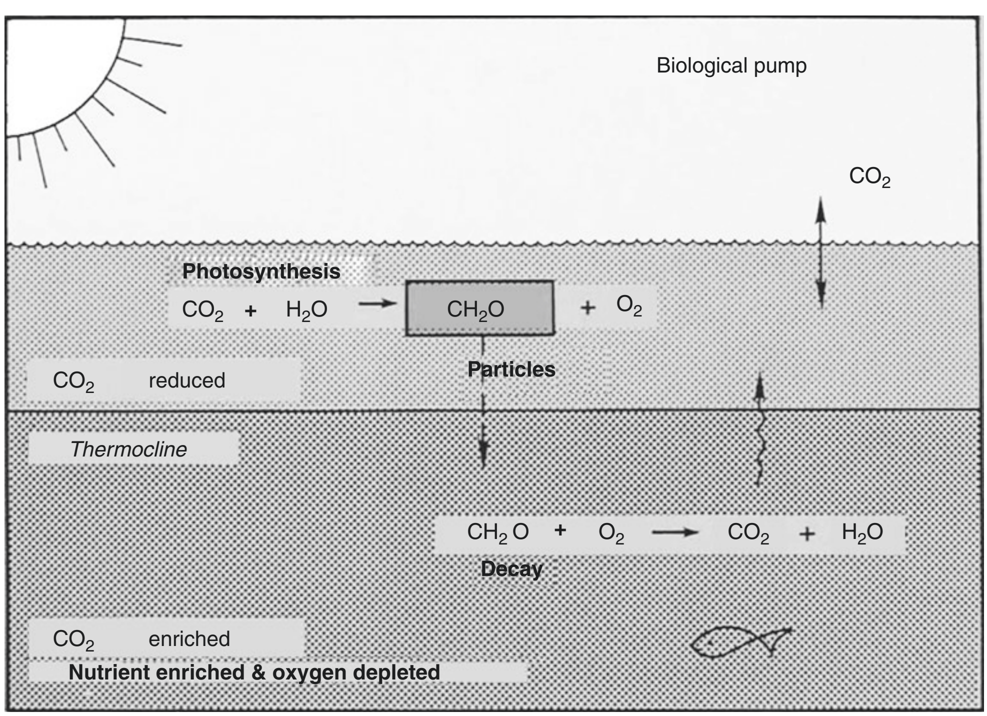

Fig. 7.5 Chief elements of the "biological pump." The pump moves organic carbon (and nutrients) into deeper waters, from the photic zone. The formula for organic matter here used (CH_2O) is an extreme simplification. In reality, the compounds in the particles are varied and have lengthy formulas

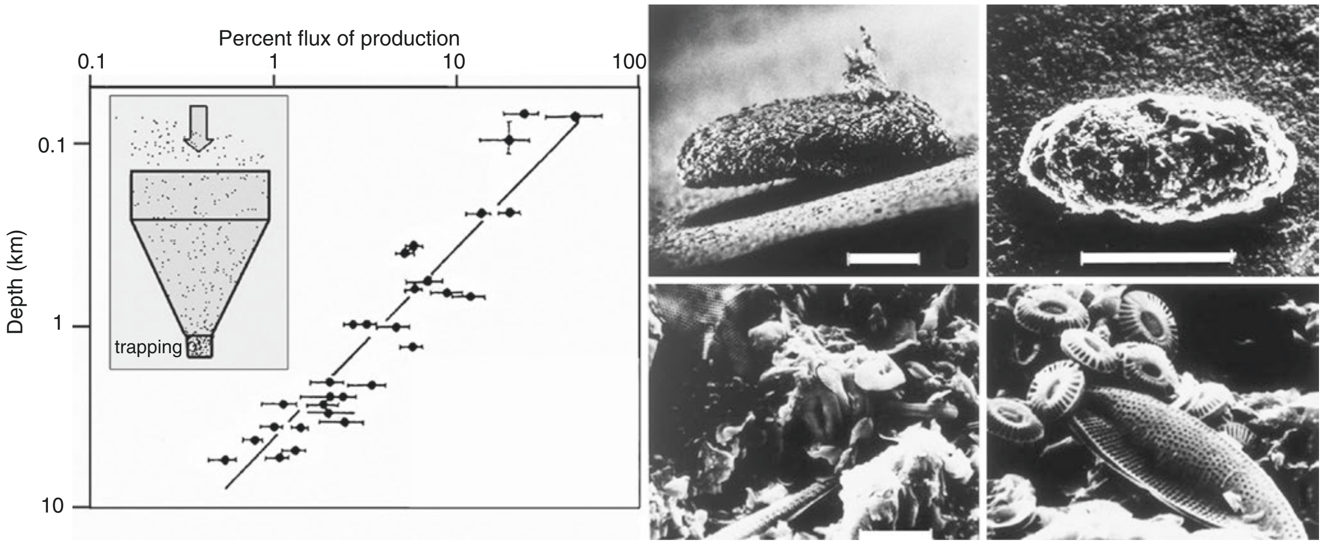

Fig. 7.6 Trapping of settling particles. Background graph (*left side*) from E. Suess [1980; Nature 288: 260; modified]. Note the general decrease of particle abundance with water depth. Sinking aggregates are loaded with shelled plankton. *Right side*: Contents of a trap set off California; fecal pellets loaded with mineral grains in addition to microfossils. Bar width 200 µm [R. Dunbar and W.H.B., 1981. Geol. Soc. Amer. Bull. 92: 212]

warm surface layer, typically in the upper 30–100 m of the water column (100–300 ft), bacteria and archaea decompose the sinking solids and regenerate dissolved nutrients (Fig. 7.1). Below the zone well flooded by sunlight (where nutrients are removed vigorously), nutrients start to pile up thanks to decomposition. Settling particles are oxidized. Thus, concentration of nutrients is high in the transition zone between the warm and the cold waters. Because nutrient-bearing particles also sink to great depths, past the toll takers in the transition layer, all of the deep waters are enriched in nutrients, compared with the sunlit water layer.

Biological pumping may be more efficient right off the coast than in the open sea, as suggested by studying the workings of the pump using traps (Fig. 7.6). In the open ocean, settling of fecal pellets is enhanced by the loading of pellets with shell material from small plankton, such as platelets falling off from coccolithophores, the so-called coccoliths (name-giving component of "coccolith ooze" and abundantly present in all "carbonate ooze"). But in the coastal ocean, sinking of fecal pellets is commonly aided by loading with sand and silt from the continents, material that is in plentiful supply from rivers and from off-land winds. Some of these particles are incorporated into fecal material, adding weight and thus accelerating sinking and enhancing the pumping action.

That there are gross differences in the transfer of particulate organic carbon to the seafloor, between coastal ocean ("green ocean") and open ocean ("blue ocean") has been appreciated for some time (Fig. 7.7). With respect to

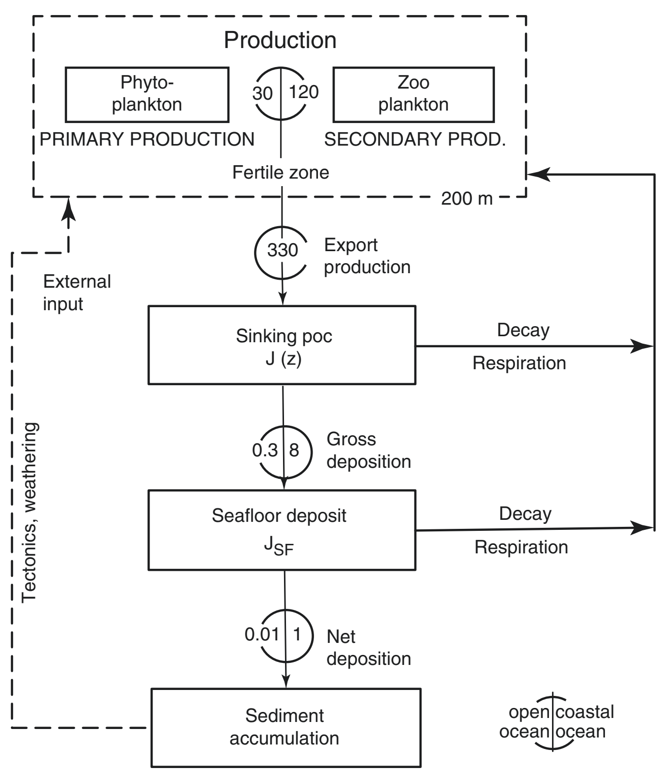

Fig. 7.7 Sketch of transfer of particulate organic carbon in the ocean, from primary production to burial in the sediment. Numbers are approximate fluxes in $gC/m^2/yr$; within each circle the value to the *left* refers to typical open ocean conditions, while that to the *right* refers to the coastal ocean [W.H.B., V.S. Smetacek, G. Wefer, 1989, in Productivity of the ocean – present and past. Wiley, Chichester]

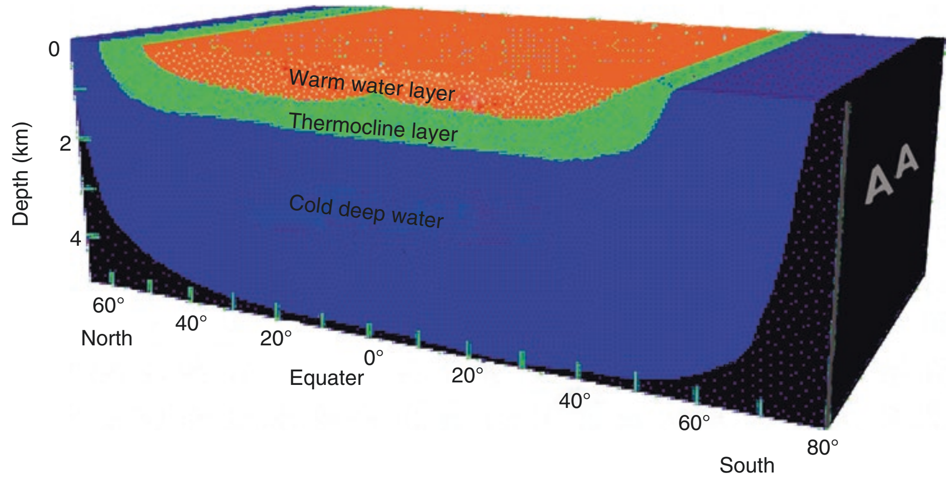

Fig. 7.8 Temperature stratification of the modern ocean (here: Pacific Basin) and position of the thermocline. AA, Antarctic continent. Note that there is no thermocline shown around the AA [Source: U.S. NOAA, slightly modified]

sediments, the observation of obvious differences was codified in the classification provided by the *Challenger* Expedition at the end of the nineteenth century. Preservation aspects of sediment particles linked to oxygen introduce yet another dimension of difference.

Presumably, a large portion of the nutrients in the seawater beyond the surf zone is brought back from the seafloor, including the limiting element iron. Spreading of the nutrients turns out to be mainly by eddies in coastal currents (seen as filaments of intense growth in the coastal ocean, from space).

Undoubtedly high nutrient supply from upwelling is one reason why the coastal ocean is so highly productive compared with the far-out open ocean. But the phenomenon of eddy formation suggests that there are mechanisms besides simple upwelling that are responsible for supplying nutrients to sunlit surface waters. The additional mechanisms have to do with the patterns of mixing intensity. Mixing is strong wherever wind-driven currents meet shallow seafloor, generating strong eddies.

The highest production in the sea occurs in the west-wind zones, in the vicinity of the equator, in certain shelf seas, along the coast, and off Antarctica. Most of the regions of high production are windy, and in any case currents are lively. It is here, too, that one finds the sea mammals and the seabirds, the fishing fleets, and benthic organisms. Because high concentrations of marine life-forms are commonly linked to high nutrient supply, which comes with low-oxygen waters, such areas are also prime sites for anaerobic developments and mass killings of fish. "Dead zones" can result from adding nutrients to seawater in sensitive regions. The origin of the nutrients may be natural or artificial; in many cases agricultural runoff is thought to be involved. Indeed, indications are that "dead zones" are expanding as a result of fertilization in agriculture.

7.3 The Dominant Role of the Thermocline

7.3.1 Thermocline and Oxygen Minimum

The transition zone between deep cold water and shallow warm water is 600–900 m thick (two to three thousand feet, approximately). It is called the *thermocline*, in reference to the rapid change in temperature with depth (Fig. 7.8). The thermocline holds the *oxygen minimum*, which results from the microbial processing (aka *rotting*) of sinking particles taking organic carbon down below the sunlit layer, as discussed. This is not the whole story, however. In all ocean basins, the oxygen minimum is strongest right next to the continents, where the thermocline waters touch the seafloor at the level of the upper "continental slope," just outside of and below the "shelf edge." This circumstance suggests that the rotting of organic matter on the seafloor is heavily involved in generating the oxygen minimum, not just the oxidation within the water.

As a corollary, the seafloor yields nutrients back to the sea while taking up oxygen. Nutrients are high wherever oxygen is low and vice versa. The implications for the deposition of organic matter within sediments are rather drastic: nutrient-rich regions (that is, the coastal ocean) produce a lot more organic material for export to the seafloor, and low-oxygen regions enhance the preservation of such material on the seafloor. (For details, see the book by Rödiger Stein, 1991.)

7.3.2 Thermocline and Productivity

With the thermocline playing such a crucial role in the productivity of the sea, we must impart to it additional scrutiny. The nature and position of the thermocline largely determines whether there is a desert or not in any given area of the sea. Obviously, then, the thermocline is among the most important geographic elements on our planet. For geologists concerned with marine sediments now found on land, the thermocline is important because it makes a lot of difference whether ancient shelf sediments were deposited in potentially stratified waters or not. A thermocline invading deep shelf areas is one way to introduce stratification to water bodies on the shelf. In consequence, a shallow shelf cover is likely to have water layers different from those of a deep one. Analogous arguments apply in the case when a thermocline is replaced by a “halocline,” with general density differences being linked to salt content of the water (rather than temperature).

Stratification presumably is destroyed in shallow waters in areas (or during times) of strong wind action. Besides keeping photosynthetic organisms from spending too much time below the sunlit layer, the thermocline acts as a barrier obstructing the flux of nutrients from deep waters back into the sunlit layer. Thus, density differences between warm and cold waters (and the great abundance of cold water in the present ocean) are fundamental elements of the production patterns of today’s ocean.

Most of the production resulting from a rich nutrient supply combined with sunlight becomes food for small animal plankton, which in turn serves as food for larger animals including fish and squid – which in turn are food for seabirds and seals and whales. Hence, shores off the Antarctic, in high latitudes in general, on the equator, and in the coastal strip in general are good places to watch for large and energetic marine animals, from many types of sharks and fast fishes to whales and seabirds. The animals are concentrated here for good reasons: they must eat lots of food to support a high-energy lifestyle. In the distant past, in a warm ocean with flooded shelves, large marine animals may have chiefly lived in shelf seas, some of which apparently were highly productive, judging from the type of sediment produced.

7.4 Food Chain Length

7.4.1 Long and Short Food Chains

In our quest to explain the ring of high coastal production in the sea, we have pointed out one important ingredient of this situation in the modern ocean: the supply of nutrients by microbes living on the seafloor, with bacteria and archaea responsible for generating an oxygen shortage. Now we must deal with yet another basic microbial factor responsible for the contrast between the green and the blue ocean: the fundamental difference in the food chain of the two environments.

Food chains start with “primary producers” and end with “apex consumers.” On land and in the sea, except that in the ocean, we largely talk about “plankton” (drifters) and “nekton” (swimmers). In the modern sea, long food chains of the central deserts start with minute prokaryotic microbes, which feed somewhat larger microbes, which feed eukaryotic microbes such as foraminifers and radiolarians and tiny zooplankton (such as copepods), which in turn feed small fish. The small fish, finally, are available for consumption by larger fish and their predators, the standard apex consumers in the sea. At each link in the chain, only some 15% of the organic matter gets transferred from one level to the next. At the sixth level in a long chain, barely one tenth of one thousandth of the original production remains after the heavy toll taking along the food chain gauntlet (or “*food chain pyramid*”). Thus, given the numbers mentioned, one readily calculates that a ton of solid carbon made by the photosynthesizing organisms has shrunk to less than 0.1 kg by the time the carbon reaches the sixth *trophic* level (that is, the sixth level in the food chain).

In contrast, a short food chain might have only four levels. For a thousand kilos produced, there is then an ultimate yield of around 4 kilos, some 40 times more than in the long chain. As the Woods Hole oceanographer John H. Ryther (1922–2006) pointed out half a century ago, the high-production areas commonly have short food chains. Such chains commonly start with diatoms, primary producers that are large compared with other phytoplankton (Fig. 7.9).

7.4.2 Diatoms

The short food chain, with diatoms as a starting base, depends on a sufficient supply of dissolved silicate, since diatoms make shells of opal (water-bearing silica). In fact, the abundance of silica within sediments (that is, of shells of diatoms and radiolarians) is traditionally used as a proxy indicator for reconstruction of the productivity of the geologic past. As discovered by the marine geologist L. Diester-Haass in cores taken on Walvis Ridge, the reconstruction of productivity from siliceous deposits does not always work very well, though. There is evidence that high wind-generated production (reconstructed from non-siliceous proxies) may co-occur with low silica content, and vice versa.

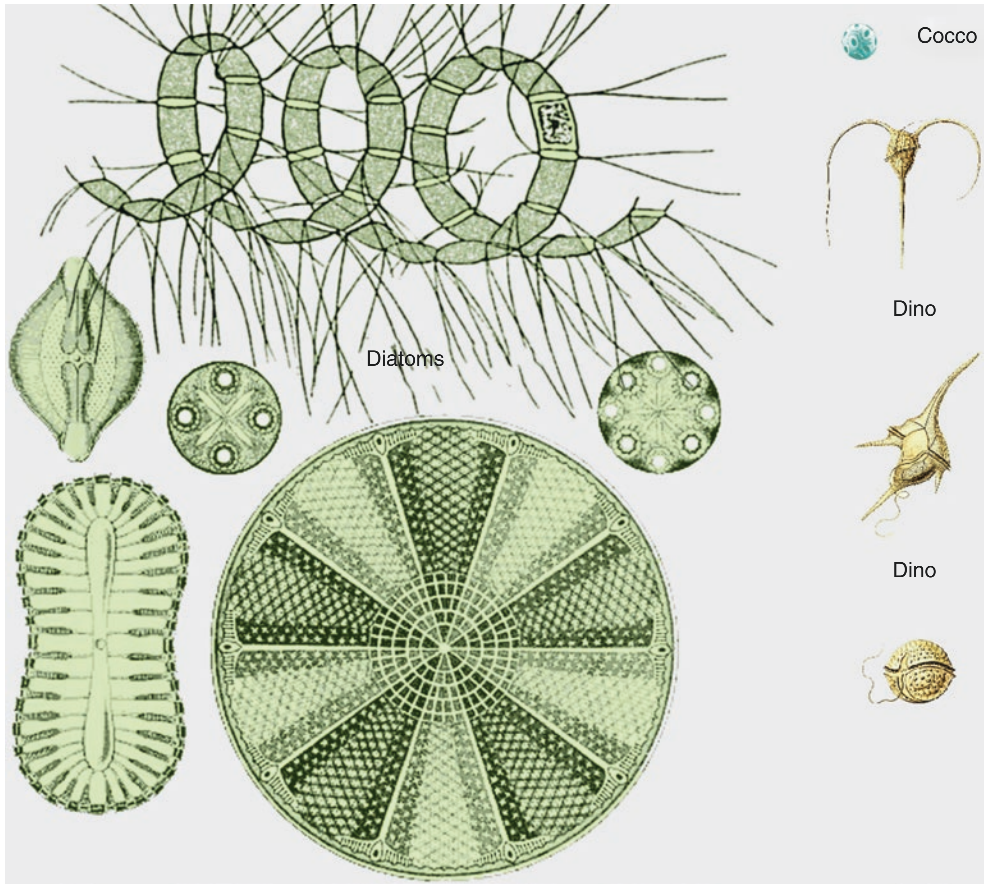

Fig. 7.9 Diatom sizes compared with those of other primary producers. "Cocco," coccolithophore; "dino," dinoflagellates (three shown). [Drawings from Ernst Haeckel, except the diatom chain with bristles (typical for high production), pictured in Sverdrup et al., 1942, probably drawn by Martin Johnson.] [The Oceans. Prentice-Hall, Englewood Cliffs, N.J., and in S.I.O. Archives. Here slightly simplified.] For scale: coccolithophores approach 20 µm in diameter (that is, one fiftieth of a millimeter, silt); large diatoms commonly are the size of fine sand

7.5 Upwelling and Its Geological Corollaries

7.5.1 Classical Coastal Upwelling

It was realized early in the twentieth century (after the oceanographer Hermann Thorade had identified the upwelling phenomenon) that the upwelled waters off California originate at relatively modest depths, a few hundred meters at most. The coastal upwelling here is fed by upper thermocline waters, as seems to be generally true for coastal upwelling. The question arises, of course, where does this upwelled water come from? Is it simply a mixture of the deep cold water and the shallow warm water, loaded with nutrients locally? Or is it water that is moved in subsurface currents over long distances from the boundary zone between warm surface waters and high-latitude regions? It might be both, in varying mixtures. If coming from afar, the thermocline waters will then bring along nutrients picked up elsewhere and modified during travel to the site of upwelling. For example, well-traveled thermocline waters made in high latitudes (why they are cold) might come with a high silicate content, which would benefit the growth of diatoms in the upwelling regions and thus the development of a short food chain.

The concept of basin-to-basin "conveyor" exchange circulation (Sect. 5.3.3) suggests that some of the silicate indeed is moved over enormously long distances, even from high latitudes to places of upwelling. The question is, is there a way to feed the nutrient pile-up around Antarctica through long-distance transport? AA-derived intermediate waters could in any case conceivably feed coastal upwelling systems south of the equator, and maybe more, making it advisable to answer the query if ocean productivity is to be understood. The modern pattern of circulation apparently started some 15 million years ago and involved the entire deep ocean as one single system (see Chap. 12, →Monterey, for discussion). According to Ryther's Principle, a short food chain could then form. This presumably generated an abundance of high-energy animals (modern fast-moving bony fish, but also mammals and seabirds) rather than favoring low-production organisms such as jellyfish. Ryther's Principle, if generally valid, obviously has important implications of the workings of the silica cycle for the nature of the sediment accumulating on the seafloor, deep and shallow.

The traditional two-dimensional view of coastal upwelling (developed within the second half of the twentieth century

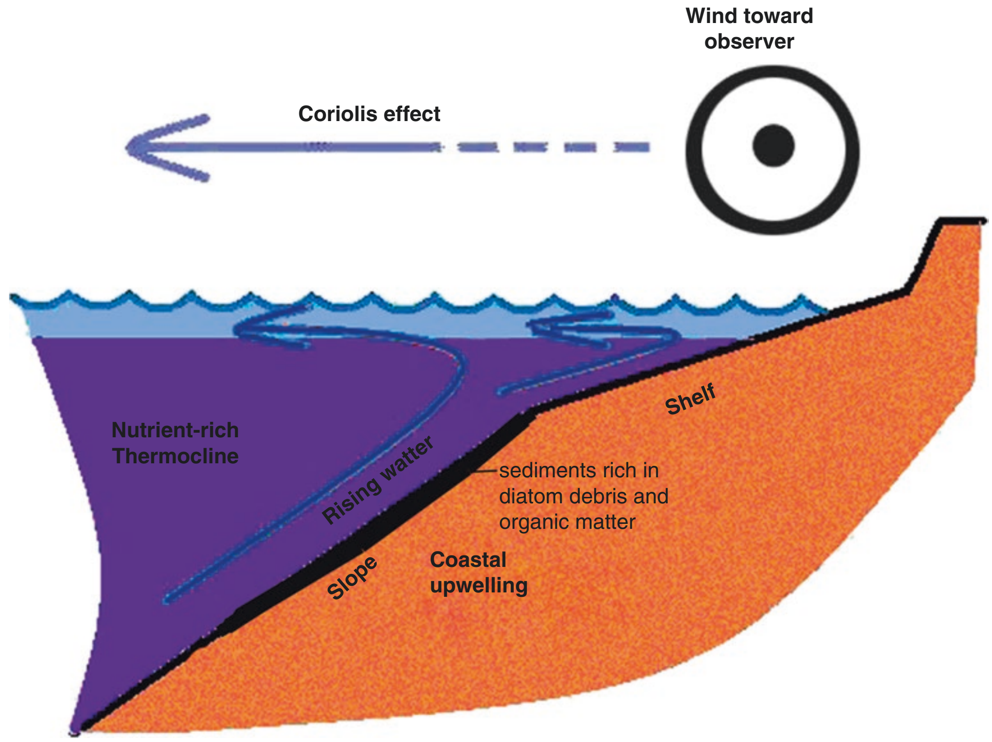

Fig. 7.10 The traditional two-dimensional view of coastal upwelling. The target symbol for the wind represents the tip of an arrow moving toward an observer looking northward off California. Coriolis deflection of motion on the northern hemisphere is to the right of the motion. The sediment below the high-production area is rich in diatoms and in organic matter [After W.H.B., 2009. Ocean. UC Press, Berkeley]

and sarcastically placed into a kind of upwelling mythology by at least one distinguished oceanographer) emphasizes the effect of longshore wind and the offshore component of the water moved by the wind (Fig. 7.10). The wind along the shores of California depends on the strength of the eastern N Pacific High, a pressure center that traditionally has developed mainly in summer over the sea.

The offshore component of the resulting current owes to the Earth's rotation. The rotation of a spherical Earth implies much slower eastward motion on the surface in high latitudes than in low ones. Thus, water moving toward the equator is left behind in the regionally ever faster moving surface. It is seen to move westward. The westward moving surface water makes room for cold water from below, which wells up to fill the void, thus, the traditional "coastal upwelling" picture. The mathematics of motion on a rotating Earth was worked out by the French engineer Gustave-Gaspard de Coriolis (1792–1843). The Swedish oceanographer Vagn Walfrid Ekman (1874–1954) applied "Coriolis Force" arguments to currents and to coastal upwelling. The appropriate coastal regions, therefore, are said to display *Ekman upwelling*.

The pile-up of organic matter below coastal upwelling is enhanced by the short travel time of particles from the productive zone to the site of deposition. This fact, along with the loading of fecal pellets with mud and with shell, makes for high rates of deposition of organic matter with important implications for the origin and distribution of petroleum and of "Earth gas" (or "natural gas," i.e., methane, CH_4). It is the prokaryotic microbial fauna (bacteria and archaea) that produces gases within upwelling sediments, notably carbon dioxide and methane, but also others containing sulfur and chemically quite similar to certain smelly gases produced in sewage.

7.5.2 Food for Abyssal Organisms

The fact that much of the organic matter reaching the seafloor is made in the coastal ocean not far from land has important implications for food supply to the deep seafloor. Animals and microbes on the seafloor are much more abundant close to continents (and large islands) than they are farther away, and the intensity of interaction between sediments and organisms is correspondingly strong offshore. This interaction is expressed both in bioturbation of the sediment (discussed in the chapter that follows) and in the uptake of oxygen. Sediments below productive areas are highly reactive, largely owing to the presence of organic matter. Thus, the commonly studied early diagenetic reactions of marine sediments on the deep seafloor are conspicuous at the continental margins below the coastal oceans and less so in enormously large regions below the blue ocean. Likewise, apex consumers and scavengers feasting on fallen animals feed largely in coastal oceans, with appropriate implications for modern marine processes of geologic interest.

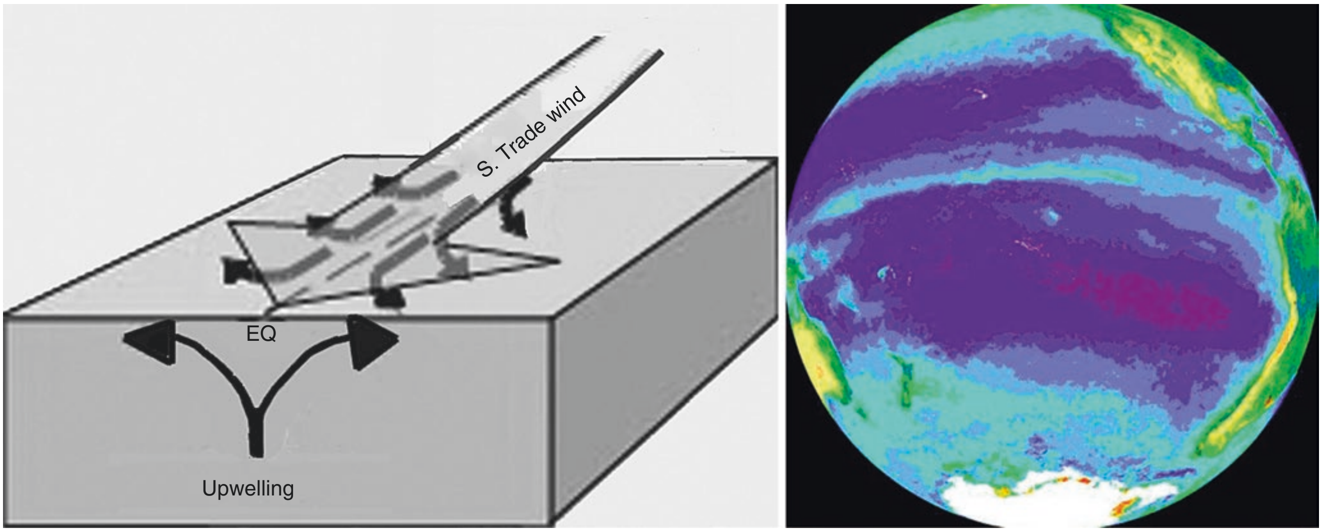

Fig. 7.11 Upwelling in the equatorial Pacific. *Left*: principle of divergence of currents along the equator, schematic after various physical oceanographers. [Wind distribution during the S.I.O. Pleiades Expedition.] *Right*: equatorial high-production belt seen in chlorophyll distribution from space (US NASA)

7.5.3 Equatorial Upwelling

"*Equatorial upwelling*" describes the divergence of surface waters along the equator as a result of the Coriolis Force working on a westward flowing equatorial current driven by trade winds (Fig. 7.11). The divergence brings nutrients to the sunlit zone. It depends on the strength of the tropical trade winds, which dominate much of the general wind field on the planet and drive surface currents in low latitudes. Since there is a slight asymmetry about the equator, with a shift of climate zones northward, it is largely the southern trades that drive the equatorial currents which produce the upwelling, as currents north and south of the equator diverge, being deflected to the right and to the left, respectively, by the Earth's rotation.

Along the ocean's equator, notably in the eastern portion of the equatorial Pacific, there is a rich collection of large marine animals. The Galápagos Islands provide a convenient site for observation. Seabirds and eared seals are abundant. Other conspicuous large animals are dolphins at sea and marine lizards on land, sunning themselves on the basaltic rocks. The marine iguanas enter the water to feed on algae growing on the seafloor in shallow areas. Their food tends to become rare when trade winds fail.

7.5.4 Antarctic Upwelling

The largest of the ocean currents runs around the continent of Antarctica. Not only does the current tap into abyssal waters, stirring them and bringing up nutrients through deep mixing, but also nutrients are supplied by large-scale addition of deep waters, especially in the Atlantic sector (Fig. 7.12). Thus, the AA current is the site of maximum upwelling. The Antarctic delivers silicate-rich waters to coastal upwelling regions in the southern hemisphere, as well, within intermediate waters. We might expect, therefore, that the southern upwelling areas have very nutrient-rich waters and are stronger and more productive than the northern ones and have more large diatoms. Apparently this is indeed the case. In any event, the sediments around Antarctica are unusually rich in diatom debris, signaling major extraction of silicate from the world ocean here. Thus, conditions around AA affect all of the sea.

The Antarctic ice mass, sea ice, and the associated strong circumpolar winds are intimately involved in the circumpolar upwelling and must be considered when reconstructing the geologic history of this signal. Absence of the Drake Passage in the Cretaceous and in the early portions of the Cenozoic must have prevented the development of the type of mighty Southern Circumpolar Current that we see today (Fig. 7.12). Also, with reference to the wind around Antarctica, if relevant winds feed off temperature gradients that are enhanced by ice on land, we should then see much less upwelling in the geologic past before the ice buildup began near the poles. We do indeed.

7.5.5 Upwelling Off Namibia

If one looks at the present situation off southwestern Africa (that is, off Namibia), one finds that coastal upwelling there started to become important sometime near 10 million years ago, at the end of the Middle Miocene. One might use that timing in general, as a legacy of winds. However, ice in Antarctica is reported from times much earlier than that. It was there, but perhaps not covering the continent as massively as it does today. Another way to approach the problem

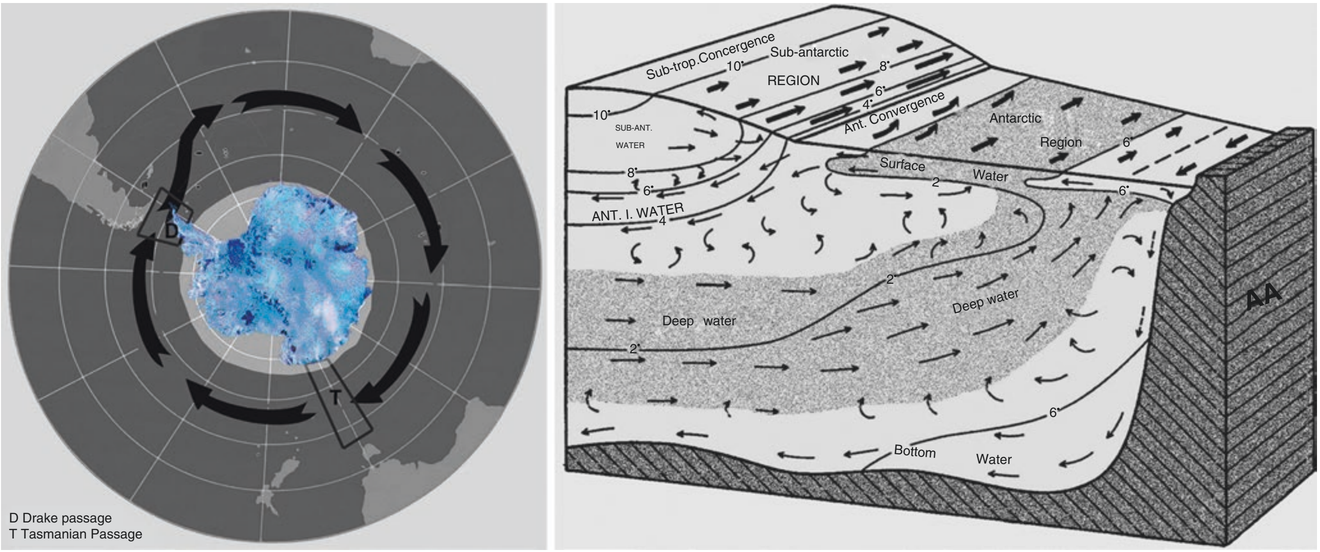

Fig. 7.12 Deep-water upwelling around Antarctica and the Circumpolar Current (Graph to the left after various authors, including the Lamont geologist J.D. Hays; graph to the right after H. U. Sverdrup et al., 1942. Modified for clarity)

of initiation of coastal upwelling (and a rather hypothetical one) is to ask at what time large consumers first found it profitable to evolve toward the presence of high production in the sea. This approach gets us back into the early Oligocene, just before 30 million years ago, when whales evolved from toothed forms, whales that have no teeth but use baleen to filter the water instead (Chap. 12).

Perhaps the opening of the Drake Passage initiated or facilitated this evolution, rather than the evolution of the wind field. The question is open. Perhaps both ice growth and the opening of Drake Passage must be considered simultaneously. Using the proposed history of productivity effects for guidance to production history may favor circular reasoning more than advisable. At present, though, this may be the only way available for a plausible reconstruction of the great Circumpolar Current. Unfortunately, the marine geophysics history of the Drake Passage region is poorly documented (pers. comm. by the geophysicist and physical stratigrapher Stephen Cande at S.I.O., in 2012). In any case, in the present book, we consider the Oligocene as the crucial period of opening Drake Passage and deepening it sufficiently for circumpolar effectiveness, rather than the Miocene, the early and middle portion of which presumably saw additional developments bearing on Antarctic circulation and production, during Monterey time (see Chap. 12).

Namibia upwelling is unusual for its great intensity, which is reflected in a strong cold-water anomaly, and remarkable release of nutrients from the underlying seafloor (Fig. 7.13). The pattern of the nutrient flux is especially effective in defining the center of high productivity, the chlorophyll color from space photography being rather variable. The patterns of release of nitrate and silicate from the seafloor to overlying waters resemble those for phosphorus release. Also, the modern phosphate release pattern agrees very well with the temperature anomalies mapped almost a century ago by the German oceanographer G. Böhnecke, based on data collected by the research vessel *Meteor*.

7.5.6 Matuyama Diatom Maximum

Strong upwelling off Namibia, as mentioned, started about 10 million years ago. Much later, within the early part of the Matuyama (geomagnetic) Chron and just before the onset of large northern ice ages, there was a time of maximum deposition of diatoms. The high rate of deposition of diatoms at the time (Fig. 7.14) may have reflected an increased supply of silicate within the thermocline layer, a layer generated at the Antarctic convergence. The ice-age decrease of the silicate supply is puzzling (see Chap. 11). Intense high-latitude extraction of silicate by diatoms may be responsible.

7.5.7 Upwelling, Drought, and Iron

Coastal upwelling and the cold anomaly it generates in surface waters and the overlying atmosphere stabilizes a high-pressure region in the atmosphere offshore, which in turn blocks the entry of moisture-laden marine air to the continent. Also, cold air holds less moisture than warm. Thus, the land behind upwelling regions is extremely dry, with the Namibian desert providing a prime example. The onset of upwelling, then, also presumably resulted in the onset of severe drought on the adjacent land, a land that is arid already because of its location in a global desert belt generated by falling air. Effects are seen in the generation of dust plumes moving offshore from desert areas.

Desert dust brings iron to the sea, an element that stimulates diatom production according to John Martin

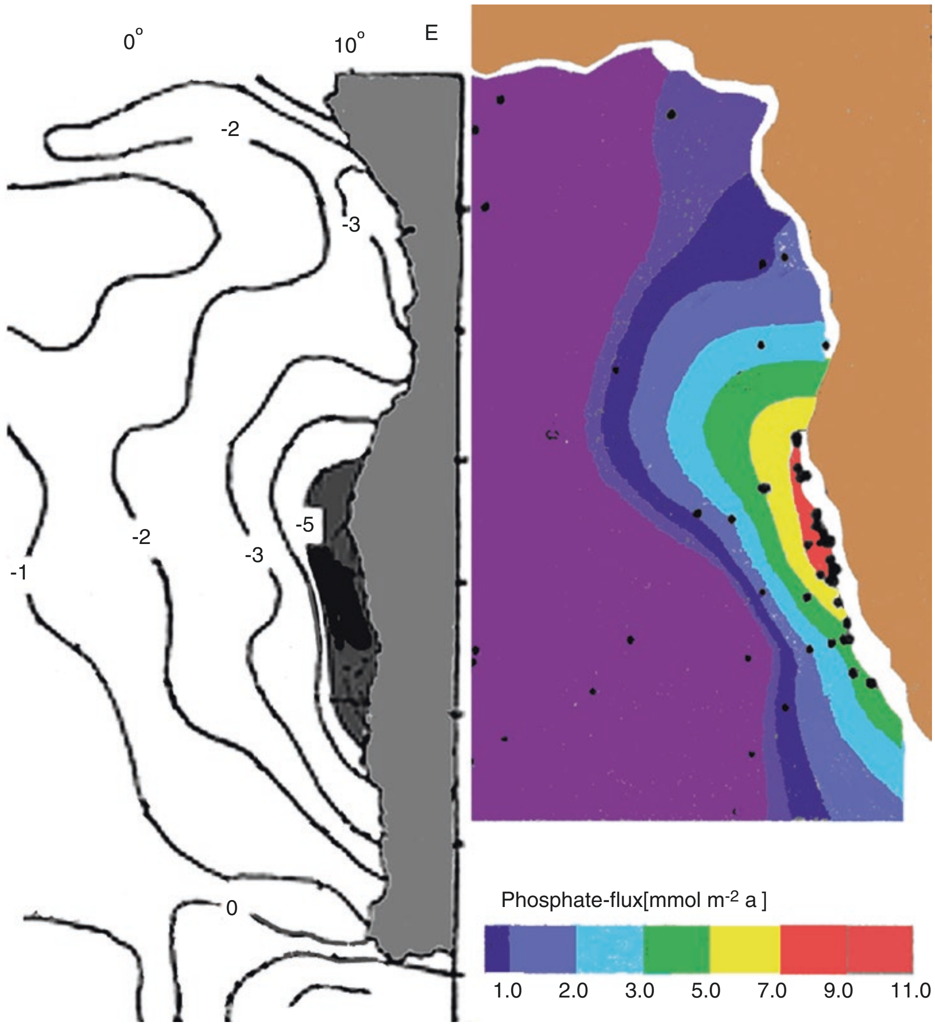

Fig. 7.13 Upwelling off Namibia and South Africa. *Left*: Cold anomalies (*center black*: more than 5 °C below latitudinal average) [After Böhnecke 1936, Meteor Werk, Bd 5; grid here removed]. *Right*: Phosphate release from seafloor [after a color graph kindly provided by M. Zabel; see H. Schulz and M. Zabel, 2000 (Geochemistry, Springer Verlag); data C. Hensen et al., 1998. Glob. Biogeochem Cycles 12:193]

(1935–1993), a US American chemical oceanographer who worked at Monterey Bay, California. Satellite images provide direct evidence that dust (and hence iron) is indeed being supplied to the upwelling areas from the landward desert. Distances of transport are remarkable and include dumping of Saharan dust in the Caribbean (dust conceivably coming with unwelcome microbes).

7.6 Early Diagenesis and Loss of Oxygen

Soon after marine sediments are deposited, there are reactions within them that change their composition. The energy-driven reactions are dependent on oxidation of organic matter. It is clear, therefore, that reactions will be most intense in highly productive regions. In today's ocean they are prominently occurring in coastal upwelling areas. Since we are largely dealing with early diagenetic reactions which involve oxygen, we typically see a marked reduction of that element in the interstitial waters of shallow-water sediments in areas of high production (Fig. 7.15). Loss of oxygen is commonly accompanied by the release of nutrients from rotting organic solids. Simultaneously, silicate is released from the dissolution of opaline skeletons (mainly diatoms). Presumably, dissolution of opaline skeletons is enhanced by the removal of organic coverings on them, coverings that keep seawater from making contact and (it is thought) greatly slow recycling of the silica.

The oxygen is used to oxidize organic carbon arriving on the seafloor; remaining organics are worked into the uppermost sediment layer by burrowing organisms for at least a thousand years. When the oxygen is used up, the microbes use nitrate and sulfate to oxidize organic matter. As a result, nitrate is being reduced and the concentration of ammonium and of elementary nitrogen rises. With the reduction of sulfate, hydrogen sulfide (the cause of the rotten egg smell) forms in abundance, and pyrite precipitates by combining the sulfide ions with iron. Iron sulfides are ubiquitous in the sedimentary record. Abundance of iron sulfide is an indicator of anaerobic conditions within the sediment at the time of formation. Fermentation sets in when the sulfate is used up, and methane ("natural gas") is then produced by bacteria and archaea, along with carbon dioxide. Thus, the study of early diagenesis implies the study of prokaryotic microbial organisms and their activities.

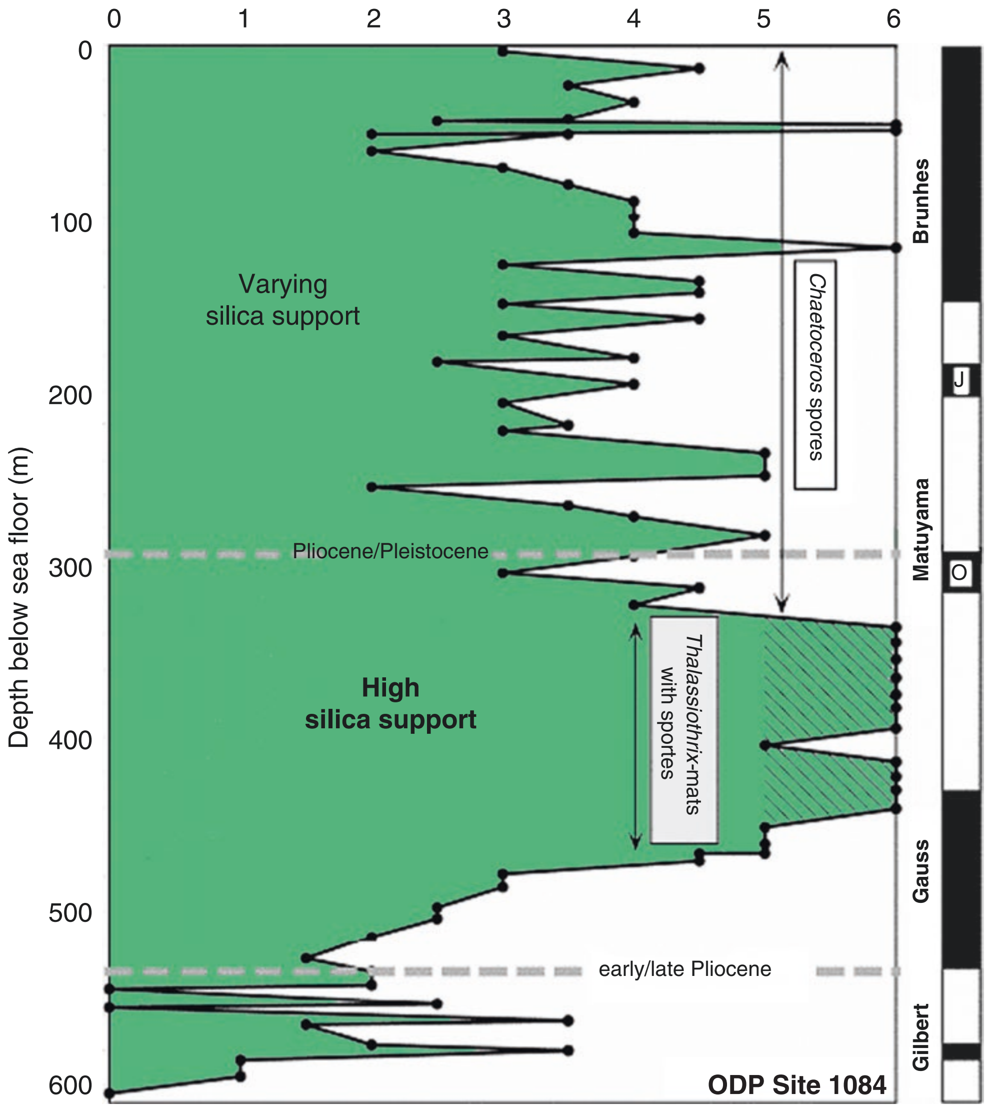

Fig. 7.14 The early Matuyama Diatom Maximum off Namibia, as seen in ODP Site 1084, Leg 175 [see also C.B. Lange et al.., 1999, Marine Geol. 161: 93]. In the paleomagnetic time scale to the right, "J" stands for Jaramillo and "O" for Olduvai. The long intervals Brunhes, Matuyama, and Gauss are labeled. The "diatom abundance index" is a semiquantitative and subjective measure and is based on inspection, on board the drilling vessel JOIDES RESOLUTION, of smear slides, using a microscope. Note that there is a major increase in diatom deposition at the end of the Gauss Chron and a considerable drop on approaching the Olduvai. The diatom maximum is in the early Matuyama

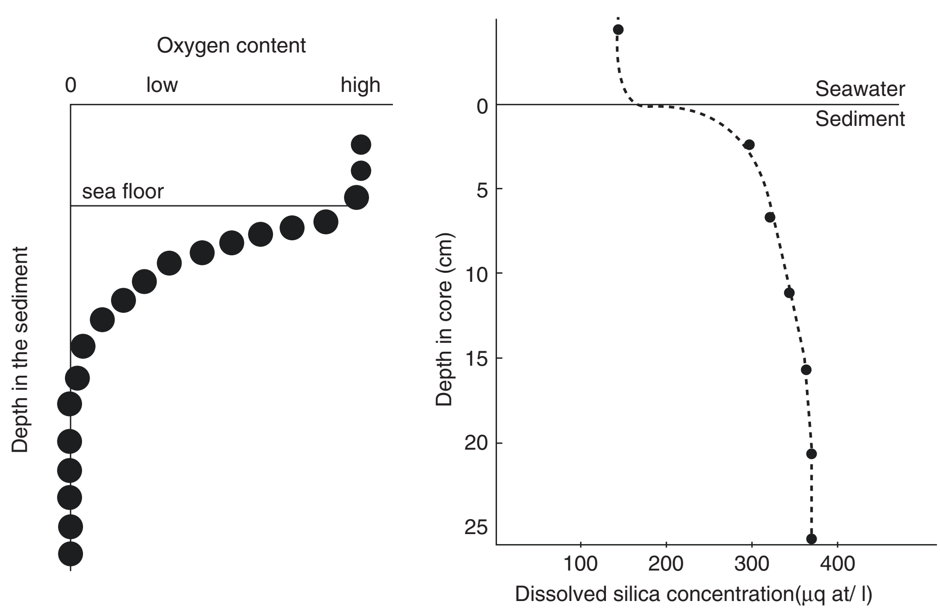

Fig. 7.15 Schematic profile of oxygen across the seafloor and evidence for nutrient reflux. The depth in the highly reactive part of the sediment, in most regions, is measured in cm ("depth in the sediment"). Where "high" oxygen in bottom waters already has very low values, it is measured in millimeters above the zero level. "High" oxygen typically is near 2–3 ml per liter in many subtropical regions just below the shelf edge. A loss of oxygen (*left*: schematic after a graph by H. Schulz, Bremen) is associated with a reflux of nutrients, here exemplified by silicate concentrations measured in the eastern tropical Pacific (*right*: courtesy of T. Johnson, then S.I.O.)

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8 Benthic Organisms and Environmental Reconstruction

8.1 On the Large Diversity of Benthic Organisms

8.1.1 A Quick Look at the Diversity of Mollusks

When discussing diversity, marine geologists commonly emphasize foraminifers. However, mollusks also are highly diverse, and many or most of them are marine organisms. Quite commonly they serve as guide fossils, especially in Mesozoic sedimentary rocks. Ammonites (cephalopods) are prominent in this (Fig. 8.1).

Producing highly diverse assemblages evidently takes long time spans. This requirement of time has been invoked by some geologists to deal with gradients of diversity of modern organisms from the poles to the tropics, which invariably shows the higher diversity in the (geologically much older) tropics. Fossil mollusks are plentiful even back in the early Paleozoic, hundreds of millions of years ago. Among the best known fossil forms are the "ammonites," extinct coiled cephalopods somewhat similar to the modern cephalopod Nautilus (Fig. 13.3) and serving as guide fossils in Mesozoic shelf rocks. While they were presumably largely pelagic, judging from distributions, we must assume that some of them were benthic, from arguments centered on morphology as well as on diversity patterns.

8.1.2 On the Contrast Between Benthos and Plankton Diversity

Snail shells are abundant on the deep seafloor in places, notably in pteropod ooze, an accumulation of pelagic shells, not benthic ones. There are much fewer species in pteropod shells than in benthic snails in general, the discrepancy being greater than a factor of 100 (i.e., hardly subtle). Since the shells of pteropods are easily dissolved in deep cold waters and in sediments rich in carbonic acid, the occurrence of pteropod shells tends to be restricted to the more shallow regions on the deep seafloor away from continents (as noted by John Murray of the *Challenger* Expedition). The fact that the pelagic shells

Fig. 8.1 The great diversity of marine mollusks, even without ammonites. Considerable diversity is evident at all taxonomic levels, including shelled forms that can guide geologists in biostratigraphy. Graph on *right* from the drawings by the German biologist Ernst Haeckel (1834–1919). *Left*: Large squid, Baja California. [Photo W.H. B]

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_8

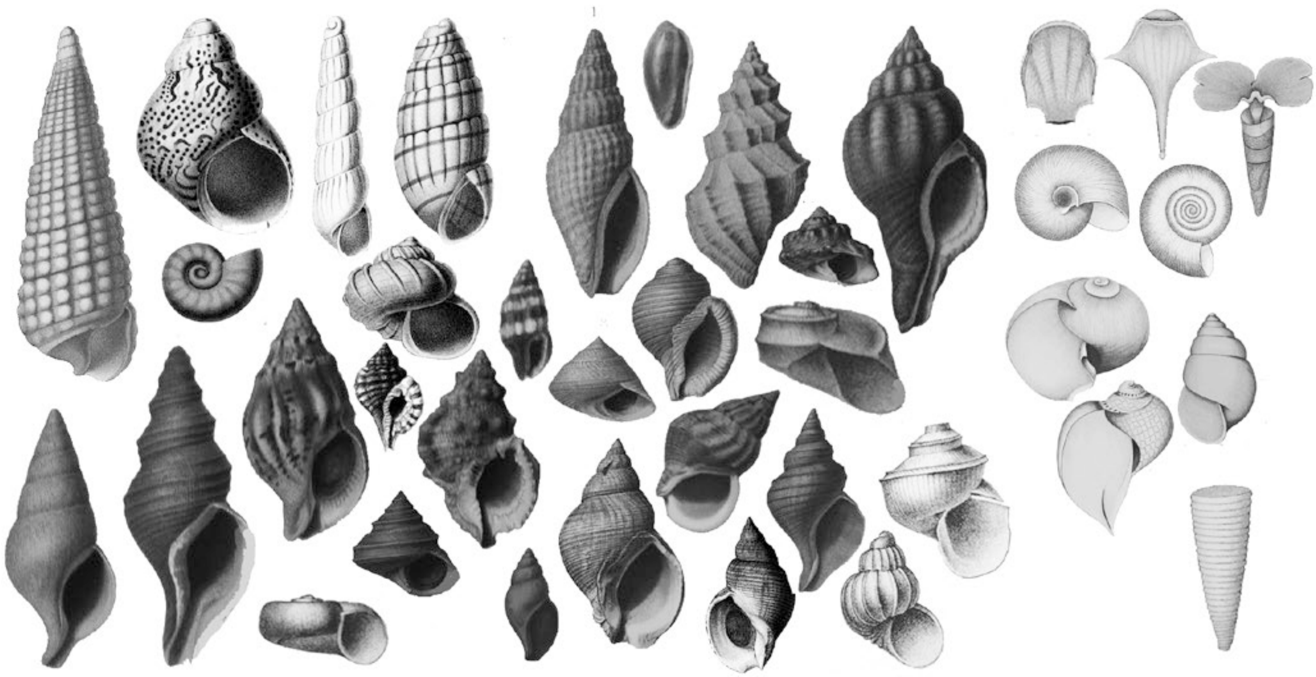

Fig. 8.2 High benthos diversity compared with plankton diversity. Marine gastropods from the North Atlantic. *Left* (dark): benthic forms (i.e., species living on the seafloor). *Right*: a selection of the light-colored and right (light gray): small plankton forms (i.e., gastropods living in the water and commonly called pteropods) [All drawings from expedition reports of Prince Albert I of Monaco, 1848–1922]

are made of aragonite, there is a likelihood that aragonite precipitation is widespread in mollusks (ammonites also had aragonitic shells). To some degree, fossil distributions reflect preservation, of course (Fig. 8.2).

As far as the latitudinal diversity change, it is well studied on the shelf of the US East Coast, where it is in striking evidence. Gastropods (snails) seem to show a greater change across latitudes than do bivalves (clams, mussels, cockles, and close relatives; also referred to as lamellibranchs or pelecypods). Many of the clams live *within* the sediment, in contrast to snails. Seasonal variability and inclement weather conditions appear to be important factors that help determine diversity patterns, presumably in addition to overall temperature patterns. In any case, the patterns observed presumably largely reflect evolutionary development, that is, history and adaptation. There are reasons why tropical species are more abundant than Arctic ones, and one reason may well be with the geologically short time that cold regions have been prominent on the northern hemisphere. The cold deep water may have to do with the fact that many benthic deep-sea organisms have close relatives on high-latitude shelves.

There is yet another consideration. Arctic environments are very demanding as regards benthic life. Paucity of species is a characteristic. One lagoon in Alaska, southeast of Point Barrow, was observed to be covered with ice for much of the year. On freezing, saltwater yields the salt that is excluded from the ice, thus producing brine. The brine is diluted again upon melting of the ice in summer, when also normal marine seawater enters the lagoon. The enormous variations in temperature and salinity attending these processes may be quite inimical to a proliferation of highly diverse shell-producing organisms.

As far as explaining the overall diversity difference between drifting organisms and their benthic relatives, we also might assume that the living space of the open ocean has fewer nooks and crannies than the seafloor. The many types of habitat on the seafloor presumably call for being filled by different species. In all benthic organisms, there are many different types of habitat to occupy for benthos, from shelf to abyss and from above the surface of the seafloor to various (shallow) depths below the seafloor. Such differences in habitat (and sediment types!) commonly are reflected in differences in taxonomy. Also in the case of benthic foraminifers, there are different ways to make shells ("agglutinating" forms use existing particles on the seafloor; calcareous forms precipitate calcium carbonate). In contrast, the various parts of the open sea are well connected, so that there is less opportunity than on the seafloor for evolution linked to specific localities.

8.2 The Seafloor as Habitat and the Task of Reconstruction

8.2.1 Depth and Food Supply

To a large degree, it seems, differences in biodiversity reflect differences in the diversity of available habitats and life styles. Perhaps the most drastic differences in benthic habitats are those arising between the shelf and the seafloor below the shelf edge. On the upper part of the shelf, there is sunlight as

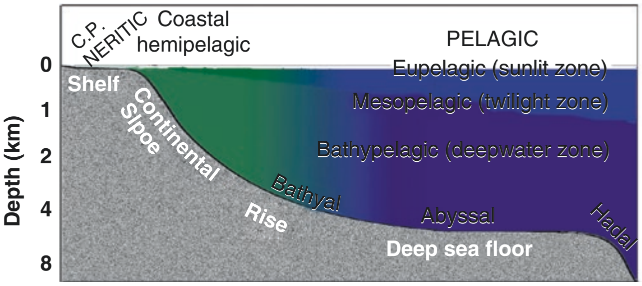

Fig. 8.3 Common environmental terms for the ocean and for seafloor studies. C.P., coastal plain. The beach forms at the upper shelf edge [Based on various texts in oceanography and geology, starting with an article by the late Californian marine biologist Joel Hedgpeth]. Not to scale

an energy source (in a layer less than about 30 m thick). In addition, nutrients are plentiful in places because of coastal upwelling and of other estuarine-type circulation. Stimulated by sunlight and nutrients, benthic primary production by marine algae can be very vigorous indeed. Sunlight limitation would imply that only 2% or 3% of the seafloor are involved in shelf benthic production. However, benthic production here can be 100 times greater than the regular background production of phytoplankton in the open sea. Kelp production in an upwelling setting, for example, has been estimated at 10,000 gC/m^{2}year (grams of carbon per square meter per year). Much benthic production is linked to shallow waters. It takes place in salt marshes and in kelp forests, but also in calcareous reefs, within the minute symbiotic algae of shallow-water coral. Some ecologists point out that higher food supply seems to come with lowered diversity.

Assuming that habitat is crucial for taxonomy, we need to focus on the environment to understand distributions of organisms. How are we to classify marine habitat in simple fashion? As is well-appreciated (see previous Chapter) productivity and the supply of organic matter to the seafloor decrease away from the shore and therefore quite generally with depth. Additional controls, especially on the shelf, are provided by temperature and seasonality in temperature. If the seasonality is between warm and not quite so warm, it is much less effective as a habitat factor than is variation that includes the formation of ice. Light is important; shallow waters have it; deep waters do not. Depth, in itself, may not be all that important, although it does imply great differences in water pressure. Nevertheless, simplicity of classification of environments (based on depth) commonly wins among the possibilities (Fig. 8.3). Food supply, while crucially important, is complicated and difficult to measure.

8.2.2 Salinity and Temperature Reconstruction

Life in the sea also depends on other factors besides sunlight and nutrients, that is, the chief controls on productivity. A large number of organisms can tolerate *salinity fluctuations* only if salinity stays between about 30 and 40 permil. Such species are referred to as *stenohaline* (*narrow salt tolerance*). Examples are among radiolarians, reef corals, cephalopods, brachiopods, and echinoderms. In general, the remains of such organisms indicate marine conditions for the sediments that contain them. However, a few representatives of the groups mentioned may have relatively wide tolerances toward salinity. Also, it is well to remember that remains embedded in sediments do not necessarily agree with the environment of deposition: they may be redeposited.

With increasing salinity, for example, in lagoons of arid regions, fewer and fewer species are able to survive, until only a few organisms remain – an *impoverished* fauna and flora. Some ostracods (bivalved crustaceans of millimeter size) can tolerate salinities of more than 10%. As such hardy forms have few competitors (or predators) in places requiring special adaptations, impoverished faunas can be very rich in individuals. Quite generally, of course, species-poor but individual-rich faunas indicate *restricted environments*. The word "restricted" in this case does not apply to space but to unusual environmental conditions that are stressful to most organisms, excluding them from thriving here.

In the open ocean, where salinities are well within the tolerance limits of all marine organisms, temperature change with depth may play a decisive role in controlling distributions. Temperature also presumably controls much of the observed latitudinal zonation (Chap. 9). In polar regions, the temperature in the water can fall to noticeably below 0 °C (when ice can form). In marginal seas in the subtropics, it can rise to well over 30 °C, as it does in the Red Sea and in the Persian Gulf. Outside the tropics, temperature can vary strongly with the seasons in the upper 100–200 m of the water column. Below this depth, however, the temperature stays rather low throughout the year. Most of the ocean is very cold: in the abyssal deep sea, the temperature stays below 4 °C, reliably so since the Neogene, with cold poles. The cold temperature in itself may not affect diversity. Instead, food supply may be the crucial factor, again. It is very low at the abyssal seafloor. Yet, there is an astounding variety of small benthic organisms. Concerning diversity, local disturbances may be important in providing different opportunities for making a living.

Can we reconstruct temperature distributions from the study of the remains of organisms?

In shallow seas and on the shelves, the oxygen isotopic composition of coral (Fig. 8.4) and also of mollusk shells, among other remains of carbonate-secreting organisms, provide records of changing temperatures within oxygen isotopes of shells.

Relevant studies by the Caltech geochemist Samuel Epstein (1919–2001) and associates introduced the methodology to marine geology. Some fossils definitely give a biased record – biased against unfavorable periods within their life histories. Clearly, whenever there is no or slow growth, there is no record or record compression. Abundance

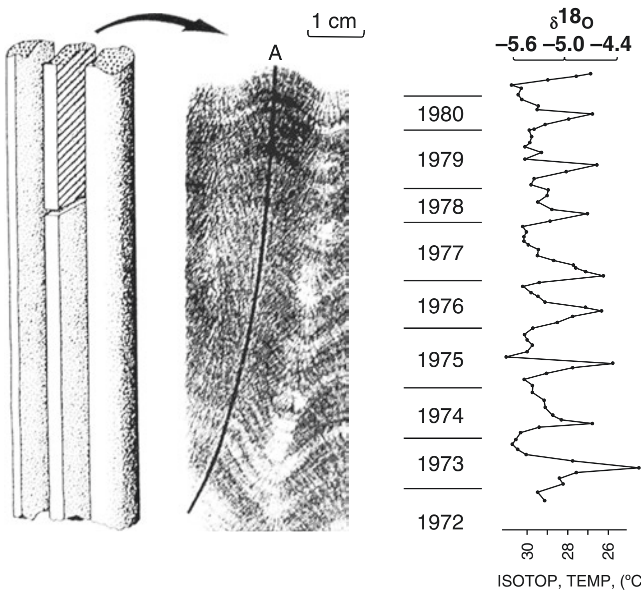

Fig. 8.4 Temperature reconstruction from isotopic variation in *Porites lobata*. In the example shown, sampling was along the profile marked *A*. A second profile (not shown here) was used to check on the width of *error bars*. Temperature varies by several degrees Celsius (colder T is to the right.). Details require additional information, e.g., on precipitation. Note the biased seasonality of the signal: Growth is largely in summer [X-ray photo and graph of oxygen isotopes courtesy of J. Pätzold, Kiel, and Bremen].

distributions of organismic remains on the seafloor change with temperature, and considerable effort has been spent on defining the appropriate statistical relationships that allow a conversion from changes in the distribution of foraminifers and other eukaryotic microbes to changes in temperature patterns, notably by John Imbrie (Brown Univ.) and his associates. For shelf seas and for enclosed basins, the reconstructions are rather more difficult than for the open ocean. At the margins, environmental factors other than temperature play an important role. Thus, a change in the faunal assemblage may reflect stress (or release of stress) in salinity, muddiness of the water, weather condition, or some other influence. One difficulty that exists for both deep and shallow environments, as far as temperature reconstruction, is the selective preservation of faunal and floral assemblages within the sediment. To separate the effects of early diagenesis from those useful for estimating the temperature of growth can be a formidable task, for any and all methods of reconstruction and throughout the geological record.

8.2.3 Oxygen Reconstruction

The content of *dissolved oxygen* in the water is an environmental factor of great importance, especially for cases in which concentrations are close to low levels of 1 ml of oxygen gas per liter of water or less. (“Normal” today is 4–7 ml/l, with the colder water holding more oxygen.) When conditions are “*anoxic*” (no oxygen) or “*dysoxic*” (extremely low oxygen), higher organisms (including shell-bearing protists) may stay away, eventually leaving only anaerobic bacteria and archaea to populate the sediment. When only microbes are present, v*arved sediments* can form, as there is then no disturbance by burrowing organisms. Much can be learned from such sediments regarding climate change on the scale of decades to millennia provided the record is continuous (Chap. 15).

The shells of certain benthic foraminifers (i.e., relatively large eukaryotic microbes) have been used to reconstruct the variation in the level of oxygenation in basins off California by the USC (Los Angeles) paleontologist R. Douglas. A pioneering effort with a somewhat similar goal of documenting the changing abundance of benthic foraminifers with water depth was the mapping of portions of an oxygen minimum layer in the Mediterranean by the US micropaleontologist F.L. Parker, in the 1950s, using samples from the Swedish *Albatross* Expedition (Fig. 8.5).

In our time of the ice ages, with oxygen tending to be abundantly present in the marine environment, we have to go to certain special areas to study anaerobic conditions and the sediment they generate – for example, certain fjords in

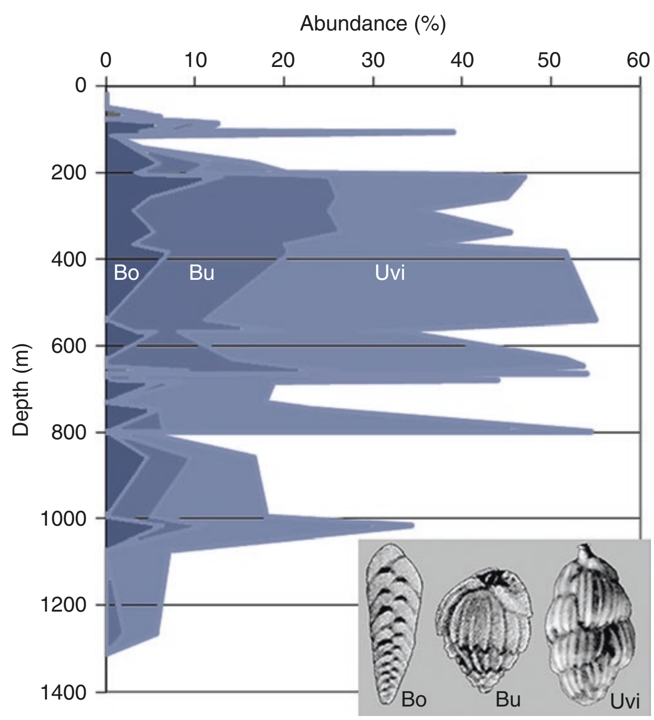

Fig. 8.5 Oxygen minimum zone seen in benthic foraminifers (genera *Bolivina*, *Bulimina*, and *Uvigerina*) in the eastern Mediterranean. Data: Study by F.L. Parker, based on Albatross samples featuring quantitative tables for fossil species (a pioneering feat itself). The three species also respond to high production [Drawings of foraminifers: H.B. Brady, Challenger Expedition]

Norway and Alaska, the Black Sea basin, and the Santa Barbara Basin off southern California north of Los Angeles. In the distant geologic past, when the poles were not icy cold and did not yet deliver oxygen-rich water to the deep ocean, conditions of oxygen deficiency apparently were much more common (Chap. 13).

Basically, oxygen deficiency arises where demand for oxygen is strong and supply is weak. For example, in the Black Sea, the saltwater filling the basin (through the Bosporus, from the Mediterranean) is covered with a layer of freshwater brought in by the Danube, the Dnieper, and other rivers, thus generating an estuarine situation typified by high production (i.e., there is plenty of organic matter). The light freshwater forms a lid on the heavy deep water, thus greatly reducing the oxygen supply by cutting down the exchange of gases with the atmosphere. As a result of the combination of high production and reduced supply of oxygen, the deepest waters of the Black Sea are entirely anaerobic.

The reconstruction of the level of oxygenation is in all cases a highly challenging task for the geologist, involving the interpretation of clues from lamination, from the presence and nature of burrowing, from mineralogy (such as habit and abundance of sulfides), and from the types of organic matter preserved. In addition, the abundance and composition of benthic foraminifer assemblages hold useful clues to both low (or high) oxygen values and correspondingly high (or low) productivity.

8.3 Living on the Seafloor

8.3.1 General Features

The greatest part of the seafloor is teeming with benthic organisms. There is benthos that stays put, called *sessile*. All sponges, corals, brachiopods, and bryozoans are sessile. And there is benthos that moves about, termed *vagile*. Vagile organisms can move rapidly, like a startled crab, or slowly, like the sluggish sea urchins, starfish, most bivalves, snails, and most worms. Both groups, sessile and vagile, have members living *on* the floor or attached to other exposed organisms (such as mussels or snails or kelp): the *epifauna*. Or they live within rocks or sediment: the *infauna*. Representatives of the epifauna are several times more abundant than those of the infauna, for reasons not known (but perhaps similar to those governing the tropical-high latitude contrast).

Benthic animals, ultimately, live on foodstuff falling down through the water or coming in from upslope along the seafloor. Both living plankton and dead *detritus* drift in the water: the *seston*. The detritus consists of organic and inorganic particles. One consequence of the dependence of the benthos on food from the sunlit zone is a pronounced decrease in benthic biomass with depth: the farther away from the productive coastal zone, the smaller the amount of nutritious rain reaching the benthic environment. Although so very little reaches the abyss (order of 1% of production or less, as suggested in Fig. 7.7), hundreds of species – tiny crustaceans and worms – and hundreds of types of benthic foraminifers make their living from the scraps coming down. They must be extremely energy-efficient to survive on such starvation diets. Exceptions to the meager nourishment are provided by falls of large dead organisms. The carcasses are quickly found by highly mobile forms or settled by large organisms growing from larvae presumably looking for this type of opportunity.

The *sessile benthos* waits for the water to bring suspended material from which to extract nourishment. The *suspension feeders* (sponges, corals, brachiopods, crinoids, bryozoans, and others) filter the water, either passively, using the natural flow of the surrounding water, or actively, moving water past their straining apparatus. *Commensals* may seek a free meal in addition to shelter provided by a *host*. For foraminifers living in sponges, it is convenient that the host provides both protection and food. Sessile benthos is especially common in food-rich agitated environments where the water is not too muddy.

The *vagile benthos* on a rocky substrate – starfish, sea urchins, gastropods, and ostracods – commonly feed on epibenthic organisms, for example, by scraping minute algae off the rocks or by preying on sessile animals. The mobile epifauna protects itself from storms and predators by hiding in nooks and crannies, by growing thick shells, and also in cases by firmly clinging to the rock, as do the chitons and patellas using their strong sucker foot. On soft bottom (in less agitated environments), most vagile benthic animals ingest sediment, especially surface detritus that contains a good portion of organic particles (*deposit feeders*). Others hunt for prey. Many use the soft bottom to make holes for hiding, as do fiddler crabs in intertidal mudflats in Southern California (Fig. 8.6) and the ubiquitous “lugworm” in the “wadden” (next section).

8.3.2 The Wadden

The most intensively studied wetlands for large organisms living within soft sediment is the intertidal flat called the “wadden” at the rim of the eastern North Sea. A great abundance of burrowing clams and a long polychaete worm (“lugworm”) is notable in these muddy and sandy deposits (Fig. 8.7). The substrate commonly is strongly influenced by the activity of benthic organisms, making bioturbation and other aspects of biological sedimentation an integral aspect of the environment.

Organisms living within the sediment are effective in hiding from potential predators. However, hiding can interfere

Fig. 8.6 Fiddler crabs next to burrows in an intertidal mudflat in Southern California. *Left*: male. *Right*: female. Hole diameter: 0.5–1 in. [Photos W.H. B]

with getting oxygen or food. Many of the infaunal organisms have means to access the sediment surface, where there is nutrition, without exposing most of the body (such as a siphon in burrowing clams). In contrast, organisms living on top of the sediment and on rocks commonly have a different problem, that is, how to fend off or avoid predators aware of their presence. The differences between infaunal and epifaunal habit are rather evident to an observer examining hard parts. But In some cases, differences are strictly behavioral, which means they tend to be cryptic for geologists, although some of the behavior of organisms in and on soft sediment can be fossilized as tracks and burrows, of course.

8.3.3 Carbonate Production Rate

One aspect of ecology that is of central interest to geologists is the *production rate* of hard parts of organisms, that is, mainly of carbonates (Figs. 8.8, 8.9, and 8.10). Availability of light is a crucial factor in carbonate production, making shallow water a favorite place for this type of production. Off Miami, the macrobenthos (large organisms) has been found to produce annually about 1000 g of carbonate per square meter in the tidal zone and between 1 and 400 g per m^2 in the deeper water offshore. In shallow areas of the Persian Gulf, a single species of foraminifer (the light-processing symbiont-bearing *Heterostegina depressa*) delivered annually 150 g of carbonate per m^2.

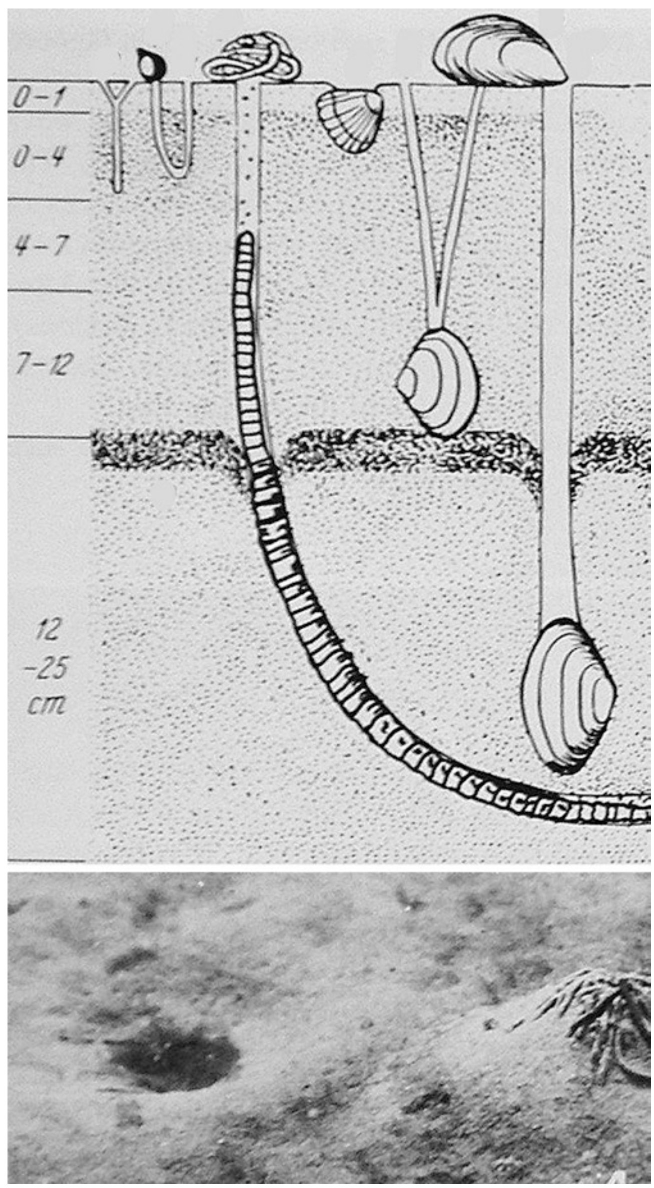

Fig. 8.7 Sediment reworking by infaunal benthic organisms in sandy tidal flats of temperate regions. One of the most abundant and active large wadden animals is the lugworm [From H.M. Thamdrup, 1964, in R. Brinkmann (ed.) Lehrbuch der allgemeinen Geologie. F. Enke, Stuttgart, modified. See Schweizerbart website for "Enke." Photo below graph: E. S]

Calcareous algae are very common contributors to the sediment made by benthic organisms. As primary producers, they need sunlight, that is, they occur in shallow waters – actually right into the surf zone in favorable habitats. Being quite accessible, they have been studied extensively for their rate of growth (Fig. 8.8), with important implications for the rates of sediment production as well. Rates of production are significant, as is the contribution of the algae to carbonate platforms and soft calcareous sediments, especially in subtropical regions.

The production of carbonate in almost all stony corals is entirely dependent on the activity of symbiotic algae ("*zooxanthellae*") living within the coral tissue. The coral species in question grow in shallow warm water, wherever temperatures do not drop below about 20 °C, that is, in tropical and subtropical waters. The organisms are colonial. Many of them make the familiar bush-like calcareous structures, and they need a rocky substrate to anchor to. The symbiosis between coral animals and their dinoflagellates allows these benthic organisms to flourish in rather nutrient-poor waters, that is, in the deserts of the sea. Nutrients are captured and recycled by the animals, and carbon is fixed by the algae. (In recent years, environmental conditions of coral growth are widely reported to have deteriorated. Much "bleaching" has been observed, a process preceding coral death. Commonly heating and acidification are cited as stressors, as well as disease of weakened corals.)

While there is tremendous biodiversity associated with reefs, the coral genus *Acropora* is clearly among past and present dominant forms within the Caribbean. Bush-like corals also dominate elsewhere (e.g., in the Great Barrier Reef in the Pacific, Fig. 8.9) and in many other places rich in reef coral. *Acropora* is a fast-growing form. Presumably the coral was selected for keeping up during the rises of sea level whenever much of the polar ice masses on the northern hemisphere melted on termination of a glacial period. Growth rates of modern individual coral can easily exceed 1 cm per year, suggesting that fast coral growth can be linked to necessity, at least on the millennial scale. Early determinations were based on coral growing on sunken ships. (This is not the same rate as that of the buildup of a reef, of course.) According to the marine paleontologist and reef scientist Jeremy Jackson (S.I.O. and Smithsonian), the coral genus *Acropora* has suffered significant reduction in abundance in the Caribbean in past decades, compared with earlier distribution, presumably largely from human impact, including fishing (pers. comm., 2000).

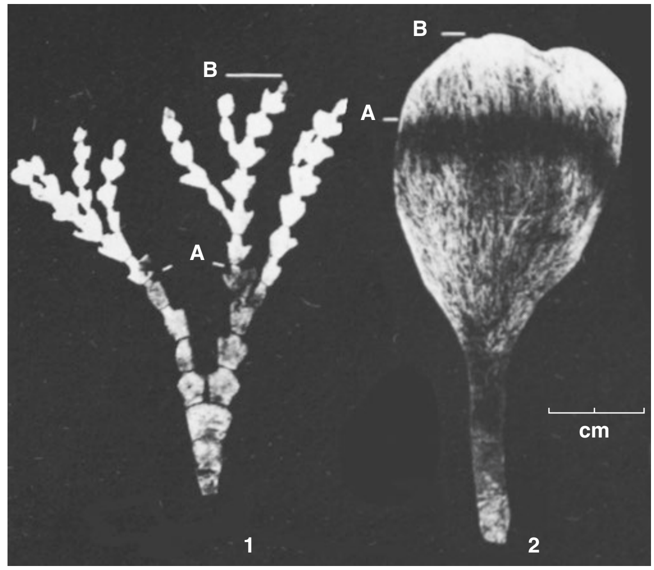

Fig. 8.8 Production of algal carbonate. The rate of production was measured in the field by staining live specimens and noting growth till collection. (*A*) Termination of staining procedure and (*B*) collection. (**1**) *Halimeda* and (**2**) *Penicillus*. Both types of algae have numerous species. The ones shown are common in Bermuda [Photos and experiments by G. Wefer, 1980. Nature 285: 323; Photos courtesy of G. Wefer]

In the geologic record, the evidence that benthic foraminifers with photosynthesizing symbionts were ubiquitous in once widespread shelf seas is very strong. Grain-shaped Fusulinids 200 million years old apparently belong into the group, as do coin- and lens-shaped relatively large benthic foraminifers from the shelf of the Tethys sea, from the Mesozoic to the early Cenozoic. Thus, large foraminifers were still dominating certain shelf environments in the early Tertiary. For example, Eocene nummulites (up to dime size and even larger) are famous for having made limestone used in building the great pyramids in Egypt. The species *N. gizehensis* bears a name that reflects the association. Quite generally, symbiosis is thought to be a major driver of evolution in foraminifers, both benthic and planktonic. In any case, the association between photosynthesizing algae and shallow-water eukaryotic foraminifers (and other shell-making fossils) results in an enormous output of carbonate minerals (Fig. 8.10).

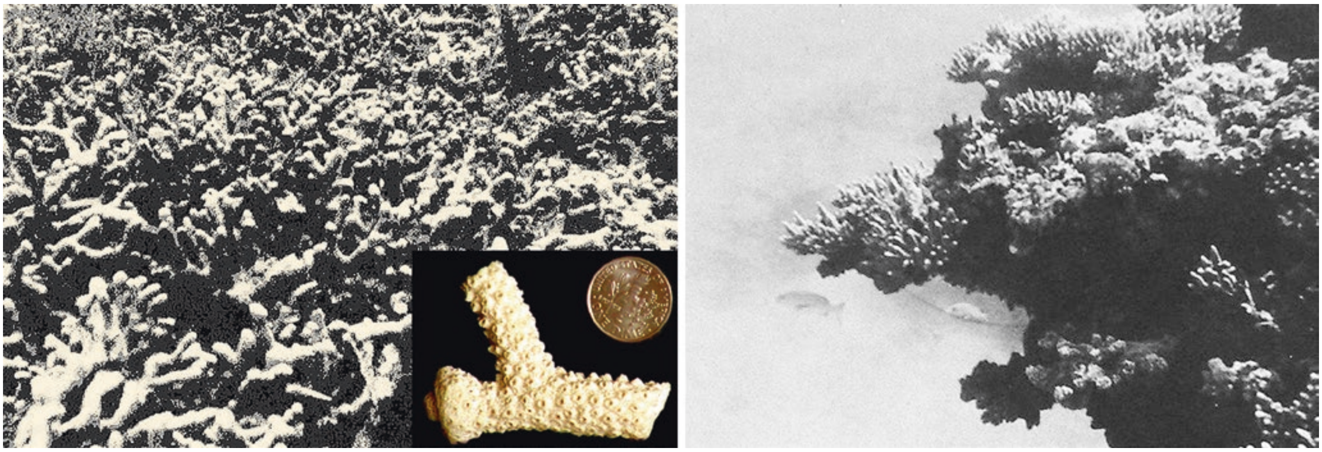

Fig. 8.9 Bushy coral growth. *Left*: *Acropora* meadow, Florida. Penny-sized coin for scale of fragment (Photos W.H.B.). *Right*: Lizard Island, Great Barrier Reef (Photo E. S)

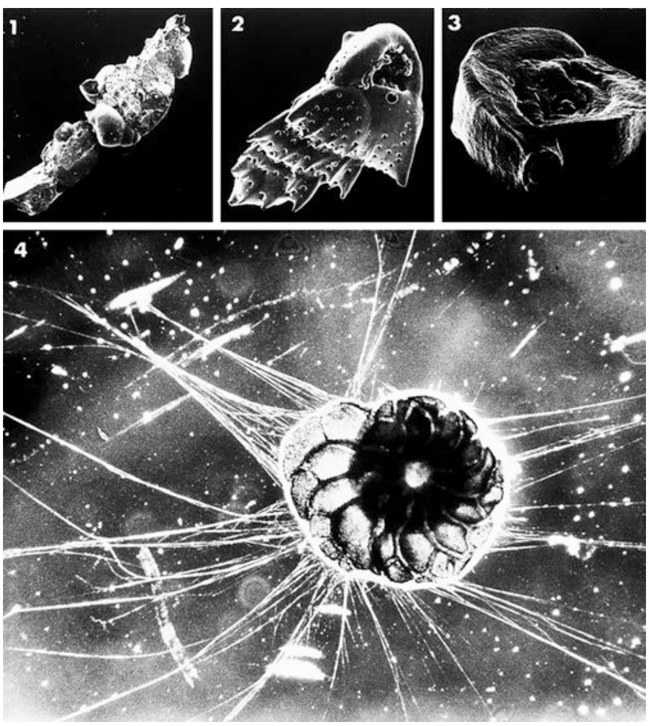

Fig. 8.10 Benthic foraminifers without and with photosynthesizing symbionts. (**1**) Arenaceous form (*Reophax* sp.) agglutinating available particles from the sea floor; (**2**, **3**) calcareous forms (*Trimosina* sp., ×150; *Spiroloculina* sp., ×50). The species of *Trimosina* shown lives below about 25 m in the Indian Ocean. *Spiroloculina* prefers shallow water and coarse substrate. (**4**) *Heterostegina depressa* (diameter of test: 0.84 mm) is a symbiotic form living in sunlit waters. The "pseudopods" radiating from the test are used in locomotion and as anchor. In other forms pseudopods can serve for catching prey. [SEM photos of benthic foraminifers courtesy C. Samtleben and I. Seibold, Kiel. Microphoto courtesy R. Röttger, Kiel]

8.3.4 Life on Bedrock

A great number of different materials can be called "rock": ancient sandstones and limestones of wave-cut platforms or along submarine canyons, basaltic scarps and manganese pavements, submerged dead coral and cemented beach sand (*beach rock*), and boulders dropped from icebergs or in moraines on the shelf. For animals normally attached to natural rock, sunken ships, and other man-made objects can serve as a substitute for attachment. All such substrates can be densely covered by benthic organisms, in some cases (if resting on deep-sea sediment or on mud) forming epibenthic "oases" on an otherwise empty-looking seafloor. The most abundant type of rocky substrate presumably is provided by basaltic outcrops on the deep seafloor, but the most visible colonies are in rocky shores of the intertidal zone and on rocks of other highly productive regions (Fig. 7.2). The abyssal basaltic outcrops are commonly rather barren (except in local spots of hydrothermal exhalation).

In shallow sunlit areas, rocky bottoms are commonly covered by algae. The algae can be *diatoms* (many of these move about on the rocks) or, in the case of macro-algae, up to several meters long soft ribbons waving in the currents and providing hideouts for fishes. Rocky areas may be good fishing grounds (even though they can be hard on a fisherman's net). Encrusting algae tend to cover rock outcrops permanently. Depending on water temperature and depth, familiar epifaunal associations are corals, tubeworms, oysters, barnacles, bryozoans, and encrusting foraminifers. Sponge thickets grow in certain nutrient-rich places that have plenty of dissolved silicate, especially around Antarctica.

The infauna on rocks consists of organisms that bore actively into the substrate (relatively soft rocks being preferred.). The remarkable ability to do so has been acquired by many types of organisms, including sponges, worms, and mollusks. Some sea urchins with unpleasantly (for human swimmers) long and brittle spines live in custom-made cavities. Some algae and some microbes also live within rock; some process sunlight within transparent rocks in shallow water. At first glance, a heavily burrowed rock may look quite solid, the entrances to burrows being small. However, in reality, the rock may look like Swiss cheese inside and thus, in the tidal zone, be very vulnerable to attack by storm surf. Shallow-water limestone is especially susceptible. In certain warm-water lagoons of the Pacific coast, some 30% of the calcareous sediment is delivered by boring sponges.

In shallow rocky areas, organisms commonly are removed periodically by wave action. Their remains collect in depressions at the base of the rock outcrops. Much of the calcareous *reef talus* deposits are of this origin. They are coarsely layered thick sequences surrounding a reef structure and are familiar from ancient reefs exposed on land (e.g., the Permian El Capitan Reef in Texas and New Mexico, Jurassic Malm reefs in central Europe, Cretaceous Albian reefs in Arizona). In the oilfields of the Middle East, highly porous reef talus typically serves as a reservoir rock for petroleum.

8.3.5 Life on a Soft Substrate

We have earlier referred to organisms living in and on mud. The benthic communities on muddy or sandy seafloor are commonly not much in evidence. Organisms living on soft sediment tend to hide by burrowing. Another factor to consider in assessing abundances is the stability of the substrate. Where the seafloor is moderately stable, sea grass can take hold and further stabilize it. Epibenthic diatoms, foraminifers, and bryozoans grow on such grass. Vagile benthos with or without burrows to return to can be quite abundant on sandy and muddy bottom; crabs and snails are common sights in the intertidal zone. A large variety of nonmarine invaders seek food here during low tide, mainly birds feeding on resident organisms. Conversely, during high tide, the invasion is from the sea. It brings plenty of shallow-water fishes digging up worms and mollusks.

The abundance of predators along with the sporadic shifting of the sand, which can suddenly expose the infauna, fosters evolution of the ability to burrow very rapidly. Some clams demonstrate this adaptation very obviously, as is well known to clam diggers. Burrowing clams have a long strong foot that is extended into the mud or sand below and then inflated by water pressure anchoring it, with the rest of the body following by muscle contraction. A smooth shell, usually quite sturdy, characterizes such clams.

Burrowing clams regularly are suspension feeders with an inhaling and an exhaling siphon. After death of a bivalve, the shells – hydraulically quite different from the surrounding sand – are sorted out and concentrated into layers of *coquina*. Presumably reflecting certain favorable conditions for clam growth, such coquina deposits are common in the geologic record and can be used as marker beds locally (i.e., assuming they represent a time horizon), when mapping.

There are some types of challenges that differ for sandy and muddy substrate even though both are soft. In consequence the faunal assemblages of muddy habitats can be quite different from those of sandy environments. As a result of the presence of fine material, the muddy substrate is somewhat firmer than the sandy one because of the high cohesion between clay particles. Thus, the mud is more difficult to burrow into than the sand, but burrows once made have a better chance of persisting. In addition, the high content of organic matter in clayey sediment makes it worthwhile for many organisms to pass the mud through their guts and extract the digestible fraction. For this reason, *deposit feeders* are commonly found on and in this type of substrate. Many of the larger burrows, though, are made by crabs and other mudflat denizens hiding away from sunlight and its predators.

Of course, *suspension feeders* would thrive on the organic matter in clayey sediment also, just as do deposit feeders. However, their filtering apparatuses tend to become clogged with the abundant clay, which is useless to them. Deposit feeders, incidentally, make it difficult for sessile benthos to find a stable foothold by constantly reworking the mud. Thus, once there is much deposit, feeding it tends to exclude sessile organisms, including many plants.

Commonly, on muddy seafloor in shelf waters, brittle stars (ophiurids) are abundant (Fig. 8.11). These are suspension feeders that are anchored within the sediment, with only their arms exposed, for catching food. Their abundance may grow with an increase in food supply (i.e., *eutrophication*). The Danish ecologist Gunnar Thorson (1906–1971) has studied shallow-water communities in northern latitudes. He suggested that such associations are similar for various upper shelf regions, albeit made up of different species (*parallel communities*). In many of them, brittle stars are important members.

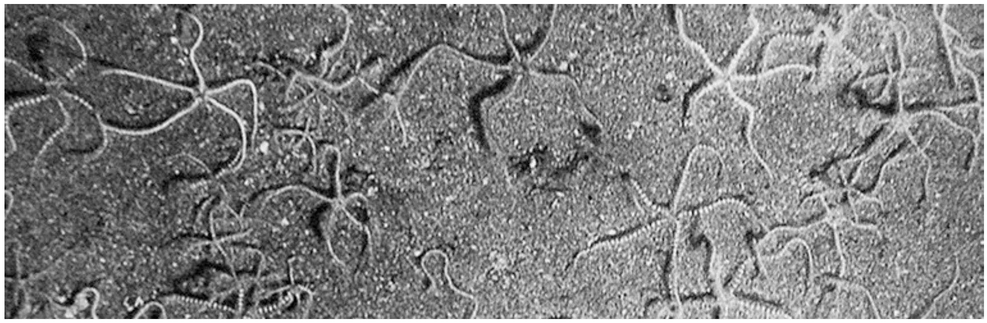

Fig. 8.11 Brittle star community on the uppermost continental slope off Senegal [Photo E. S]

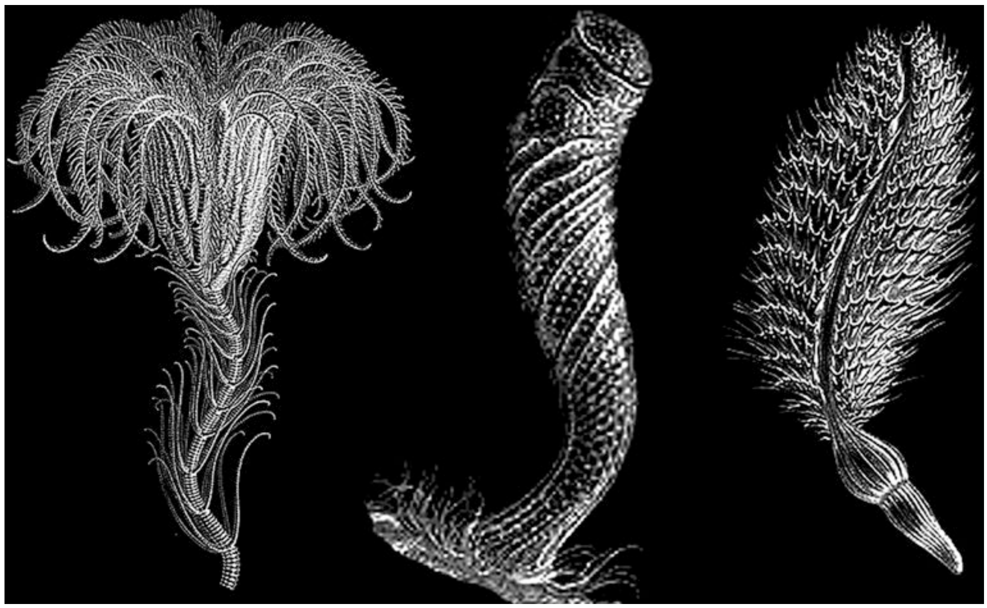

Fig. 8.12 Sessile benthos on the deep seafloor: crinoid, sponge, and sea pen (soft coral) [After E. Haeckel]

On the muddy upper slope off Senegal (NW Africa), abundant brittle stars were seen on the very top of the seafloor, presumably filtering the water. The abundance there of these animals suggests "restricted" conditions unfavorable for diversity (lack of oxygen?) rather than a "regular" Thorson community.

Muddy bottom, it turns out, largely consists of fecal material, many times recycled. This is also true for the deep seafloor, although here the cycling takes thousands of years rather than months as in tidal mudflats. The low level of benthic activity on the deep seafloor and the relatively greater importance of the infauna is illustrated by the fact that on more than 100,000 photographs from 2000 different deep-sea stations, only about 100 visible animals were counted. As B. Heezen and C. Hollister pointed out in their book on the deep seafloor, what is seen in deep-sea photographs largely belongs to the echinoderms (e.g., as is true for holothurians). Some of the animals seen are sessile suspension feeders found on the continental slope: crinoids, sponges, and sea pens (Fig. 8.12). There is, however, plenty of indirect evidence for the activity of vagile burrowing infaunal animals in barren-looking "Red Clay" (Fig. 8.13).

8.4 Trails and Burrows

8.4.1 Trace Fossils

An entire branch of geology – *ichnology* – has grown from the study of the tracks, trails, burrows, and other sedimentary disturbances made by organisms. Such disturbances include *bioturbation*, that is, the mixing that disturbs the orderly recording of events in all sorts of sediment, from which we hope to extract detailed sequences reflecting history. Effects of bioturbation are always annoying but especially so in marine deposits with a high potential for resolution on a century scale. On the deep seafloor, even just

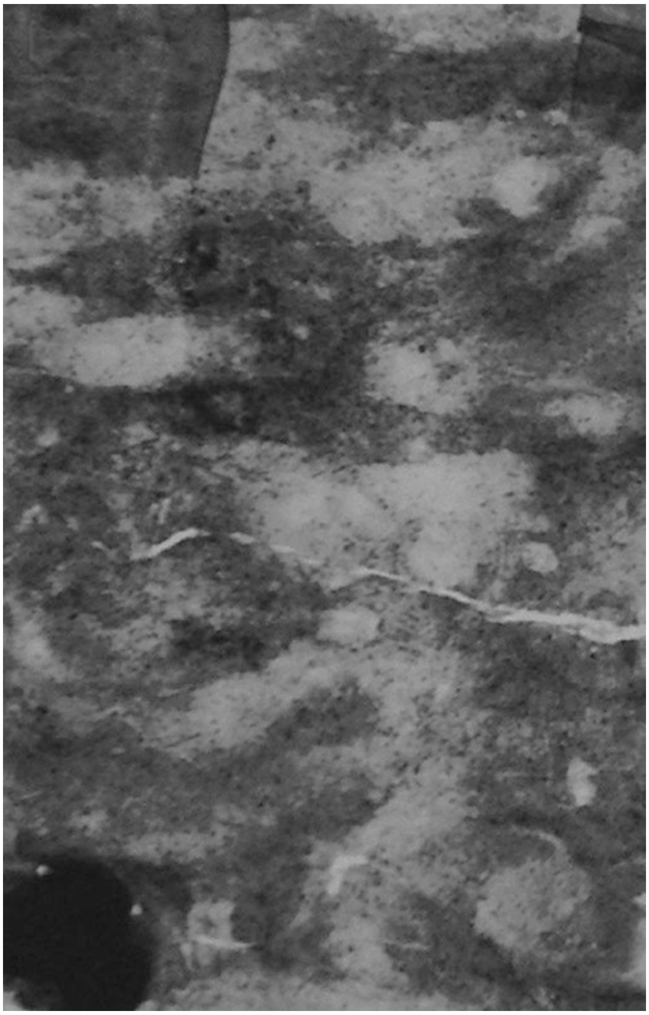

Fig. 8.13 X-ray radiograph of a slab of deep-sea sediments. The core was taken about 1000 km SE of Hawaii in a water depth of roughly 5 km. The image is a positive. It shows abundant burrowing within the deep-sea clay ("Red Clay") during the millennia represented by the sediment [Photo courtesy of F.C. Kögler, Kiel]

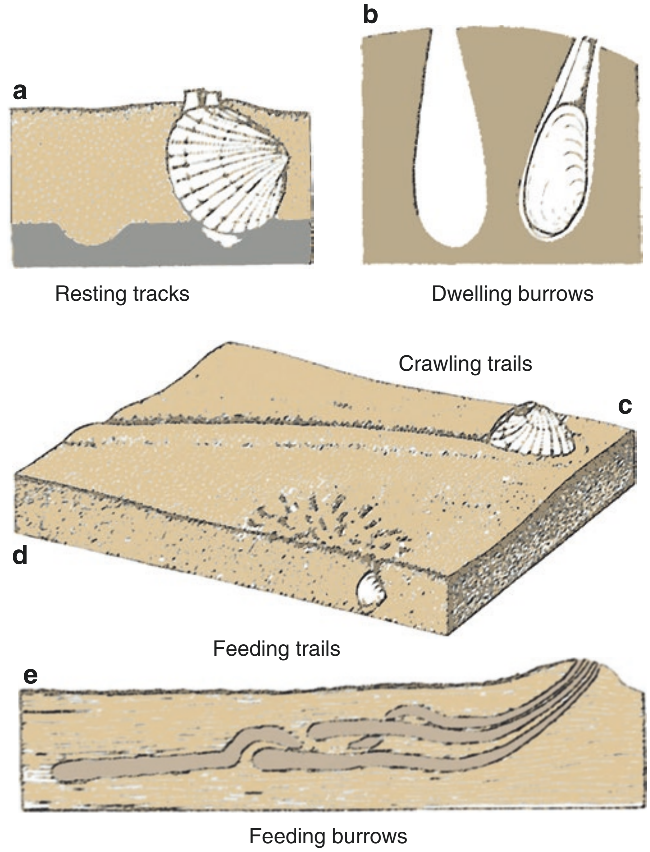

Fig. 8.14 Various types of "lebensspuren." (**a-d**) Bivalves; (**e**) burrowing worms [A. Seilacher 1953. Neues Jahrbuch Geol. Palaeont. Abh. 96:421 and 98: 87. See Schweizerbart website. Minor modification; color here added]

attaining a millennial resolution may require the *stacking* (averaging) of several records to reduce idiosyncratic errors from bioturbation by large animals that move sediment.

There is a great variety of traces that eventually become trace fossils: trails, feeding tracks, fecal strings and mounds, burrows stuffed with fecal matter, burrows used for housing (and later filled by washed-in sediment), and other legacies of animal activities (Fig. 8.14). Various kinds of worms, snails, bivalves, crabs, holothurians, sea urchins, and starfish make such tracks and burrows, and it can be very difficult indeed to assign the likely maker to a given *ichnofossil*. Crabs and shrimp make the longest and deepest burrows. In certain cores from off Northwest Africa, vertical burrows more than 3 m long were found! Such "*lebensspuren*" have been and are being studied intensely by marine geologists in Wilhelmshaven on the North Sea, in Kiel (Schleswig-Holstein, next to Denmark) and in Tübingen (Baden-Württemberg, next to Switzerland). Thus, the German word *Lebensspuren*, which is "life traces" in English, has been adopted as a technical term in geology.

8.4.2 Lebensspuren

Tracks and burrows preserve environmental information. Unlike shells, lebensspuren cannot be transported by currents but stay in the sediment where they were made. Thus, they definitively indicate in situ conditions, whether it be food supply, oxygen availability, sediment stability, or water motion.

Just like shells (which are inanimate particles after death of the maker), tracks and burrows can transmit information pertaining to physical geology rather than to ecologic reconstruction. In the deep sea, for example, vertical burrows do

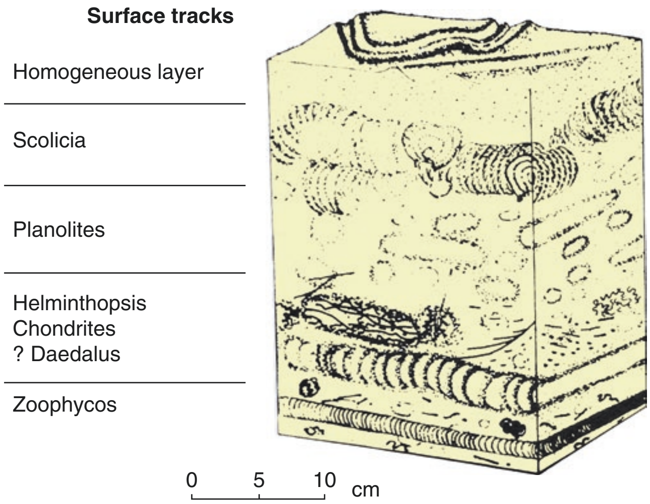

Fig. 8.15 Biogenic sedimentary structures in deep water off NW Africa (at about 2000 m). Tiered arrangements of active burrowers reach about 30 cm into the sediment. Bioturbation homogenizes the uppermost 3 cm of the sediment, destroying surface tracks and smoothing the record. Various burrowers make traces within the sediment. These tend to persist but likewise affect the reliability of the historical record negatively [A. Wetzel, 1979, Ph.D. Thesis, Geol. Inst. Kiel; slightly modified; color here added]

not last if the sediment has a tendency to shear horizontally by moving downslope. The effect is seen in deep areas, where dissolution of carbonate makes the sediment soft. Sediment presumably creeps downhill whenever shaken by earthquakes, a common occurrence in areas with volcanic activity. The motion destroys vertical burrows. Such downhill gliding may also be important on many continental slopes in areas of a muddy seafloor substrate, notably in volcanic areas or sedimentary slopes with high gas pressures.

Lebensspuren show different degrees of preservation, of course. A delicate trail made by a starfish or a mussel has but little chance of entering the record. On the other hand, a 30-cm-deep *Arenicola* or *Mya* burrow or especially a 2-m-deep crab burrow has an excellent chance, since much sediment would have to be moved to obliterate it once the burrow is made. In general, surface tracks can only be preserved by catastrophic deposition such as by flood or turbidity current. In the case of continuous accumulation, when sediment builds up gradually and there is a mixed layer from bioturbation, surface tracks are destroyed, and only burrows penetrating the mixed layer are preserved in the record (Fig. 8.15).

8.5 Partial Preservation and Bioturbation

8.5.1 Gaps in the Record

Only a selected portion of the biological community can be fossilized. Of Thorson's parallel communities of shallow water environments, for example, we can expect to find some of the shells and some of the burrows he identified but not all of them. Nothing readily recognizable will be left of most worms or shrimp. Nothing much remains of the fish and other predators that come during high tide to feed on intertidal organisms, and nothing virtually of the predaceous birds that take advantage of low tide to capture food. Nevertheless, predation from flatfish, rays, and birds presumably exerts considerable control on the composition of the nearshore communities. For example, muddy sand and deeper waters are commonly settled by the *Syndosmya* community. The name-giving clam is a favored food item of certain flatfish – something that certainly would be hard to tell from the fossil record, putting a crimp in reconstructing the evolution of both the clams and the flatfish.

8.5.2 Bioturbation and Radiocarbon

Bioturbation prevents the preservation of thin layers, whether thin turbidite layers or layers produced by contour currents (contourites) or some other layer. The result is a loss of resolution in the sediment record. In times past, however, when the ocean was much less well oxygenated, burrowing organisms may have been much less abundant, so that the process of homogenization would have been less widespread than today. In ancient shelf seas, we quite commonly see finely layered sediment in the geologic record on land. Perhaps only since the planet acquired ice caps has there been plenty of oxygen supplied to waters in contact with the seafloor. In summary, possibly bioturbation was possibly much less vigorous before there was ice.

In finely layered sediments, each layer has a message somewhat different from adjacent ones. Scientific questions are usefully centered on the length of time for which an average is made. Clues are derived from the *dating* of the sediment. Appropriate detailed studies have been made for deep-sea box cores with some promise for (geologically) detailed paleoclimatic reconstruction (Fig. 8.16).

For calcareous sediments younger than 30,000 years, radiocarbon-age dating can deliver excellent results. The *radiocarbon* (^{14}C) enters the sedimentary record within the calcium carbonate of calcareous shells. A certain proportion of the carbon dioxide in the air contains the radioisotope, which is produced from ^{14}N, by bombardment with cosmic particles. The carbon-14 produced decays back to nitrogen, half of it being gone in 6000 years. This is called the "half-life." The short half-life (6000 years being geologically "short") implies a severe limit on dating: after 42,000 years, only a tiny fraction of the original radiocarbon is left ($1/2^7$ or 1/128). At such low concentrations of radiocarbon (<1% of the original ^{14}C in a sample), contamination problems become huge on a planet (ours) with lots of carbon. Other problems also arise, and they affect the entire radiocarbon record. For example, an assumption has to be made about the

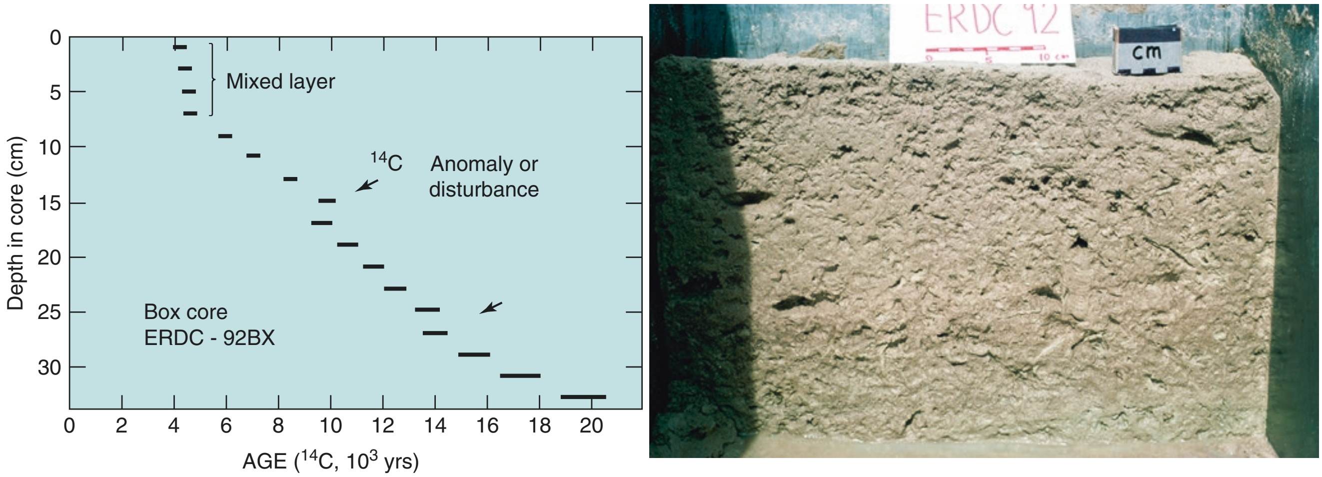

Fig. 8.16 Radiocarbon stratigraphy of box core ERDC-92 from Ontong Java Plateau in the western equatorial Pacific [Graph and data in T. H. Peng et al., 1979. Quat. Res. 11:141]. *Right*: ERDC 92 shortly after recovery and washing of face [Photo W.H.B. and J.S. Killingley]

original concentration of the radiocarbon where a shell was made. The assumption introduces guesswork to otherwise perfectly quantitative methods.

The radiocarbon stratigraphy of the deep-sea box core shows the signal of a mixed layer for the uppermost several centimeters, suggesting that assuming a homogenizing layer of 2 cm is quite conservative (Fig. 8.16).

In the example shown, there seem to be two major disturbances in the radiocarbon record after the last glacial maximum, in the millennia during the transition from glacial to postglacial conditions. The nature of the disturbances suggests sudden deposition of upslope material. The timing of the deeper disturbance is close to the age of a proposed impact of a bolide from outer space on the surface of the Earth. If it hit close to a continental margin, such a bolide would probably have set off tsunamis (either directly or via earthquakes and landslides), which disturbed sediments on top of submarine mountains such as the one where the core was taken. However, the sediment is heavily burrowed here, so the disturbance may owe to bioturbation rather than impact. Alternatively, it may be the result of redeposition of shallower ooze, for example, following a quake on the plateau. What is most puzzling is that the events seem to mark a change in long-term sedimentation rate. This is unexplained.

8.5.3 Mixing and "Stacking" of Deep-Sea Cores

On cutting a vertical face of a deep-sea box core, one realizes that the record is severely disturbed, on a scale of centimeters. Given a sedimentation rate between 1 and 2 cm per 1000 years, a resolution of better than a 1000 years would seem impossible to obtain directly from any one record. Possibly, resolution can be improved by *stacking* several records, thus diminishing the weight of aberrant values introduced by bioturbation. However, a limit for resolution of about 1000 years will persist. "Stacking" averages the input data. Averaging degrades resolution, although it can improve overall results by reducing extremes that may be spurious.

8.6 The Deep Biosphere and Methane

The exciting discovery of hot vent microbes feeding hot vent tube worms (Fig. 1.9) may have somewhat eclipsed the equally interesting and significant discovery, by deep-ocean drilling, of abundant microbial life forms in the sediments of the deep sea, especially in those accumulating on continental slope and rise. Even at great depths within the sediment (hundreds of meters), there are signs of biological activity, albeit decreasing markedly with depth in the sediment (Fig. 1.11). At great depth within sediments, the effects from a rise in temperature become noticeable as one approaches deeper layers of the crust. It has been suggested that this rise in temperature (typically between 20 and 30 °C for each additional kilometer) may be ultimately the limiting factor for the distribution of life at depth within sediments. Interestingly, archaea make up a substantial portion of microbes at great depths, more than one half of the mass in places. Presumably, their relative abundance is linked to the extreme conditions prevalent in this environment (high temperature, low food supply). In addition, it is assumed, archaea are responsible for the production of methane from organic matter within sediments. If so, this would mean that much or all of the "natural gas" below the seafloor is really an indicator of microbial activity there, rather than indicating supply from the crust or mantle.

A host of questions, largely unanswered as yet, arise in the context of prokaryotic microbes living deep within marine sediments. How long do these microbes live? What do they live on? How did they get to where they are? What are their closest relatives on the surface, and how do they differ from them genetically? Answers will come from microbial investigations, including studies from biogeochemistry focused on the carbon cycle.

Suggestions for Further Reading

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Imprint of Climate Zonation on Marine Sediments

9

9.1 Main Factors and Overall Zonation

9.1.1 The Input of Energy

The main factor in all processes on the surface of Earth, including geological ones involving sedimentation, is thought by many scientists to be the energy from the sun, which provides for *exogenic forcing*. However, major changes in geography (such as the distribution of land and sea and elevations) are wrought by *endogenic forcing*, that is, forcing that comes from the mantle and drives tectonics, including plate motions. In addition, there is astronomic forcing, which works through the Earth's orbital features, through impact, or by cosmic radiation. Milankovitch forcing is orbital and modifies solar forcing. Apparently it is responsible for much of the ice-age fluctuations. Exogenic (solar) forcing directly involves the *radiation balance* (which is part of the everyday climate system), that is, the equilibrium between incoming solar heat and outgoing radiation from the Earth, an equilibrium that keeps the planet at a given average temperature (Fig. 9.1). The latest (and most rapid) of changes are from human activities, presumably to be counted as "exogenic" since they work by enhancing solar input (i.e., blocking output of heat radiation to space). The changes are commonly (quite properly) categorized as "anthropogenic," emphasizing human causation.

About 340 W/m^2 of energy arrive from the sun (making up 100% in the illustration shown). Reflection adds up to roughly 100 W/m^2; the rest is absorbed by ground, ocean, and atmosphere, warming the planet. Balance is achieved by infrared radiation back to space. "Greenhouse" gases trap some of the heat (infrared radiation) preventing it from going out into space. The fact that our habitat is distinctly warmer, on average, than the average of the surface of the moon (with the same average distance from the sun as the Earth) is largely a result of the presence of "*greenhouse*" gases in the atmosphere, foremost water vapor and carbon dioxide. This is not some esoteric hypothesis but is based on measurable observation and has been known for more than a century.

As a consequence of its effect on the radiation balance and thus on temperature, the rapid rise of carbon dioxide in recent decades feeds concerns about global warming. The melting of ice is expected to produce a substantial rise in sea level in the future. Just how much and in which time frame is difficult to predict because the climate system is extremely complicated, with lots of feedback elements, both "positive" (favoring runaway) and "negative" (opposing change). Marine geologists can contribute, in principle, by reconstructing ele-

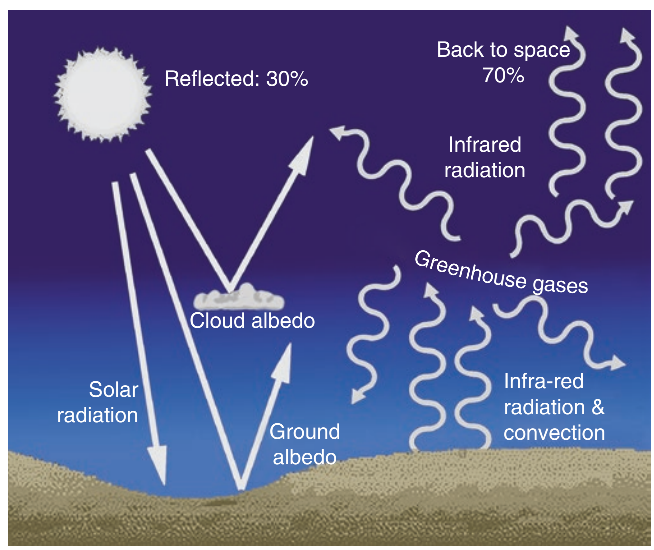

Fig. 9.1 Schematic NASA-inspired representation of the radiation balance, featuring solar input radiation, reflection from clouds and from the ground ("albedo"), and infrared radiation of a warm Earth into space. The presence of the atmosphere and the greenhouse gases within it ensures that the effective back radiation is from a level within the atmosphere rather than from the ground. As a result, the ground below that level is warmer than it would be without the atmosphere [After US NASA, in W.H.B., 2009. Ocean. Univ. Calif. Press, Berkeley. Color here added]

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_9

ments of climate history and especially the response of ice to long-term warming. What happened and what is measurable establishes what can happen, with great confidence. Historical examples from geology illustrate the size of natural variation; they do not necessarily facilitate prediction.

9.1.2 The Zones

The amount of energy received from the sun varies with latitude – it is relatively large at the equator and much less in polar areas. Thus, we have coral reefs in low latitudes and ice in high ones (Fig. 9.2). Apparently it was simply just cool in high latitudes for much of geologic history. Ice may not have been present during much of geologic history.

A coarsely latitudinal zonation of the oceans reflects the existing situation (Fig. 9.3). It employs the categories *tropical*, *subtropical*, *temperate*, and *polar*. Some zonation schemes distinguish, in addition, *subpolar*, which represents the transition between temperate and polar. The boundaries between the zones are somewhat arbitrary and do not follow latitudes very precisely (in Fig. 9.3, note the large distortion introduced by the Gulf Stream). Yet, zonation is sufficiently well represented by the distribution of planktonic organisms to be a useful concept both in oceanography and in marine geology.

The tropical zone has an excess of heat, much of which is exported to cooler regions by winds and currents. Seasonal fluctuations are relatively modest, with temperatures typically near 25 °C and sustained open ocean maxima close to 30 °C. Near the equator daily rainfall along with cloud cover and weak winds usually provides for excess precipitation over evaporation. Note that this zone is the largest identified in Fig. 9.3. (True areal relationships are obscured in many maps, owing to the widespread use of the Mercator projection in map-making.) The tropical zone and the subtropical

Fig. 9.2 Climatic extremes in today's marine realm. *Left*: Reef rim around island ("fringing reef") and lagoons. Huahine, Society Islands, tropical Pacific [Photo courtesy D. Eicher, Boulder, Colorado]. *Right*: Tabular iceberg, Antarctic Ocean [Photo courtesy T. Foster, S.I.O]

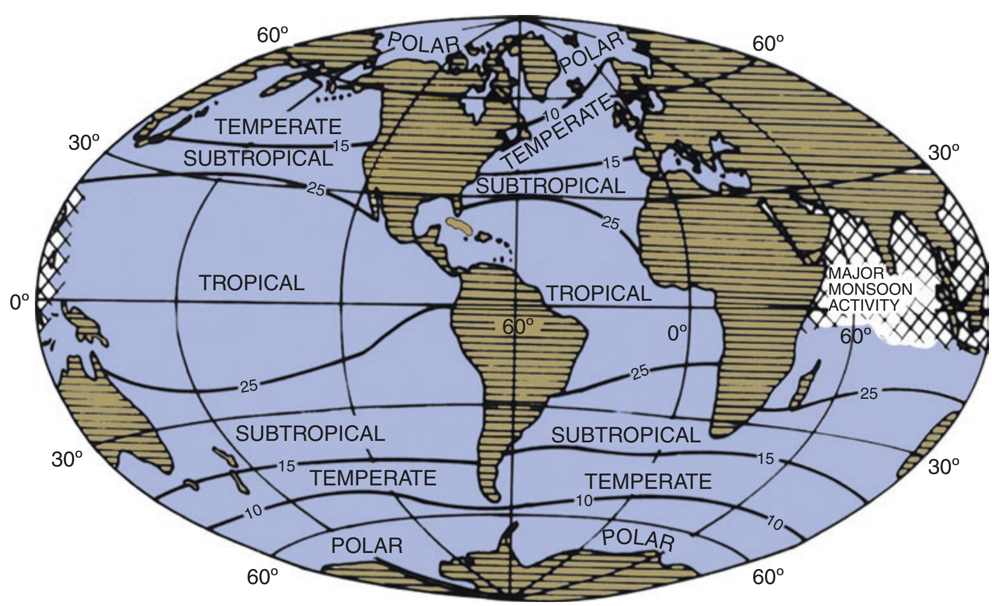

Fig. 9.3 Climatic zonation of the oceans. Zone boundaries tend to follow latitudes. Temperature, seasonality, and water budget vary across zones. Approximate temperatures of surface waters are given in °C. (Boundaries are seasonally variable.) "Monsoon" describes wind systems setup by continent-ocean temperature gradients. Such gradients vary seasonally

one together make up more than one half of the surface of the planet, by simple spherical geometry.

The subtropical zone includes the desert regions both in the sea (low-productivity gyre centers) and on land (drylands). Also, this zone includes the areas of major coastal upwelling (with cold water invading otherwise warm regions and re-enforcing drought conditions along the coast). Cloud cover is low, evaporation rates are high, and in the shelf seas, high salinities can be attained, therefore, favoring anti-estuarine circulation and carbonate deposition. This zone is invaded by the neighboring climate regimes depending on seasons. Annual temperature ranges can be quite high, as can the strength and direction of winds. Correspondingly, associated upwelling activity can vary greatly, seasonally and between years.

The temperate zone is strongly influenced by seasonal change. Rainfall generally exceeds evaporation and salinities are correspondingly reduced, especially in shelf seas. At the margin, estuarine circulation (deposition of organic matter and diatomaceous sediments) is common. As this zone represents the transition between the warm and the cold regions of the planet temperature gradients are strong and so are the zonal winds that they produce, with implications for the organization of currents and weather systems. The winds stir the water, bringing nutrients to the surface. Thus, whenever the sunlight becomes plentiful, such as early in summer, productivity is greatly enhanced, producing a corresponding dump of organic matter and shells on the seafloor (a dump resulting from what is called the *spring bloom*).

The polar areas, finally, while taking up relatively small regions, are of prime importance in harboring feedback drivers of climate change. The rim of the sea ice fixes the zero point of the latitudinal temperature gradient, a gradient that informs high-latitude winds, currents, and evaporation-precipitation patterns. Familiar patterns may be changing, making experience a questionable guide to predicting the new patterns that are replacing the old ones.

9.1.3 Temperature and Productivity

Temperature distribution and the marine carbon cycle (that is, productivity and associated sedimentation) are chief agents of control of climate. Not surprisingly, production is high wherever both nutrients and light are plentiful. In low latitudes, this is commonly the case in coastal settings, especially in the eastern part of ocean basins (Fig. 9.4). The influence of temperature on productivity, by itself, apparently is very modest. However, commonly temperature distribution is linked to nutrient supply, with cold water having a relatively high nutrient content (Chaps. 7 and 8) and a cold planet favoring the development of strong winds, as pointed

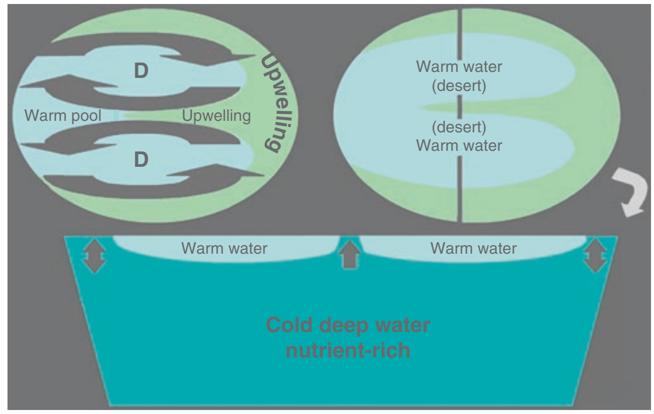

Fig. 9.4 Schematic representation of the zonal elements of the oceanic climate system pertaining to productivity and biogenous sedimentation in the sea. Note the controlling importance of wind belts in generating great gyres, chiefly the westward-moving trades and the eastward-moving westerlies in high latitudes. Also note the role of the thermocline separating the warm water from the deep nutrient-rich water [W.H.B., 2009. Ocean. U. C. Press, Berkeley, modified. Also, color here added. *D*, desert]

out by Gustaf Arrhenius of the Albatross Expedition in the 1950s. The depletion of tropical and subtropical surface waters with nutrients in a stratified ocean (nitrate, phosphate, silicate, iron) is a crucially important phenomenon. As mentioned (Fig. 7.9), the vertical density gradient (light warm water on top of cold heavy water) hinders upward diffusion of deep waters that are rich in recycled nutrients (Fig. 9.4). Generally, therefore, production is relatively low except in areas of upwelling. Upwelling and vertical mixing are at a minimum in the centers of the great central gyres that dominate the distribution patterns. Thus, we find the great deserts of the sea in the northern and southern gyre centers of the corresponding subtropical zones (Fig. 7.3). In high latitudes light is seasonally absolutely limiting to growth.

Temperature imprints itself on sediment patterns in several ways, either directly (through coral reefs, e.g., or through sediments left by ice) or indirectly (through correlation with productivity). Both direct and indirect imprinting play a role in the production of shell remains, including those of the sand-sized planktonic foraminifers (Fig. 9.5).

The relationship between temperature and productivity is rather subtle. To grow, algae in the plankton need to stay in the sunlit zone, in the uppermost waters. To get nutrients, however, deep mixing has to occur. Such mixing, of course, also moves photosynthesizing organisms into the dark, below the sunlit zone. At any one location, then, a succession of nutrient delivery and stable stratification provides the most favorable conditions for growth and hence for biological sediment production.

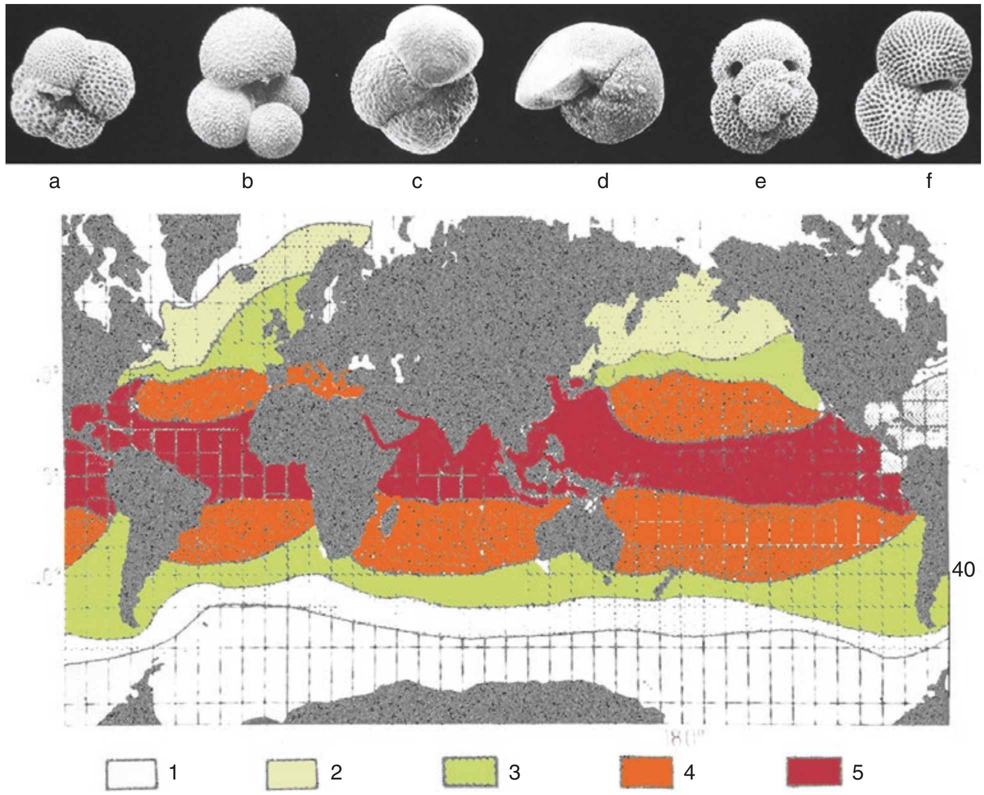

Fig. 9.5 Major distributional zonation of planktic foraminifers, from polar (*1*) to tropical (*5*). Dominant foraminifers (cold to warm zones): (*a*) N. pachyderma (polar, left coiling; subpolar, right coiling), (*b*) *G. bulloides*, (*c*) *G. inflata*, (*d*) *G. truncatulinoides*, (*e*) *G. ruber*, and (*f*) *G. sacculifer*. [Map after A.W.H. Bé and D.S. Tolderlund, 1971, in B.M. Funnell and W.R. Riedel, eds., The Micropalaeontology of Oceans. Cambridge Univ. Press, Cambridge, modified. Color here added. *1*, *2* ,*3*, *4*, *5*, climate zonation from polar to tropical. Examples of fossils (SEM photos) courtesy of U. Pflaumann, Kiel]

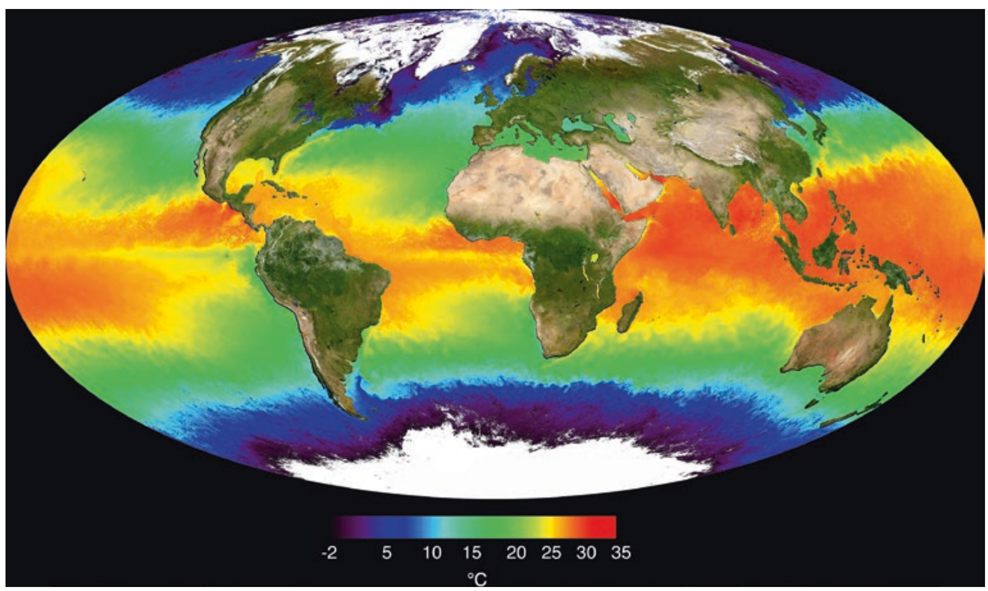

Fig. 9.6 Global temperature distributions according to a satellite-based NASA compilation by Jacques Descloitres of the MODIS Response Team. Note the large warmwater anomaly in the North Atlantic and the cold anomaly along the world's equator in Atlantic and Pacific

9.1.4 Large Anomalies

Areas of large anomalies hold special interest because they are likely to be especially sensitive to climate change. The very fact that there is an anomaly suggests the presence of uncommon conditions. Cold anomalies within the subtropics and along the equator denote regions of upwelling, that is, of high production. Production fluctuates in response to climate variation. One indicator of such sensitivity is strong variability of environmental conditions, commonly with a strong link to the wind field. Both cold anomaly and high variability are evident in all the major upwelling areas – off Peru, off California, off northwestern Africa, off Namibia, and also in the Arabian Sea (here with a strong role for monsoonal winds). On a global temperature map (Fig. 9.6), such regions can be recognized by their anomalously low temperature for the latitude where they occur.

The largest of all temperature anomalies is not a cold anomaly, however, but a warm one. It resides in the northern North Atlantic. Warm surface waters invade the region thanks to the Gulf Stream extension. The Gulf Stream water brings enormous amounts of heat northward, as warmish water. Some of the northward moving water, after cooling, sinks and makes North Atlantic Deep Water (NADW). The process is responsible for much of the formation of the deep water of the present ocean. Deep-water distribution patterns, in turn, greatly influence the distribution of calcareous ooze on the deep seafloor, controlling it via production of preservation patterns, for example, by helping to determine the elevation of the "CCD" (whose origin is discussed in the next chapter). The link between Gulf Stream and the distribution of calcareous oozes nicely illustrates some of the less obvious complications involved in tying wind-driven circulation to deep-sea sedimentation.

9.1.5 Heat Transfer and NADW Production

At present, heat from surface waters in the Indian Ocean may end up in the northern North Atlantic in an enormous global heat transfer proposed decades ago by the Woods Hole oceanographer Henry Stommel (1920–1992). In places in the Atlantic, the transfer is measured in "petawatts," that is, 10^{15} watts. A commonly used name for a thousand billion is "trillion" (10^{12}). The number 10^{15} denotes a thousand trillion, that is, a million billion (a 10 with 15 zeros).

The global transfer of heat is a chief corollary in the "conveyor belt" hypothesis of W.S. Broecker and G.H. Denton, a hypothesis that deals with the possible origin of and likely effects of a major cold spell (the "Younger Dryas") during the last deglaciation. Thus, it deals with the role of oceanic heat transfer in climate change. In their hypothesis major meltwater influx into the northern North Atlantic interfered with NADW production, as seems likely since freshwater floats on top of saltwater, and production of deep water implies sinking of surface waters, which is interfered with by the melting of ice supplying freshwater to the surface.

It is the production of North Atlantic Deep Water that is responsible for the transfer of heat into the northern Atlantic. This production is linked to the Gulf Stream, the major conveyor of heat into the Arctic realm (Fig. 5.14). This aspect of heat transfer was well appreciated even back in the nineteenth century, foremost by the Scottish geologist James Croll (1821–1890) who wrote intelligently about climate and the ice ages and about the heat transferred by ocean circulation. In modern times the Lamont oceanographers Arnold Gordon and Wallace Broecker have written extensively and memorably on global patterns of heat transport by circulation. There is concern (recently forcefully expressed by S. Rahmstorf, Potsdam) that global warming is interfering with the traditional heat transport patterns.

9.1.6 The Great Oscillations

The large amounts of heat imported into the northern Atlantic provide for a considerable anomaly at high latitudes (Fig. 9.7). The heat, among other things, feeds the Icelandic Low, creating a feedback system that increases variability of climate in the northern Atlantic. (The variability is linked to

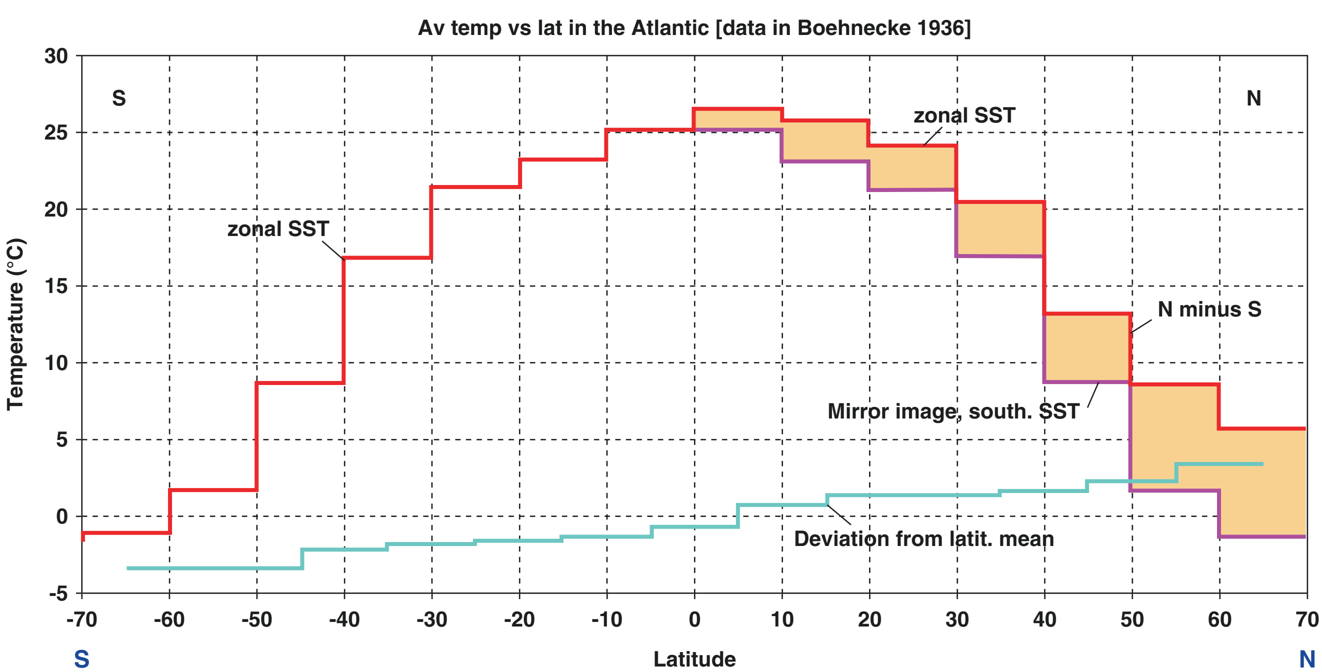

Fig. 9.7 Temperatures of surface waters in the Atlantic (*heavy red line*) and the amount by which the northern ones exceed the southern ones by latitude (*orange fields*) [Data from G. Boehnecke, 1936, cited in Sverdrup et al., 1942]

the North Atlantic Oscillation and thus to rainfall in and around the coast of Norway.) The observed phenomena agree with a suspicion that areas of temperature anomalies are prone to oscillation (presumably from lagged negative feedback). A likely source of variation is wind strength and direction.

In the central Pacific, the El Niño Southern Oscillation (ENSO) is the dominant source of interannual climate variation. The oscillation has global significance. Rather than being linked to cross-latitudinal heat transfer, it is linked to east-west transfers generated by variation in the strength of trade winds. The trades, blowing from east to west in the tropics after leaving the shores, pile up warm water in the tropical western Pacific (the great tropical "warm pool"). In marine geology and marine ecology, the warm pool is the site for maximum reef growth and diversity of coral and reefal fishes. The warm pool grows whenever the trades are strong but shrinks when they are weak. When the trades fail, the warm water they piled up moves eastward. The most likely time for this to happen is toward the end of the year, hence "El Niño," meaning "Christ Child."

Peruvian fishermen and their families suffer during El Niño, since upwelling and fish production are greatly decreased in the equatorial Pacific during El Niño events. Some years are worse than others in this regard, and some are outright awful. Concern has been expressed regarding the possibility of an increase of El Niño conditions in response to anthropogenic global warming. The ENSO dynamics are poorly understood, however. Information on a long time scale is difficult to obtain from the marine record since sediments commonly have low resolution owing to bioturbation, as mentioned (Figs. 8.13, 8.14, 8.15, and 8.16). The step in moving scientific insights pertaining to oceanography from a geological thousand-year scale to more of a human scale of decades is enormous. The history of coral growth may provide applicable evidence from marine geology (Fig. 8.4), although the possibility of coral growth shutdown or slowing during "bad" conditions (i.e., biased reporting) may be a problem.

9.1.7 High-Resolution Records in the Marine Realm

Obviously, being linked to the winds, the great oscillations – ENSO, NAO, and others – are influenced by many factors. In some areas they are clearly influenced by monsoonal changes. On the continents away from the sea, seasonal contrasts are exaggerated compared with temperature ranges offshore (hence the term *continental climate*). Around the large and high land masses of East Asia, seasonal contrast affects the entire northern Indian Ocean (hachured area in Fig. 9.3). During winter, cold air masses flow from the Asian highlands to the sea. The falling continental air brings drought, and the wind causes upwelling in the Arabian Sea. During summer the circulation is reversed: the southern monsoon brings rain from the ocean to northern India. If it fails to come, there is a serious shortage of food.

Coral skeletons, certain sponge deposits, and shell growth rings are one possible source of detailed information on marine climate oscillations, comparable to tree rings on land. Varved sediments (lacking bioturbation, a lack usually caused by lack of oxygen for the organisms capable of moving sediment) are another. They can be quite informative (Fig. 9.8). At present, such sediments are quite rare, though.

In the varved sediments of the basin off Santa Barbara in southern California, basin-wide ENSO signals are seen in the deposition of microfossils indicating warm water and low productivity. In the recent past, the link between El Niño conditions and low-productivity offshore California has been quite reliable and strong. The situation may have shifted, though, in recent decades, so that the experience of current

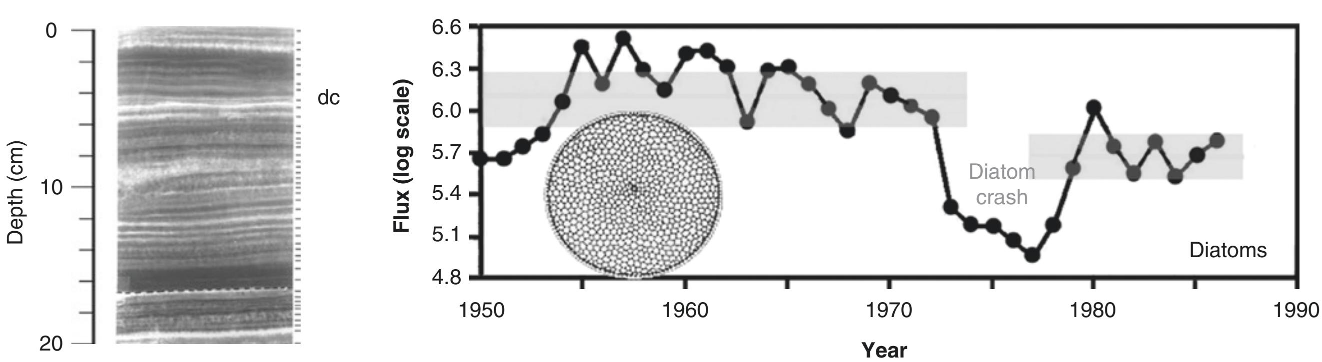

Fig. 9.8 The diatom crash of the 1970s, as seen in the sediments of Santa Barbara Basin. *Left*: X-ray radiograph of the sediment (Courtesy of Arndt Schimmelmann, then S.I.O.). [Abundance of diatoms: Lange et al. 1990, Clim. Change 16: 319, redrawn, diatom example added.] *dc* position of diatom crash in the sediment. Note log scale for abundance (base 10, *y*-axis; *tick marks*, factor of two); *shaded regions*: typical flux (per year, $\log_{10}$ scale)

observers may be losing relevancy. In many attempts at describing change, present conditions are compared with the average of several past decades. The method minimizes apparent change, since the "natural background" shifts along with the target of study. This type of shift is well known to many marine ecologists, who refer to the phenomenon as "baseline shift."

9.2 Paleotemperature and Climate Zonation

9.2.1 Reconstruction of Temperature History

Biogenous sediments can be excellent indicators of climate change. Many organisms have narrow ranges of tolerance in temperature and productivity, and the presence and absence of their remains readily yield clues for climate reconstruction. The potential for quantitative reconstruction using statistics on deep-sea assemblages of planktonic foraminifers was recognized in the 1960s by the geologist John Imbrie and his associate Nilva Kipp (both then at Brown University, Rhode Island). Earlier, material from short cores from the central Atlantic recovered during the *Meteor* Expedition (1925–1927) had been used by the German geologist Wolfgang Schott (1905–1989) to identify and document the great change from glacial to postglacial conditions in the ocean, based on planktonic foraminifers.

Planktonic foraminifers live mainly in the uppermost 150 m of the water column, as shown by depth-specific towing of nets that can be opened and closed. There are about 20 abundant species and as many rare ones. Each climatic zone has one or more characteristic species and a typical abundance pattern of species within it. Some of the more common temperature-sensitive species are shown in the SEM photos of Fig. 9.5. The general agreement of the climate zonation with the faunal zonation allows for the reconstruction of one from the other, using mathematical statistics, including *calibration*, that is, a matching of physical conditions to elements of the fossil assemblages.

One important problem always arises, however: the environmental variable being reconstructed may not be entirely in control of the patterns observed. Other controlling factors may be at work also, and their correlations with the targeted factor (and with the main factor of control) may change through time. Also, there are the problems of admixture of older material and the presence of material of poor preservation, that is, problems that haunt all paleoecologic reconstruction. Commonly, when preservation problems are indicated, the more delicate fossils have been removed, and the thick-walled ones have been concentrated in the remaining assemblage.

One cannot assume, of course, that the abundance patterns of resistant forms are just like those of the more complete set including delicate fossils. In addition, the preferences of fossils found in the sediments may be unlike those of the closely related fossils that were calibrated (and this may not be obvious), so that the information match may suffer both from preservation and adaptation. One possible answer to such problems is to employ a number of different sets of fossils or methods of reconstruction.

For example, in addition to reconstruction of past temperatures from the abundances and the chemistry of planktonic foraminifers (or related fossils made of different materials such as silica), one can use the abundance patterns and the chemistry of the most abundant and minute calcareous fossils on the seafloor, the coccoliths (Figs. 4.7 and 4.11). Also available are the oxygen isotopic composition of fossils (cold water favors precipitation of the heavier isotope), the ratios of homologous elements, and the growth rings. In other words, any fossil abundance or sediment property that is influenced by temperature also can deliver clues to past temperatures. However, the uncertainties (up to 2 °C in the reconstruction of the temperatures of the last ice age and presumably even more for the more distant past) cannot be made to vanish. Some remnant uncertainty always survives. For once, there are the seasons, which introduce a range even into present conditions.

To exemplify using the organic chemistry of the plankton flora, the late geochemist Peter Müller (Kiel and Bremen) and his associates have made temperature estimates that are quite precise as far as the calibration. However, as Müller has stated in private conversation, the quality of paleotemperature estimates based on chemistry very much depends on the correct reconstruction of the chemical environment in the distant geologic past, a tall order indeed.

9.2.2 Climatic Transgression

The precision of temperature estimates is not invariably the most important issue in marine paleoecology. Of equal or greater significance is the change of patterns, which contains information about changes of currents and other items of interest in the context of rapid climate change. Conditions in climate during the last glacial maximum in the North Atlantic, on approaching the end of the Pleistocene, provide a striking example (Fig. 9.9). The information is retrieved from planktonic foraminifer abundance patterns from numerous sediment cores distributed throughout the region. The reconstruction of climate patterns via Imbrie-Kipp-type statistical temperature estimates was done by the CLIMAP group. Conditions are seen to have differed greatly from the present during the last ice-age maximum some 20,000 years ago, no matter the precise temperature assignments.

Today the polar front (isotherm slightly closer to the pole than the 0 °C surface isotherm) is just south of Greenland. During the glacial maximum, the front ran from New York to the Iberian Peninsula (Fig. 9.9). At that time Norway and even England were largely cut off from the warming influence of the Gulf Stream extensions, with dramatic climatologic consequences (e.g., Scotland covered by ice; sea ice over large parts of the northern North Atlantic, with corresponding sediment patterns). The motion of the polar front in the North Atlantic, as reflected in the record in deep-sea cores, is a prime example for climatic transgression, as reconstructed in some detail by the marine geologists Bill Ruddiman and Andy McIntyre then at the Lamont-Doherty Geological Observatory of Columbia University, New York, commonly referred to as "Lamont."

9.3 Coral Reefs: Markers of Tropical Climate

9.3.1 Global Distribution

In geology, the tropical climate zone is commonly defined by coral reef deposits with their unique fauna and flora but with organisms slightly different from those of today's reefs, even just a million years ago.

In the present ocean (and presumably throughout the Neogene), the distribution of coral reefs closely follows the presence of 20 °C water in the coldest month of the year (Fig. 9.10). Where temperatures drop below this limit, tropical corals do not grow. For this and other reasons, corals and their associates avoid regions of coastal upwelling. Presumably, even though averages of oceanographic maps may not show it, the temperature can drop below the coral limit at least in pulses here. A high nutrient content in the

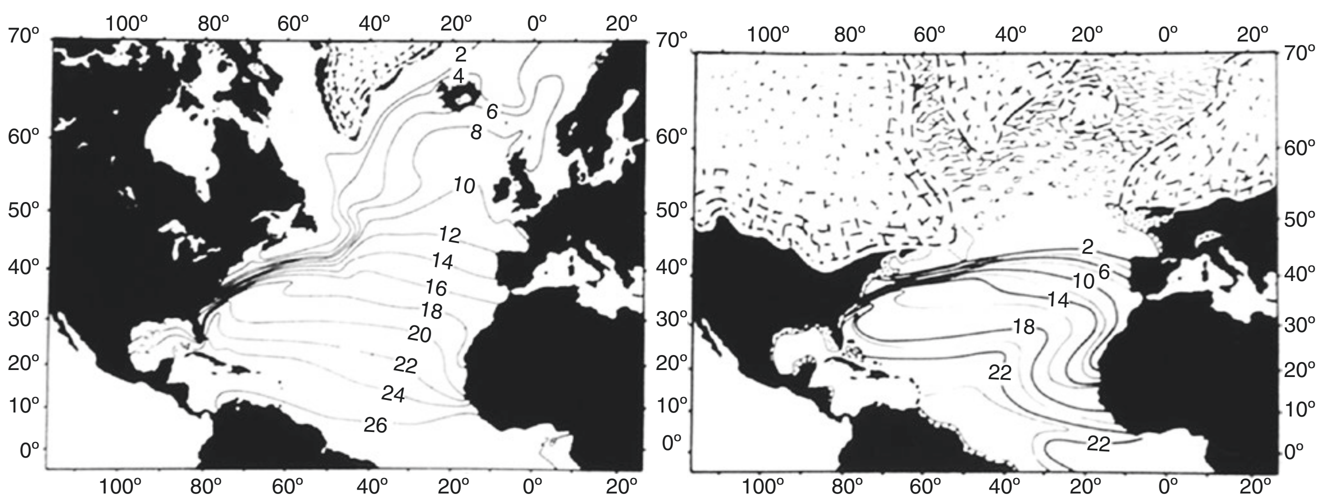

Fig. 9.9 Reconstruction of climate transgression in the North Atlantic: comparison of last glacial maximum conditions (*right*) with the present situation (*left*). For reconstruction by the CLIMAP group using abundance patterns of planktonic foraminifers (see 1976 Science 191:1131; and 1976 Geol. Soc. Am. 145:1–464) [Graphs shown are from E.S., 1975. Naturwissenschaften 62:321]

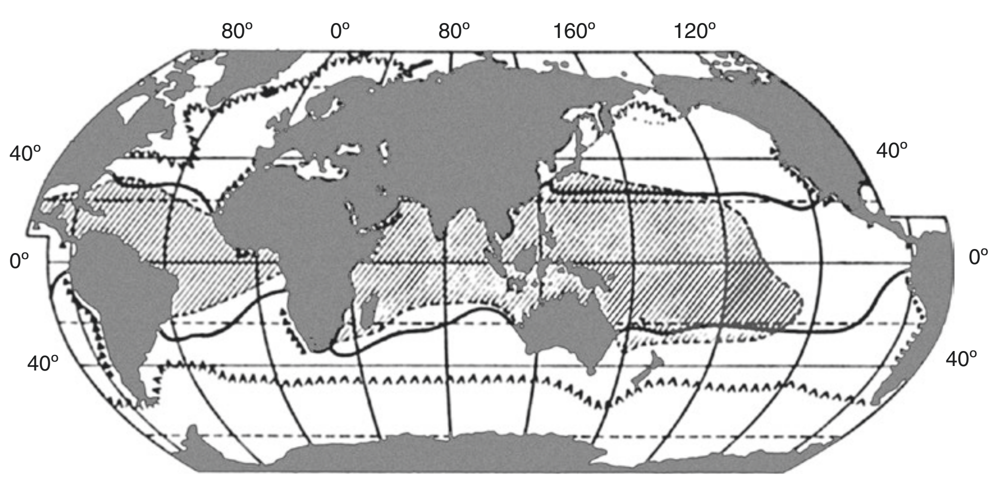

Fig. 9.10 Distribution of tropical coral reefs according to the New York paleontologist J.W. Wells (1907–1995; *hachured areas*). *Heavy line*: coldest month has 20 °C. Areas of coastal upwelling (*filled triangles*) inhibit coral growth. Maximum extent of icebergs: v's (northern hemisphere) and inverted v's [From E.S. 1964. In: R. Brinkmann (ed.) Lehrbuch der allgemeinen Geologie. F. Enke, Stuttgart. See the Schweizerbart web site for Enke]

water is inimical to coral growth also. A pulse of high nutrient supply is commonly accompanied by a distinct drop in temperature. Corals are desert organisms living in symbiosis with minute algae. Just as we would not expect healthy cactus resulting from heavy watering, we should not expect corals in high-production areas. Owing to the fact that reef building commonly involves light-using organisms, most reefs are in shallow water on shelves. For ancient reefs, therefore, continental drift has to be considered and plate tectonics.

9.3.2 On the Nature of Coral Reefs

The greatest diversity of stony corals within tropical coral reefs is found and around the western tropical Pacific warm pool, on the Indonesian shelf and off New Guinea. Northeastern Australia with its Great Barrier Reef comes in as a close second to the leading sites as far as diversity. Remarkably, a large number of the dominant species apparently are fast-growing forms somewhat similar to the common Caribbean species *Acropora cervicornis*, that is, they are bushy, with thin stalks, and they grow fast in very shallow water, as pointed out by Harvard's R.A. Daly early in the last century. This abundance of bush-like coral with finger-like extremities may go against the grain of Darwinian selection for hard-to-eat coral (that is, flat and solid). There is a great abundance of hungry parrot fish grinding up accessible coral skeletons like rabbits eating carrots. However, the possibility that coral growth rates were extremely important during deglaciation events in the ice ages must be kept in mind.

We may be looking at adaptations to conditions ruling the environment thousands of years ago and long gone. The message, presumably, is that evolution and attendant diversity distributions reflect long-term (geological) conditions and their variations in much of the fossil record. Observable evolution commonly happens on a time scale much longer than the human one of decades and centuries, causing incredulity regarding evolution in many people (contrary claims about evolutionary rates in certain insects exist, e.g., the observation that certain mosquitoes are resistant to DDT. Short-term claims also were implied for domesticated animals by Darwin).

That evolutionary history is likely to be extremely complicated is implied in the presence of a great number of different types of coral on reefs (Fig. 9.11). Despite this focus on coral in coral reefs, perhaps the most important organisms are microbes, many of which are photosynthesizing symbionts. The structures known as "coral reefs" are in fact built by a great number of carbonate-secreting organisms. Corals do not necessarily deliver the bulk of the material. Up to 10 kg of calcium carbonate can be produced annually for each square meter, much of it made by algae or mollusks, with rates of up building of up to 1 cm per year. By comparison, calcareous ooze takes 500 times longer to grow upward by that amount.

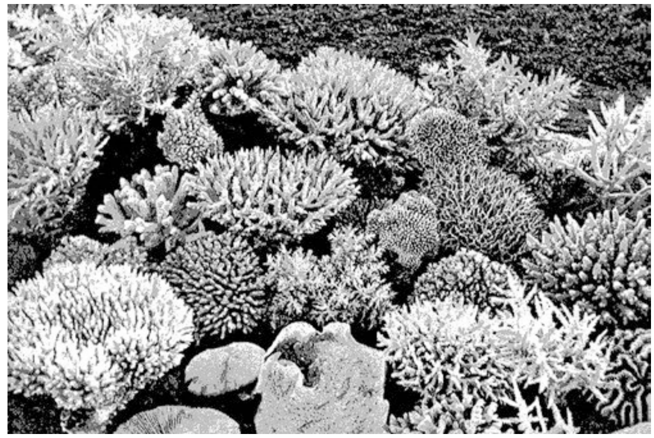

Fig. 9.11 Tropical stony corals from the Great Barrier Reef, western Pacific, documenting the large diversity in this realm (ca. 90% of maximum). The maximum (ca. 400 species) is associated with the western warm pool [Photo from R. A. Daly, 1934. The Changing World of the Ice Age. Yale Univ. Press, New Haven]

About one half of a typical reef consists of solid matter; the rest is empty space (that is, pore space filled with water). Much of a reef is rubble, or else of algal origin or sand or silt made by parrot fish and other organisms (including sponges). We find the material along certain sinking continental margins as well as rimming or topping sunken volcanoes. When considering these structures, we have to differentiate between the *framework* of the reef (i.e., the supporting structure) and the *fill,* much as in building construction. Fill can provide up to nine tenths of a reef mass in places. Calcareous precipitates, notably coral skeletons, are broken up in various ways. Much of the solid material is crushed by fishes and crustaceans.

Many skeletons of reef organisms are held together by organic material. When the organic matter decays, such skeletons disintegrate and make lime mud (*micrite*). Also, waves and tidal currents work on the relatively soft calcareous matter, grinding it up into micritic carbonate (carbonate made of minute solid particles). The "fill" thus generated largely ends up in the interstices within the reef structure. Some of the debris trickles down the sides of reefs and becomes part of the reef flanks. Dissolution of aragonite and recrystallization and cementation of the debris start very soon after deposition. The relevant processes are influenced by microbial activity as well as by interaction with rainwater. Already a slight drop in sea level would expose large areas of carbonate platforms and reef flats to rainwater, as emphasized by the US geologists E.G. Purdy and E.L. Winterer. During the ice ages, of course, sea level varied through hundreds of feet, which meant that "phreatic processes" (karstification and cementation involving rainwater) were extremely important. R. A Daly early on realized the great importance of sea-level variation to reef growth.

There are other features made of skeletal carbonates that are referred to as "reefs" besides the familiar tropical reefs. We find them even in Arctic waters, and they are made of red algae or of mollusks or of ancient carbonate rock. Some Norwegian fjords have cold-water coral reefs made of species of the genus *Lophelia*. Representatives of this genus of stony coral are abundant in deep cold waters all over the North Atlantic.

9.3.3 On the Origin of Atolls

Atolls, that is, palm-studded rings of coral and shell debris enclosing a lagoon and sheltering it from the pounding surf are common landmark islands in the South Pacific. Charles Darwin made an attempt to explain their origin more than 150 years ago. He was followed (and emulated) by his admirer, the US geologist, mineralogist, and coral expert James Dwight Dana (1813–1895). Darwin's hypothesis of a sinking volcanic edifice being adorned by coral reef growth was long the dominant explanation for atolls, reproduced routinely in texts of marine geology (including earlier editions of this one). It was said to be supported by finding thick carbonate rubble and coral on top of volcanic material upon drilling into atoll islands in the middle of the twentieth century.

However, the sinking of the base of the reef is exceedingly slow and may well be irrelevant to coral growth. Instead, the point made by the Canadian-American Harvard geologist R. A. Daly (1871–1957) that the atolls owe much of their appearance to the fluctuations in sea level during the northern ice ages surely must be taken very seriously, perhaps in combination with the Purdy-Winterer scheme of karst development in atoll floors whenever such floors are well above sea level during glacial times, collecting rainwater.

9.3.4 The Great Barrier Reef

The 2000-km-long *Great Barrier Reef*, lying 50–250 km off northeastern Australia, is the most impressive reef structure in the world today and demonstrates how biological sedimentation can increase the size of continents. The reef grows on subsiding crust. It is most massive in the north and in the tropics (around 10°S), reaching a thickness of about 1500 m. In the south (in the subtropics, at 24°S), it is only one tenth as thick. Many geologists have linked the reef's position and thickness variation with the northward migration of the Australian continent. However, the bulk of the reef off Queensland apparently is less than a million years old, based on drilling results. If this is so, much of the cause for reef buildup has to do with climate change in the ice ages rather than with plate tectonics. Also, the observed differences in reef thickness may be owing to differences in growth rate of the species involved in reef building, with those closer to the equator growing faster. There are plenty of species to choose from. Different species inhabit different environments, and we may be confident that coral reef communities changed in response to changes in temperature regimes and sea-level position throughout the ice ages and that reef building was affected. We may be equally confident, based on numerous studies, that existing reefs respond unfavorably to the ongoing ocean warming and acidification. In fact, concern has been expressed by coral experts that the GBR is in trouble owing to global warming and various other changes ascribed to human activities, including fishing, some of it illegal.

9.3.5 Reefs Under Stress

Coral reefs are changing before our eyes, with undesirable consequences. They apparently respond to temperature changes, to nutrient pollution, to carbon dioxide increase (pH drop, i.e., acidification), and (according to the Smithsonian reef expert Jeremy Jackson) to removal of certain key reef organisms that control the outcome of competition between coral and large algae. Similar contests involve other reef members, and the outcome (one assumes) favors the "weedy" species (i.e., the ones more tolerant to a change in conditions). Temperature change (that is, warming of surface waters) is widely identified as a stressor producing "*bleaching*" in coral. Bleaching results from a loss of (colored) symbionts. Possibly the coral colony itself sheds the symbionts, which absorb sunlight and thus contribute to local warming, or various coral species find that their symbionts are not sufficiently productive to warrant hosting (a possibility raised by the marine biologist Nancy Knowlton in private conversation) or the symbionts decide to leave, being disappointed in the performance of the host, or a combination of these various reasons applies. Alternatively, there are altogether different reasons, some linked to microbial diseases, which are perhaps more common when corals are stressed.

What we know is that if summer temperatures are uncommonly high for some time, bleaching sets in. According to certain research results, dense human populations add to the ambient stress level, which seems important for the coral response to unusual warming. Most corals tend to recover on cooling, but they may be more vulnerable to disease after having experienced bleaching. If the initial state is not entirely attained, recovery is partial only, and then stating a recovery time is but a futile exercise, involving some arbitrary impression of what looks like "recovery." The challenge is to define the prestress condition of the reef organism in question. This task is likely to be difficult.

Generally, much concern attaches to the fact that changes in the environment are too fast at present for organisms to adapt to in traditional ways. Thus, the fact that there were reefs in the distant past, when the planet was very warm, does not hold any particular promise on a rapidly warming planet with icy poles. The range of present-day natural conditions (including those of the ice ages) is what counts. It is what extant organisms are adapted to.

It is surely alarming that many knowledgeable marine ecologists consider that reef dynamics may be suffering greatly from the interaction of a great number of stressors and that the end result might be a cessation of delivery of reef ecosystem services, affecting coastal protection, fisheries, and tourism among other things.

9.4 Climate Indicators Other than Coral Reefs

9.4.1 Geochemical Climate Indicators

In addition to tropical carbonate reefs, there are a number of other indicators useful in delineating climate belts. Salt and dolomite deposits, for example, have been used to delineate climatic belts. Salt indicates excess evaporation and is characteristic for subtropical conditions in horse latitudes, therefore. Also, salt deposits presumably are indicators of shallow water – although the possibility of a deep-water origin has been emphasized in connection with the discovery of salt within latest Miocene sediments of the Mediterranean, during Leg 13 of the Deep-Sea Drilling Project. (The hypothesis of deep-water salt is widely ignored, however). As mentioned in the chapter that follows, deep-sea geologists commonly are faced with clay-size climate indicators washed in from land, some resulting from deep chemical weathering of soil. Such deposits, independent from their ultimate resting place, indicate climate conditions at the region where they came from, of course. This is very commonly true for nonbiological sediments, requiring much study to recognize the exceptions.

9.4.2 Oxygen and Carbon Isotopes

The isotopes of oxygen and carbon are among the most widely used chemical markers in marine sediments in the last several decades. The word "isotope" ("same place") commonly denotes involvement of an atom of a weight slightly different from that of the most common atom of a given element taking up a certain place in the periodic table. The arrangement of elements in the periodic table is based on their number of electrons and associated chemical properties. The differences in weight in the same element (oxygen has weights of 16, 17, and 18; carbon of 12, 13, and 14 (the latter being known as "radiocarbon") owe to differences in the number of neutrons in the nucleus, the number of protons being equal to that of the element-defining electrons). The existence of different atom weights accounts for the "isotopes."

The ratio of oxygen isotopes in a calcareous shell indicates the temperature of precipitation, as shown by the physical chemist Harold C. Urey (1893–1981) at the University of Chicago. For the determination of temperature from shells, one uses the abundance ratio between oxygen atoms of weight 18 to atoms of weight 16. Values of the measure are listed as deviation from an arbitrary standard (hence "delta") in "permil," that is, in tenths of a percent. A change of 1 °C in the sea produces a change in the delta value of roughly 0.2 permil. Deep-sea benthic foraminifers typically living at temperatures some 20° colder than those of warmwater planktonic species have delta values typically heavier by around 5 permil (0.5%) than those of tropical planktonic ones. Changes in ice mass and in salinity also are correlated with changes in the oxygen isotope values. This can pose a problem when attempting to extract only temperature information from the isotopic ratio (or only ice mass; Fig. 1.5).

The ratio in stable carbon isotopes ($^{13}C/^{12}C$) is linked to the carbon cycle in the sea and to the metabolic activity of the shell-secreting organisms. Fast-growing organisms and those hosting symbiotic photosynthesizing algae precipitate shells with delta values that may be very difficult to interpret. Other shells, however, may yield information about the isotopic ratio of carbon atoms in the seawater they grew in. This is the case for many planktonic foraminifers. Their isotopic carbon ratio has information about the intensity of "*biologic pumping*," especially in adult forms. The matter was elucidated by the Lamont geologist and chemical oceanographer Wallace S. Broecker. (See Chap. 7 on biological pumping.) "Radiocarbon" is the radioactive isotope among the carbon atoms. Its half-life is near 6000 years. As mentioned earlier, it is of questionable value for determining ages beyond seven half-lives owing to problems arising from contamination once the resident radiocarbon has reached a very low abundance level.

9.4.3 Physical Geologic Indicators

Shelf sediments across the latitudes, from tropics to high latitudes, reflect climate regimes in their physical properties

(Fig. 9.12). In high latitudes there is the influence of ice on the sediment delivered, and in temperate latitudes it is rivers and ice products from times past, that is, from glacial periods.

Many high-latitude shelves (e.g., off Siberia, off northern Alaska) were (or still are) covered seasonally by sea ice. Early in summer, when flooding sets in, river water may stay on top of the ice. When penetrating through cracks in the ice, the freshwater can dig deep potholes on the seafloor. Also, sea ice can pile up compressed by ocean currents, as so vividly described by Fridtjof Nansen (1861–1930), the great Arctic explorer from Norway. On the shelf, moving ice can produce deep and long ravines. Furrows produced by drift ice during the last glacial period have been described from Scottish and Norwegian shelves and are likely present elsewhere in high latitudes.

Among the more interesting high-latitude marine deposits is the *ice-rafted debris* or *IRD* carried to the site of deposition by ice. Icebergs are made by calving glaciers entering the sea, and the glaciers commonly are loaded with morainal material, at the base of the ice or in cracks (Fig. 9.13).

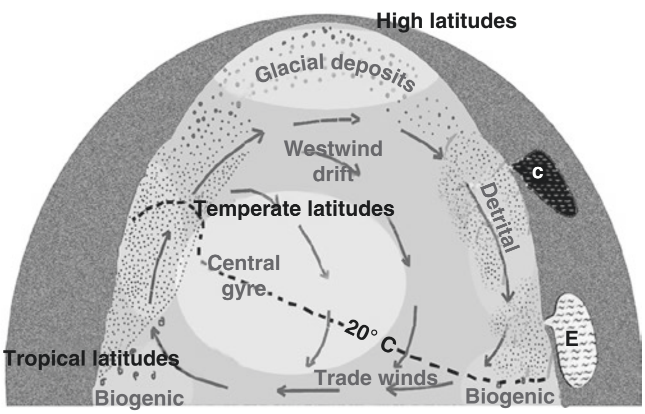

Fig. 9.12 Schematic of the distribution of shelf deposits in relation to climatic zonation in the northern half of an idealized ocean basin. Restricted shelf seas are marked as collectors of organic matter (*C*, for *carbon*) and evaporites (*E*), depending on climate zone. "Biogenous" material on shelves (mollusks, foraminifers, calcareous algae) occurs in all zones but is concentrated in tropical areas (here there are also stony coral) [Mainly after K.O. Emery, 1969, Sci. Am., modified]. Not to scale

9.5 Climatic Clues from Restricted Seas

9.5.1 Exchange Patterns of Marginal Basins with the Deep Ocean

When contrasting tropical reef and ice-carved shelf, we compare the warm and cold extremes in the present global ocean (i.e., the ocean of the interglacial ice ages of the last 600,000 years). There is quite generally yet another important contrast, the one between regions of excess precipitation and regions of excess evaporation (Fig. 9.14). On land, these conditions are known as "humid" and "arid." The ocean being water, the words "arid" and "humid" might seem to make little sense in describing the corresponding climatic belts in the sea. However, we retain the terms as convenient descriptors of the balance between evaporation and precipitation. Most of the precipitation, incidentally, is from vapor generated at sea.

Climate-generated differences in sedimentation are greatly amplified in restricted shelf seas, so that it is here we find potentially extremely interesting modern examples of making characteristic rock types found in the geologic record for the entire Phanerozoic.

The seas in the arid belt, where excess evaporation determines conditions, commonly have a typical exchange pattern with the open ocean, called *anti-estuarine* (Fig. 9.15). The opposite is the *estuarine* circulation. Prime examples for anti-estuarine circulation are the entire Mediterranean, the Persian Gulf, and the Red Sea. Since the Atlantic Ocean takes in surface waters from elsewhere and sends North Atlantic Deep Water south, it too is characterized by anti-estuarine circulation, especially in the North Atlantic, and it is here in the open ocean that we find calcareous sediments most widespread on the deep seafloor (as discussed in the chapter that follows).

Anti-estuarine circulation in a shelf basin is maintained by excess evaporation, which increases the salinity of incoming shallow water, so that it sinks and fills the basin, leaving it over the sill when reaching that level. The outgoing deep water takes the nutrients regenerated from fallen organic

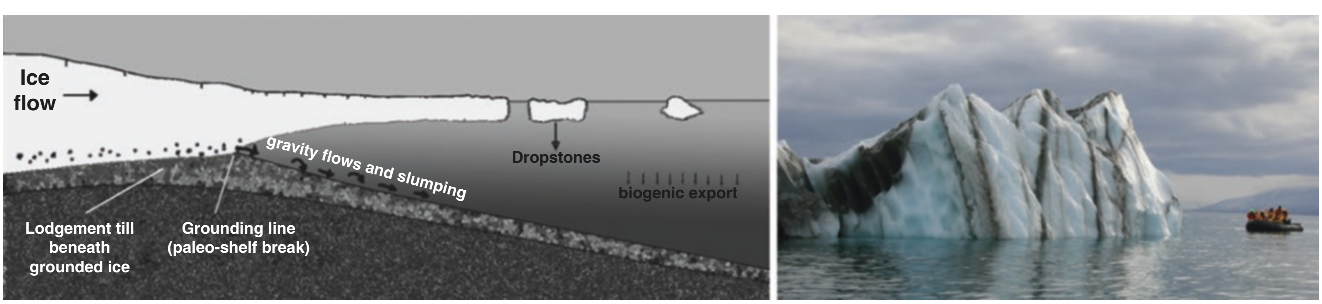

Fig. 9.13 Ice-rafted debris. *Left*: Sketch of origin of ice-rafted debris in high latitudes [After A.K. Cooper et al., 1991. (USGS); see Marine Geol. 102:180; modified]. *Right*: Arctic iceberg bearing sediment in a fjord in eastern Greenland. Rubber boat for scale [Photo courtesy K.J. Berger]

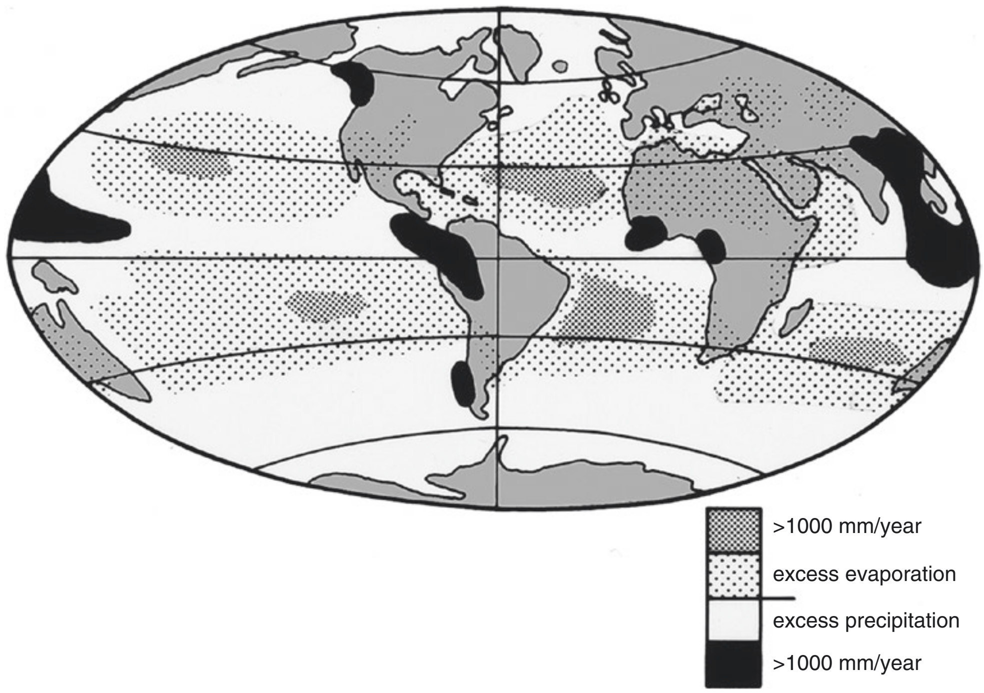

Fig. 9.14 Sketch of the major patterns of precipitation and evaporation on Earth. *Dark areas*: large regions of distinct excess precipitation. Continental areas marked *gray* [E.S., 1970, Geol. Rdsch.60:73; also see G. Dietrich and K. Kalle, 1957]

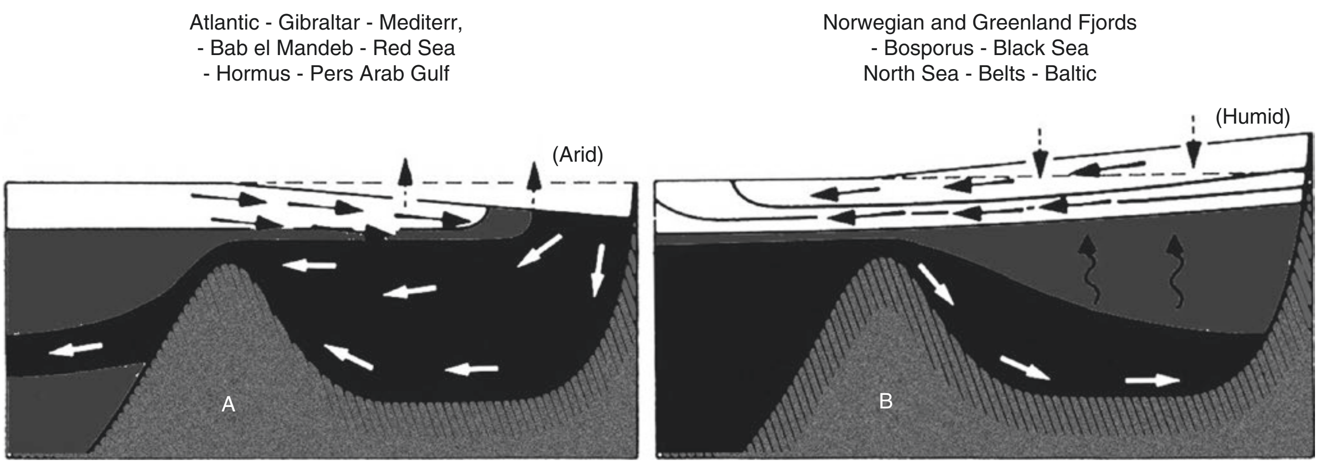

Fig. 9.15 Circulation in basins based on the arid-humid exchange paradigm [Graph after G. Dietrich and K. Kalle 1957, Borntraeger, Berlin, modified; names of examples added] Permission from Schweizerbart. For copyright-relevant information, see the Schweizerbart web site

particles with it, with the result that anti-estuarine basins typically have low nutrient content, low productivity, and a low supply of organic matter to the seafloor. Instead, they are carbonate traps.

Where rain and river influx exceed evaporation, the water level rises, and the freshened water on top of the water body forms a lid, being *lighter* than the open ocean water entering the basin at depth. The deep water rises in response to the outflow of brackish lid water (which runs downhill and out of the basin, taking deeper-lying seawater with it). The rising deeper water, rich in nutrients in comparison with the outflow, stimulates production. In essence, we have an upwelling situation. Examples for this type of circulation (deep water in, shallow water out) are the Black Sea, the Baltic, and the fjords of Alaska and Norway. This *fjord-like circulation* is common at the mouths of rivers and in other estuarine situations (such as, e.g., in Chesapeake Bay, one of the largest estuaries in the world). High precipitation is typical for subpolar to temperate latitudes and for the equatorial region.

9.5.2 The Baltic Sea as a Model for Humid Exchange

In the Baltic Sea, an estuarine basin in the northern rain belt, surface waters are less saline than the deep water. The high

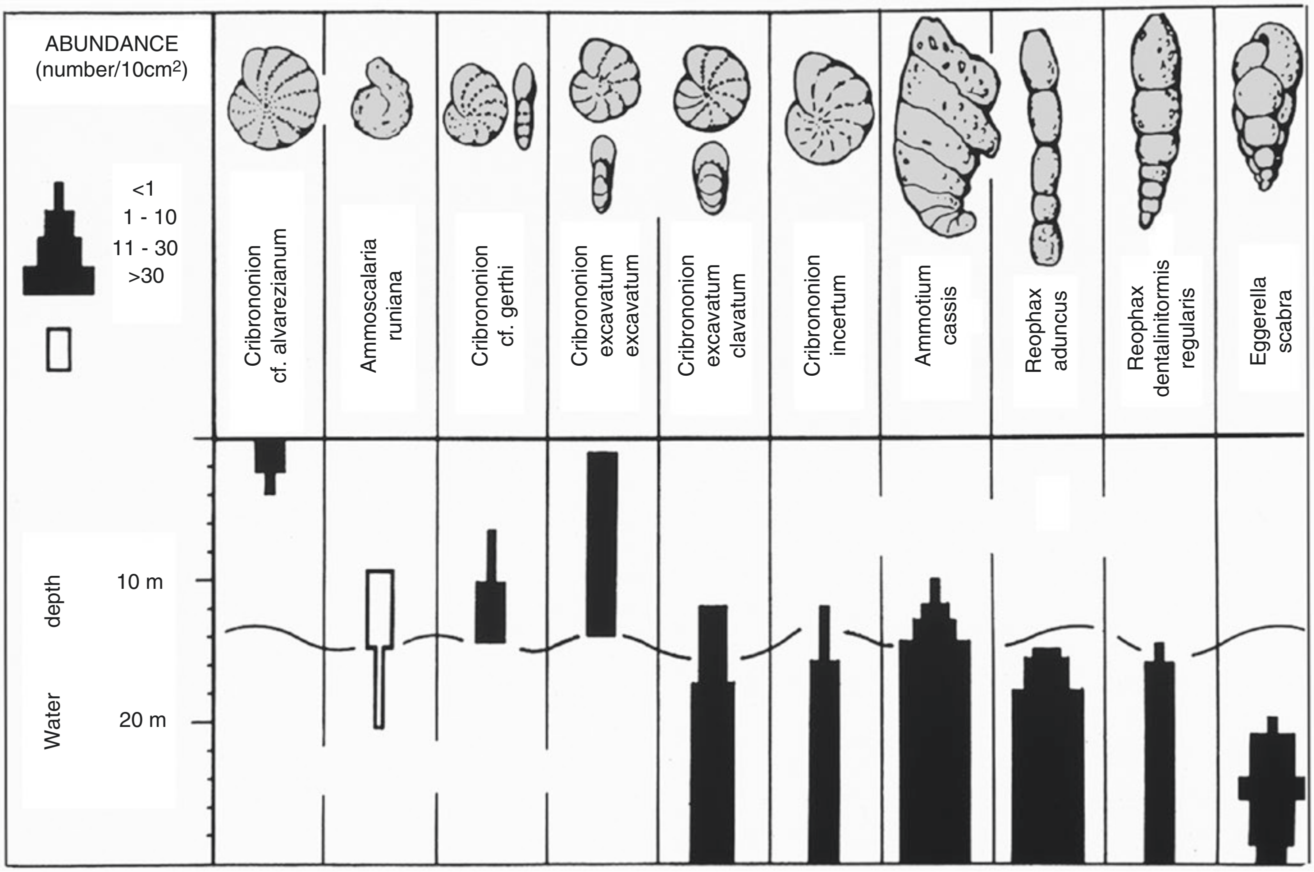

Fig. 9.16 Shallow- and deep-water benthic foraminifers in the western Baltic: coincidence with salinity change (wavy boundary). Width of *black symbols* denotes abundance [G.F. Lutze, 1965. Meyniana 15:75 (Kiel)]

supply of freshwater from rivers entering the Baltic makes the entire sea rather brackish, with salinities about one half of those in the Persian Gulf, a typical arid exchange basin. In the Baltic, the saline deep water enters the ice-carved depression from the North Sea through the straits between Denmark and Norway.

The high degree of stratification from the density contrasts owing to differences in salinity between surface waters and deep waters in the Baltic Sea is readily seen in the distribution of benthic foraminifers (Fig. 9.16). The stratification cuts down on vertical mixing, thus keeping oxygen out of deep water. In addition, oxygen is rapidly lost in the overall upwelling conditions that characterize the basin. Both factors contribute to oxygen deficiency, resulting in values of less than 10% oxygen saturation in many places. Wherever all oxygen is used up, the foul-smelling H_2S develops from microbial sulfate reduction. In the geologic record, the former presence of the gas is witnessed by minerals of iron sulfide, notably the common mineral pyrite ("fool's gold").

When oxygen is entirely absent, there is no macrofauna on the seafloor, and no burrowing occurs. At a concentration of less than 1 ml O_2 per liter of water, shelled organisms become rare. At such low concentrations, few metazoans remain: mainly worms (annelids, nematodes) and some crustaceans. Certain snails have been observed grazing on top of an anaerobic seafloor; the siphons extended into a water layer slightly above the seafloor and presumably with slightly more oxygen. Below 0.1 ml of oxygen gas per liter of water, only various bacteria and archaea survive and certain foraminifers.

Since nutrient supply is involved in producing oxygen deficiency (by the upwelling connection, via generation of plenty of organic matter) could human activities causing high production increase the occurrence of such deficiencies? The answer is "yes." Deforestation and all sorts of agricultural activities (including the use of fertilizer) increase nutrient input into nearshore basins. Sewage and industrial waste is yet another source. Thus, the spreading of *dead zones* at the continental margins and the intensification of *stagnation* in marginal basins such as the Baltic Sea are indeed favored by human activities. However, sediment cores from the central Baltic basin show that oxygen sporadically disappeared from the seafloor even well before human factors could play a role, resulting in finely *laminated*

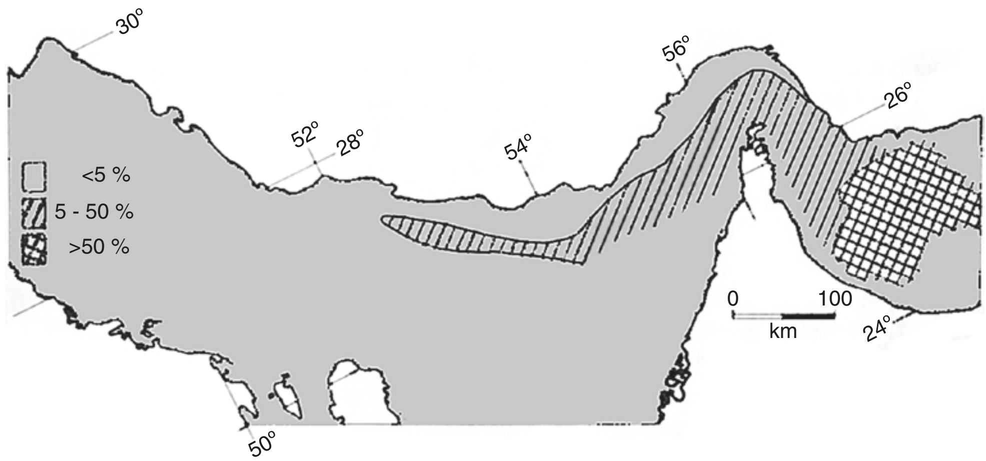

Fig. 9.17 Evidence for open ocean surface water influx in the Persian Gulf: diminishing abundance of planktonic foraminifers (ratio pf/total f) in surface sediments [M. Sarnthein 1971. Meteor Forsch.-ergeb. Reihe C 5:1]

sediments, that is, conditions without burrowing metazoans, hence presumably oxygen-free. While the loss of oxygen in the central Baltic basin during the last 50 years is ominously linked to human activity, the record suggests that it can be extremely difficult to tell natural background fluctuations from human influences.

9.5.3 The Persian/Arabian Gulf as a Model for Arid Exchange

The sediments of the Gulf proper are very different from those of the Baltic. Organic content of sediments (less than 1%) is about five times lower, and carbonate (>50%) is more than ten times higher. Also, in the Gulf benthic macrofauna is found at all depths and at all times, and any lamination is quickly destroyed by bioturbation. Heavy metals (notably iron and manganese) do not normally reach the surface of the sediment for release to the bottom water but are precipitated as oxides on reaching the mixed layer within the sediment. The incoming surface waters from the open ocean are loaded with plankton; some of it survives as the salinity increases within the Gulf as a result of excess evaporation; the remainder dies off. In the sediment one finds a record of the processes (and their variation) within the assemblage of planktonic foraminifers (Fig. 9.17).

None of the conditions typical for the Baltic Sea develop in the Persian Gulf, because plenty of oxygen is being supplied to the seafloor by sinking saline waters from the surface, and productivity is generally low. The situation is characterized by downwelling rather than by upwelling. The incoming surface waters from the open ocean are low in nutrients, as are warm surficial seawaters just about everywhere. The Persian Gulf is nutrient starved. One place where anaerobic conditions can develop is in hypersaline lagoons. Sulfate reduction then occurs within the sediment and organic-rich layers with iron sulfide can develop. However, such layers are commonly distinguishable in the geologic record from the stagnant basin layers reporting humid conditions, among other things by the organic carbon and carbonate content.

Suggestions for Further Reading

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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,
DOI 10.1007/978-3-319-51412-3_10

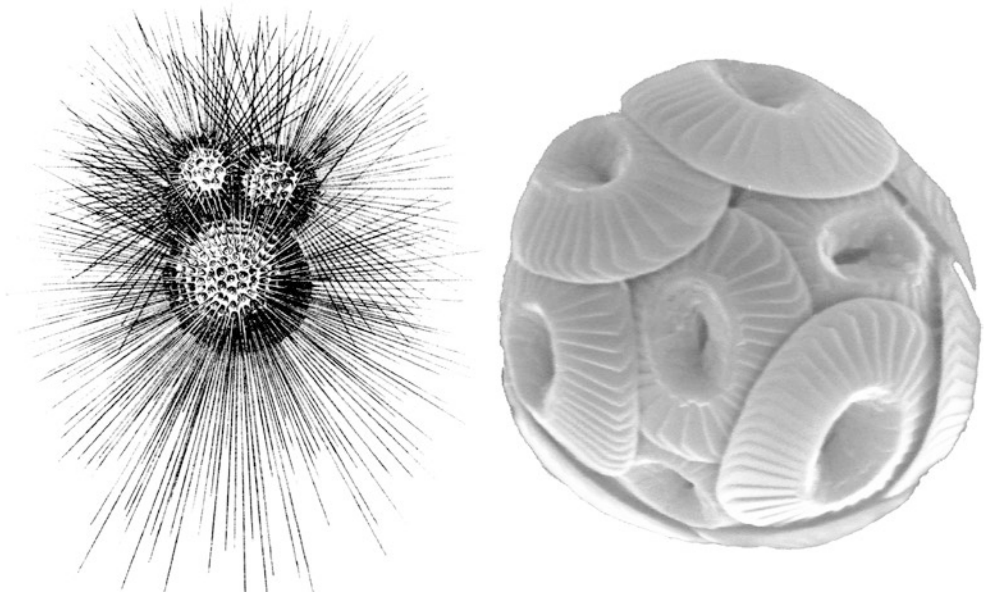

Fig. 10.1 Dominant sediment particles in calcareous deep-sea sediments. *Left*: Planktonic foraminifer (in the water); from H. Brady, Challenger Expedition. *Right*: coccolithophore, SEM graph (Courtesy R. Norris; images not to scale, the coccosphere being roughly ten times smaller than the foraminifer)

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).

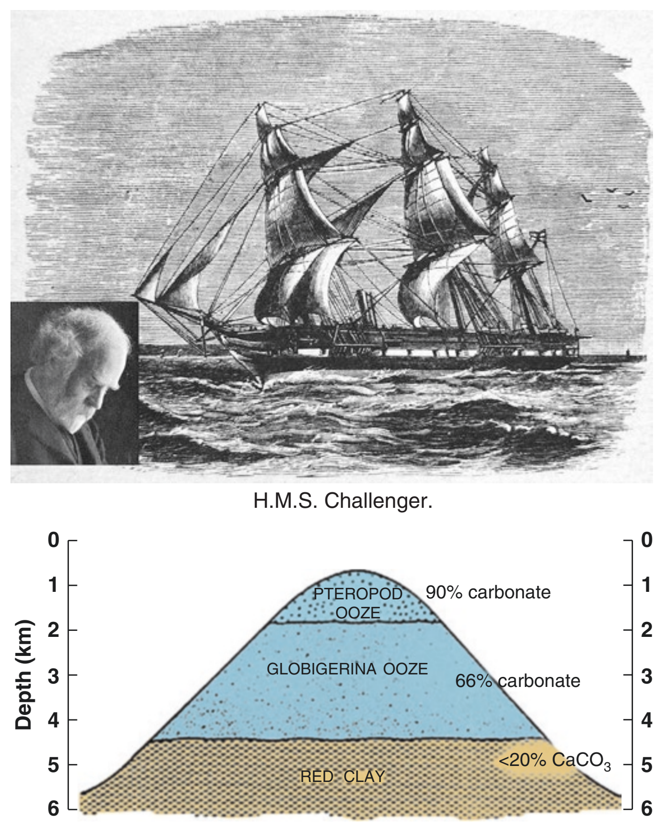

Fig. 10.2 A major result of the nineteenth-century *Challenger* Expedition: the CCD. Dominant types of deep-sea sediment are carbonate "oozes" and "Red Clay" (the residue after dissolution of carbonate ooze, commonly at depths well below 4 km). The calcareous oozes cover almost half of the deep seafloor (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

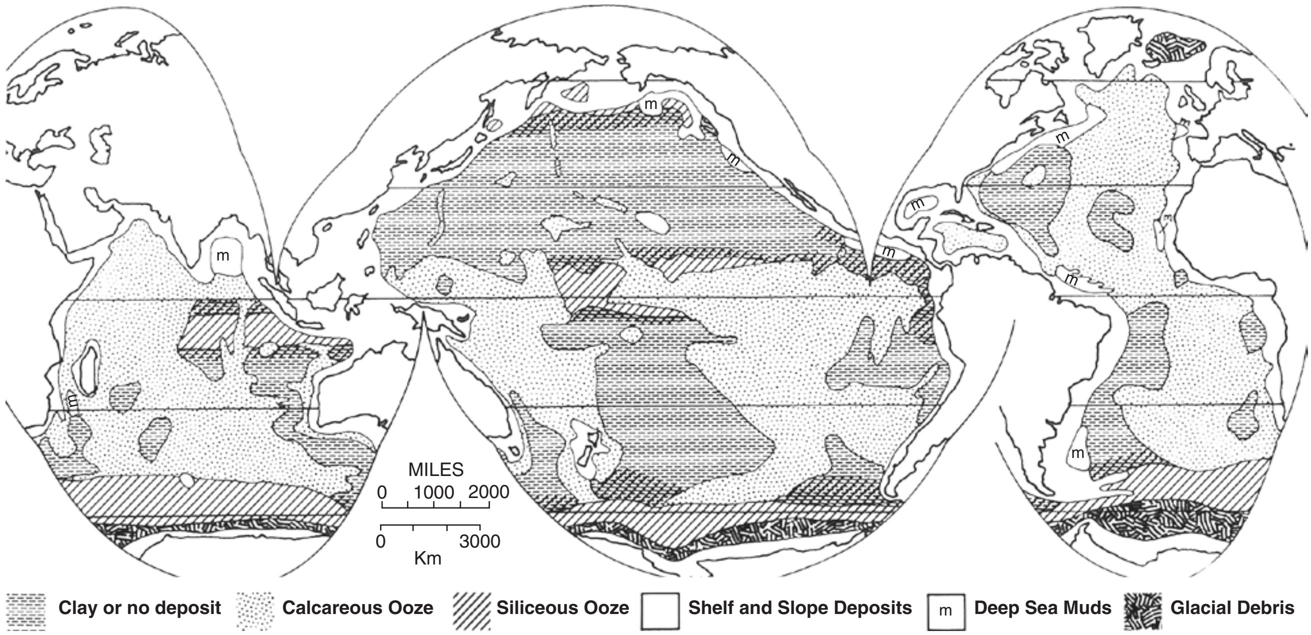

Fig. 10.4 Sediment cover on the deep seafloor. Compiled from many authors (Source as for Table 10.1)

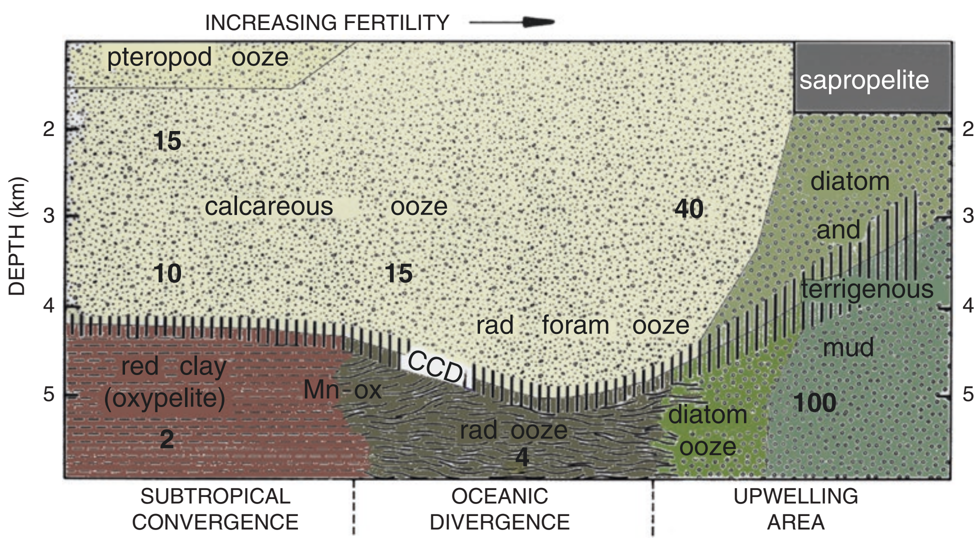

Fig. 10.5 Distribution of major facies in a depth-fertility frame, based on sediment patterns in the eastern central Pacific. Numbers denote typical modern sedimentation rates in mm/millennium (m/million years) (W.H.B., 1974, in C.A. Burk and C.L. Drake (eds.). The *Geology of Continental Margins*. Springer,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

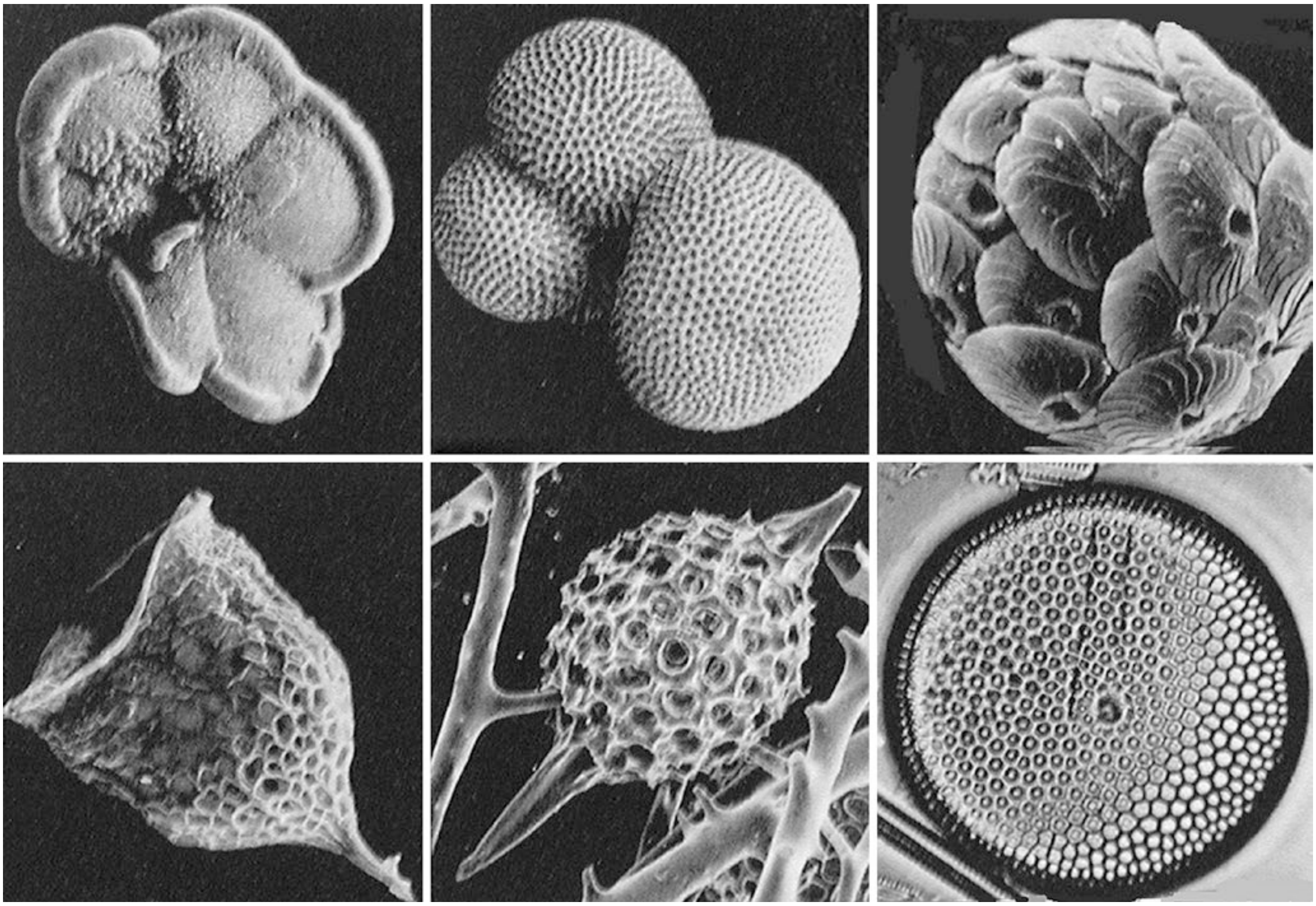

Fig. 10.6 Sediment particles made by shell-bearing plankton. *Upper row*: calcareous forms (two foraminifers, one coccolithophore). On the latter, note the interlocking platelets covering the organism. The platelets are abundant in calcareous ooze. *Lower row*: an organic-walled tintinnid and two siliceous 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

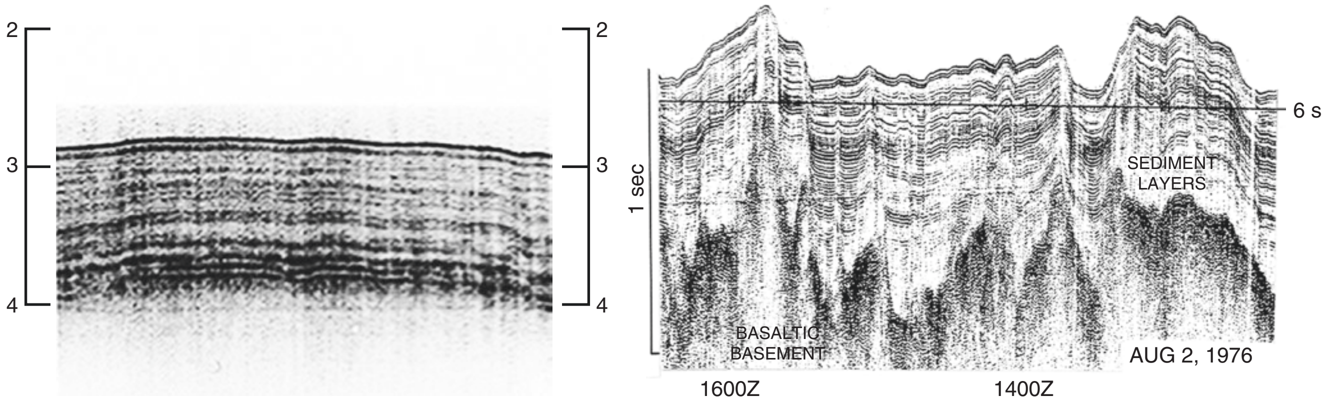

Fig. 10.7 Evidence from seismic echo surveys that deep-sea sediments are surprisingly thin. Thickness scale: echo return time (sound velocity in sediment, 1500–2000 m per s). *Left*: shallow part of Ontong Java Plateau (ca. 100 million years; thickness in excess of 1 km in places); *right*: eastern equatorial 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

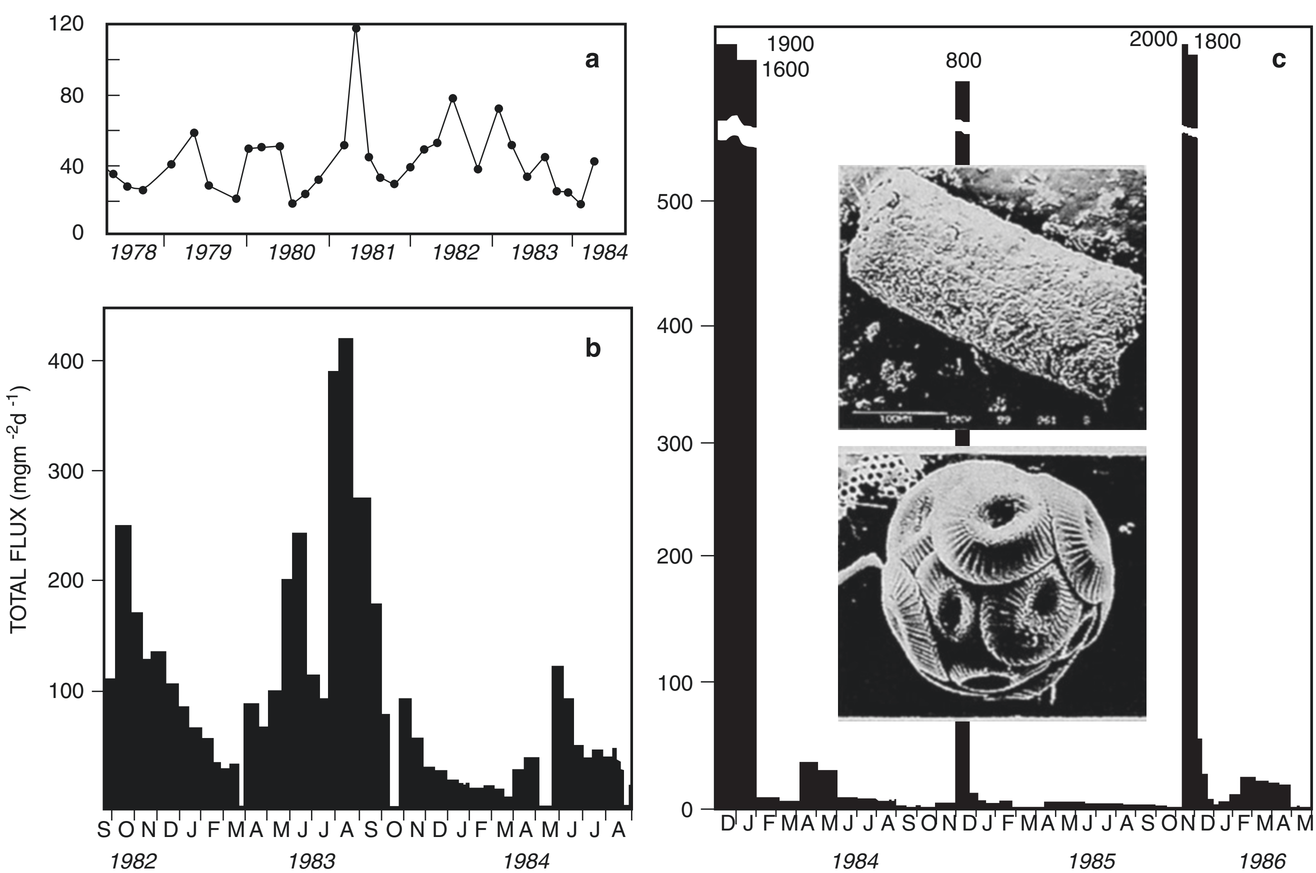

Fig. 10.8 Seasonal and interannual variation in particle flux as observed in traps. (**a**) Sargasso Sea (W.G. Deuser, Woods Hole); (**b**) Gulf of Alaska (S. Honjo, Woods Hole); (**c**) Bransfield Strait, Antarctica (G. Wefer et al., Bremen) (W.H.B. and G. Wefer, 1990. *Global and Planetary Change 3* (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

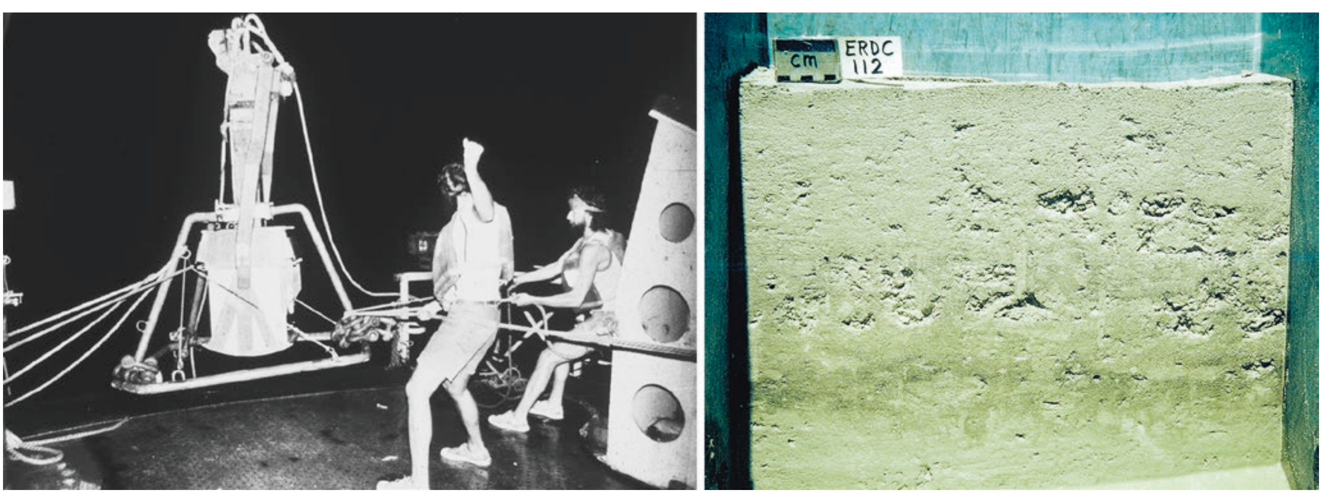

Fig. 10.9 Recovery of calcareous deep-sea sediment by box corer. *Left*: operation of the device, developed from a similar one used by H.E. Reineck (erstwhile director of the Senckenberg Museum in Frankfurt) in the wadden of the North Sea. Lines attached to the equipment are reducing the swinging of 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.

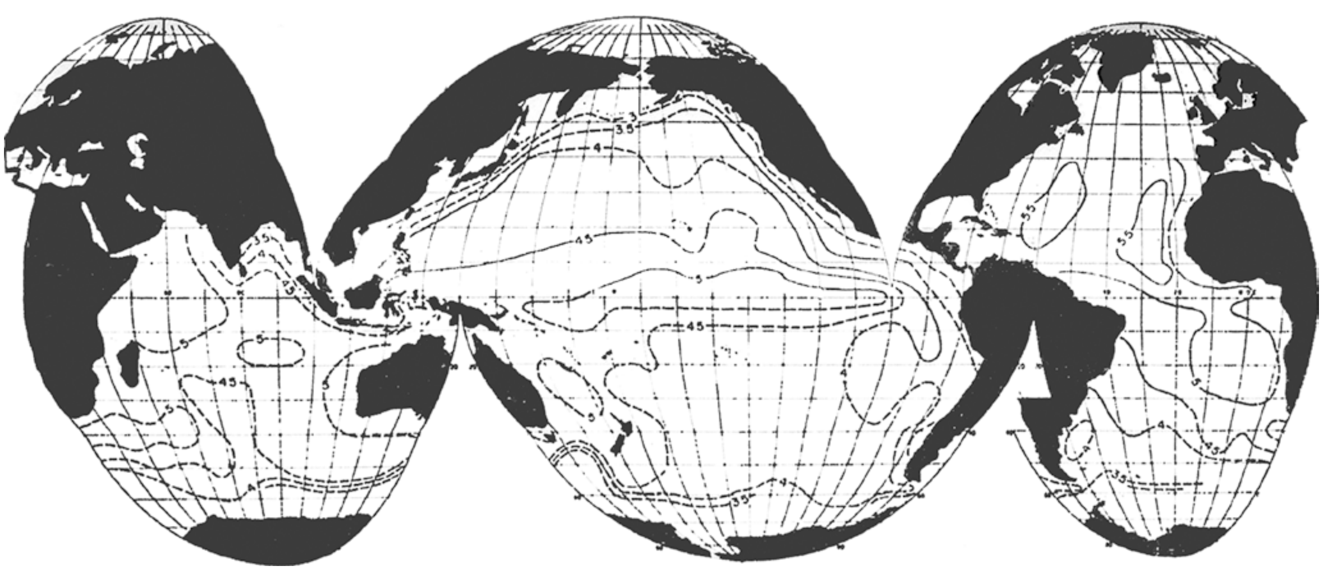

Fig. 10.10 Topography of the carbonate compensation depth (CCD). The seafloor facies boundary (in places extrapolated) between calcareous sediments and sediments with no or very little carbonate. Numbers: depth in km (W.H.B. and E.L. Winterer, 1974. In: K.J. Hsű and H. Jenkyns (eds.) Pelagic Sediments 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.

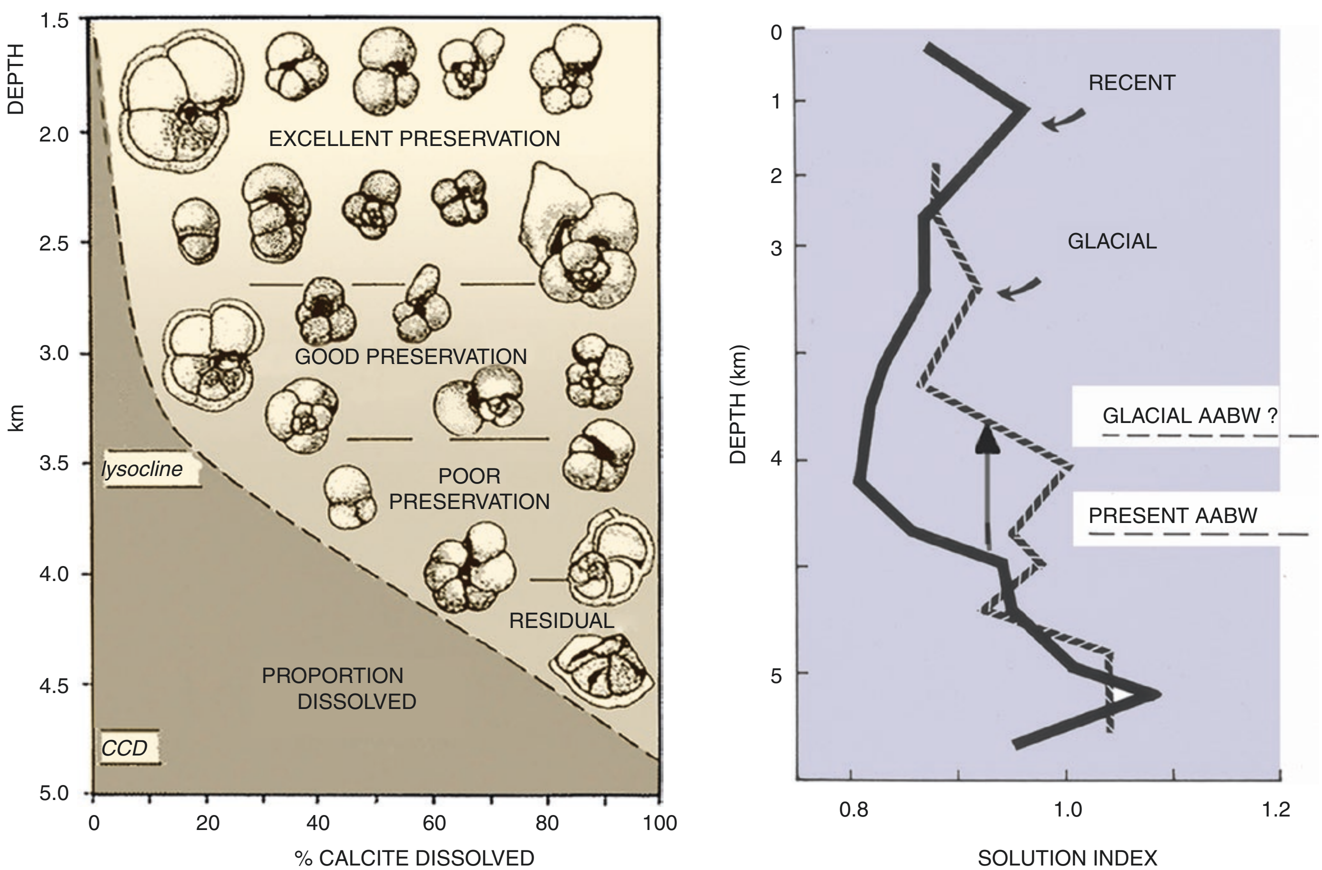

Fig. 10.11 Preservation patterns for planktonic foraminifers on the deep seafloor. *Left*: generalized sketch of distribution of shells in the central Pacific. Drawings courtesy F.L. Parker, S.I.O. (W.H.B., 1985. Episodes 8:163). *Right*: Foraminifer dissolution index in the South Atlantic, in the Holocene 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

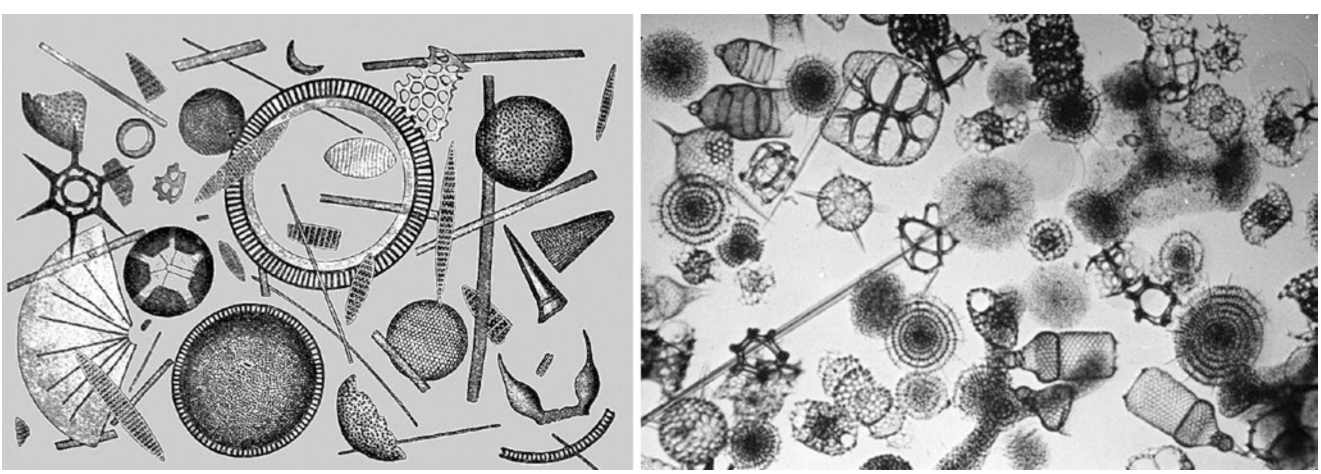

Fig. 10.12 Sand-sized microfossils in siliceous ooze. *Left*: diatom ooze [image from the marine biologist C. Chun via the textbook author O. Krümmel, 1907]; *right*: radiolarian ooze (Microphoto W.H.B., guided by W. Riedel, then S.I.O.)

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.

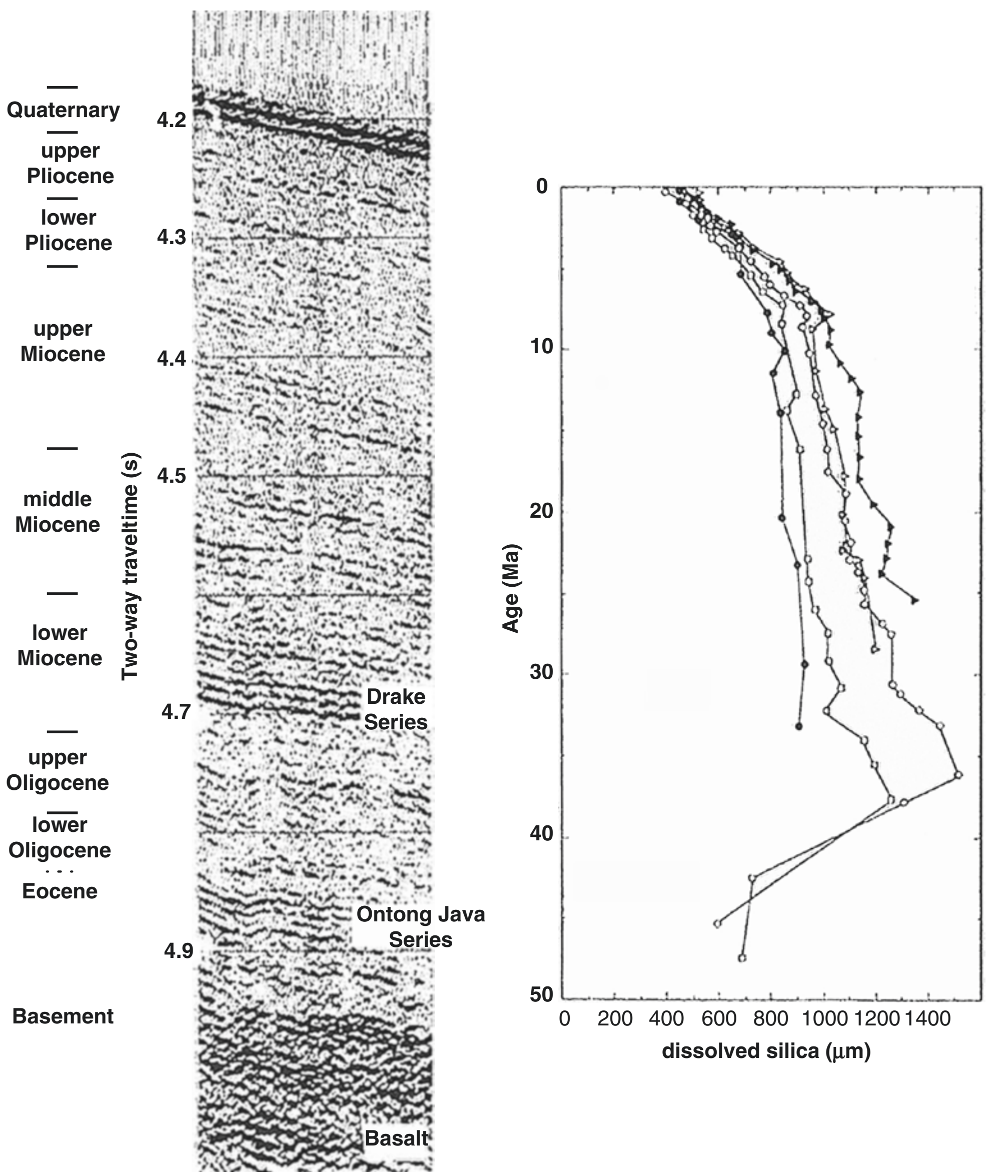

Fig. 10.13 Physical and chemical properties related to silica deposition at ODP Sites on Ontong Java Plateau, western equatorial Pacific. *Left*: acoustic reflectors at Site 805 (vertical scale: two-way travel time of sound, in seconds). *Right*: interstitial water content in Site 805 and several nearby 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-

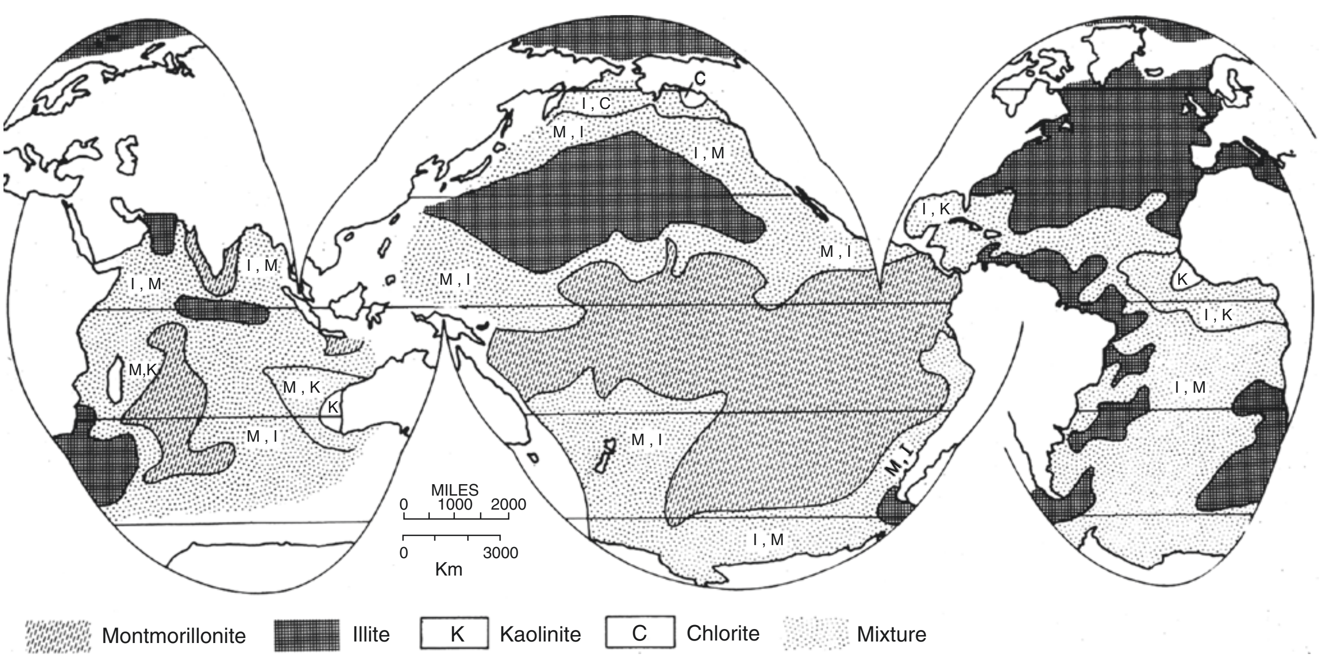

Fig. 10.14 Distribution of the dominant clay minerals in "Red Clay" on the deep seafloor. Compilation from data in works of P.E. Biscaye, J.J. Griffin et al., E.D. Goldberg and J.J. Griffin, D. Carrol, and H.I. Windom (W.H.B., 1974. In: CA. Burk and C.L. Drake (eds.). *The Geology of Continental Margins*. 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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11 Geologic History of the Sea: The Ice-Age Ocean

11.1 Background

11.1.1 The *Albatross* and the Rise of Paleoceanography

Paleoceanography, that is, the study of ocean history, emerged with the investigation of the record of the ice ages in cores from the deep seafloor. Initial efforts by W. Schott (1905–1989) in the 1930s were based on very short cores. Thus, it was the Swedish Deep Sea Expedition (1947–1949) that launched the new science, thanks to long cores retrieved from many parts of the world. In essence, the Swedish expedition played the same role in launching the science of ocean history that the British *Challenger* Expedition had played some 70 years earlier for deep-sea sediment types. The Swedish expedition, led by the radiochemist and physicist Hans Pettersson (1888–1966), took the four-masted research vessel *Albatross* from Gothenburg all around the world, bringing back several hundred long cores taken with a device developed by the oceanographer Börje Kullenberg, who invented the original version of modern "piston corers" for the *Albatross* Expedition.

Kullenberg's device typically recovered cores of a length of 7 m or so, with the oldest sediment commonly having ages between 0.3 million and 1 million years. Many of the cores unfortunately were disturbed, the sediments showing signs of having been "sucked" into the barrel taking the sample by the powerful forces invoked by the piston that facilitated the entry of sediments into the coring tube. Many of the cores turned out perfectly usable, however. They opened up an entirely new way of looking at geologic history, with remarkable time resolution. Devices modified from Kullenberg's invention but retaining the piston principle became the workhorse instrument for gathering ice-age sediments from the deep seafloor (Fig. 11.1). Developments include coring by the drilling ship (coring ahead of drilling) thus ensuring the long-lasting contribution of Kullenberg's engineering.

The scientists reporting on the *Albatross* cores became the founders of paleoceanography. In this fashion paleoceanography became linked to piston coring. The *Albatross* pioneers defined fundamental questions about

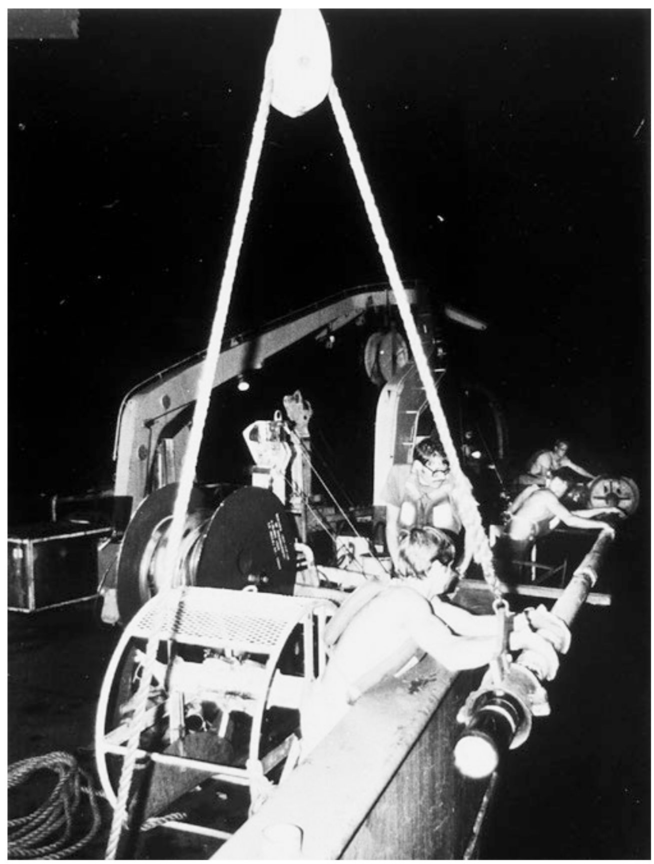

Fig. 11.1 Recovering the Pleistocene record by piston coring. The corer is a wide-diameter model; note the white sediment in the core nose. Other equipment seen on deck: deep-sea camera frame (with protective grid), hydrophone for seismic profiling (wrapped on a spool), and box corer (aft) (SIO Eurydice Expedition, 1975; photo Tom Walsh; taken at night)

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_11

the Pleistocene history of the ocean. The questions included ice-age fluctuations in productivity (G.O.S. Arrhenius, then Stockholm), ice-age changes in plankton distributions and surface currents (F.B. Phleger and F.L. Parker, then Massachusetts), ice-age changes in surface water temperatures and in ice volume (C. Emiliani, then Chicago), and ice-age changes in deep circulation patterns (Eric Olausson, Göteborg). Since then the questions raised by the pioneers, as well as related ones, have been pursued vigorously at the major oceanographic institutions in several countries. Perhaps the most significant early contributions to deep-sea-based ice-age lore were by scientists working on a large number of cores raised on Lamont's research vessel *Vema* (Columbia University).

The outstanding pioneer working on the ice-age record of the deep seafloor was Cesare Emiliani (1922–1995). Emiliani (not a member of the *Albatross* Expedition, but a scientist who used *Albatross* samples he obtained from H. Pettersson) was an Italian-American Chicago-trained nuclear chemist who introduced isotopic studies to deep-sea research. In addition, he was a paleontologist familiar with foraminifers, with a relevant doctoral degree from the University of Bologna, Italy. For much of his career, he worked in the Caribbean, from Miami, Florida.

11.1.2 Support for Milutin Milankovitch (1879–1958)

The cyclic variations of isotopic composition of foraminifers that Emiliani discovered provided crucial support for Milankovitch Theory, that is, the notion that ice ages are the planet's climate response to solar forcing linked to orbital variation (Figs. 1.2 and 11.2). Such variations are strictly cyclic, and their timing can be calculated with great precision. In pursuing such calculations, the Serbian civil engineer and astronomer Milutin Milankovitch (1879–1958) faced formidable obstacles, though. In the geologic record first studied on land, the ice-age sequence was poorly defined and not readily recognized as related to Earth's orbit. The available astronomical background calculations were unsatisfactory by today's standards. Also, Milankovitch's calculations, unaided by computing equipment, were extremely time-consuming.

To show that the course of the ice ages owed to Milankovitch forcing, reliable sequences of deep-sea sediments needed to be reliably dated. Radiocarbon dating (often cited as an important ice-age tool) was poorly suited for the long-term dating necessary to document Milankovitch cycles. Appropriate long-term dating was (eventually) achieved by radiochemical dating of volcanic products on

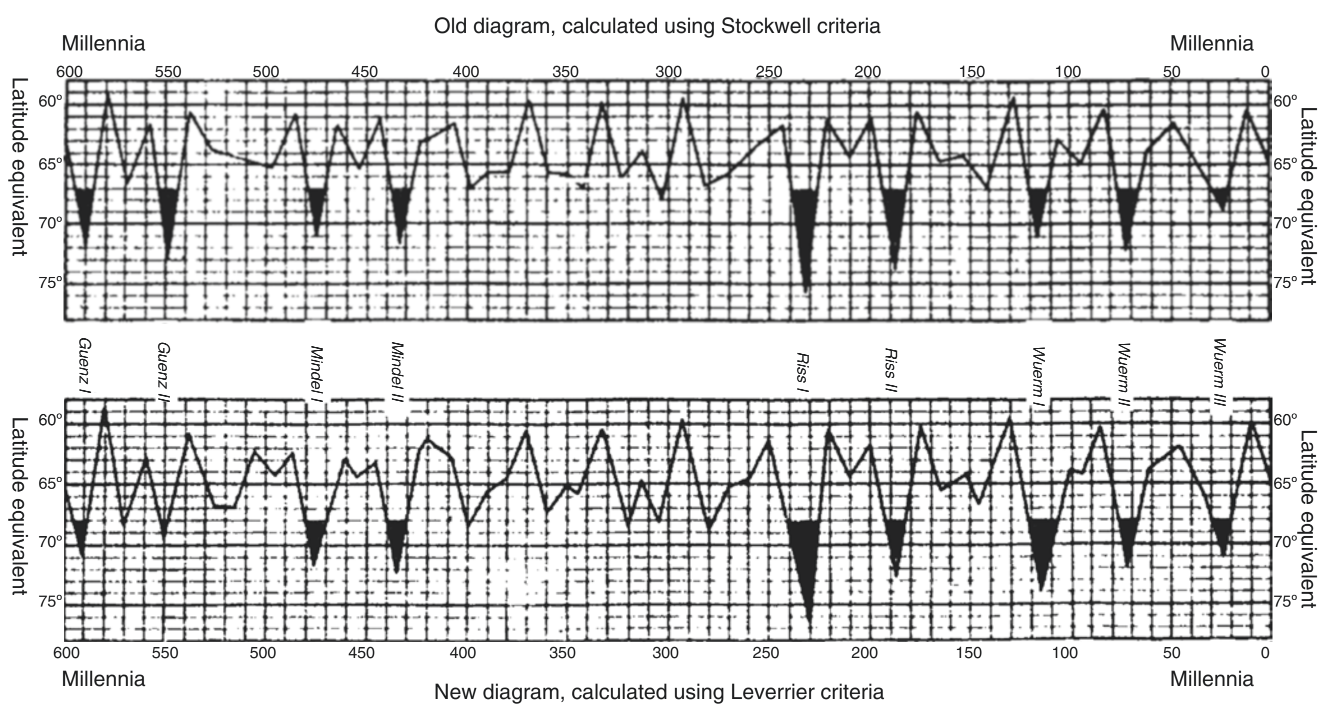

Fig. 11.2 Milankovitch diagram using solar input in high northern latitudes (y-axis) as the driver of major climate change in the ice ages (cold periods: filled valleys in the graph). [The data were sent to the Russian-born German botanist and climatologist Wladimir Köppen by Milankovitch and were soon published by Bornträger (in 1924) (See Schweizerbart web site for "Bornträger")]

land and correlation of such dating into deep-sea sequences with the aid of magnetic stratigraphy (Fig. 6.10). Emiliani's original time scale turned out to be quite incorrect, illustrating the difficulties encountered by the pioneers. The task of refining orbital forcing was tackled by the Belgian astronomer and climatologist André Berger and his associates who reconstructed the various wobbles of planet Earth and its orbit in detail. The template he provided greatly facilitated the matching of deep-sea records to Milankovitch forcing, even bringing the entire Neogene into the range of millennial resolution. Matching the oxygen isotope record to astronomic forcing (Milankovitch "tuning") was the favorite dating tool of the British geophysicist N.J. Shackleton (1937–2006). Shackleton made much use of oxygen isotope stratigraphy in Pleistocene research, usually isotope sequences that were produced in his own laboratory. The third problem, the great labor of making appropriate calculations, was resolved by the rapid development of computing devices and the application of Fourier mathematics (Fourier methods are available since Napoleon but were not routinely used in ice-age research until the 1960s).

Beginning in 1968, the year when the drilling vessel *Glomar Challenger* left port in Galveston (Texas) to initiate scientific drilling in the deep ocean, many new dimensions in the interpretation of marine sediments were bound to emerge in paleoceanography. For once, the length of time that became available for detailed study increased from about 1 million years to the past 100 million years! But more to the point for ice-age studies: thanks to the technical advances in coring on the *JOIDES Resolution* (the drilling vessel that took over from the *Glomar Challenger* in 1985), ever more detailed studies could be done on the full 2 million-year ice-age record of the deep sea, taking advantage of a new (millennial) time resolution based on Milankovitch tuning (i.e., using Milankovitch Theory as a dating tool).

11.1.3 Milankovitch Cycles Explained

A brief explanation of Milankovitch Theory is in order. Unsurprisingly, when the disk of the sun is large in the sky (whenever our planet is close to the sun, i.e., in "perihelion" position), more sunlight is received than when the sun appears small. According to Milankovitch, if the disk is large in summer in high northern latitudes (i.e., perihelion in northern summer), melting of northern ice masses can occur, but if small (i.e., perihelion in northern winter), northern ice buildup proceeds. The seasons migrate along the orbit (relative to perihelion), completing a cycle roughly every 21,000 years; this defines the climate-relevant *precessional cycle*. Thus, precession is a matter of the *eccentricity* of the orbit discovered by Johannes Kepler (1571–1630). In addition, the tilt of the Earth's axis changes through a range of somewhat less than three degrees on a cycle near 41,000 years (the present tilt is intermediate, at 23°27′). The tilt (*obliquity* of the rotational axis) determines how high the sun can rise during noon, in northern summer. A higher position translates into higher insolation in high latitudes, adding to any precession effect.

And that is all that is to it: the "precession" effect controls the apparent size of the sun through the seasons, and the obliquity or tilt of the Earth's axis determines how high the sun rises at noon at the different latitudes. One more thing: the precession effect has opposite signs in the northern and the southern hemisphere, while the orbital tilt effect is the same in both hemispheres (the sign of seasons being opposite).

The ice-age fluctuations in both hemispheres being parallel (more or less) and the available record apparently mainly reflecting northern irradiation patterns, we must assume with Milankovitch that the northern hemisphere takes the lead in pacing the ice ages. There is more land at crucial latitudes in the northern hemisphere (affecting land albedo from snow cover), and ice buildup is farther from the pole than in the south. As a result, northern ice masses are a lot more sensitive to change (positive albedo feedback) and vulnerable to destruction by solar variation than southern ice. The climate information of the northern ice ages is readily made global by changes in sea level and in carbon dioxide content of the atmosphere. Neither effect is restricted to one hemisphere, but the northern one is taken as dominant in originating changes.

It is obvious that the presence of Milankovitch forcing in the ocean's climate should be reflected in appropriate cycles of sedimentary particles in deep-sea sediments, and this is indeed the case, according to a study by the Lamont geologist J. Hays and colleagues, in 1976. For clarity, the study did show that Milankovitch forcing is extant; it did not document that this forcing is the only one at work.

11.1.4 Melting Not Ice Buildup

As the time scale for the ice ages evolved, it supported Milankovitch's emphasis on summer insolation in high northern latitudes (i.e., melting) in preference to earlier ideas that focused on the building of ice caps. The climatologist Wladimir Köppen, who advised and supported Milankovitch, had been right. His intuition in backing the thoughts of the confident young Serbian proved correct. Milankovitch was right on when he implied that the problem is *not* how to make ice, as assumed earlier by the brilliant geologist James Croll and various other scientists linking astronomy and ice ages; instead, it is how to *get rid of the ice in an unusually cold world*. Contrary ideas on this point still surface on occasion, for example, in the hypothesis of the UC physicists R. Muller and G. MacDonald; a hypothesis that purports to explain the 100-kyr cycle of the ice ages by rhythmically obscuring the sun, a notion that answers the ancient question about cooling and buildup of ice mass, rather than worrying about melting, as did Milankovitch.

11.2 A Search for Lessons

11.2.1 The Ice Ages as Information Resource

We live in a period of northern ice ages, that is, a succession of periods of enormous ice buildup in northern latitudes, especially on the North American continent. The great extent of the ice here during the last glacial maximum was mapped by R.A. Daly in the first half of the last century and subsequently by R.F. Flint in the 1950s. One of the most appealing compilations is from the second half of the last century, done under the tutelage of the late John Imbrie of Brown University (Fig. 11.3). The book by J. Imbrie and his daughter K.P. Imbrie (published in 1979) is a special treat in this context. Our current situation is characterized as "postglacial." The last glacial period ended about 15,000 years ago with vigorous melting setting in. Sea level ceased rising some 7000 years ago. It did rise for nearly 10,000 years and by some 125 m. Remnant ice near the North Pole (foremost on Greenland; Fig. 11.4) represents but a small fraction of the former northern ice mass.

Albedo feedback is crucial in ice age theory. Ice forms largely in high latitudes. At sea, the formation of sea ice (itself reflective and a base for snow) provides for sudden and large albedo change. In the mountains elevation is of prime importance for ice formation, so that the time since unloading continental crust and letting it rise is clearly of great relevance to ice growth. Lately, mountain glaciers the world over have been shrinking. Examples are seen throughout the Rocky Mountains of western America (some rangers fear the disappearance of glaciers in Montana's Glacier Park on the scale of

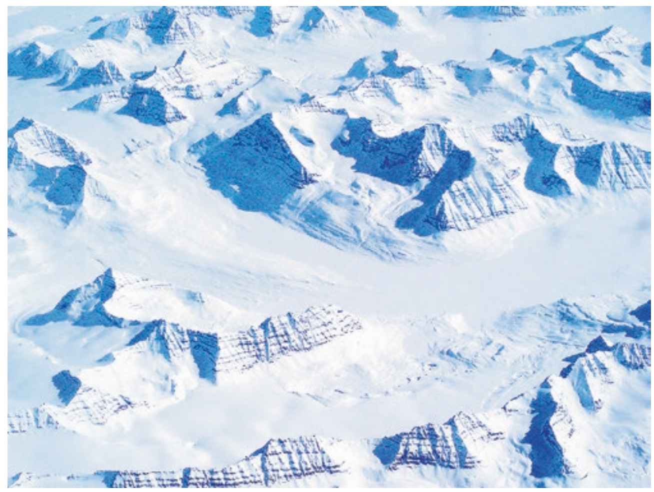

Fig. 11.4 Our place in geologic time: remnant ice in Greenland (Air photo W.H.B)

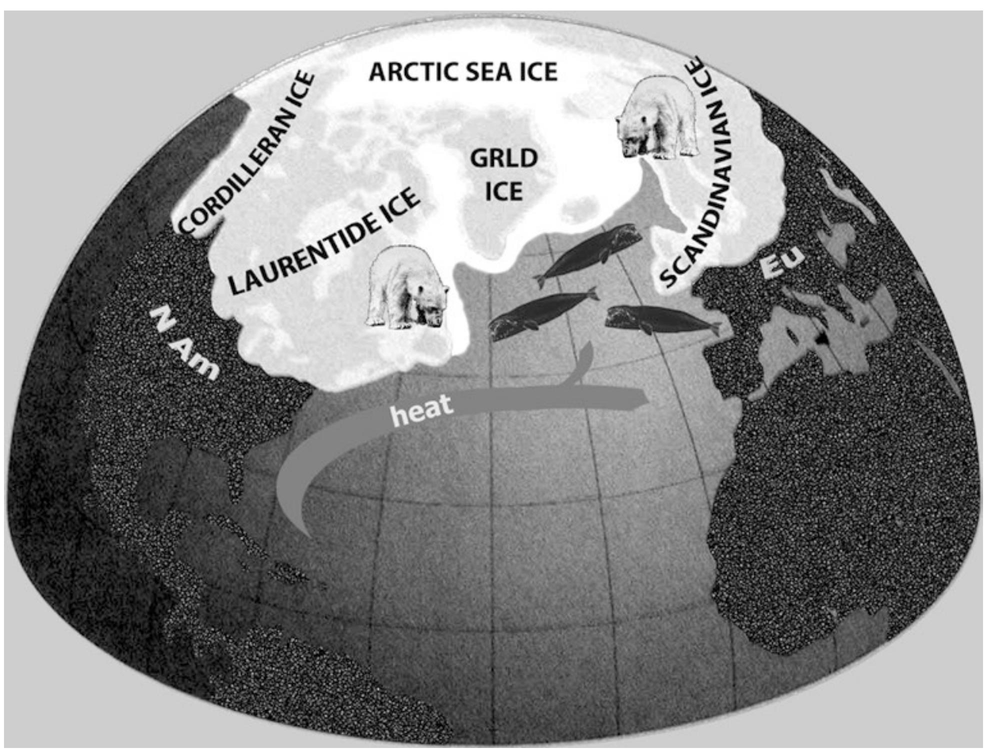

Fig. 11.3 Sketch of the northern hemisphere (Atlantic and Arctic) in the last ice age. Note the asymmetric distribution on both sides of the Atlantic (Map after J. Imbrie and associates. See J. Imbrie and K.P. Imbrie, 1979. Ice Ages, Solving the Mystery. Enslow, Short Hills, N.J. Polar bears from a drawing by F. Nansen; whale shadow-grams (placement hypothetical) after US NOAA)

decades). Such developments, and the ongoing rise in sea level, have motivated increasing interest in the history of the northern ice ages with a view to lessons obtainable for generating expectations and hence for planning.

11.2.2 Ice Ages and Positive Feedback

The ice ages do have interesting information on climate change, of course, some of which is relevant to present concerns and discussions of the topic, despite the great difference in time scale between ice-age history and human life (factor of 100). Regarding geologic history, the ice-age time scale (measured in millennia) is more than a hundred times more refined than the regular geologic scale, with important implications for understanding rapid change. The ice ages have geological lessons, because an appreciation of positive feedback (mainly albedo and carbon dioxide, perhaps methane) is crucial for understanding abrupt climate change. Positive feedback in the climate system is implied in Milankovitch Theory in the first place. When using rather subtle changes in the distribution of solar energy to drive major climate change, the question of albedo change becomes paramount. Snow reflects sunlight and snow-free areas do much less so; thus, small changes in solar forcing are readily magnified wherever snow is seasonal, that is, mostly in high and in moderately high latitudes (sea ice included) and in high elevations (Fig. 11.5).

The term "ice age," incidentally, is used in various somewhat confusing ways. The sense in which the term is used in any given situation has to arise from the context. One convention applies the term to the entire period since the buildup of large ice masses on the northern hemisphere, that is, roughly the last 3 million years (here referred to mainly as "the (northern) ice ages"). Another type of use restricts the label "ice age" to periods of glaciation, as defined by the ice covers in Canada and in Scandinavia. Finally, there is a third use of the word train. It is employed, on occasion, as an abbreviation for the term "last ice age," which had its peak some 20,000 years ago. Many prefer the expression *last glacial maximum* (LGM) for that particular concept. The "ice age" that is congruent with the "Quaternary" (yet another use) is here taken as comprising the last 2 million years. If the last 10,000 years (the "Holocene") are exempted from the "Quaternary," one obtains the "Pleistocene," which is a formal technical term and hence avoided here.

The study of the ice ages cannot but improve one's understanding of climate change. However, expectations for elucidation of relevant processes are easily exaggerated. As geologists we are largely in the dark about the future. The future in fact has no analog in the past, as far as this can be determined. We are moving from an unusually warm interval within a long succession of ice-dominated times into a period of ever greater warming, a movement of a type that is without precedent in the last several million years. Also, at present it presumably occurs at a highly unusual speed. Geologic history reliably defines what is possible (what *can* happen) in the future, that is, in the attempt to predict, based on observation and experience. It is not a reliable source of information about what *will* in fact happen.

11.2.3 Useful Insights

Among the firm insights that emerge from the study of ice ages is the one that albedo feedback is a prime mover in rapid climate change. The behavior of snow and ice (including sea ice) matters. A second fundamental insight – this one from polar ice core studies rather than from marine geology – is

Fig. 11.5 Albedo feedback from ice and snow. *Left*: Sierra Nevada; peaks are higher than 4000 m. Note brightness of snow, darkness of forests, implying large seasonal change in albedo. *Right*: Svalbard 80°N. Note reflective snow on ice; relatively *dark water*

that warming and cooling over the last million years was invariably accompanied by a natural increase or a decrease in the concentration of carbon dioxide in the atmosphere. To what degree this change in concentration of greenhouse gas is a driver of the change in climate and to what degree it is merely an expression of the climate change is a famous chicken-or-egg problem and a topic for much interesting academic debate.

Actually, any postulated either-or scenario may not address what is happening. Effects can develop into causes – a situation highly relevant in all of geology and well appreciated by all scientists and engineers working with evolving systems. When studying the ultimate cause for the appearance (or elimination) of an ice age, we come up against the problem of evolving systems, that is, systems whose change produces more change. The link is called "feedback," and it works both for the onset of ice ages and the subsequent cycles (which characterize the transitions from being ice-free to having eternal ice and vice versa).

Traditionally the question about the origin of ice ages has been about the cooling that is necessary to make lots of ice. Geologists commonly proposed at least one central force resulting in cooling: mountain building and uplift of the land. Uplift can increase the reflectivity of the ground rather suddenly, as when a large area goes above the snowline or a shelf emerges, exposing white carbonate rock. Also, uplift and warming have for more than half a century been recognized as a control of volcanogenic carbon dioxide, some of which is used up in weathering, by becoming a component of carbonate. Deep mechanical weathering (fostered by the buildup of ice) can enhance uplift by unloading mountains, that is, leaving the upward push by mountain roots less opposed by materials covering the mountain. There is a reason why one of the modes of elevation (Fig. 2.1) is well above sea level. It suggests that uplift is ubiquitous. It is up to geologists to document the underlying process. Another important insight derives from the observation that whenever fast melting occurred during deglaciation, much of the required energy appears to have been delivered by the gravitational instability of the ice itself, rather than solely from the heat in the surrounding environment.

11.3 The Last Glacial Maximum in the Sea

11.3.1 General Patterns

It is generally agreed that *surface currents during glacial periods were stronger than now*. It is obvious why this should be so: surface currents are driven by winds, and the strength of relevant winds depend on horizontal temperature gradients at sea level. With the ice rim and polar front much closer to the equator, the temperature difference between ice rim (0 °C) and the high tropics (25–30 °C) was compressed into a much shorter distance than now. Hence, the temperature gradients were greater, winds were stronger, and so were the ocean currents generated. As a consequence of stronger surface currents, *equatorial upwelling* was intensified (as first suggested by G. Arrhenius during the Swedish Albatross Expedition), as was *coastal upwelling* (as seen, e.g., in the benthic foraminifer sequence off Namibia). Thus, despite the fact that the productivity of the sea must have decreased in very high latitudes because of growth of sea ice cover, the LGM likely led to an overall increase in production owing to an increase in mid- and low latitudes because of intensified mixing and upwelling. In consequence of increased meridional temperature gradients, we must assume that winds during glacial periods were more zonal, that is, more predictable than in warm periods.

Also, it is generally accepted that the glacial-time *ocean surface was cooler*, on the whole, than today. With a substantial part of northern continents covered by ice and with sea ice greatly extended, the Earth reflected the Sun's radiation more readily (had a higher *albedo*) than today. As a result, it absorbed less of the incoming radiation, and its atmosphere was cooler. In addition, concentrations of the greenhouse gas carbon dioxide were lower during the last glacial maximum (a fact documented by laboratories in Grenoble and in Berne, in polar ice cores). Carbon dioxide in the last glacial maximum was but 2/3 of the postglacial natural background concentration. As a result, the lower atmosphere held less water vapor (the most potent common greenhouse gas) than now. Other possible feedbacks have to be considered also when discussing a cooling of the planet, for example, from changes in plant cover on land affecting albedo and from changes in chlorophyll content in surface waters at sea (also affecting albedo, as variously pointed out by professional students of plants on land and at sea).

An attempt to precisely determine the amount of cooling on an ocean-wide scale (as in the 18 k map by the CLIMAP group), while confirming the concepts mentioned (e.g., strengthening of surface currents), turned out to be an extremely difficult task, involving plankton ecology. An attempt to precisely reconstruct ocean temperatures from fossils in cores may easily fail in finding the correct historical drop over large areas. An overall cooling of around 5 °C for tropical surface waters (roughly twice the general CLIMAP value) now seems acceptable to many workers in the field. True, the difference of past to present temperature seems to change considerably with latitude. However, when using organic remains for reconstruction of precise temperature history, one must be aware of the ability of organisms (especially tiny ones with rapid reproduction cycles) to adapt to climate change on a millennial scale and thus for fossils to show less change than might seem appropriate for differences in conditions. In addition, organisms may react to elements in the changing conditions that are quite different from those assumed to be controlling the fossil abundances. These types of problems of fossils are pervasive in all of historic reconstruction.

A fourth agreement among students of the ice ages is particularly important: a substantial *drop in sea level* results from the buildup of glacial ice. Roughly 125 m or so is the generally accepted range for the last 20,000 years, for which changes in sea level are dated and documented in great detail (Fig. 6.5). These changes represent a phenomenon with an enormous number of important implications for geology and for climate change (including changing the sites of deposition of carbonate: shelf seas tend to trap carbonate; dry shelves cannot do so).

11.3.2 A Millennial Perspective and the Task of Correct Dating

A millennial perspective is appropriate when discussing ice-age climates: the resolution of many of the available deep-sea records is limited roughly to a thousand years. The ice record may offer greater resolution, but what can be cored to great depth is restricted to high latitudes (or high elevations). The record on the seafloor is not so restricted, but it has other problems. The ocean mixes on a time scale of about one millennium, and great ice masses take millennia to build and also to melt. Even the timing of the last glacial maximum is in some doubt on the millennial scale.

A resolution focused on one millennium might seem a bit coarse for many purposes. However, for many geologists an ability to separate one millennium from a neighboring one is commonly referred to as "high resolution." It took some time to get there, actually. Dating used to be quite fuzzy even a few decades ago, with geologic age estimates routinely off by some 10 or 20% (i.e., by tens of thousands of years or even by millions, depending on the age being discussed). Absolute dating (assigning ages in terms of years before present) using radioactive isotopes other than radiocarbon (mainly certain types of *uranium* and its decay products, or "daughters") has brought relief from the contamination problems.

The exploration of paleomagnetic sequences and their introduction to deep-sea cores has fundamentally changed the earlier limitations on dating deep-sea sediments by correlation to widely used time scales on land. These developments have allowed expansion of millennial assignments to the million-year scale, based on Milankovitch tuning. Orbital cycles can serve as guides back to many millions of years ago because – according to experts concerned with the history of the solar system – the planets of our solar system seem to retain their current patterns of travel over many millions of years.

11.4 The Pleistocene Cycles

11.4.1 Background

It is commonly safe to start any essay on pioneers in any geological subject whatever with the British barrister Charles Lyell (1797–1875), erstwhile vice-president of the Geological Society of London and prolific textbook writer. His opus "Principles of Geology" (first published in the 1830s) provided a scientific framework for doing geology at a time when the Holy Bible was still widely used as a geology text even by some scientists. In later editions of his work (e.g., Lyell, 1868), we find that what he had to say about the ice ages, though appropriately vague in places, is quite interesting. For example, he quotes John Herschel (1792–1891), son of the famous astronomer William H., as invoking changes in the brightness of the sun as a possible cause for climate change on Earth – an early version of change of radiation balance. More in line with later reasoning, John Herschel already invoked effects of orbital elements on climate, notably eccentricity. Almost of equal interest is what Lyell does not say, implicitly dismissing various early notions regarding the origin of ice ages.

It was the brilliant self-taught Earth scientist James Croll (1821–1890), member of the Geological Survey of Scotland, who followed up with astute calculations on the notion of orbital forcing floating about within the scientific community of the time. He argued that the effects of changing eccentricity of Earth's orbit on climate would be substantial. Obviously, such change would affect the contrast between seasons as Earth changed its distance to the sun, with the closest approach sometimes in northern summer and sometimes in northern winter. Croll's theories (summarized in a book published in 1875) bravely addressed the challenge posed by multiple ice ages. Unfortunately for Croll, multiple ice ages were not yet generally recognized as a reality of geological history. They were proposed by his colleague, the distinguished Scottish geologist James Geikie (1839–1915) at the time Croll pondered the matter, and practically by no one else.

Multiple ice ages were a tremendous discovery, of course. The origins of this discovery are not entirely clear. Albrecht Penck, the leading ice-age geologist of his time at the start of the last century, gave credit to James Geikie. Penck himself (with his associate E. Brűckner) postulated four large glaciations, which he labeled with the names of rivers draining the Alps, rivers that bore rubble in their banks from ancient floods presumed to have been associated with glacial activity (i.e., the melting of ice). The postulated reasoning was acceptable at the time, but Penck-Brűckner assignments are in doubt since the 1960s and have been abandoned decades ago. Instead, there is the Milankovitch scheme of orbital forcing of climate change, a scheme that Penck thought bound for failure. However, the isotopic geochemistry introduced by Cesare Emiliani attained its prominence because of its relevance for Milankovitch Theory. The Theory attempted to explain the sequence of multiple ice ages. And thanks to Emiliani, Pleistocene geology acquired a multitude of numbered climate excursions, many more than Penck's measly four.

Milankovitch triumphed. Ironically, his goal was to explain the multiple glaciations as interpreted by Penck and Brűckner (see Fig. 11.2). The success of his theory (well after his death) in fact stems from observations on deep-sea sediments, observations that suggested that Penck's scheme could be misleading when applied globally. The chief problem arising with Milankovitch, though, has to do not with the Penck target he pursued. Rather, it has to do with the observed prominence (in deep-sea sediments) of a cycle near 100,000 years, a climate cycle that dominates the time period studied by Milankovitch and apparently was not identified by him. The origin of the long cycle is still not clear.

However, regardless of the doubts arising with respect to the 100,000-year cycle, Milankovitch Theory is now the most valuable part of the toolbox of ocean historians, as emphasized by Nicholas Shackleton of the UK; André Berger of Louvain, Belgium; Lamont's James Hays; and Brown's John Imbrie, among other ice-age experts of the last several decades. In fact, Milankovitch Theory has achieved textbook status. It is a tool without peer when the task is to date ice-age sediments from the deep seafloor or to determine sedimentation rates of such sediments. Apparently the theory works for sediments that were deposited well before the onset of the northern ice ages but carry information from orbital cycles, even in ancient Cretaceous deep-sea sediments (Chap. 13). Milankovitch Theory has profound implications for all of climatology and the Earth sciences in general, because of its emphasis on solar system astronomy, which provides for known forcing (although not feedbacks!). The theory is a revolutionary force in natural philosophy: it successfully emphasizes external astronomical factors in the determination of geologic processes on the surface of the planet.

11.4.2 Carbonate and Productivity Cycles

The first hint from deep-sea sediments that there was a long succession of cycles (as called for by Milankovitch Theory) was delivered by the carbonate cycles of the eastern equatorial Pacific (Fig. 11.6, left panel). The cycles were described

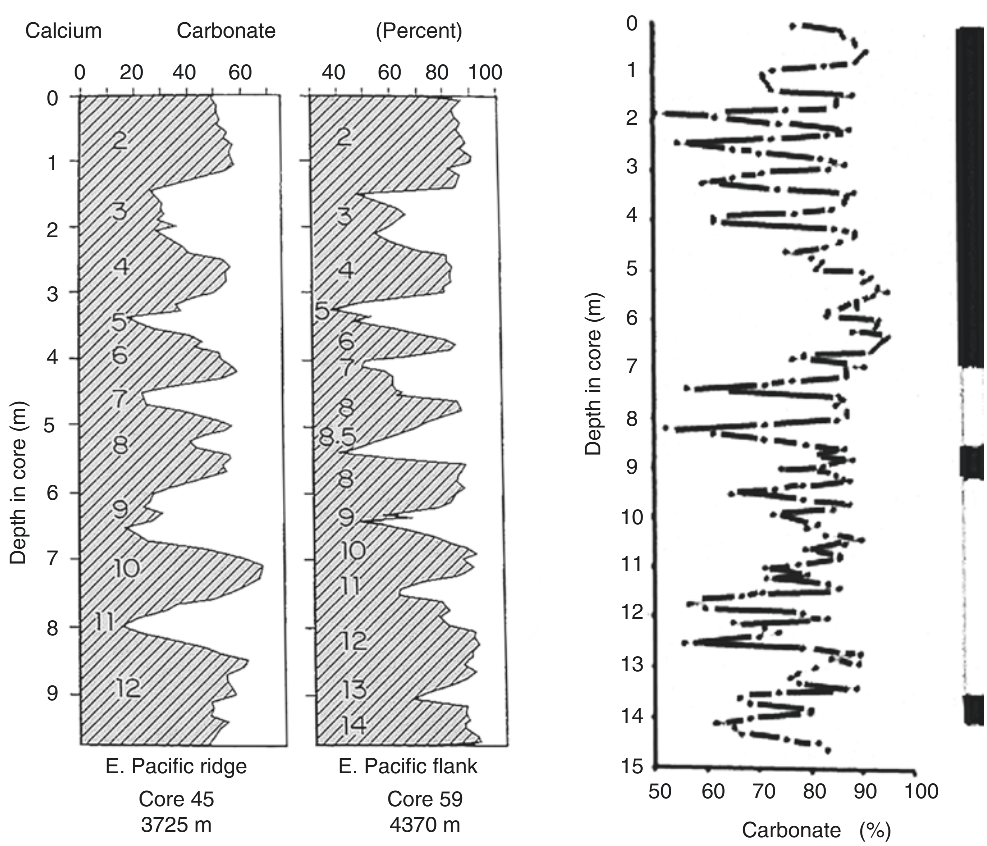

Fig. 11.6 Carbonate cycles of the eastern tropical Pacific. *Left*: cycles as recorded in the Albatross report (After G.O.S. Arrhenius, 1952. Reports of the Swedish Deep Sea Expedition). *Right*: dated cycles as reported by Lamont scientists (J.D. Hays et al., 1969. Geol. Soc. Am. Bull. 80:1481). Note the control of timing by paleomagnetism

by the Swedish-American geochemist G.O.S. Arrhenius, member of the *Albatross* Expedition. How the cycles are made has been the subject of much academic discussion. The cycles are now commonly interpreted as *dissolution cycles*, with high dissolution of carbonate in interglacial time intervals (low carbonate values) and low dissolution in glacial periods (high carbonate values), rather than chiefly as evidence for varying production in the equatorial zone of the Pacific, as originally suggested.

That the carbonate cycles are in fact Pleistocene in age and therefore represent conditions in the ice ages is readily seen in a long piston core studied at Lamont by the paleoceanographer-geologist J. Hays and the magnetism expert N. Opdyke (then Lamont, now in Florida) and colleagues (Fig. 11.6, right panel). The Lamont core is dated by several magnetic reversals. The first one (going down in the core) is the Brunhes-Matuyama reversal, the date of which is roughly 780,000 years, as confirmed by radiochemistry on land. The patterns suggest that the *Albatross* cores end in the earliest part of the Brunhes Chron, making the sedimentation rates come out near 1.3 cm/millennium. A cycle length of 1–1.5 m consequently suggests a carbonate cycle close to eccentricity, when assuming Milankovitch forcing.

Even though the carbonate cycles may not chiefly result from variations in production, this does not mean that they cannot run parallel to such variations. That plankton productivity is higher during glacials than during interglacials in the equatorial Pacific now appears well established, and the associated variation in carbonate output must therefore contribute to the carbonate cycles observed. Whether the change is large enough to explain most or all of the observed range of variation of carbonate ooze, however, is another question and is seriously doubted by many geologists.

Convincing evidence on Arrhenius-type productivity variations along the eastern Pacific equator was presented by A. Paytan and collaborators, using barite concentrations in ice-age sediments. H. Perks and R. Keeling found evidence on Ontong-Java Plateau in the western equatorial Pacific that glacial-time production exceeded interglacial production. Foraminiferal composition in a long core studied by M. Yasuda and unpublished box core results from that area point in the same direction (Fig. 11.7). Thus, effects of productivity cycles along the Pacific equator on deep-sea carbonate sedimentation are well established. Obviously, the question, when discussing effects from dissolution cycles versus effects from productivity variation in making carbonate cycles, is one of proportional importance. It is not a question of the either-or type.

Evidence that ice-age productivity fluctuations were global in nature comes from a record in the central Atlantic (Fig. 11.8). P.J. Müller and E. Suess used organic matter as an indicator in their reconstruction. Such effects can be difficult to assess, due to problems arising from changes in dilution with inorganic matter and in preservation of organic matter. Regarding glacial productivity, Müller and Suess suggested that it exceeded the interglacial one by a factor of 2.3, very similar to the factor of 2 determined for the eastern tropical Pacific by A. Paytan and colleagues. Increased glacial productivity in coastal upwelling regions, presumably thanks to increased winds (in agreement with the Arrhenius mechanism), seems to hold true in many places all over the world.

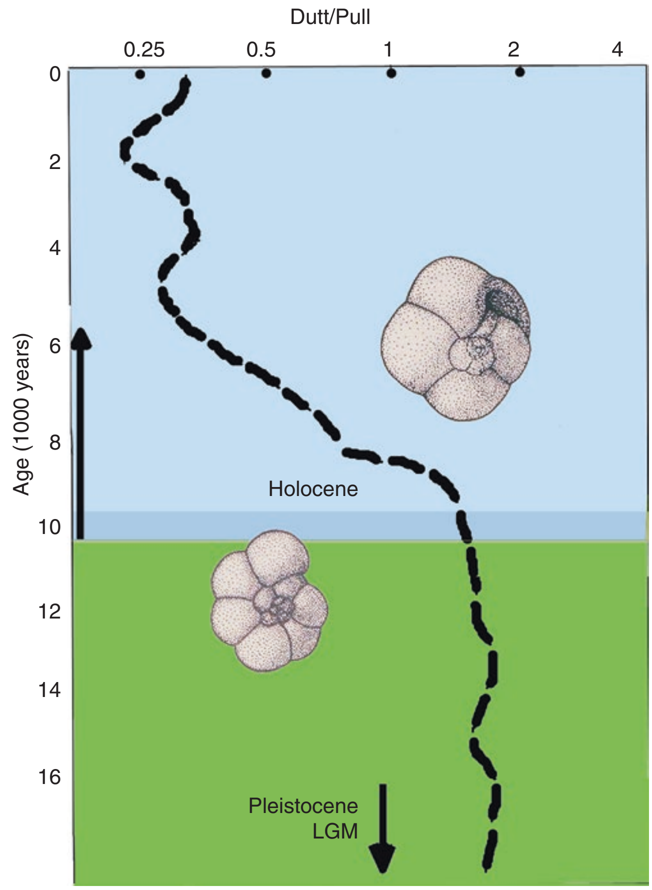

Fig. 11.7 Evidence for a sharp decrease in productivity in the Holocene in the western equatorial Pacific, on Ontong-Java Plateau, based on a drop in abundance of the planktonic foraminifer *N. dutertrei* (a high-production indicator) and a rise in the species *P. obliquiloculata* (a low-production species) in a box core bearing a well-preserved fossil assemblage. The indication is that productivity was considerably higher during the last glacial maximum than within the interglacial period following deglaciation (Drawings of foraminifers by F.L. Parker, S.I.O., here shaded. Data: unpublished notes of W.H.B.; ages based on radiocarbon determinations)

Productivity cycles imply faunal and floral cycles, which are indeed pervasive in the ice ages. These fossil cycles invite reconstruction of temperature changes, given present biogeographic patterns. The most striking result of the effort to link eupelagic fossils mathematically to surface water temperatures is the famous 18 k map of ice-age surface temperatures, a map that was widely used as a target for computer modeling of ice-age conditions of the sea.

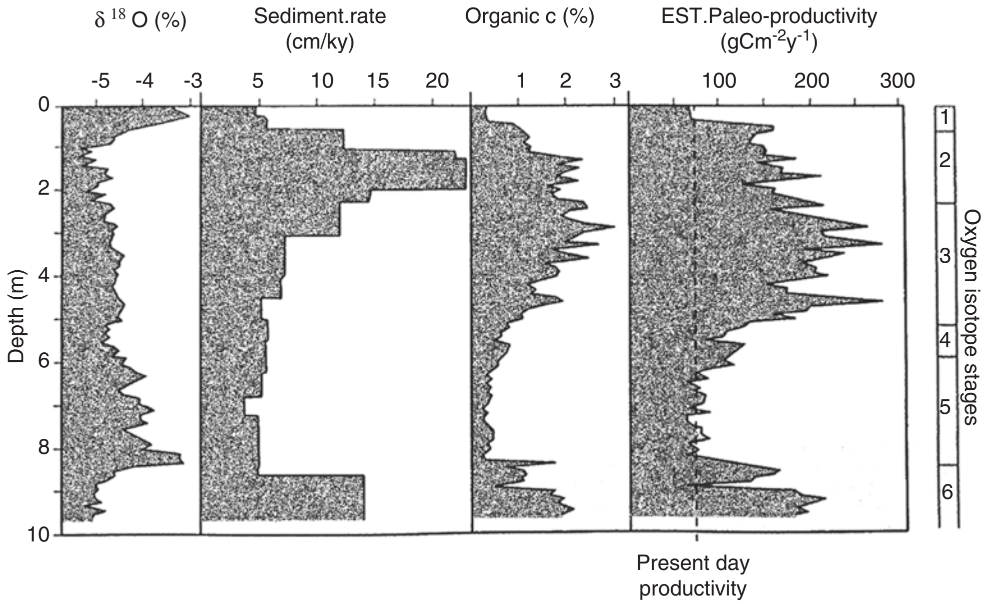

Fig. 11.8 Ice-age variation in productivity off NW Africa, based on organic matter content in Meteor Core 12,392 (P.J. Müller and E. Suess, 1979. Deep-Sea Res. 26A:1347; also P.J. Müller et al., 1983. In: J. Thiede and E. Suess, Coastal Upwelling, Part B, Plenum Press, New York)

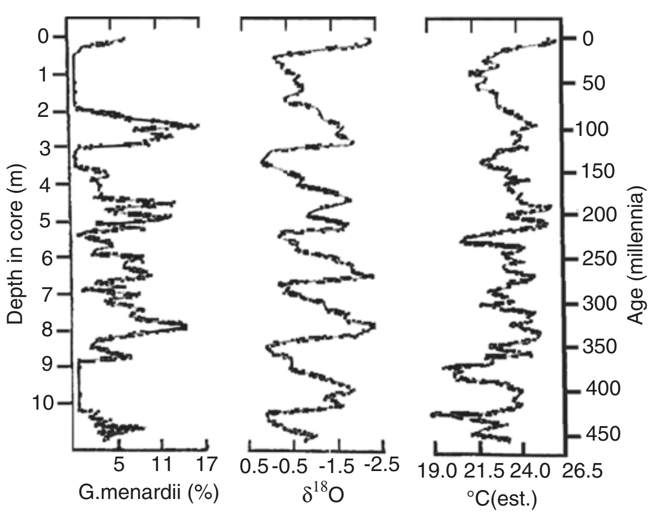

Fig. 11.9 Ice-age cycles in the deep-sea sediments of the Caribbean Sea. Left: pulses of *Globorotalia menardii*, a warm-water planktonic foraminifer that disappeared entirely from the central Atlantic during the last glacial period and also 400,000 years ago (not a cold spell); middle, oxygen isotopes in foraminifers; right, temperature estimates from abundance distributions of planktonic foraminifer shells (Imbrie et al., 1973. J. Quat. Res. 3:10)

11.4.3 Faunal and Floral Cycles and Some Open Questions

That some species are not particularly reliable as recorders of paleotemperature was established early in the experiments involving planktonic foraminifers (Fig. 11.9). In a record from the Caribbean Sea, the foraminiferal cycles (reflected in temperature estimates) rather closely follow the associated oxygen isotopes, but without duplicating them, supporting the notion that a comparison of the two curves (temperature estimate from faunal composition and from oxygen isotopes) might be useful when attempting to separate the various controlling factors. In some disagreement with expectations, one important tropical species (*G. menardii*) is missing from the Atlantic both during the last cold period and also during Emiliani Stage 11, a warm interval 400,000 years ago. Evidently there are controls still to be discovered; temperature alone seems a fickle guide to faunal change.

11.4.4 Walvis Silica Cycles

A strong worldwide increase in glacial production apparently left its mark on the chemistry of the sea: in many places glacial output of opaline shells (diatoms and radiolarians) was diminished rather than increased by high production (the *Walvis Paradox*, first described by the marine geologist L. Diester-Haass, Saarbrücken, Germany). Major changes in the chemistry of the sea also are in evidence in the variation of carbon dioxide in the atmosphere. The first description of the Diester-Haass effect was from sediments off Namibia. It was confirmed by drilling there, during ODP Leg 175, as being valid for older Pleistocene sediments studied, as well as the younger ones assessed by Diester-Haass. Apparently, productivity increased, while the supply of diatoms did not, calling for different controls on the two types of phenomena, productivity and diatom abundance.

11.4.5 Oxygen Isotope Cycles: H.C. Urey, Sam Epstein, and C. Emiliani

Oxygen isotopes, determined for the calcareous shells of planktonic and benthic foraminifers, have become the master signal of ocean history and especially of ice-age history, comparable in importance to the ocean's temperature distributions in oceanographic studies.

Aware of the central importance of temperature in the reconstruction of oceanic conditions, Cesare Emiliani introduced a concept to deep-sea studies he labeled *isotopic temperature*. It is based on the discovery by the physical chemist Harold Urey (then Chicago) that calcite ($CaCO_3$) in shells precipitated in equilibrium with seawater are enriched in ^{18}O relative to seawater but less so at higher temperatures. The equation relating temperature of precipitation to the oxygen isotopic composition of shells was found for mollusks by the Canadian-US geochemist Sam Epstein (1919–2001) and his coworkers in Chicago, including H.C. Urey. It was adopted by Emiliani for the foraminifers. The measure reported is the δ-value. It is the difference in the isotope ratios of sample and standard, as permil of the standard ("permil" is ten times percent). The standard usually is taken as "PDB," named for a now vanished belemnite from the Pee Dee formation in South Carolina that was originally analyzed by Urey's group. For the correct interpretation of the changes in δ-value, one needs to know the δ of the seawater within which the shell was precipitated. This value is usually not known and must be guessed at, before the temperature of precipitation can be calculated.

Emiliani was well aware of this effect, as well as of other complications in pursuing his measure of "isotopic temperature." In his initial studies, he carefully listed the various problems that interfere with reading the oxygen isotope record in terms of the history of temperature (the ice effect, geographic variation in isotopic composition of seawater, vital effects, seasonal growth of shells, and growth of foraminifers at various depths in the water). But once he settled on the two preferred planktonic species for analysis (*G. rubra* and *G. sacculifer*), he presented his data as indices of "isotopic temperature" in his graphs. Thus, he implied that the various other interfering factors can be captured by a single proportion. His assessment that the ice effect (and others) may be taken as a constant proportion of the overall change in oxygen isotopes was widely adopted. However, his estimate of the ice effect was too low, making his guesses for temperature differences automatically too high. The correct value for the ice effect was first suggested by the Swedish marine geologist Eric Olausson (1923–2010) and subsequently confirmed by the British geophysicist N. Shackleton (1937–2006). The delta value for the sea is taken to be near 1 permil (Fig. 1.5).

Emiliani's studies introduced fundamentally new ways of reconstructing ocean history. Unfortunately, however, the time scale he employed was quite incorrect, just like his guess on the ice effect was. Valid cycles with a (nearly) correct interpretation were first presented not by Emiliani using samples obtained from Hans Pettersson (leader of the *Albatross* Expedition) but by Nicholas Shackleton and the Lamont geophysicist Neil Opdyke, who used a core taken by Lamont's research vessel *Vema* and employed paleomagnetics for dating by correlation (Fig. 6.10).

The link to Milankovitch Theory that emerged once the time scale for the oxygen isotope variations became a lot closer to the truth than the one offered by Emiliani represented strong evidence for the theory and invited its use for precise dating. Milankovitch Theory, by providing accurate wavelengths for the climate cycles contained in the deep-sea record, delivered a standard sequence based largely on averages of oxygen isotope series in several cores from the deep Atlantic that has remained useful for Pleistocene sediments back to 650,000 years since first proposed in 1984 by John Imbrie and associates (Fig. 11.10a). Subsequently, it was shown (by N. Shackleton, A. Berger, and R. Peltier) that Milankovitch Theory is applicable in finding an age for the Brunhes-Matuyama magnetic reversal. In a very convincing demonstration, these scientists produced an age for the beginning of the Brunhes Chron identical to one obtained by radiochemistry. The discovery allows for revision and extension of the standard isotope curve for the Quaternary ice-age fluctuations (Fig. 11.10).

Despite the successful application of Milankovitch Theory, it is important to realize that the theory is not in fact capable of describing all of the observed variations in ice mass and temperature in simple fashion. For clarity: the Theory may be unexcelled as a dating tool, but it is incomplete as a mechanism for explaining ice ages.

11.4.6 Sea-Level Cycles and Limiting Feedbacks

Assuming that phase shifts between forcing and response are constant, we can use the detailed information from the timing of the forcing to reconstruct rates in sea-level change. One typically obtains maximum rise rates between 2 and 3 m per century (the 3 m/century value being extremely rare). Falling sea level moves more slowly than rising one, suggesting a contribution to (fast) sea-level rise from instability in the ice (see also Chap. 6). The fastest sustained rises of sea level are seen within the times of glacial *termination* (= *deglaciation*; = major melting), that is, whenever the large northern ice masses on North America and Scandinavia were melting vigorously (see section on *deglaciation*). The last one of these ice-age events ended about 7000 years ago and took almost 10,000 years. The rise of sea level was approximately 125 m, that is, the average rise rate, sustained for many millennia, was just above 1 m per century (Fig. 11.5). For any one millennium within that set, of course, an average rise per century could have been much greater than the typical millennial value. Likewise, for any one century, within the ten centuries making up a millennium within deglaciation, the rise rate could have greatly exceeded the average. The available data do not show such short-term pulses.

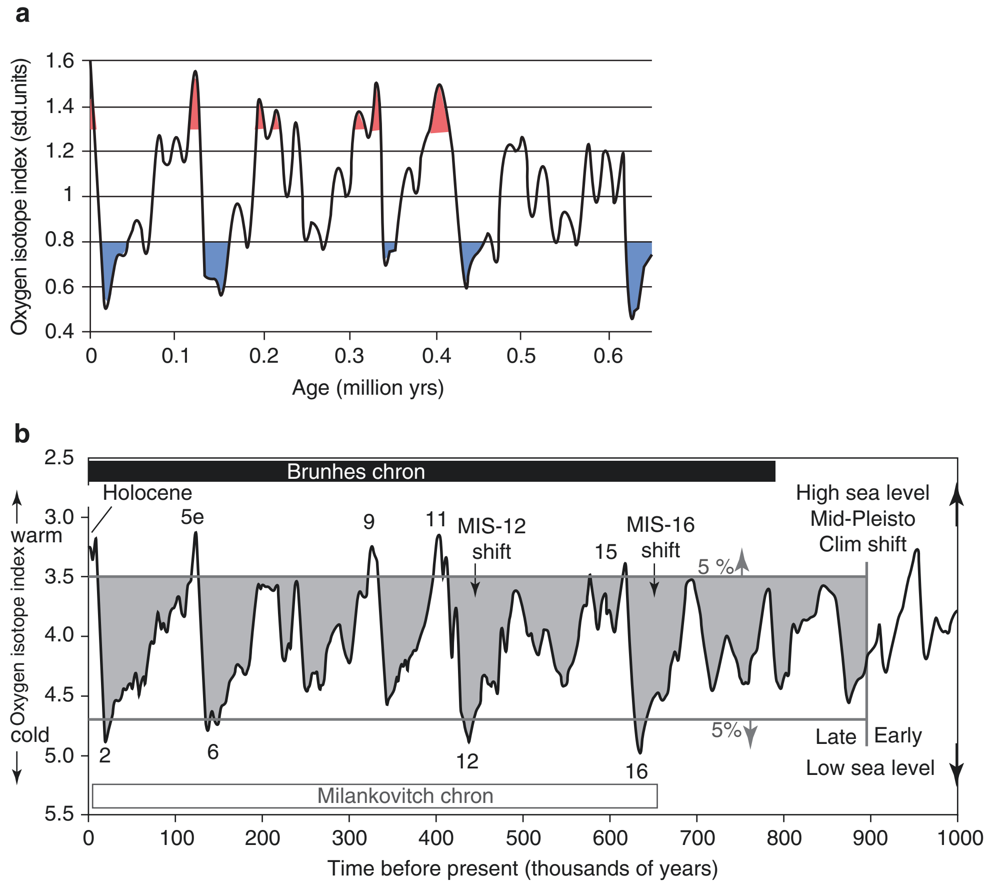

Fig. 11.10 Quaternary isotope standards based on Milankovitch dating. (**a**) Useful portion of the Imbrie et al. (1984) data ("SPECMAP"; *red*, warm peaks; *blue*, cold peaks); the Holocene appears to be missing. (**b**) . Revision and extension (to 1 million years) of the "SPECMAP" standard based on subsequent isotope data. (Holocene added from box-core information.) Numbers are Emiliani stages. Large ice-age events are followed by shifts in climate systems

In evaluating the confidence to be placed in the results of this statistical exercise (and others like it), it is well to remember that the Milankovitch Chron (the time span studied by Milankovitch, the last third of the Pleistocene) has plenty of major melt events. There are only very few of those in the earlier portion of the ice-age record.

What is intriguing is that there seems to be strong resistance in the ice-age climate system to sea level dropping below a certain maximum low-level zone (Fig. 11.11; also see Fig. 6.10). Upward boundaries are less well defined, but there does seem to be a resistance for late Pleistocene sea level to climb above a certain high level zone, here interpreted as evidence for negative feedback at the boundary for the "normal" range. The suggestion is that the onset of strong negative feedback at both lower and upper boundary zones kept ice mass variations within more or less regular limits in the last million years. To what degree such observations can be applied to present concerns is unknown, of course. The ice ages offer illustrations of response to millennial-scale changes in solar forcing; one cannot assume that they also reflect fast response (on the human scale) to fast changes in external forcing. Applying findings in one-time scale to another always involves much guesswork unless all details are known and amenable to calculation.

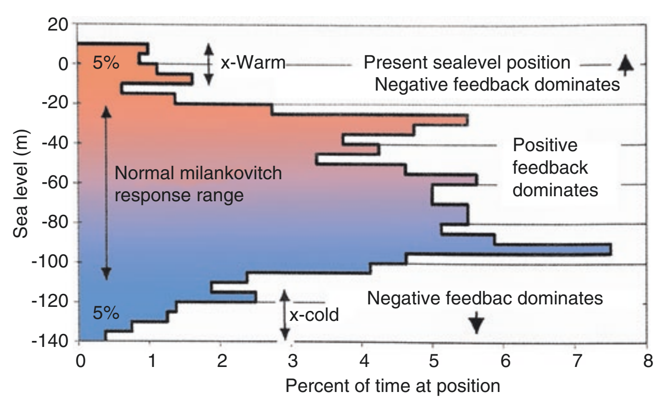

Fig. 11.11 Histogram of estimated sea-level positions for the last million years, based on oxygen isotope measurements in foraminifers of deep-sea sediments. Note maximum position near −100 m (W.H.B., 2008. Int. J. Earth Sci. 97:1143; color here added)

11.4.7 Carbon Cycles

It may seem obvious that ice-age cycles should generate carbon cycles. The link is implied in the propositions re the Arrhenius carbonate cycle and in the Mueller-Suess measurement of variations in productivity (the latter measured as grams C per square meter per year). (The marine component in the variations of organic matter within sediments in principle is accessible through *biomarkers*, that is, organic substances bearing on the origin of the organic carbon within sediment.) In addition, there are the carbonate cycles, presumably strongly influenced by the changing availability of shelves, the preferred sites of carbonate deposition. A move of carbon back and forth between organic matter dominance in glacial periods and carbonate dominance on a shallow seafloor during warmish interglacials must influence changes in atmospheric carbon dioxide (although to determine the size of the signal independent of ice core data is extremely difficult and yields unreliable information).

Small changes in the chemistry of the sea, leading to a small proportional change of carbon chemistry in the ocean, can potentially have large effects on the atmosphere, since the ocean reservoir available for exchange with the air is relatively very large. The question is how to track such changes in the marine realm. One commonly used proxy is the $\delta^{13}C$ signal, describing the changing ratios between the isotopes ^{12}C and ^{13}C. The $\delta^{13}C$ signal emerges together with $\delta^{18}O$ when analyzing carbonate. Biological pumping tends to remove the ^{12}C slightly more efficiently from the photic zone than ^{13}C, because ^{12}C is more readily incorporated into organic matter during photosynthesis than is ^{13}C.

The effect of the slight difference in reactivity is that *carbon-13 (or ^{13}C) is enriched in surface waters* relative to deep waters. Thus, the $\delta^{13}C$ (which is recorded in carbonate plankton) is a measure of the intensity of the pumping, as pointed out by the Lamont geologist and chemical oceanographer W.S. Broecker several decades ago. The maximum $\delta^{13}C$ value in any period depends on the nutrient content of deep-ocean waters, which controls the amount of isotopic fractionation that can be achieved.

11.5 Deep-Ocean Drilling and the Ice-Age Target

11.5.1 Advantages of Drilling

While much of the material recovered for ice-age studies in deep-sea sediments was recovered by piston coring, the contributions from drilling were momentous. Technically, they involved obtaining excellent records after piston coring was adapted to drilling by taking cores ahead of the rotating pipe end, before the sediment was touched by the drill. There are several additional reasons that drilling became central to ice-age studies. One is that drilling allows the sampling of the entire ice-age sequence, starting with the onset in the middle of the Pliocene. In areas of high sedimentation rate, where the record is potentially very promising, plain piston coring may retrieve a record that falls short of completeness. Any drilling into the seafloor, of course, likely has to penetrate ice-age sediments to get to a target older than the ice ages. Thus, ice-age sediments became the chief product of deep-sea drilling, not necessarily with explicit intent.

The main result of studying ice-age sediments was a deeper understanding of climate change and the role of the ocean in it. This type of understanding became ever more desirable as the importance of climate models gained momentum in the discussion of global warming.

A rather pragmatic approach to the advantages of drilling was illustrated by Cesare Emiliani. He pointed out that on land confusion is likely to reign with regard to ice-age history thanks to the prevalence of erosion and especially owing to the fact that subsequent ice-age cycles tend to destroy the evidence left by previous ones. In contrast, in the deep sea, one might expect a long undisturbed record of the ice ages, with drilling even recovering the evidence for the onset of ice ages in routine fashion. The onset, of course, follows a general cooling. It is part of Cenozoic history therefore, and will be discussed in the chapter that follows.

11.5.2 The Mid-Pleistocene Revolution and Milankovitch Theory

As mentioned, a long ice-age record is a common product of drilling. One implication is that the study of lengthy time series from the deep seafloor ceased to be a privilege of successful piston coring and became routine instead. Through such routine work, it was confirmed in the western equatorial Pacific that some 900,000 years ago, the ice-age cycles changed rather drastically (Fig. 11.12). While the nature of the ice-age cycles changed, that of the astronomic forcing remained as before. In addition to the appearance of seemingly unforced long cycles arising near 700,000 years ago, there is another related enigma: variations in the quality of response to forcing, which may be addressed as changes in the quality of listening of Earth's climate system to Milankovitch forcing in general. There seem to be periods of defective listening, and they are as yet poorly documented or understood.

The Mid-Pleistocene Climate Shift (*MPR* in Fig. 11.10) that separates late Pleistocene long climate oscillations larger than 70,000 years from shorter ones before about 900,000 years ago, dominated by obliquity cycles (41,000 years), should be seen in acoustic profiles, since the echo structure is bound to change at that level within the sediment. There are indeed indications that this is so, in slope sediments off Angola and Namibia (cf. Fig. 3.6). The phenomenon fitting expectation was discovered during preparations for drilling during ODP Leg 175. Quite generally, drilling has had benefits through expanding the need for preparation (and thus exploration of poorly known seafloor conditions).

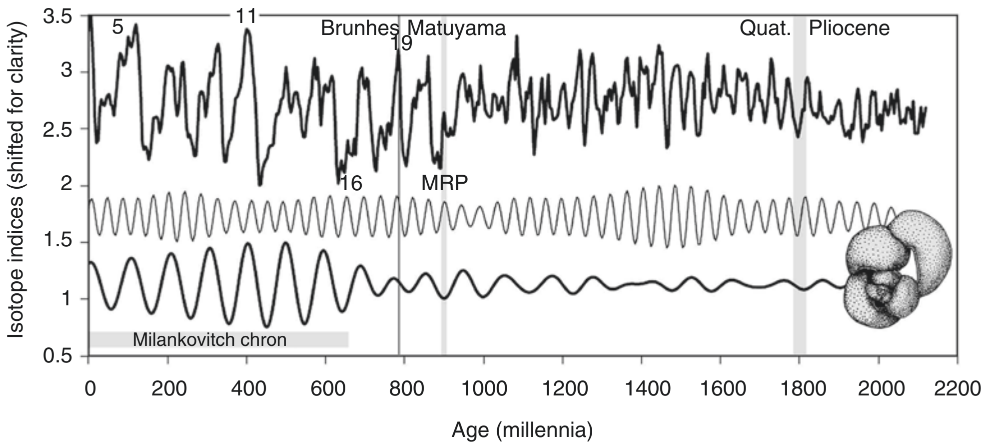

Fig. 11.12 Oxygen isotope record of the planktonic foraminifer *G. sacculifer*, ODP Site 806, western equatorial Pacific, 41,000 y- and 100,000 cycles extracted by Fourier mathematics. 5, 11, 16; Emiliani isotopic stages. The Milankovitch Chron starts with Stage 16 (Marine Isotope Stage 16). It is dominated by 100,000-year cycles, as shown. The Brunhes Chron starts with Stage 19. The *MPR* (mid-Pleistocene Revolution) is at Stage 22, apparently the first very large glaciation in the Quaternary (W.H. B. and G. Wefer, 1992. Naturwissenschaften 79:541)

11.6 Deglaciation

11.6.1 Background

The discovery of the relatively young age of the last glacial maximum implied a short time span for moving from glacial conditions into postglacial ones (Fig. 6.5). It is an achievement that owes much to radiocarbon dating of deep-sea sediments during pioneer time (1930s–1980s), with strong connections into the study of oxygen isotopes in foraminifers. Naturally, the great mass of meltwater delivered during transition time (some in tremendous floods such as the famous Columbia River Flood studied by the U. Washington geologist Harlan Bretz in the 1920's) had implications for the ocean's stratification and circulation, which stimulated much discussion and speculation. Discussions on this topic may have lost vigor in the past few years, with other issues taking center stage, but the problems identified half a century ago are by no means solved.

The last deglaciation is but one example (albeit the closest one in time) of a dozen rapid climate change events associated with major melting in the last million years. The appearance of these events (called "terminations" by W. Broecker; Fig. 11.13) presumably signaled an increase in potentially unstable ice mass. In any case, W. Broecker and his student J. van Donk in 1970 boldly drew the fast deglaciation events ("terminations") on top of Emiliani's isotope stratigraphy, thus introducing this very fruitful concept into the thinking about the ice ages.

11.6.2 The Younger Dryas

One major enigma arising in the last deglaciation is the problem of the Younger Dryas cold episode (the name is taken from a plant fossil, an Arctic flower, Fig. 11.14). The episode is a 1000-year relapse into glacial conditions in the northern hemisphere first seen in Greenland ice cores studied by the Danish physicist Willi Dansgaard (1922–2011). The Younger Dryas occurred in the middle of the last deglaciation interval, which followed considerable warming. The reversal of postglacial warming and its halt during the Younger Dryas has led to much discussion, including speculations about changes in deep circulation. Many of the suggestions regarding the effect of the Younger Dryas on the deep-sea environment are based on information from the chemical stratigraphy of benthic foraminifers (the shells of which carry clues about ventilation of deep waters in their chemistry). Other constraints are based on modeling of climate change and inferred responses of the circulation.

In recent years the problem of the Younger Dryas cold spell has been linked to a postulated impact from space some 12,800 years ago. An impact event could perhaps explain drastic change at the time of the onset of the Younger Dryas, including megafauna extinctions, which have been linked to both climate change and to human over-hunting. But an impact too (like so many other possible explanations) would leave unexplained the preceding drastic warming of the first melting step (the "Alleroed" warm spell that presumably started the melting and thus the "deglaciation" process). As long as the Alleroed warming is unexplained, it seems difficult to claim that the deglaciation record is "understood."

11.6.3 On Sudden Mammoth Extinction and Other Unexplained Deglaciation Topics

Discussions of possible reasons for the extinction of the mammoth and other large mammals (also large birds) have

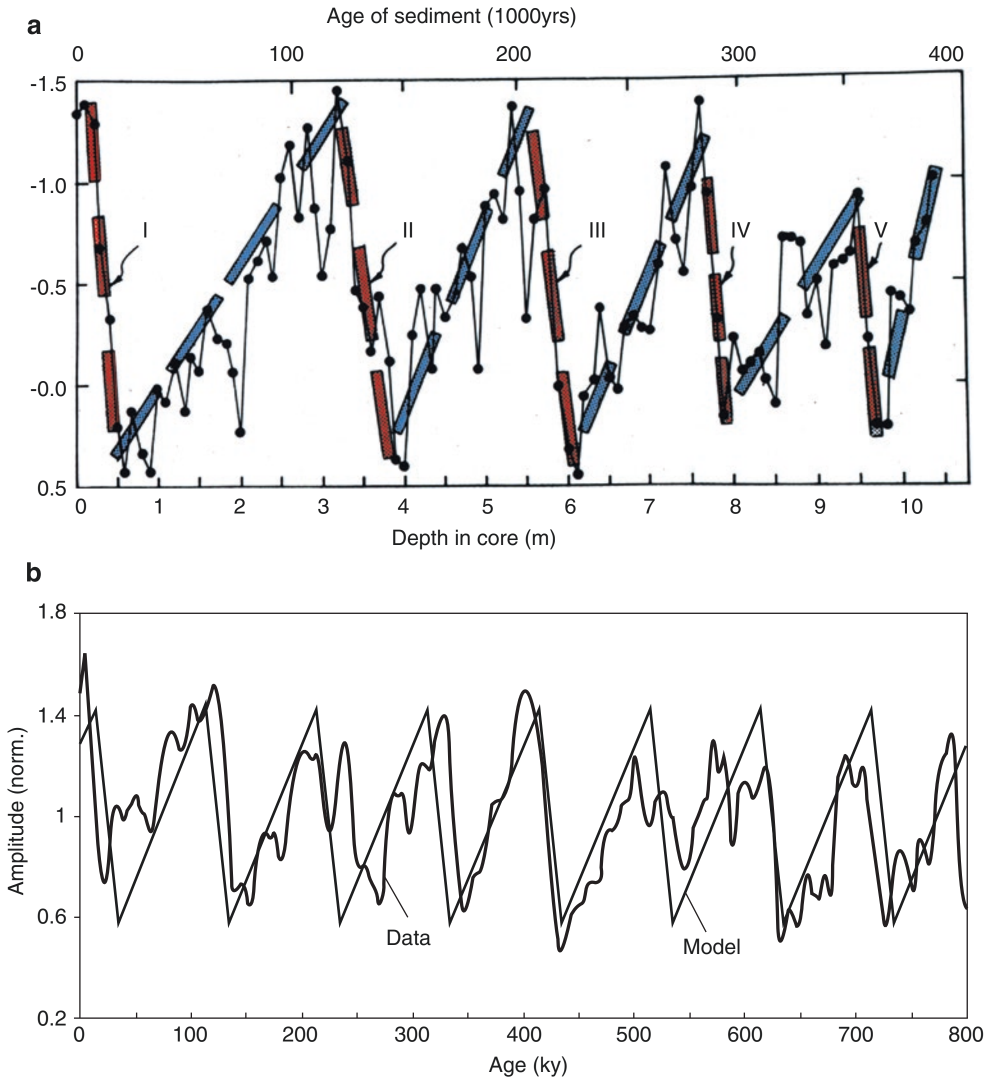

Fig. 11.13 (**a**): Terminations I to V in the sawtooth interpretation of Emiliani isotope data by W.S. Broecker and J. van Donk, in 1970. [Rev. Geophys. Space Phys. 8:169] *Red*: fast melting; *blue*: buildup of ice; colors here added. (**b**): Remarkable agreement between data series (ODP Site 804) and a simple termination model with a period of 100 ky (W.H.B., presentation in Berne, 1999) Phase difference between data series and termination model was set to zero at an age of 400 ky

been vigorous among geologists after Baron Cuvier showed that the mammoth is extinct (i.e., that extinction is for real).

In this context (as well as other ones related to abrupt change in climate) climatologist geologists would like to know why the melting of the last deglaciation started when it did and what precisely controlled the rate of destruction of glacial period ice masses. The assumption is that a rise in temperature, while important, is unlikely to be the whole story. The fact that *terminations* exist suggests that large ice masses can become unstable. Dansgaard-Oeschger Oscillations and Heinrich Events with their evidence for sudden cooling appear to have a similar message. The notion of unstable ice is not new, actually: regular instability of ice has been urged by the US glaciologist T. Hughes (University of Maine) for many decades. "Heinrich Events" (pulsed IRD delivery), among other evidence, seem to bear him out.

Modern concerns revolve around the striking presence of methane ice, much of it below sea level, in high-latitude northern regions (known as "permafrost" and in the past stable enough on land for bearing houses and roads and telegraph poles). Melting such ice releases the powerful greenhouse gas methane (exceeding effects from CO_2 on a century scale by more than 25 times). Some of the methane may escape destruction by oxidation for a number of years. In historical marine geology, the methane problem appears when discussing an uncommonly strong spike of ^{12}C-rich carbonates at the end of the Paleocene, a time of maximum warmth, within the early Cenozoic. The Cenozoic is the topic of the chapter that follows.

11.6.4 Mediterranean Sapropels

The Swedish *Albatross* Expedition retrieved a large number of cores in the Mediterranean Sea, many of which contain the black organic-rich layers called *sapropels*. Black sediments commonly are addressed as "sapropels," a term

Fig. 11.14 *Dryas octopetala*, the Arctic flower that gives its name to the 1000-year cold spell during the last deglaciation, the "Younger Dryas" (implying the existence of an "Older Dryas") (Photo W. H.B., taken near Joestedal Glacier, Norway)

related to the Greek word for rottenness and thus presumably referring to a disagreeable smell involving hydrogen sulfide and related compounds that can be associated with an oxygen-free environment. Burrowing commonly was suppressed when the dark layers were deposited (Chaps. 8 and 13).

The debate about origins of these sapropel layers has focused largely on the question of whether the organic-rich layers in the Mediterranean reflect a lack of oxygen (at depth) or a marked spike in production (within surface layers). There is some doubt that these are separable causes. The responsible factors may be aspects of the same change in overall circulation within the basin, that is, a shift from anti-estuarine to estuarine deep circulation lasting several thousand years. Identifying climate change (precipitation, freshwater influx) as the main factor links the sapropel origin there to the oceanography of deglaciation. The spacing of sapropel layers provides a base for an astronomical time scale back into the Pliocene, as documented by the stratigrapher Frederik Hilgen in Utrecht. Drilling recovered material there that allowed extending the sapropelic orbital time scale deep into the Tertiary.

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Cenozoic History from Deep-Ocean Drilling

12

12.1 Elementary Considerations and Leitmotif

12.1.1 Cooling: The Leitmotif of the Tertiary

Our planet has plenty of ice in high latitudes (Fig. 12.1). In fact, the central theme of climatic evolution in the *Cenozoic* (the time after the demise of ammonites and dinosaurs) is an overall cooling (starting with the Eocene and culminating in the northern ice ages) and an associated overall drop of sea level (making ice on land takes water; sea ice does not change sea level: the water stays in the ocean). Sea ice, like land ice, does affect albedo, though, by providing a base for snow.

The overall cooling trend has been known for some time from the study of fossils on land. It is reflected in the evolution of animals and plants whose offspring exists today. Modern evidence comes from oxygen isotopes of benthic foraminifers in deep-sea sediments recovered by drilling (Fig. 3 in Preface). It has become general knowledge, along with an appreciation for increased climate variation with ice buildup. Cooling occurred largely in steps, presumably a reflection mainly of sudden and large changes in albedo whenever snow fields or sea ice underwent notable change in extent, invading new areas (Fig. 1.7).

Changes in ocean productivity in the Cenozoic presumably resulted from intensification of the wind along with a change in the nature of the thermocline. The thermocline, the boundary between warm and cold water, became shallower and better defined as the cold water layer at depth grew thicker. A shallow and distinct thermocline means decreased productivity over large oceanic areas, with low nutrient content in surface waters. In contrast, coastal ocean productivity increased noticeably in the Neogene, especially the late Miocene. There was an increase in the range of productivity, with large areas becoming deserts and opposite extremes focused within upwelling regions. The result of a strong and shallow thermocline presumably is an enhanced depth stratification of plankton, a development noted in sediments since the beginning of the Neogene right after the somewhat chaotic Oligocene.

12.1.2 Thermocline and Diversity: Problem Corollaries of a General Cooling

Strengthened winds (polar ice increases the temperature gradients that drive winds) presumably increase upwelling, which (it is thought) lowers diversity locally. Thus, while we can be sure that there is a relationship between an overall cooling and a change in diversity of plankton and other organisms, we have problems specifying details on a regional level. The available information is not unambiguous, with the desert specialists (the nannofossils) having an evolutionary trajectory quite different from those of diatoms and foraminifers. In both of these forms, abundance and diversity seem to increase during the Cenozoic, a statement not applicable to nannofossils and to radiolarians (Fig. 12.2).

12.1.3 Evidence from Oxygen Isotopes

Evidence for the great cooling trend in deep-sea sediments was first provided by the oxygen isotopes in benthic foraminifers sampled by deep-sea drilling (Fig. 12.3). The relevant graph by Ken Miller (Rutgers University) and associates (published in 1987) has been referred to hundreds of times. A similar compilation by J. Zachos and associates (global, rather than Atlantic focused; published in 2001) shows that the major features recognized in the Atlantic by Miller et al. are in fact global in nature. Both compilations show two remarkable ramps of cooling lasting for more than 10 million years – one in the Eocene and the other within the Miocene. In each case, at the end of the gradual cooling, there is a striking acceleration, as though the cooling reached a critical level resulting in vast ice buildup and albedo change (positive feedback). We might guess that an end-of-Eocene ice buildup

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_12

Fig. 12.1 Ice in high latitudes. *Left*: broken off shelf ice off the Antarctic continent (Gerlache Strait); *right*: iceberg calved from a mountain glacier with tidal termination in a fjord in eastern Greenland (Photos W.H.B.)

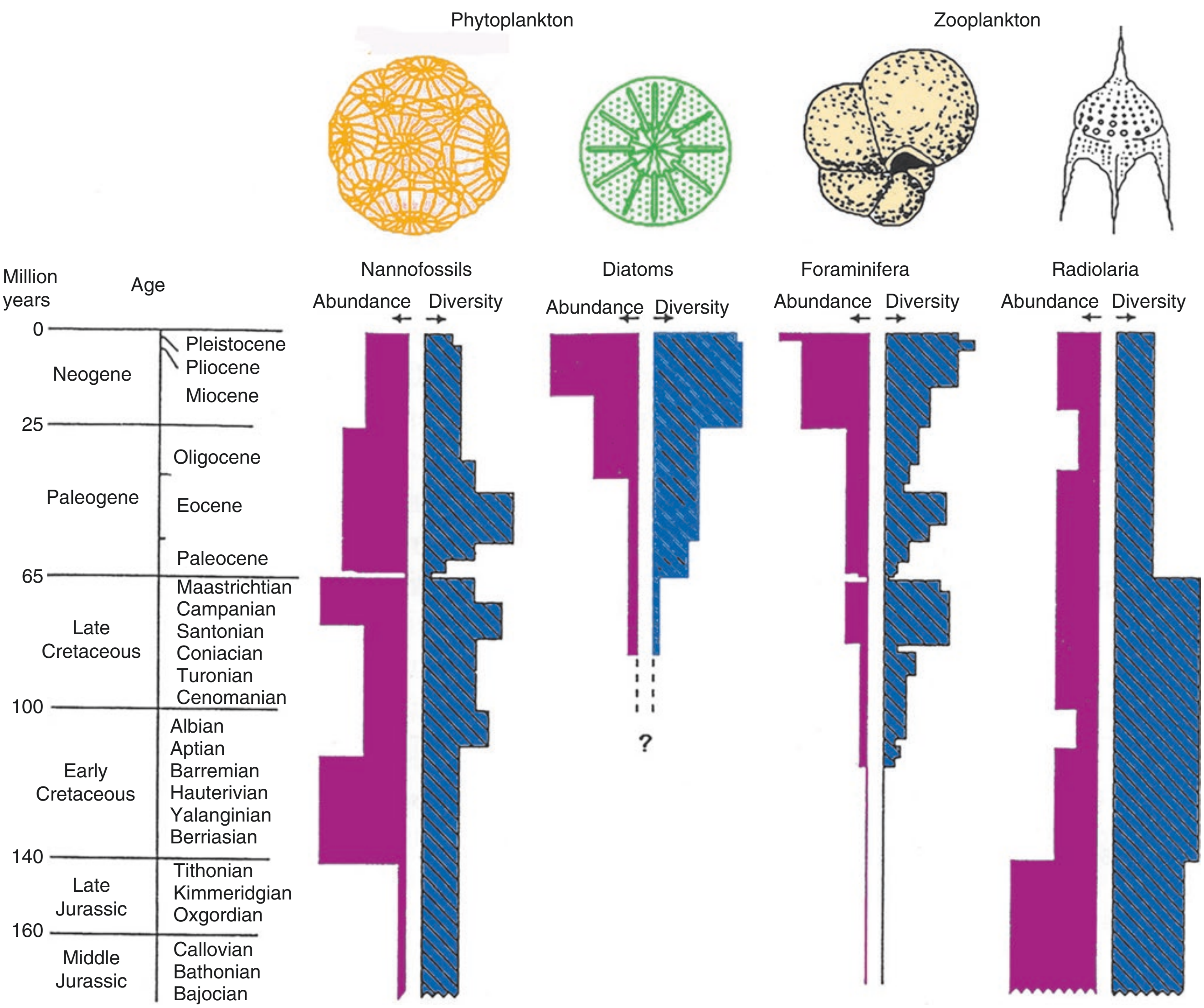

Fig. 12.2 The plankton organism diatoms and foraminifers show a strong increase in diversity within the Cenozoic, along with increased abundance. Nannofossils and radiolarians (likewise shelled plankton) do not show those trends (H.R. Thierstein et al., 1988. In: G.B. Munsch (ed.) Report of the Second Conference on Scientific Ocean Drilling COSOD II. European Science Foundation, Strasbourg; color here added)

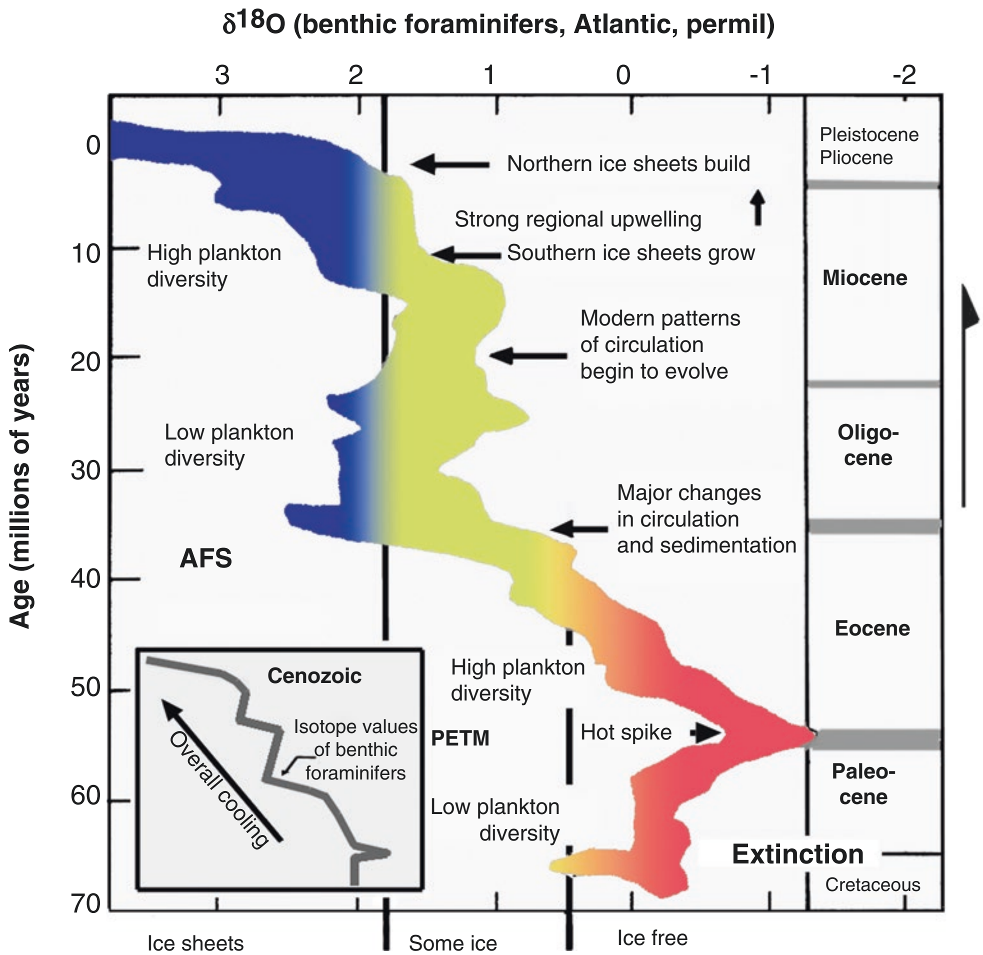

Fig. 12.3 The Cenozoic temperature history as seen in oxygen isotopes of benthic foraminifers in material recovered by deep-sea drilling. Modified after K. Miller, R.G. Fairbanks, and G.S. Mountain, 1987. [Paleoceanography 2: 1] (Compare Fig. 3; colored band: generalized distribution of data point scatter) *AFS* Auversian facies shift: drastic change in shelf sedimentation in the latest Eocene, *PETM* Paleocene–Eocene Temperature Maximum, presumably linked to sudden methane release. *Blue* color: ice present on the planet; *red* color, planet ice-free

took place rather readily on Antarctica, as the place where one of the poles is located centrally on an enormous continent with cold winters during the dark season. The continental ice buildup of the late Neogene, around the Arctic Sea, may be considered a lot more difficult and hesitant; the solid base for land ice is being removed from the pole. A mid-Miocene ice buildup is commonly assigned to Antarctica, the IRD (ice-rafted debris) in northern latitudes being as yet rare or missing at the time.

The essential trends and steps within the general trends were worked out soon after the Deep-Sea Drilling Project started, by two teams combining the skills of paleontology and isotopic geochemistry. The pioneers were the US geologists R. Douglas and S. Savin and the NZ-US geologist J. Kennett with the British geophysicist and geochemist N. Shackleton (later Sir Nicholas). Their observations showed that cooling was especially vigorous in high latitudes, being obvious in benthic species (bathed by high-latitude bottom water) and in planktonic species from the subantarctic realm. Also, their data indicated that the cooling occurred in a few major steps, suggesting the operation of nonlinear feedback mechanisms (i.e., the uneven response to forcing and the action of positive feedback are indicated, such as expected when freezing is involved).

12.1.4 Evidence from Strontium Isotope Stratigraphy

An important aspect of the overall cooling trend, and one that touches on the search for causes of the cooling, is the stratigraphy of strontium isotopes. Chemically, strontium is an element homologous to calcium, that is, it acts somewhat like calcium and is present in calcium carbonate, therefore (even if not very abundantly). This means it can be found in calcareous fossils, whether they are remains of benthic organisms or pelagic ones, shallow or deep. The ratio between the isotopes ^{87}Sr and ^{86}Sr (both less than 10% of naturally occurring Sr) reflects (among other items) changes in the sources supplying strontium to the sea. The common sources are erosion of continents (i.e., river input) and supply from volcanic emissions. The long-term trend

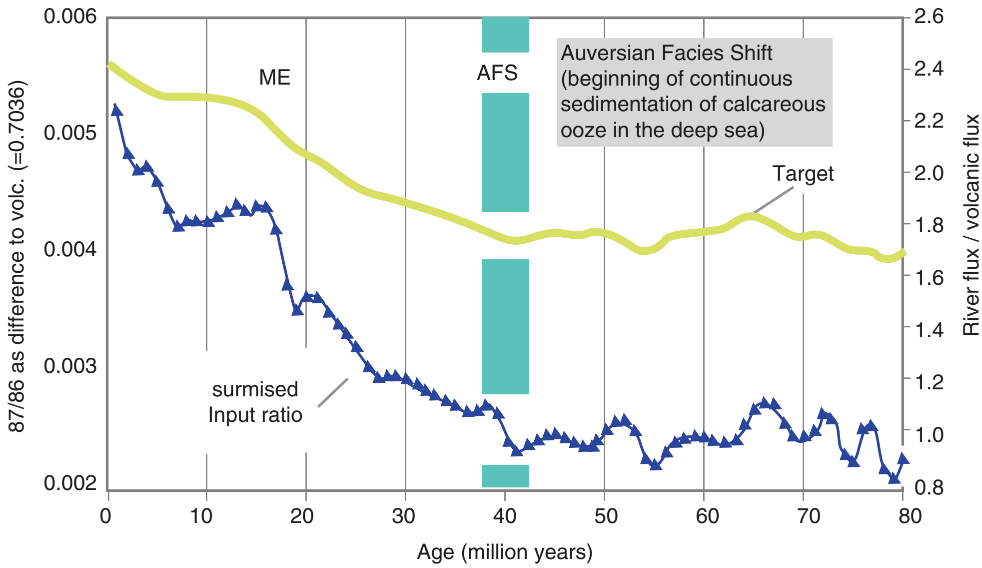

Fig. 12.4 Cenozoic stratigraphy of the Sr^{87}/Sr^{86} ratio. The curve of the "input ratio" assumes a linear mixing of endpoint Sr isotope values and a 4 million year residence time (W.H.B. and G. Wefer, 1996, modified. In: G. Wefer et al. (eds.), the South Atlantic. Springer, Berlin and Heidelberg. Target data from (H. Elderfield, 1986. Palaeogeogr., Palaeoclimat., Palaeoecol.,57)) *AFS* Auversian facies shift, a drastic change in marine sediment types. *ME* Monterey Event, major shift in nutrient distribution and increase in coastal upwelling

in the ratio shows a steadily increasing influence of continental erosion relative to volcanic activity, as expected for cooling (Fig. 12.4). The trend distinctly accelerated about 40 million years ago and again 16 million years ago. Erosion of continents, of course, is tied to uplift and regression (water-covered continents are not eroded). But there is evidence that volcanic activity changed also throughout the Cenozoic, complicating the interpretation of shifts in the strontium isotope ratios.

When focusing on cooling, a change in volcanism is not helpful: It potentially produces a change in ratios of Sr isotopes that is independent of cooling. ("Potentially" independent because ice buildup results in "loading" of continental crust which presumably stimulates volcanic activity, which would have opposed any evidence for increased erosion of continental crust.) As far as the observed trend, it pretty much is as expected for cooling. It accelerates from the end of the Eocene to about 16 Ma, an interesting period when the overall tendency of cooling may have given way to signals from increased volcanism (J.P. Kennett published Neogene DSDP ash layer abundances in 1982, p.384). The overall trend expected from cooling then starts up again at about 7 Ma, a time of major uplift in the Asian highlands (Tibet) and a volcanically relatively quiet period. Sea level apparently dropped low enough to help isolate the Mediterranean from the world ocean. (Thus, cooling pulses may have been an important cause for a repeated drying up of that sea at the end of the Miocene.) Toward the end of the Neogene, some 4 million years ago, there are indications of major volcanic activity again in DSDP cores.

The apparent long-term increase in weathering of continental crust suggests a long-term geological trend in the reduction of carbon dioxide, by the *Urey mechanism* (=absorption of the carbon in carbon dioxide into carbonate when weathering calcium-rich silicates). To this mechanism, we must add burial of organic carbon when considering the abundance of carbon dioxide. The removal of carbon dioxide from the atmosphere engenders cooling, according to atmospheric physics. The direct effect can be measured, but the total effect on climate change is difficult to assess, because of various feedback processes. The least understood of these feedbacks involves cloud formation, which has important implications for the evolution of planetary albedo.

12.2 Reconstruction of Conditions in the "Tertiary" (or *Cenozoic*)

12.2.1 On the Crucial Role of Deep Ocean Drilling and Biostratigraphy

Deep-ocean drilling (Fig. 12.5), beginning with the Deep-Sea Drilling Project (DSDP, *Glomar Challenger*) and continued with ODP (and a new vessel, dubbed *JOIDES Resolution*), profoundly changed the understanding of geologic history for the last 100 million years. The grand steps of cooling in the Cenozoic and the anaerobic periods in the Cretaceous were among the most stupendous findings of scientific deep-ocean drilling. The steps were a surprise for most geologists, as was the fact that the cooling is mainly a high-latitude phenomenon. These were, however, only two of several major discoveries based on using the technology of drilling, a technology borrowed from the oil industry and adapted for academic use.

Without assignment of ages and environmental interpretation after inspection of microfossils and nannofossils, the

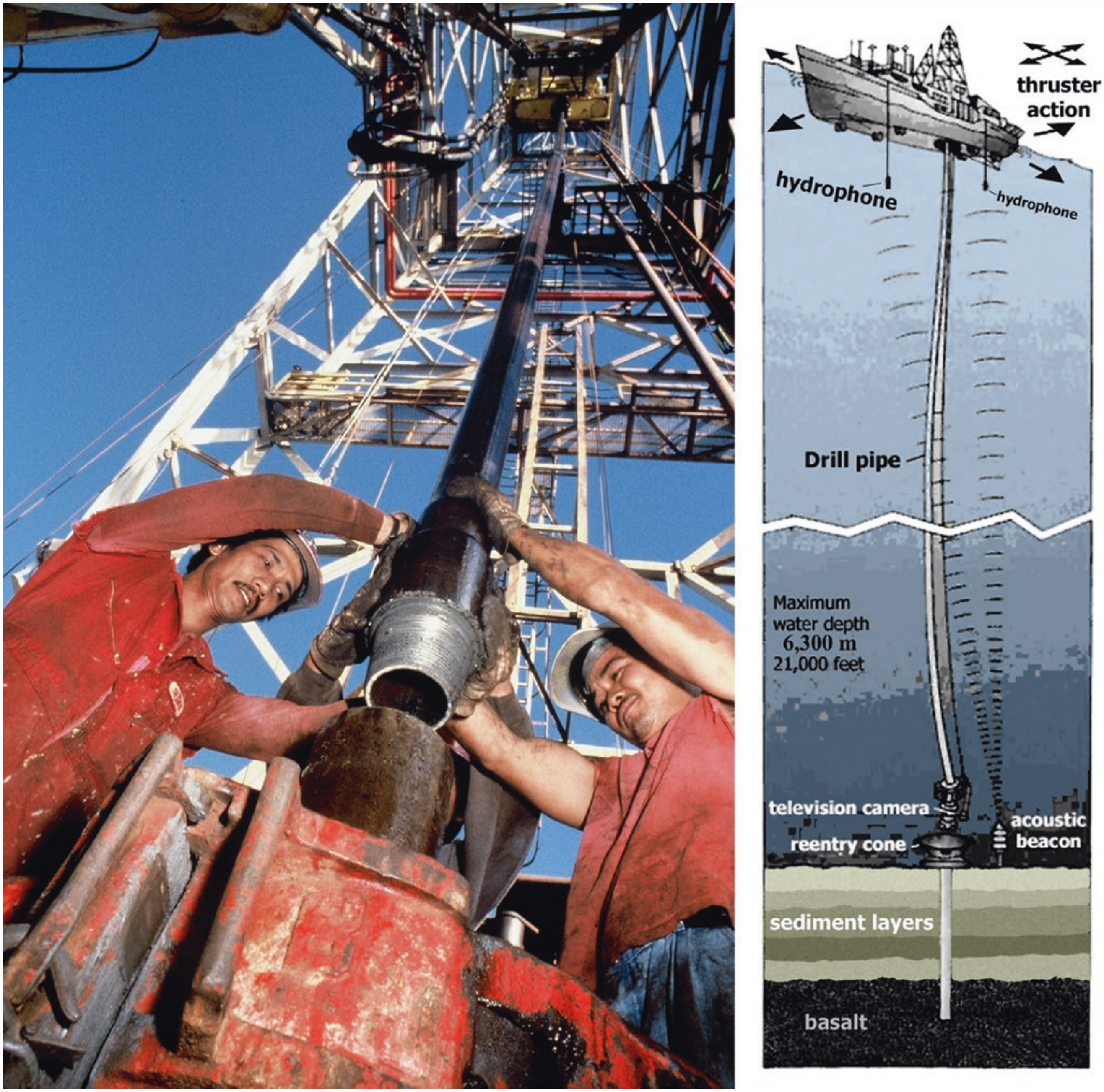

Fig. 12.5 Retrieving information on Cenozoic history from the seafloor. The chief source of information is drilling from a vessel (here: the JOIDES Resolution is a vessel originally used by BP in commercial drilling and greatly modified for scientific drilling) (Source: *ODP* Ocean Drilling Program, College Station, Texas)

available information would not have yielded the relevant insights: To some degree, the fossils (Figs. 4.7 and 12.6) allow keeping track of the changing environment, thanks to evolution, that is, thanks to the fact that they are the remains of organisms adapted to their habitat. Much more reliably, however, they allow assignment of geologic ages because they differ from one period to the next. Age control is greatly refined and turned into absolute age numbers, using correlations into radioactively dated land sections, correlated with sediments on the seafloor based on magnetic reversal stratigraphy and to some extent on isotope stratigraphy, that is, changes in physical and chemical changes in sediment properties, providing probable ages for biological changes observed.

The major cooling steps identified tend to be associated with major changes in both the marine flora and fauna. A very

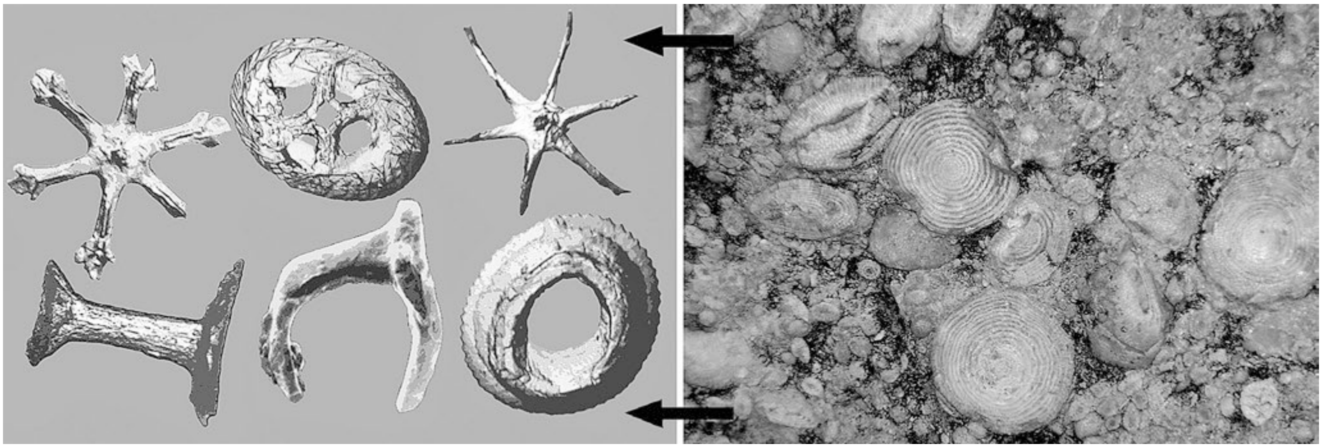

Fig. 12.6 The great change at the end of the Eocene. As sea level dropped, many benthic shallow-water foraminifers (Girona, Photo W.H.B.) went extinct. *Left*: nannofossils (examples retrieved during Leg 1 of DSDP, J.D. Bukry and M.N. Bramlette, 1969)

big change was at the end of the Eocene (Fig. 12.6). There was a general dying off of ancient benthic species, as the water became too cold for them (or else shallow-water forms on shelves found themselves without any water). Planktonic species also suffered: a rich Eocene flora and fauna disappeared, and a rather dull Oligocene one appeared, with comparatively little diversity. At times, in the Oligocene, there were enormous blooms of *Braarudosphaera*, a nannoplankton genus. It is a form now mainly found in stressful conditions in shallow water. The deposits of these strange blooms have long been a source of puzzlement to geologists.

One possible response to questions arising with regard to the Oligocene blooms is to invoke vertical water column instability in the Oligocene (which is bad for phytoplankton production). Well-developed stratification may have developed only sporadically, thus favoring the blooming of stress-indicator species. There is an interesting message in the Oligocene blooms: (1) Biostratigraphy may routinely reflect strange events and (2) the biostratigraphy of ancient deep-sea sediments may reflect times of unusual production, rather than recording regular conditions. While the resulting sequences might not affect dating, they might render questionable the reconstruction of "typical" past conditions.

12.2.2 On the Origin of Cooling Steps

The cooling steps are commonly referred to as products of ice-related feedback (calling on albedo changes from the spreading of snow and ice). If the assumption is correct, then an important aspect of albedo feedback is the associated sea-level change. Sea-level change is a complicated issue (see Chap. 6). It may respond to tectonic forcing without any involvement of ice, making interpretations difficult. In any case, however, cooling steps tend to be associated with a drop in sea level. Even modest changes in this level have potentially large effects because much land is close in elevation to the level of the sea (Fig. 2.1). Whenever sea level stands high, sediments are trapped in great expanses of shelf seas, and the deep sea receives relatively less sediment. When sea-level stands are low, the reverse is true. The same pattern holds for carbonate rocks: A high sea level results in buildup of limestones on the continents; a low sea level moves carbonate to the deep sea, where it accumulates as shell sand and – silt replete – with plankton fossils. Changes in ocean productivity, though, can greatly complicate the pattern observed.

When correlating the cooling steps into land sections, we tend to find evidence for mountain building. In answer to the question "whence the stepwise cooling," the suggestion comes up, therefore, that mountain building and loss of shelves are to blame (Fig. 12.7). As with ice-driven changes, albedo is part of the story. Shelf sediments are commonly light colored; the sea is not. In addition, the wearing down of the mountains removes the greenhouse gas carbon dioxide from the atmosphere. One might expect, as well, that less water vapor (a powerful greenhouse gas) is delivered by a sea that has lost its shelves. The cooling itself then develops its own dynamics: A cold ocean takes up carbon dioxide, for example, diminishing carbon in the atmosphere (and presumably in other reservoirs communicating with the ocean and the atmosphere).

Carbon burial in upwelling regions is another aspect of those internal dynamics, a phenomenon especially important in the Neogene with its striking history of changes in upwelling and increased burial of land-derived sediments, some presumably replete with organic matter, as well. In addition to minding evidence for the uplift of the Himalayas (Fig. 12.7), then, we must contemplate the masses of erosion products in the Bay of Bengal and elsewhere (and the composition of those masses) when discussing global cooling.

The high-latitude cooling indicated in the oxygen isotope trends shows two major steps in the last 50 million years

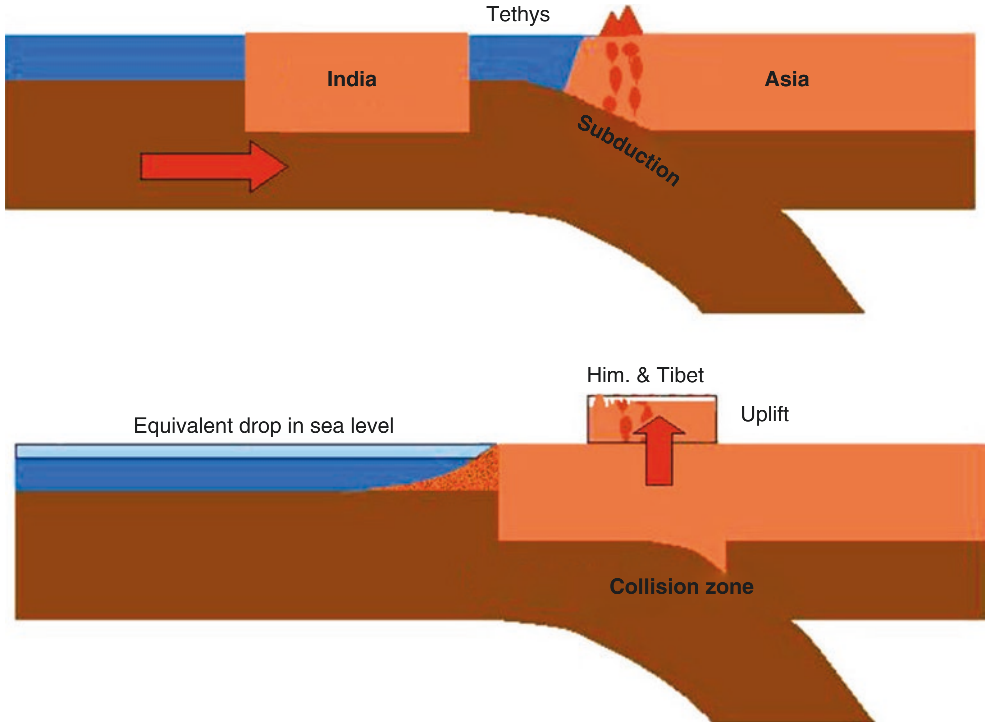

Fig. 12.7 Sketch of the connection between mountain building and sea-level drop, exemplified by the uplift of the Tibet and the Himalayas in the Cenozoic. Not to scale for the various geologic elements involved. Note the partial subduction of the Indian subcontinent

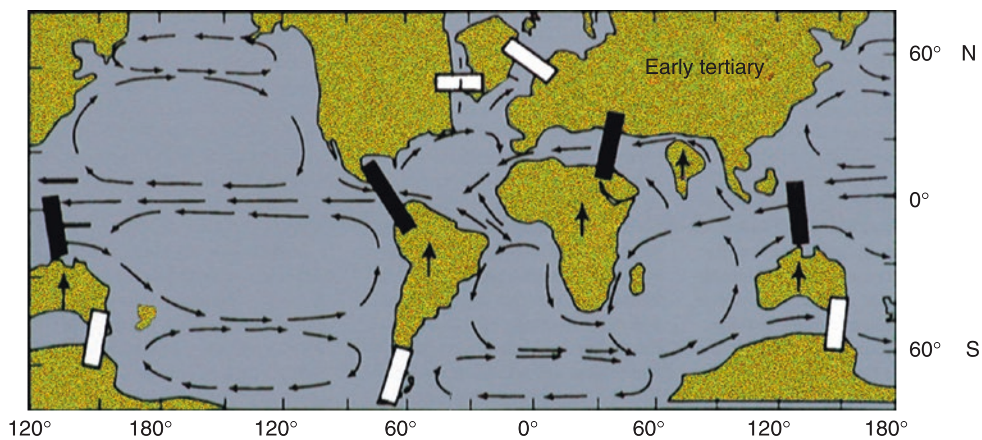

Fig. 12.8 Sketch of geography of the middle Eocene (ca. 45 Ma) and critical valve points for ocean circulation. Tropical valves are closing (*black rectangles*, Tethys); high-latitude valves are opening throughout the Cenozoic (*white rectangles*) (After B.U. Haq, 1981. Ocean. Acta 4, Suppl.: 71, with some minor modifications)

before the last one that is associated with the start of northern ice ages (Fig. 12.3). The first one roughly coincides with the AFS and the collision of the Indian subcontinent with Asia, a collision that initiated the closure of the ancient tropical seaway "Tethys." The AFS is near the end of the Eocene and thus was preceded by millions of years of cooling (presumably related to mountain building and attendant erosion). The second large cooling step is in the middle Miocene, associated with a fundamental change in deep-sea sedimentation and preceded by evidence for carbon dioxide drawdown (signaling involvement of the carbon cycle).

We do not know why the steps occurred and exactly when they did. Perhaps we see the effect of reaching a critical threshold along a trend. Or perhaps, the physical geography changed at that very time, barring or opening certain ocean passages and thus redirecting the heat carried by ocean currents and the nutrients they transfer (Fig. 12.8). Of course, the two types of possible causes – *internal feedback and change of geography* – are not mutually exclusive. We know that both are at work. The question is how important they are in each given situation.

The time when summer snow could first stay on the ground after sufficient cooling (first on Antarctica and then on continents surrounding the Arctic) must have been crucially important. Similarly the appearance and varying extent of sea ice carrying a snow cover must have provided significant positive albedo feedback to cooling (note the full reversibility of this positive snow feedback). Details in the timing may have been influenced by the sequence of closing gateways in the Tethys and opening gateways in the Southern Ocean. Without these changes in geography, ice ages presumably would have come anyway, thanks to mountain building. However, they might have come at somewhat different times and with different dynamics.

12.2.3 On Opening the Drake Passage

One way of looking at the problem of where in time a step should be placed is to note the signs of a corresponding large reorganization of the ocean's geochemistry (and hence planetary climate). According to strontium stratigraphy, one such reorganization took place around 40 Ma close to the end of the Eocene (labeled "AFS" in Fig. 12.4, for "Auversian facies shift.") The "Auversian" is an obsolete technical term for a time span in the late Eocene, defined in the region of Auvergne in France. In Auversian time (in modern terms, Bartonian-Priabonian time), the deposition of calcareous sediments in the deep ocean greatly increased, a phenomenon usually referred to as a drastic drop in the CCD at the end of the Eocene. Also, from the Oligocene on the types of deep-sea sediments were rather strictly segregated, with Murray-type carbonate ooze accumulating on the elevated parts of the deep-sea floor, and siliceous sediments rich in the remains of diatoms and radiolarians restricted to zones of upwelling. It is equally possible, of course, that the main change occurred at the end of the Oligocene. The fact that some time is needed for the Drake Passage to become highly effective (through deepening) suggests the Paleogene-Neogene boundary as the correct choice for the event defining an oceanographically effective passage between the Patagonia and the Palmer Peninsula.

The drop in the CCD may signify that the shelves around the ancient tropical seaway called "*Tethys*" ceased to function as efficient shallow-water carbonate traps. There must be a reason why the epoch following the Eocene, the Oligocene, was not recognized by the perceptive land geologists studying ancient shelf deposits for many years after the Eocene had been identified and named. One candidate reason is a scarcity of shelf carbonate or shelf rocks in general. In addition, the deep ocean may have experienced depletion of silicate after the AFS, as a result of upwelling, and the attendant explosion of diatoms. The result might have been a disappearance of large siliceous sponges from shelf seas.

If we place the opening of the Drake Passage (a rather poorly known event from the point of view of plate tectonics) at the end of the Eocene, we may choose the beginnings of a long-term process. In fact, the effective opening of the passage is a matter of controversy and learned discussion. Today the Drake Passage is responsible for the major mixing in the sea in high Southern latitudes and for the existence of the largest of all currents there, the one circling Antarctica, flowing from west to east (formally, the "Antarctic Circumpolar Current"). The appearance of the current may have been rather sudden (as far as geologic history), but simple geologic reasoning suggests that there must have been at least one element favoring gradualism: a gradual deepening and widening of the tectonically opening Drake Passage. In fact, the response of benthic foraminifers, displaying gradual turnover of species uncomfortable in the cooling bottom water of the late Eocene into the early Oligocene, may well reflect the expected gradual trend linked to the deepening of Drake Passage after opening.

12.2.4 A Matter of Time Scale

Moving from one time scale to another can be confusing. There are really three such scales that need to be considered: First, the human scale, which includes decades and even centuries, that is, personal career changes, for example, and grandchildren. Also, the "Renaissance" (a period in human history) belongs into this scale. The second important scale is the intermediate one, which is focused on one to several millennia and includes the mixing time of the ocean and (at the long end) ice ages and Milankovitch Theory and radiocarbon dating. The third scale is the very long geological time scale, measured in units of a million years. As we go back in time to exemplify processes, we are likely to encounter geological time. There is one great exception: Impact time presumably is very short (Chap. 13). Many of the "official" boundaries defining epochs presumably are linked to impacts large and small (This is a matter of discussion among experts).

The Drake Passage and the associated Circumpolar Current are features that belong to all scales, but the source of silicate for upwelling may be seen differently, depending on the time scale employed. On the geologic scale, the current maintains a reasonably high value of the nutrient in the ocean, feeding high-production systems. On the human scale (the one at issue in much of the climatology and oceanography of global warming), the complicated network of subsurface currents carrying nutrients is at focus in the attempt to understand productivity. Human-relevant information from marine geology is being contributed on the intermediate and the short scales, not so much on the long scale.

Thus, the actual histories of the Cretaceous and the Cenozoic are of great interest when reconstructing marine resources, but as far as analogs have very limited application to some of the major problems facing humankind. The study of history, however, is bound to increase understanding of processes relevant to those major problems, as mentioned.

In many cases, we become aware of "tipping points," situations of reaching critical levels along a trend. For example, at

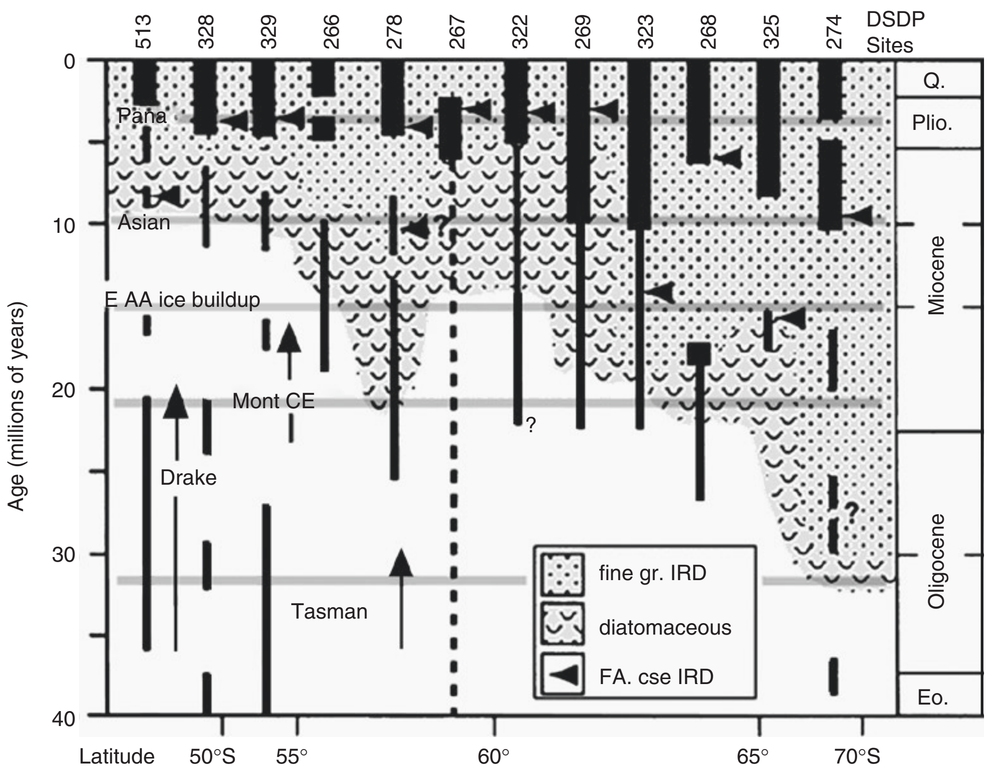

Fig. 12.9 Pattern of increasing IRD and of diatom deposition around Antarctica, as an indicator of global cooling. *FA* first appearance, *cse* coarse, *gr*. grained, *Tasman* approx. time of Tasmanian Strait opening, *Drake* approx. completion (?) of Drake Passage opening, *Mont. CE* Monterey Carbon Excursion, *Pana*. Panama Isthmus closes. *Arrows* on thick vertical lines: IRD appears in coarse fraction (after Deep-Sea Res. II (54:2399); see P.F. Ciesielski and F.M. Weaver, 1983. Init. Reps. DSDP 71:461)

some crucial level of sea-level drop (e.g., from mountain building), shelf seas can no longer serve as carbonate repositories. Instead, karst-making begins, which implies delivery of carbonate from the shelves to the ocean, for deposition at depth. (The dissolution of shallow-water carbonates on an exposed shelf produces a landscape called "karst.") The shortage of silicate in the world ocean (chert becoming less abundant) suggests that wind-driven upwelling starts vigorously, presumably around Antarctica, robbing the rest of the sea floor of opaline deposits. A corresponding shift in sediment types into siliceous ooze and mud around Antarctica is seen in deep-sea sediments of late Paleogene to the early Neogene age (Fig. 12.9).

12.2.5 On the Rise of Diatom Abundance at the End of the Paleogene

Diatoms may be considered the typical Neogene plankton microfossil, in preference to foraminifers or nannofossils (Fig. 12.2). This observation suggests that the cooling trend must have greatly affected diatom abundance. It is clear that cooling started in the early Eocene (Fig. 12.3). One infers that the appearance of Drake Passage may have intensified any ongoing cooling, following B. Haq. The capture of silica by diatoms of the Antarctic margin presumably contributed to the purity of Oligocene deep-sea carbonates over much of the seafloor (a purity noted by K.J. Hsü in the DSDP Leg 3 report). Today's sediments around Antarctica are exceedingly rich in diatoms and other siliceous fossils, as has been pointed out more than a century ago by many workers (Fig. 10.11). The high silica deposition is a presumably legacy of an open Drake Passage and the presence of ice on Antarctica. The latter generates enormous circumpolar winds, winds that drive the great ring current and generate its deep mixing aspect (i.e., bringing up silicate and other nutrients for recycling). Without the Drake Passage, world-wide productivity might be lower. It certainly would be differently distributed geographically.

On land, a great extinction occurred at the Eocene-Oligocene boundary. It is called *Grande Coupure* (Great Break) in Europe with reference to a change in land mammals. In the sea, the turnover includes a host of microfossils. The Australian biostratigrapher B. McGowran in his 2005 text on ocean history and biostratigraphy shows a prolonged benthic break at the end of the Eocene which he links to a cooling step ("Chill II"), presumably "Oi-1" in the oxygen isotope stratigraphy of K.G. Miller and colleagues. It fits the picture of major reorganization of ocean circulation and climate. Some geologists have emphasized purported effects of impacts of bolides from space in generating extinctions on land at the time. Impacts are important, but whether they play a role in this time of rapid evolution at the end of the Eocene is not clear.

12.2.6 On the Origin of Baleen Whales

There are good reasons why whale teeth are rarely found on the seafloor. One is the obvious one that they are in short supply, being delivered by animals high on the food pyramid. Also, large teeth are rare because all the "*great whales*" except the (tooth-bearing) sperm whale are baleen whales, that is, they gather food by filtering the water. They have no teeth. According to Tom Demeré, curator of marine mammals at the San Diego Natural History Museum, the first fossil baleen whales probably appeared in the Oligocene. Apparently, in spite of having no teeth, they retained many features of their ancestral stock, such as a double blowhole. The modern-toothed whales (including dolphins) on the other hand, while retaining teeth, have evolved a single blowhole, presumably to facilitate echo location. Their echo hunting targets are squid and certain fishes. Evolving a new way of hunting would seem to imply a change in prey type and abundance, which in turn supports the idea that we are seeing a change in the food chain.

12.2.7 On the Importance of Gateways

The Drake Passage (a "gateway") serves as a prime example of the importance of plate motions for global patterns of circulation, as pointed out by the Swedish-trained Pakistani-US American geologist Bilal Ul Haq. The once dominant connection between major water bodies and the large circumtropical passages forming the *Tethys* have long ago closed and disappeared. We now have cul-de-sac basins connected by the cold-water ring around Antarctica (i.e., by the "Southern exchange").

According to Haq, the difference between the present ocean and the Eocene one of about 45 million years ago reflects strongly the closing of tropical gateways and the opening of passages around Antarctica and in high latitudes in general (Fig. 12.8). In detail, each of the changes associated with openings and closings holds a wealth of interesting stories. One of the most fascinating, for example, is the drying out of the ancestral Mediterranean at the end of the Miocene between about 6 and 5 million years ago, as discovered during Leg 13 of the Deep-Sea Drilling Project (with W.B.F. Ryan and K.J. Hsü as cochief Scientists).

Huge amounts of salt were deposited at that time in the Mediterranean Tethys basin, so much in fact that the salinity of the global ocean must have been reduced by several percent ("La crisi di salinità" of Italian geologists). In the case of the Mediterranean desiccation, one assumes that a drop in sea level owing to ice buildup was at least partially responsible for isolating the ancient sea from the world ocean. Messinian gypsum-bearing rocks are seen on land, as well, uplifted. Drying the ancient Tethys basin (that is, the ancestral Mediterranean Sea) must have raised sea level elsewhere, by some 10 m, briefly reversing the general regression and perhaps thereby slowing ice buildup, which arrived around 3 million years ago in northern high latitudes on land, rather than during the time of the crisi salinità, as one might expect if cooling accompanied regression.

12.2.8 On the Grand Asymmetries in Circulation and Sedimentation

As concerns the heat budget, one corollary of the cooling leitmotif of the Neogene ocean is a displacement of the *intertropical convergence zone* (ITCZ, *the heat equator*) to north of the geographic equator. There are several reasons for this. One is the whitening of the Antarctic continent. This has the effect of increasing wind speeds in the southern hemisphere and pushing southern climatic zones northward. Another is the northward movement of large continental masses which set up monsoonal regimes favorable to the northward transfer of heat. The uplift of Tibet and the growth of the Himalayas as a Neogene consequence of the Paleogene collision of the Indian subcontinent with the Eurasian Plate had climatic consequences: a strengthening of monsoons and an increase in weathering. Another factor favoring northern heat piracy is the peculiar geographic configuration of margins in the Atlantic and to a lesser degree in the Pacific basin, configurations that provide for the northward deflection of westward-flowing equatorial currents. As a consequence, the Gulf Stream in the Atlantic and the Kuroshio in the Pacific are strengthened. The end result is that the southern hemisphere loses out on heat: Glaciers in Southern New Zealand at present are in walking distance from the seashore. At that latitude, vineyards occur around Bordeaux in the northern hemisphere, not glaciers.

One important aspect of this planetary heat asymmetry is the fact that the North Atlantic tends to deliver deep water to the ring current around Antarctica. On the whole, the North Atlantic receives shallow water in return. Thus, the northern Atlantic system represents a heat pump, with warm (shallow) water in and cold (deep) water out. Besides heat, nutrients are involved, making the North Atlantic "anti-estuarine." It entered this state in the middle Miocene, as seen in the "silica switch" of the Swiss-American geologist G. Keller (Princeton) and the USGS geologist J. Barron (California) (see Fig. 12.10). Incidentally, they favor an opening of the Drake Passage as linked to the Miocene silica switch, based on evidence from hiatus stratigraphy. Their suggestion certainly emphasizes the importance of the two events, the mid-Miocene "silica switch" and the "Drake Passage opening."

When the North Atlantic turned anti-estuarine and the North Pacific correspondingly estuarine (with high silicate

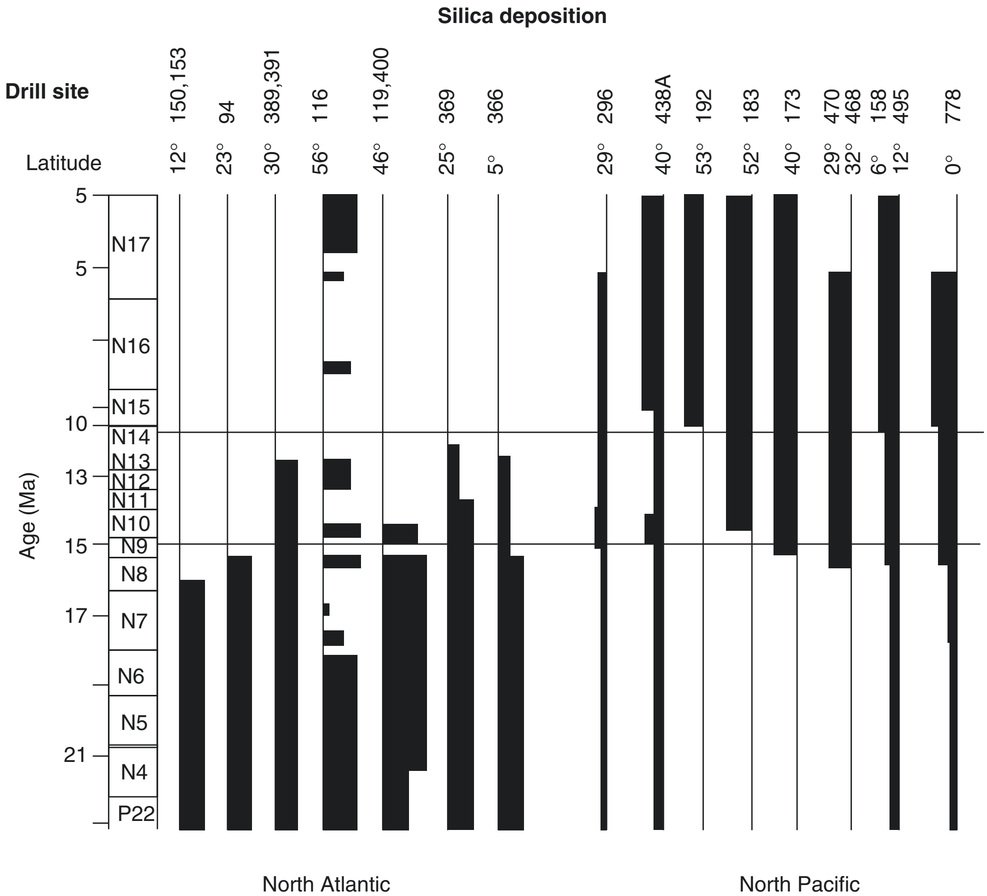

Fig. 12.10 The middle Miocene "silica switch" from the North Atlantic to the North Pacific suggesting the timing of turning on of the North Atlantic heat pump and anti-estuarine status. Silica-rich sequences marked in black (After F. Woodruff and S.M. Savin, 1989. Paleoceanography 4:87; based on a compilation of deep-ocean drilling results by G. Keller and J.A. Barron, 1983, Geol. Soc. Amer. Bull. 94:590)

values in deep waters), the present North Atlantic heat pump pattern was established. It happened sometime near 15 million years ago, judging from sedimentation patterns (actually, within the interval of 16–10 Ma, judging from the various site results and from fossil diversity). The timing is not sharply defined. The crowding of *N-zones* (based on Neogene planktonic foraminifers of biostratigraphic importance) suggests large environmental change in the middle Miocene. The isotopic changes in benthic foraminifers employed by F. Woodruff suggest an age of 14 million years (early in the transition). Using changing ratios of cadmium to calcium, M. Delaney and colleagues show a switch just after 16 Ma, one that increases substantially toward the end of the middle Miocene near 12 Ma.

12.2.9 On the Middle Miocene Cooling Step

Significantly, the "silica switch" occurs close to a major cooling step, presumably accompanied by an ice buildup and a substantial drop in sea level. The step may be the largest one in the Miocene. It is called "Mi3" in the oxygen isotope stratigraphy of K.G. Miller and colleagues. B.U. Haq and associates, in their 1987 Science article on sea-level variations through geologic time, placed a major drop in sea level at the end of the middle Miocene, based largely on seismic information in sediment stacks on continental margins but also on subsurface data from drilling.

The cooling step and major sea-level drop in the middle Miocene (presumably a large icing over in Antarctica) were associated with major upwelling, as seen in the record of margins in upwelling regions, for example off Namibia. The foraminifers of that time display a major excursion of carbon isotopes (toward positive values) indicating the burial of organic carbon (Fig. 12.11). The carbon isotope excursion is the largest in the last 50 million years. It was discovered in the earliest drilling sites, even in the 1970s. The second largest in this time span is at the Eocene-Oligocene boundary, in the Paleogene, an isotopic event somewhat reminiscent of the one in the Monterey in being closely followed by cooling. Choosing "the last 50 million years" for a discussion of cooling avoids consideration of the (negative) carbon isotope excursion at the Paleocene-Eocene boundary, which was associated with major warming. A large release of methane from the deep seafloor to the environment has been postulated for that time (the position of the event and the warming are in the name PETM, Paleocene-Eocene Thermal Maximum).

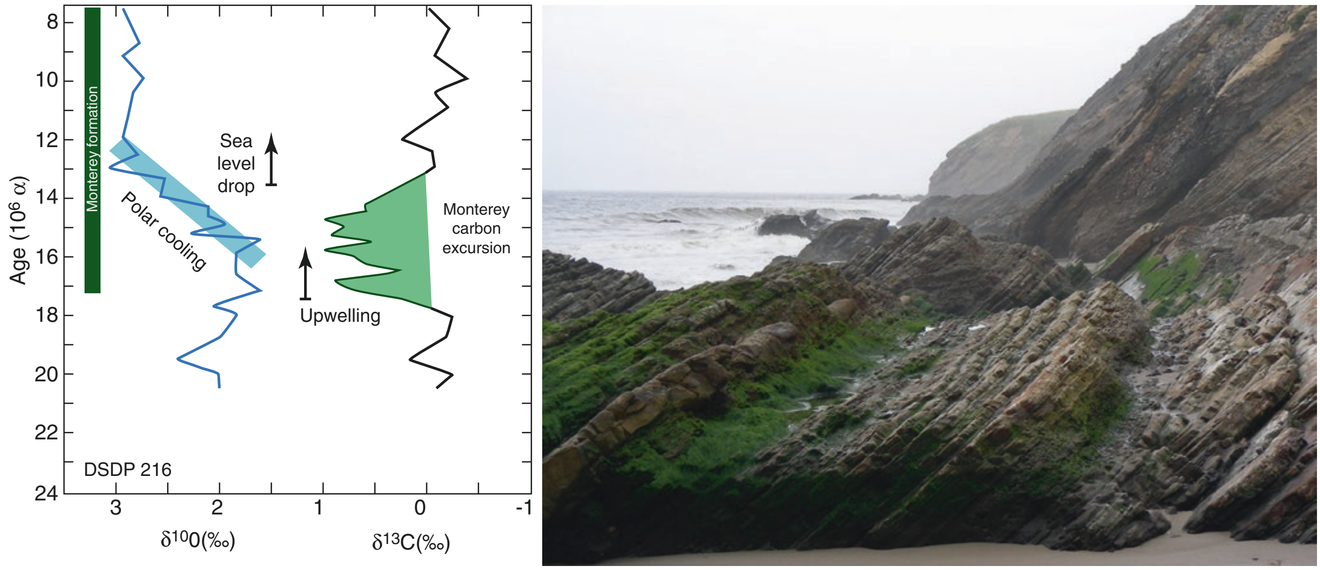

Fig. 12.11 The Monterey Event (From W.H.B., 1985, Episodes 8, 163; based on E. Vincent and W.H.B., 1985. In: E.T. Sundquist and W.S. Broecker (eds.) The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present. Geophys. Monogr. 32: 455. *Right*: Monterey Formation near Santa Barbara; Photo W.H.B.)

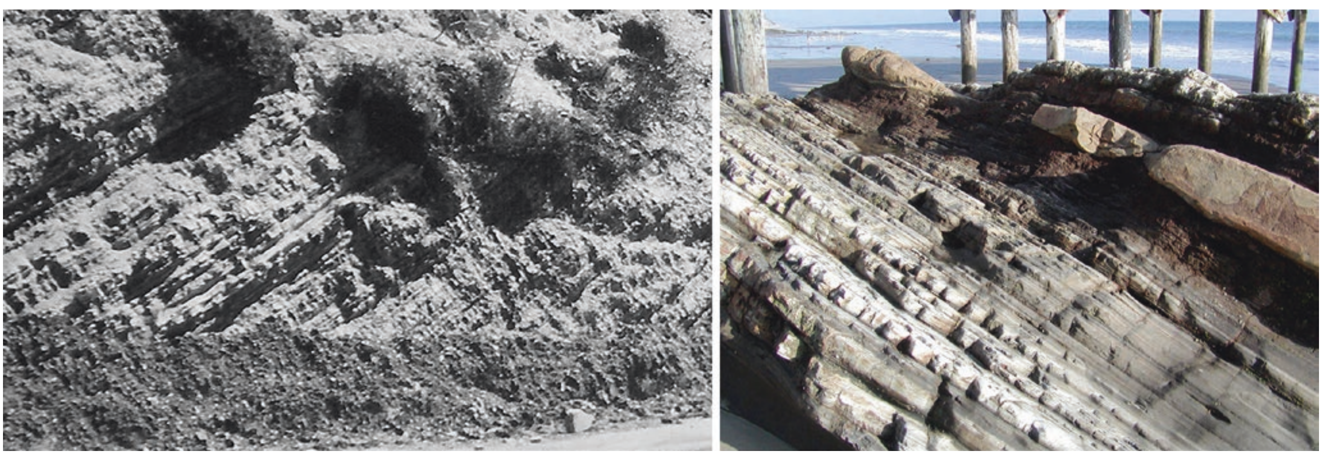

Fig. 12.12 Miocene Monterey Formation in southern California. *Left*: uplifted section on the road near Lompoc (Photo Univ. South. Cal., courtesy E. Vincent, S.I.O.) *Right*: On Gaviota Beach, near Santa Barbara (Photo W.H.B.). Fine layering is caused by lack of burrowing owing to a scarcity of oxygen, typical below upwelling areas

The French-American marine geologist and biostratigrapher Edith Vincent (then at S.I.O.) used the label "Monterey Event" for the Miocene carbon isotope excursion. The label implies drawing a parallel of the carbon isotope excursion to the evidence for increased upwelling at the time of the origin of the Californian Monterey Formation. Some of the organic carbon buried globally (and generating the characteristic carbon isotopic signal in benthic deep-sea foraminifers prior to the great middle Miocene cooling) could conceivably be of terrigenous origin, as emphasized by the marine geologist L. Diester-Haass (Saarbrücken). But much of the signal presumably has a marine origin, judging from the sediments delivered by upwelling in the eastern Pacific margin (i.e., the Monterey Formation). Much of the sediments of the Monterey Formation are finely layered and also have plentifully phosphate and chert content, the latter presumably generated by diatoms and other siliceous export from the sunlit zone and suggesting high marine productivity (Fig. 12.12).

Strong upwelling apparently started only about 10 million years ago in most coastal systems, but high-coastal production presumably started considerably earlier. Additional upwelling intensity likely set in with each cooling step. Thus, ocean productivity presumably did benefit from the cooling all through the Cenozoic, especially in post-Eocene time, and earlier than suggested by strong coastal upwelling.

12.3 Culmination: Onset of the Ice Ages

12.3.1 The Timing

We have earlier referred to the drying up of the Mediterranean at the end of the Miocene. It was not to be the culmination of Neogene cooling; the northern ice ages were. Presumably the Mediterranean desiccation was indeed facilitated by a sea-level drop caused by ice buildup, some of it on the northern hemisphere. But in the time that followed the last desiccation pulses (apparently including the earliest Pliocene), the planet experienced a relatively calm and warm period, lasting some 2 million years. This warmish time in the Pliocene was terminated by the onset of the northern ice ages, some 3 million years ago.

The onset of northern ice ages looks like expected, in principle, from the general Cenozoic cooling trend (see Hodell et al., 2002. ODP Leg 177 synthesis). For the onset to materialize, one would think, continued uplift had to keep moving the climate toward an increasingly cold state, so that at times, even the mighty Gulf Stream failed to bring enough heat northward to stop an ice age from developing. In the late Pliocene, when the signs were right (low summer insolation as postulated by Milankovitch Theory), the ice started growing on Canada. The climate then entered a 3 million year period (lasting throughout the Pleistocene) when relatively small changes in summer sunlight in high northern latitudes were translated into large-scale changes in ice mass and sea level (see Chap. 11). That northern ice buildup started around 3 million years ago within the late Pliocene was documented by the Woods Hole geologist and biostratigrapher W.A. Berggren during one of the early DSDP legs, based on ice-rafted debris (Fig. 12.13). The event not only resulted in cooling within the Gauss Chron but also provoked the onset of major climate fluctuations, as suggested in a sudden increase of isotope variations in benthic foraminifers (*Cibicides* spp.) and in planktonic ones seen off Antarctica, as documented by D. Hodell and associates, by deep-ocean drilling (see in J.P. Kennett and D.A. Warnke, eds., *1992*, *The Antarctic Paleoenvironment: A Perspective on Global Change*. AA Res. Ser. 56).

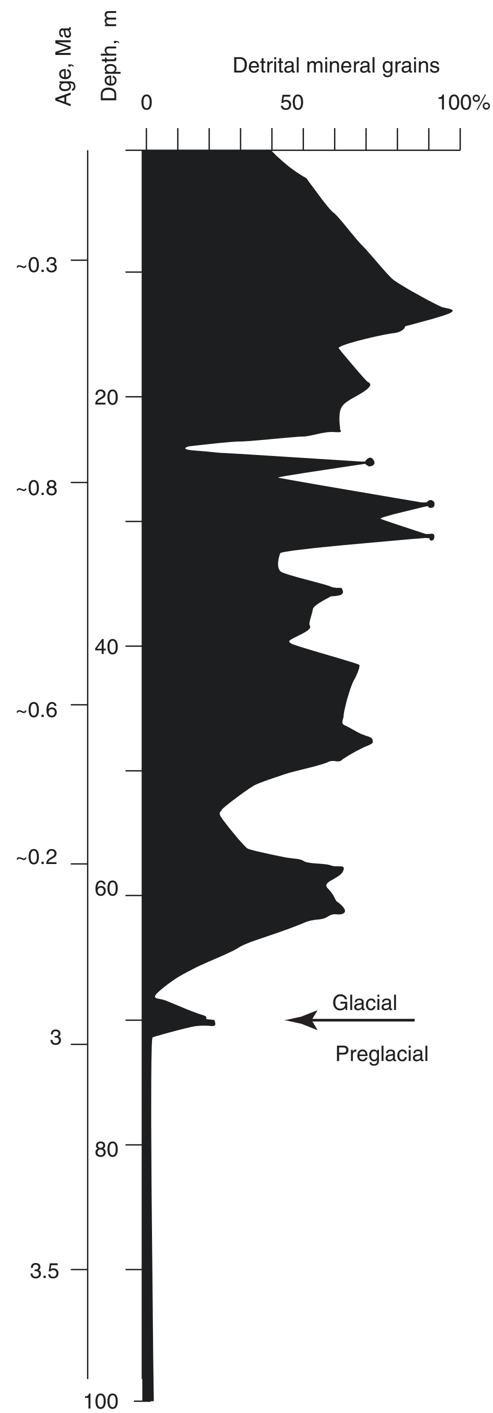

Fig. 12.13 Onset of the northern ice ages. W. Berggren, 1972 evidence (based on erosion-produced mineral grains and biostratigraphy) from the NW deep Atlantic, DSDP Site 116 (Leg 12). Onset placed roughly 3 million years ago, in agreement with geologic evidence from the Sierras

12.3.2 Panama Paradox

Several distinguished marine geologists have insisted that the timing of the onset of the ice ages is linked to closing the Panama Strait, which ended the loss of heat to the Pacific by westward currents from the Atlantic. Supposedly, moisture for making snow was then available to high northern latitudes, moisture brought northward by warm Caribbean waters (e.g., Gulf Stream, Fig. 5.14). The implication is that availability of moisture is limiting to northern polar ice growth, not just cooling. The concept, as an explanation for the origin of northern ice ages, has been questioned by W.H.B. and G. Wefer, whose "*Panama Paradox*" hypothesis calls for attaining a critically low temperature within the long-term cooling trend rather than a link to the closing of the Panama Straits, as the crucial element of the onset of the ice ages. Clearly, however, moisture is involved also, whenever making snow and ice, as evidenced by modern distribution patterns of ice in Scandinavia. Thus, the question about the role of moisture supply in starting the ice ages must be considered unresolved.

12.4 Plate Stratigraphy and CCD Fluctuations

12.4.1 Backtracking and CCD Reconstruction

To interpret the meaning of a given sample from the seafloor, the sediment has to be placed in its original depth and location at the time of deposition. This is done by "backtracking" the path that a site has taken as it aged. Without taking proper account of this path, the sediment sequences in drill cores cannot be interpreted correctly, especially if substantial changes in depth or in latitude are involved. For example, when a site moves down the East Pacific Rise in the Neogene and northward across the equator, it necessarily collects a number of different types of sediment in sequence (calcareous ooze, siliceous clay, siliceous calcareous ooze, siliceous clay, Red Clay). The resulting vertical stack of different sediments (leftmost panel, Fig. 12.14) can look like evidence for changes in the sedimentary record of the ocean's climate system, when in fact it was produced by plate motion.

Horizontal backtracking (a focus of the work of Y. Lancelot, 1937–2016) refers to geographic position, using the appropriate pole of rotation for reconstruction of latitude and longitude. *Vertical backtracking* (Fig. 12.14) addresses the reconstruction of depth by considering the subsidence of a piece of seafloor, most commonly from a cooling of the crust and uppermost mantle. Vertical backtracking is a necessary procedure whenever depth-sensitive properties are involved, such as carbonate content, sedimentation rates of oozes, or preservation of calcareous fossils, on time scales for which tectonic subsidence matters.

The position of the CCD, by definition, involves both depth and carbonate content of deep-sea sediments, and its path through the Cenozoic, therefore, can only be accurately described when taking the general subsidence of the seafloor into account. To obtain the path of the CCD

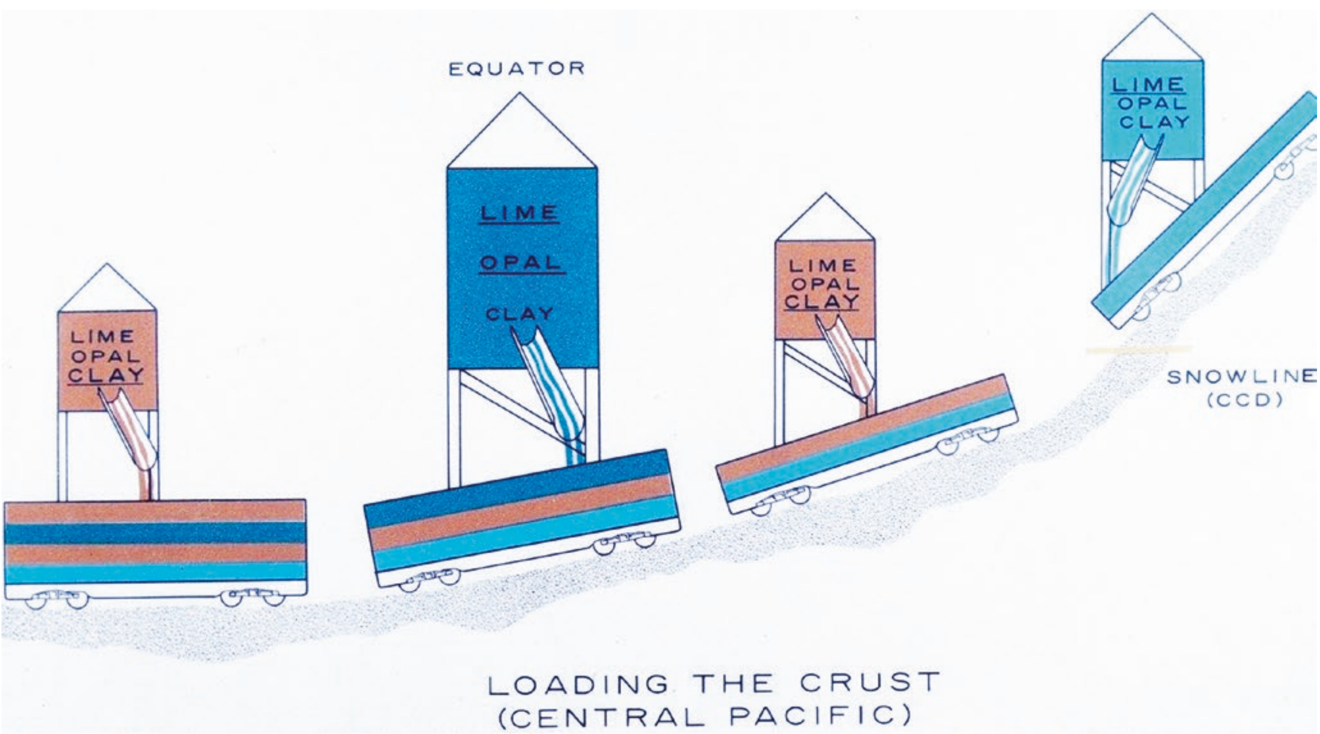

Fig. 12.14 Schematic of plate stratigraphy in the eastern equatorial Pacific: latitudes and depths. The hypothesized seafloor originated above the present CCD. The high production along the equator persisted throughout the Neogene. Sharp separation of facies (sediment types) began with the Oligocene (After concepts of W.H.B. and E.L. Winterer, 1974. [Int. Assoc. Sedim. Spec. Pub. 1:11])

(an important element of all conditions that relate to the marine carbon cycle), we must reconstruct the depth levels at which the CCD was crossed for a large number of drilling sites. Reconstructions vary with the quality of dating of the basement and of sediments, but there is general agreement (since 1985) on the major features of the Cenozoic behavior of the CCD. The most notable features are the drastic drop of the CCD at the end of the Eocene and the short-lived but remarkably substantial excursion to shallow depths during the "*carbonate crash*" of Mitchell Lyle (ODP Leg 138) at the end of the middle Miocene (Fig. 12.15). The first feature is linked to the end-of-Eocene Auversian facies shift (the "AFS"); the second (the "Crash") is tightly linked to a cooling step and a sea-level drop producing a large hiatus.

An overall parallelism in CCD variation between Atlantic and Pacific basins is remarkable. It suggests that the geochemistry of the marine carbon cycle is global in nature rather than carrying only information from each ocean basin. Comparison of sea-level reconstructions and $\delta^{18}O$ stratigraphy suggests that periods of high sea level are characterized by a shallow CCD and by warm high latitudes and periods of low sea level by a deep CCD and cold high latitudes (and cold deep waters). Why should a relatively warm ocean have a *shallower* CCD than a cold one? Is not cold water *less* favorable to the preservation of carbonate than warm water? The fuzziness of possible answers to such simple questions reflects the depth of ignorance surrounding the discoveries having to do with the history of the CCD.

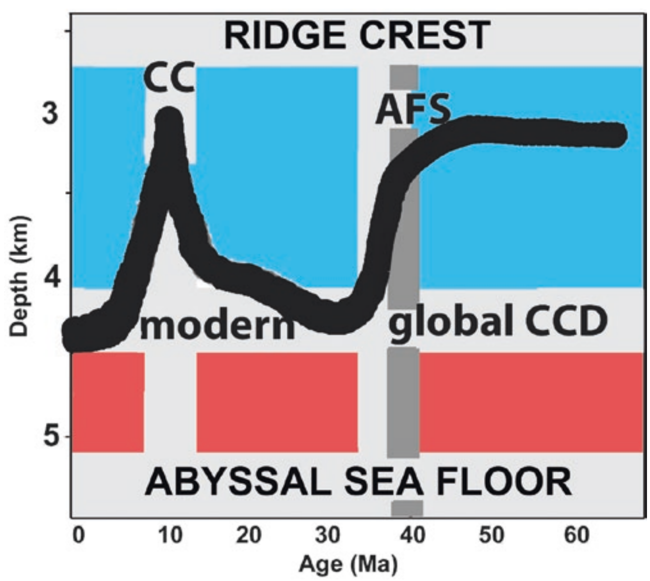

Fig. 12.15 Schematic of the Cenozoic variation of the CCD, based on numerous authors (with special kudos to K.J. Hsü and R. Wright, 1985. In: K.J. Hsű and H.J. Weissert (eds.) South Atlantic Paleoceanography, Cambridge Univ. Press, Cambridge UK.) "*Ridge crest*," depth level near −2500 m; "*modern global CCD*," depth-level variable between and within ocean basins, typically between −4 and −4.5 km; "*abyssal seafloor*," depth level below −5000 m; "*AFS*" Auversian facies shift, late Eocene, near 40 Ma; "*CC*" "carbonate crash" ca. 11 million years ago (end of middle Miocene)

12.4.2 Possible Causes of CCD Fluctuations

A simple hypothesis linking sea level to CCD fluctuations is the concept of *basin-shelf fractionation*. As mentioned earlier, shelves covered by seawater are carbonate traps. As they remove $CaCO_3$ from the ocean, they tend to starve the deep seafloor of carbonate and hence raise the CCD. Conversely, bared shelves supply carbonate to the deep sea. In this very simple bookkeeping hypothesis, variations in temperature, while important in the thermodynamics of carbonate deposition, are incidental, while changes in sea level are close to basic forcing – and to geological time scales (Fig. 6.1).

There is good reason to believe that the mass balance hypothesis of CCD fluctuations in fact does not account for all the relevant observations. It is compatible with the CCD drop at the end of the Eocene but not necessarily with the other major feature in the CCD history, that is, the carbonate crash (CC). The CC is reminiscent of the dissolution event observed at each onset of a glacial period during the last million years. It may be due to increased production and provision of organic matter producing carbonic acid upon oxidation. Available data bearing on the questions are still quite incomplete.

12.5 On the Cenozoic Methane Ice Problem

The production of methane in and on the continental margins of today's seafloor by certain archaea (once called *methane bacteria*) is pervasive. It may be considered a result of the cooling that brought us the ice ages and apparently is linked to high export production in offshore waters ("upwelling"). Methane seeps are especially abundant in the continental margin off Oregon, where they were studied by the marine geochemists E. Suess, G. Bohrmann, A.M. Tréhu, and M.E. Torres, among others.

Some of the methane is locked up in methane ice (technically "methane clathrate"), given high pressure and cold water temperature in the surrounding environment (Fig. 4.5). Methane is a powerful greenhouse gas. Vast deposits of methane ice are reported from the present seafloor in the coastal ocean in seismic profiles and in sediments retrieved

Fig. 12.16 Methane ice (methane clathrate) embedded in black mud (DSDP Leg 84; photograph from S.I.O. archives)

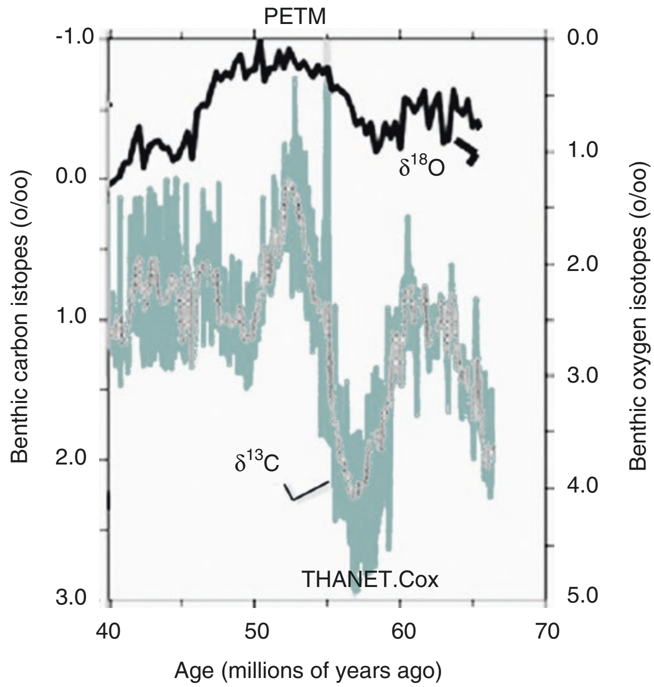

Fig. 12.17 Oxygen and carbon isotope stratigraphy in Paleogene benthic foraminifers from various drilling sites in the world ocean (Data compiled by J. Zachos and colleagues, 2001. Science 292:686; here interpolated) *PETM* Paleocene-Eocene Thermal Maximum, *THANET. C ex.* Thanetian carbon isotope excursion toward heavy values

by drilling (Fig. 12.16). The likely response of methane clathrates to global warming is as yet obscure, but the threat is palpable. What we do know is that the methane ice, once exposed to temperatures warmer than the thermodynamic value appropriate for stability, will melt readily at modest pressures (Fig. 4.5). Also, methane ice melts as confinement is removed (low pressure). In consequence of warming and pressure release, methane is released in profusion when methane ice is brought on board and can then be ignited by those holding the melting ice in hand (Fig. 1.10).

As discussed, the middle Miocene Monterey Event starts with a large excursion of carbon isotope values toward high values (i.e., with an unusually large amount of ^{13}C in the ratio of $^{13}C/^{12}C$ in calcareous fossils). The most parsimonious explanation for the excursion involves the buildup of C^{12}-rich stores of biogenic carbon during the time, including methane ice.

The most impressive methane ice event may have happened in the Paleogene. According to the co-discoverers of the carbon isotope anomaly marking the time of the event, J.P. Kennett and L.D. Stott (ODP Leg 113), the event was responsible for major extinction of benthic foraminifers in the deep sea. A release of massive amounts of methane from melting of clathrate has been proposed as an explanation of a (negative) carbon isotope excursion at the very end of the Paleocene. The very same explanation was used to address the thermal maximum (the "*PETM*") associated with the isotope event. The warming during the PETM, it is widely assumed, largely reflected an excess abundance of carbon dioxide generated by oxidizing the methane.

Interestingly, the PETM-associated carbon isotope excursion toward negative values is preceded by a large and distinct excursion toward heavy isotopes several million years earlier, in the "Thanetian" (Fig. 12.17). A buildup of organic matter deposits, including methane ice, could have produced this excursion.

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13 Cretaceous Environments and Deep-Ocean Drilling

13.1 Background: Mesozoic Rocks and Fossils

13.1.1 Mesozoic Marine Rocks

Marine Cretaceous rocks are familiar sights on land. Their age (roughly between 65 and 145 million years) is great enough so that even slow uplift measured in mm per millennium can raise them high above the surroundings in places. An unusually high stand of sea level at times within the geologic period flooded shelves widely, greatly expanding the seafloor on continents. Today, we find the legacy, sedimentary rocks: limestones, sandstones, and black shale (Fig. 13.1). In La Jolla, Southern California, upper slope sediments were transformed into stone and can be studied as uplifted cliffs at Boomer Beach (Fig. 13.1d).

13.1.2 The Carbon Energy Connection

To many geologists, the Cretaceous is familiar as a source of fossil carbon energy (Fig. 13.2), including coal (from swamps along shallow seaways and basins) and petroleum of marine origin. Middle East oil sources are mainly of Cretaceous age, as are those of North Africa, for example. Carbon-based energy has become highly problematic, because the chief waste product when using this energy source is carbon dioxide, a *greenhouse gas* (i.e., an infrared-trapping gas) that warms the lower atmosphere (by simple rules of physics). Also, carbon dioxide turns into carbonic acid (when reacting with water) and thus acidifies the sea, with deleterious effects for a host of carbonate-producing organisms.

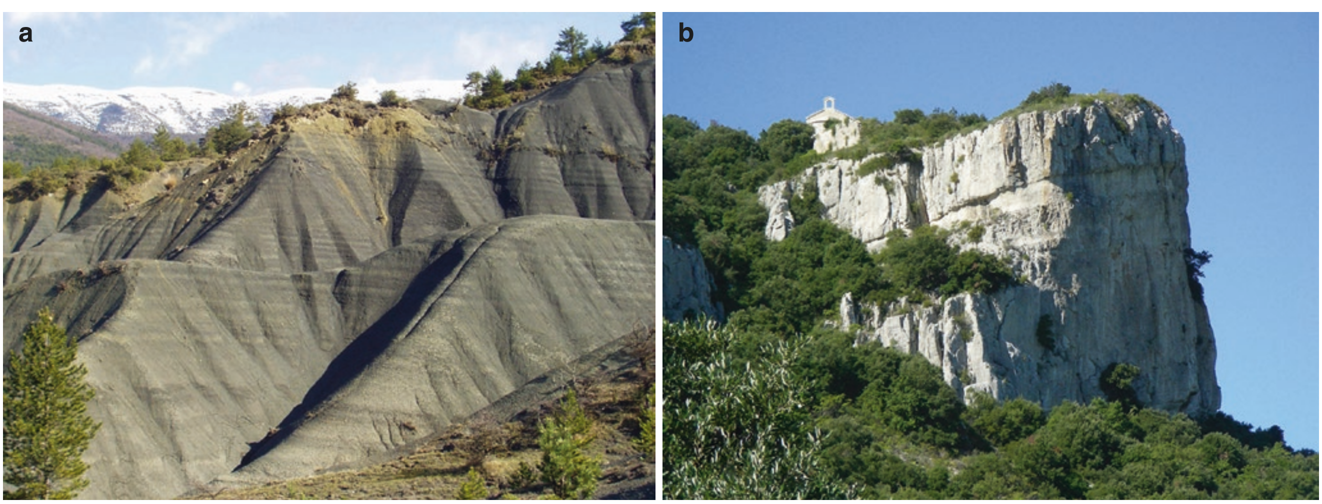

Fig. 13.1 Landscapes of marine sedimentary rocks of Cretaceous age. (**a**) Black shale west of the Jura Mountains, SE France; (**b**) limestone near La Ciotat, SE France; (**c**) sandstone layers of the Mesa Verde group, SW Colorado, (**d**) Sand- and siltstone cliff, Boomer Beach, La Jolla, San Diego (Photos W.H.B.)

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_13

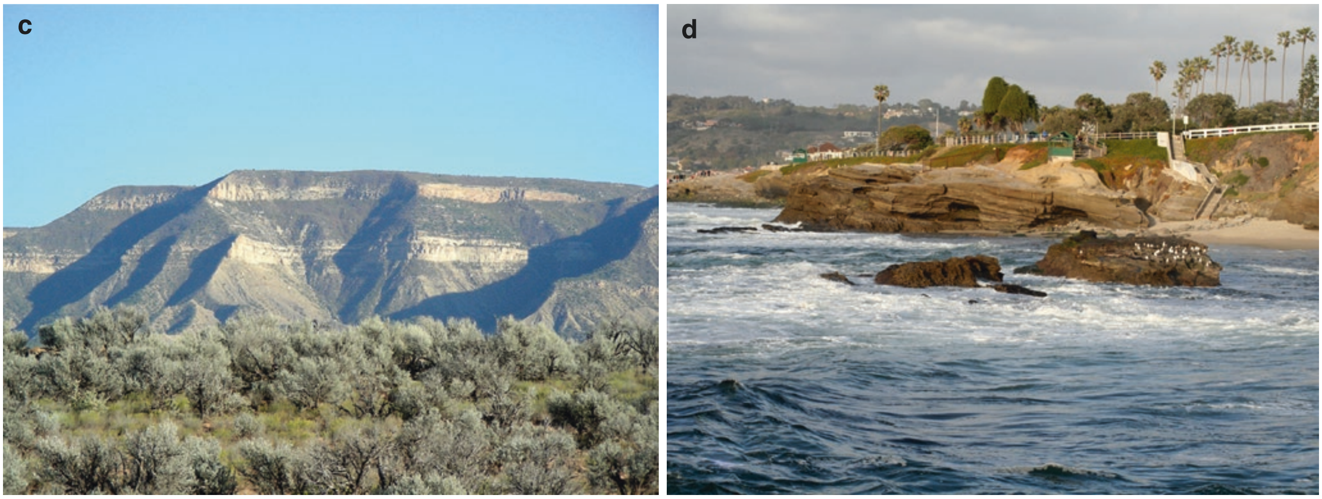

Fig. 13.1 (continued)

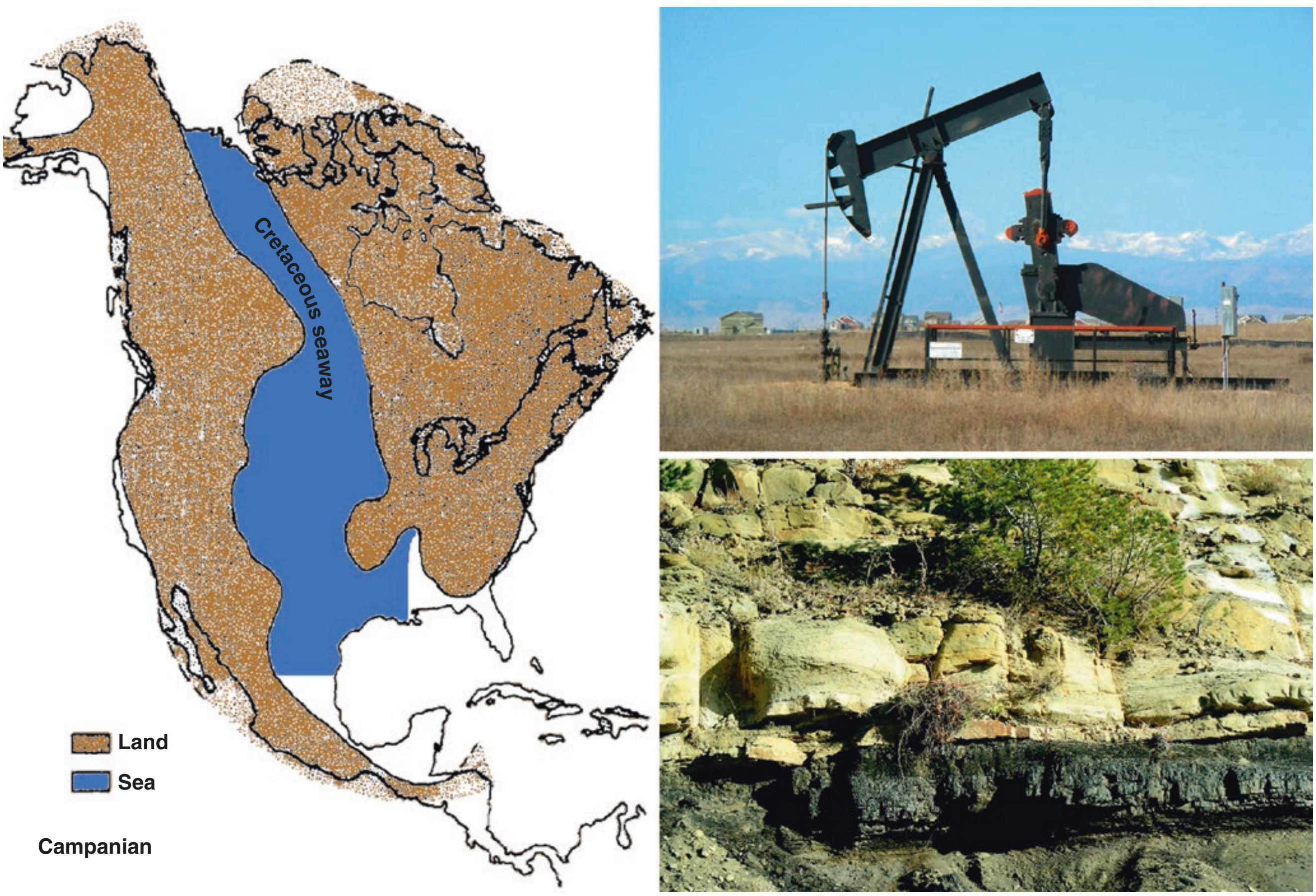

Fig. 13.2 Petroleum in the Cretaceous, North America. *Left*: approximate extent of the Campanian (upper Cret.) seaway in North America (after M.E. Donselaar and H.L. Levin). *Right*: petroleum pump near Fort Collins, Colorado, and coal layer from swamp at the edge of the seaway, Colorado (Photos W.H.B.)

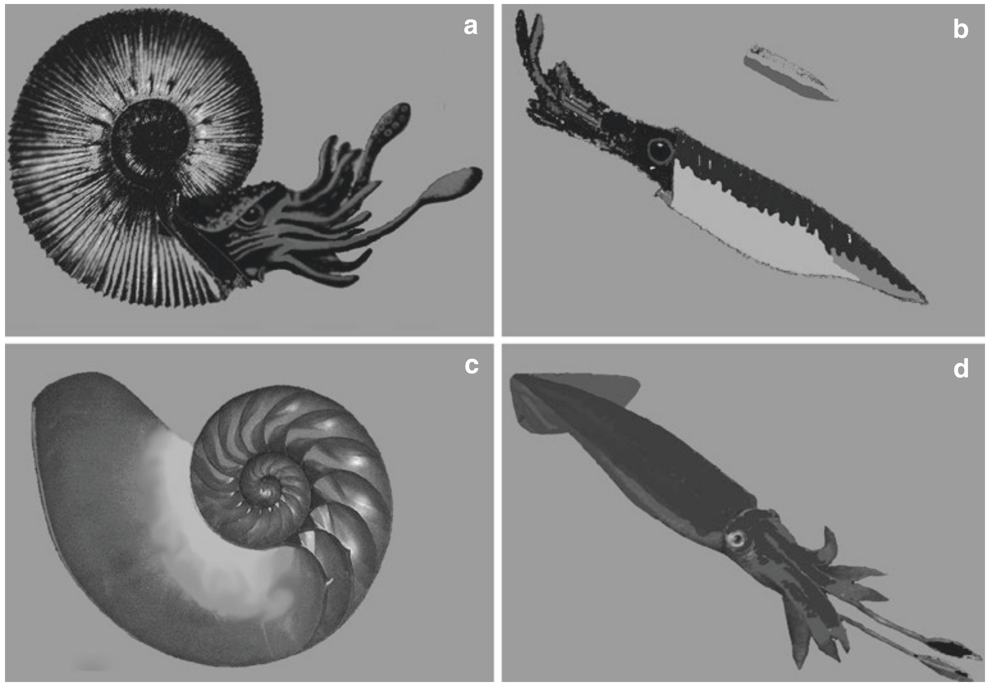

Fig. 13.3 Mesozoic cephalopods (reconstruction of an ammonite and a belemnite, (**a**, **b**); the cigar-shape rostrum is commonly all one finds of a belemnite) and two modern-related forms (Nautilus shell and *Loligo*, (**c**, **d**), not to scale) (Sources of information: M. Neumayr (**a**), US NOAA (**d**), and various museum exhibits (**b**, **c**), notably in Long Beach (California) and near Copenhagen (Denmark))

13.1.3 Mesozoic Marine Fossils

Marine sedimentary rocks commonly have abundant fossils, which invite reconstruction of the environment based on their distribution. Of course, reconstructing the preferences of extinct organisms is a hazardous undertaking and commonly provokes much discussion. This being the Mesozoic, we have plenty of ammonites and belemnites among the macrofossils (i.e., remains of extinct cephalopods) (Fig. 13.3). Ammonites, judging from their shape and their modern relatives, were slow and had to rely on depth and perhaps on life environments lacking oxygen to escape agile marine predators. Unfortunately, for any ammonites hiding in oxygen-poor water, though, air-breathing hunting reptiles would not have cared about the oxygen content of the water. Many belemnites apparently were fast swimmers, somewhat like the modern squid (but yet others may have been slow like the modern cuttlefish). In any case, we must assume that these organisms were among the more intelligent ones in the sea, judging from present behavior of cephalopods.

There are many other types of marine Cretaceous fossils, of course, especially mollusks and echinoderms, but none more prominent than the bones of dinosaurs in river and swamp deposits and bones from their marine relatives such as ichthyosaurs and plesiosaurs. The ichthyosaurs and the plesiosaurs lived in the Cretaceous and before, within the Mesozoic. They now are extinct, as are the once ubiquitous ammonites and belemnites, the terrifying mosasaurs, large and toothy marine predatory lizards of the late Cretaceous. Today the only marine lizard is the modest-size alga-eating diving iguana on the Galapagos Islands on the equator in the eastern Pacific. It braves what some might consider “cool” waters (down to 20 °C or even slightly less) but likes to hang out in shallow rock pools in the black Galapagos basalt where the water is very warm. It is roughly the size and color of a gray squirrel. It looks much like its (yellowish and larger) land-living cousin. The marine species has conspicuous defensive armor on its back (spines); the land-living larger cousin does with somewhat less obvious armor. It would be a mistake to consider marine reptiles as almost extinct, even though fossil remains are exceedingly rare. In Indonesian waters and elsewhere in the Indo-Pacific tropics, there are sea snakes, and off Australia, there is a large marine crocodile. Sea turtles are ubiquitous in tropical and subtropical waters. One sees them in Hawaii, in the Caribbean, and in the Sea of Cortez, for example.

Fossils are the stuff of biostratigraphy. At sea much of the biostratigraphy of the Cretaceous is based on foraminifers, that is, shelled microfossils identified under a microscope (Fig. 13.4). Samples from deep-sea drilling also are routinely examined for nannofossils (Fig. 12.6).

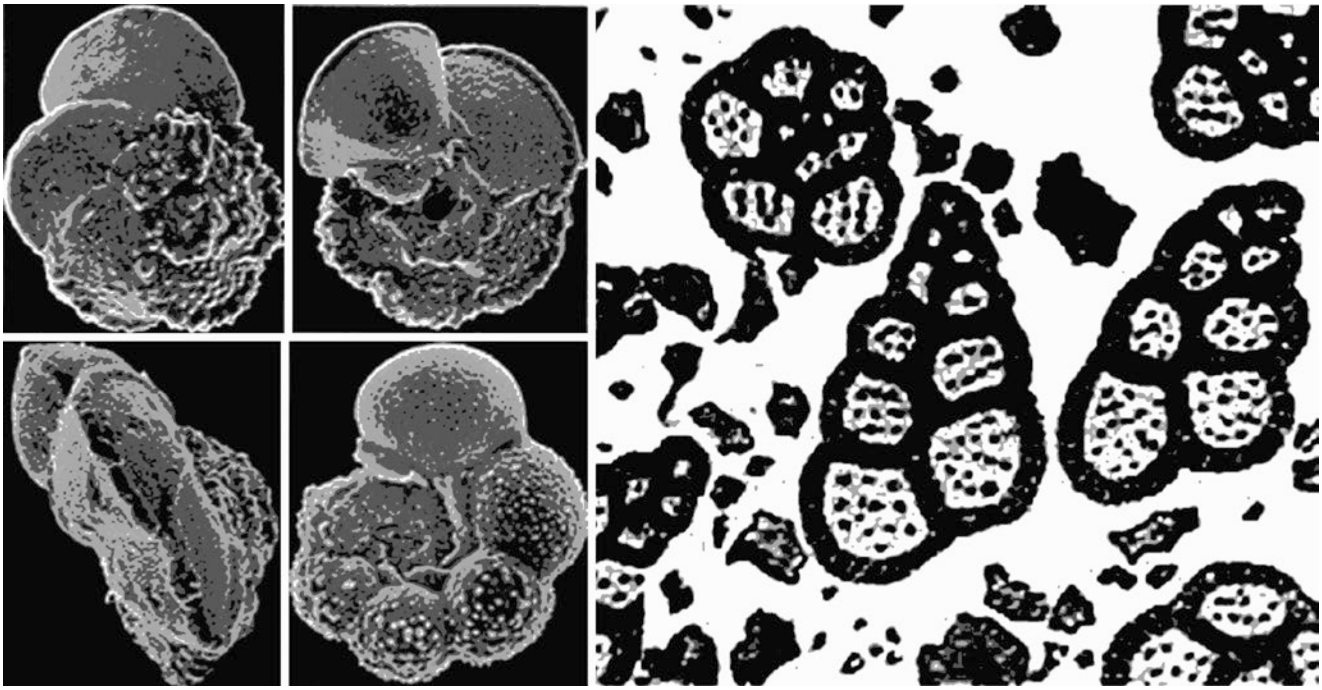

Fig. 13.4 Cretaceous planktonic foraminifers. *Left*: from deep-sea sediments (SEM graphs E. Pessagno, Leg 1 of DSDP); *right*: contents of chalk at the English Channel (From the nineteenth-century textbook of M. Neumayr, greatly enlarged)

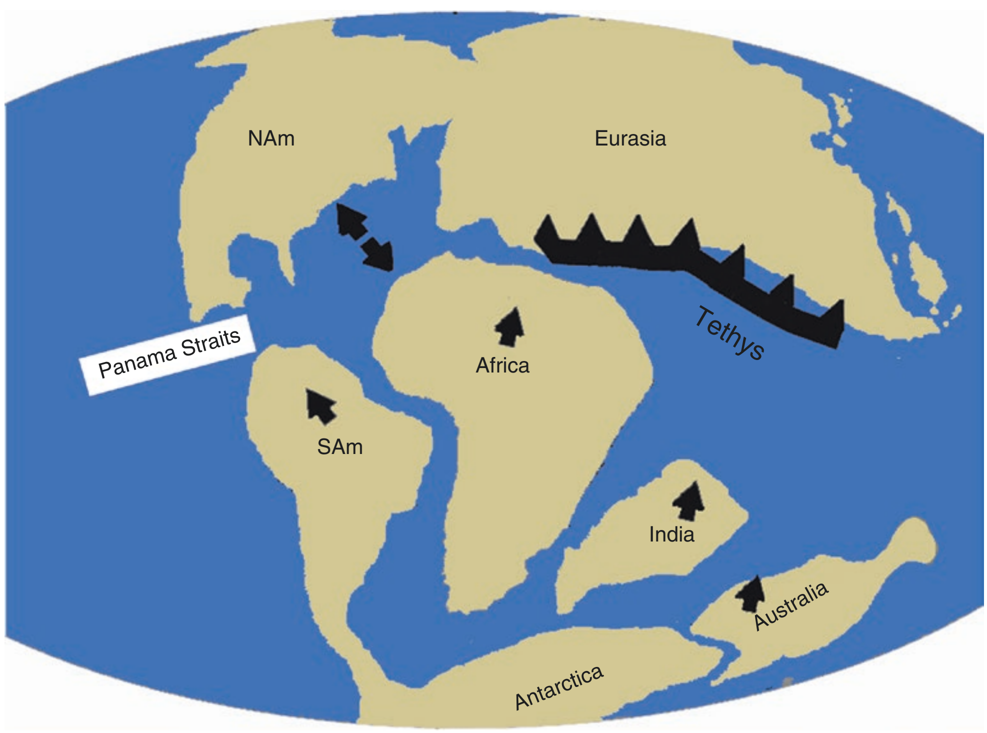

Fig. 13.5 Rough sketch of continental positions 100 million years ago. The continents on the whole are moving northward. The Tethys is closing (trench teeth point toward closure), and the North Atlantic is opening by seafloor spreading (*arrows*). "Panama Straits" will turn into "Panama Isthmus" in the Pliocene. The rapidly drifting micro-continent "India" is noteworthy. Its slamming into Eurasia (toward the end of the Paleocene) signals the closure of the Tethys (Map largely after E.J. Barron and C.R. Scotese, modified)

13.1.4 On the Abundance of Mesozoic Seafloor

About one half of the seafloor is Mesozoic of age. When contemplating seafloor of that age, we are dealing with a geography that has been greatly changed by continental drift and seafloor spreading over many millions of years. Thus, in middle-Cretaceous marine rocks on land, we find evidence for the tropical seaway "Tethys" where India is now docked against Eurasia. Also, we find a much smaller Atlantic than now, opening vigorously in the northern hemisphere (Fig. 13.5). Obviously, much of the oceanic crust generated in the Atlantic is Cretaceous in age, since spreading started well before that period. In the Pacific, it is the far western region that has the oldest basaltic crust. Characteristically, the seafloor is very deep here: The lithosphere had time to cool. In the Indian Ocean, the Indian subcontinent, with the northern portion as yet then exposed, was migrating northward carried by Cretaceous seafloor toward its encounter with Eurasia. The encounter of India with the Eurasian continent took place after the end of the Cretaceous, in the Paleogene. The history of the Indian Ocean seafloor is complicated. In any case, the distributions of magnetic anomalies and elevations suggest that somewhat less than one half of the sea floor is Cretaceous in age here.

13.2 A Warm Ocean and a Dearth of Oxygen

13.2.1 The Discovery of Black Shale and Some Implications

A warm *ocean*, such as prevailed in the Cretaceous, implies a low-oxygen ocean. Naturally, what we have learned about organic-rich sediments in the present ocean (a cold one, not a warm one) looms large in the interpretation of black organic-rich shale. A realistic approach considers black shale a common product of an oxygen-poor sea, that is, a sediment type commonly occurring with deposits rich in pyrite and other sulfides resulting from sulfate reduction. In this view, it is the late Cenozoic shelf sediments that are unusual and not the Cretaceous ones. These older sediments of a warm ocean (documented for the deep seafloor by drilling during DSDP Leg 11, see Fig. 13.6) reflect conditions that apparently prevailed in the sea during much of the Phanerozoic, judging from fossil-bearing rocks on land. Certain fossils, though (very large flying insects of the Carboniferous), suggest a high concentration of oxygen in the atmosphere, according to J.B. Graham (1941–2011, physiologist-biologist at S.I.O.). However, according to H. Craig (1926–2003, geochemist, S.I.O.), an upper limit for the oxygen is set by the need of the forest to escape spontaneous combustion if there is to be coal. As is not uncommon for ancient rocks, environmental clues may be difficult to interpret.

Regionally, within the Cretaceous, the oxygen content apparently dropped low enough to prevent burrowing organisms from establishing a mixed layer on the seafloor. Thus, laminations are common in pre-Tertiary marine sediments. We must assume, from the abundance of laminations in sediments, that the availability of recycled nutrients for marine production was quite limited compared with a situation where sediment is stirred. It seems reasonable, therefore, to invoke low-oxygen content and reduced loss of organics to explain the wide distribution of high organic concentrations, rather than unusually high supply of organic carbon. This does not mean, of course, that upwelling might not be responsible for local oxygen shortage. It does suggest, though, that a common occurrence of black and pyritic sediment of shelves must call on a more general causation than offered by regional upwelling and increased regional productivity.

A search for modern analogs of Cretaceous environments is a somewhat problematic endeavor in that the analogs sought may not exist. Nevertheless, it is instructive, of course, to define the conditions producing black sediments today. Many of the processes found presumably also apply in a warm ocean.

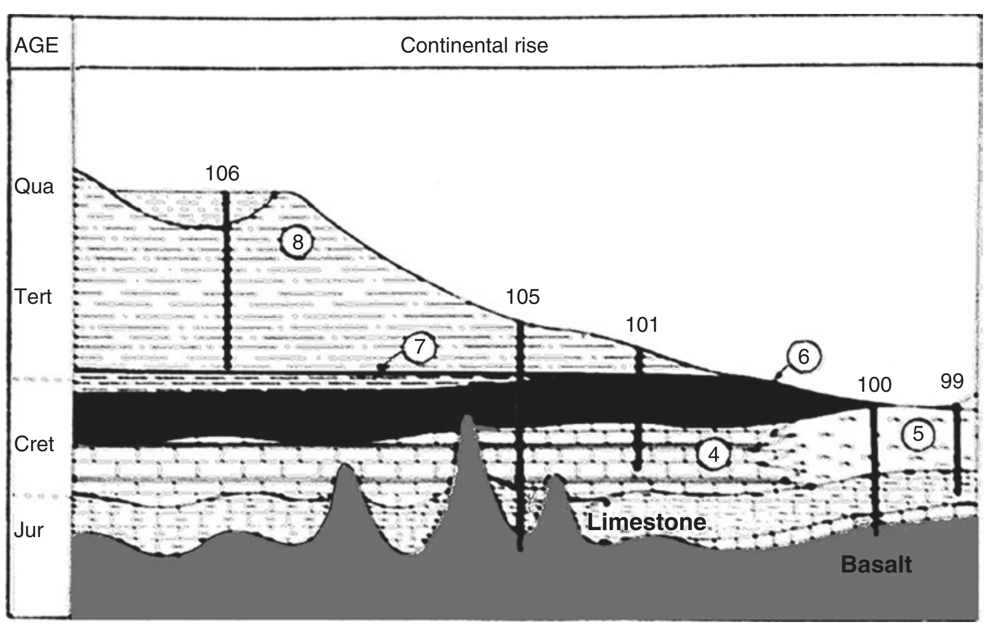

Fig. 13.6 Schematic diagram of sedimentary sequence off the eastern US Coast found by drilling, including Cretaceous black clay (*6*) on limestone (*4*) and chalk (*5*) and followed by multicolored clay (*7*) and hemipelagic mud (*8*). Numbers: DSDP sites (From Y. Lancelot et al., 1972. DSDP Leg 11)

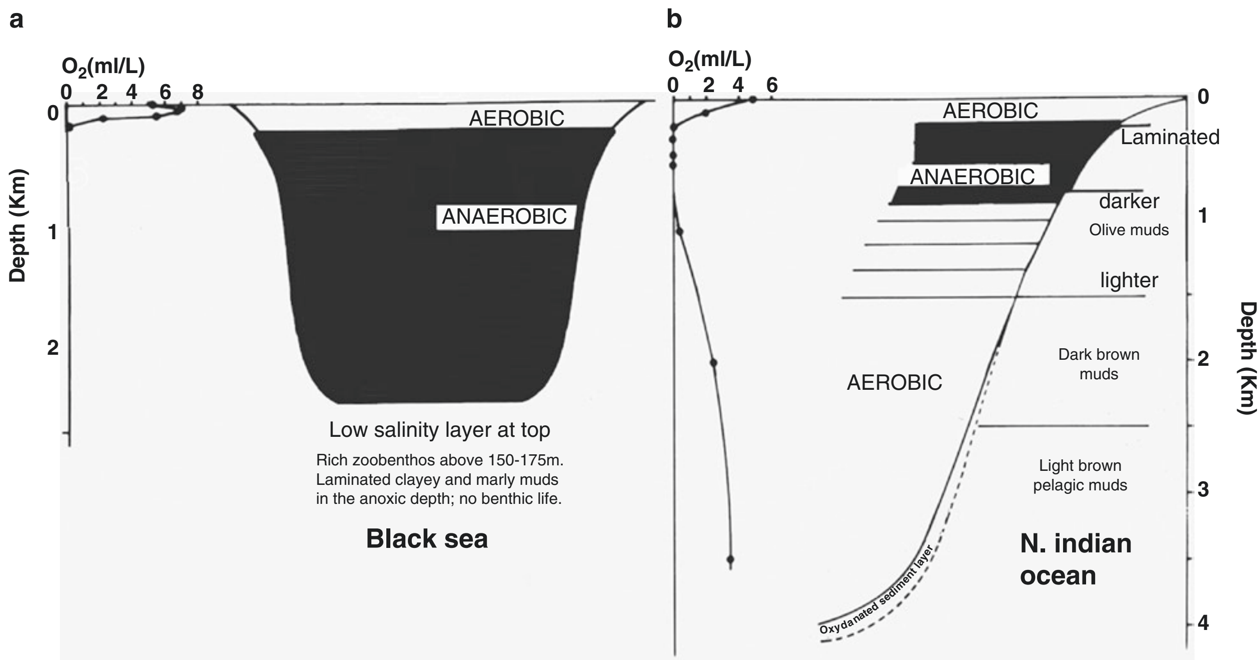

Fig. 13.7 Modern analogs of black shale deposition according to J. Thiede and Tj. van Andel. (**a**) Black Sea, with strong salinity stratification preventing vertical overturn and thus blocking the supply of oxygen to the deep layers. (**b**) Intersection of an oxygen minimum with a continental margin (After J. Thiede and Tj. H. van Andel, 1977, simplified. Earth Planet. Sci. Lett. 33:301)

13.2.2 Modern Analogs for Black Sediment Deposition

There are two situations where sediments with high organic content are being deposited today: (1) partially restricted basins with estuarine circulation and (2) the oxygen minimum zone on the upper continental slope, especially in areas of upwelling. The Baltic Sea and the Black Sea are examples for the estuarine circulation in somewhat restricted basins; the Gulf of California and the slopes of Goa (Indian Peninsula) and of Namibia exemplify the oxygen minimum situation (Fig. 13.7). Common to both environments is a high supply of organic matter and a relatively low supply of oxygen. To expand the likelihood of obtaining "black" deposits, these observations suggest that we can increase productivity or decrease the oxygen supply, or do both simultaneously.

13.2.3 Oceanic Anoxic Events

Several time spans have been identified as being especially rich in anaerobic sediments in the Cretaceous. The periods were labeled "oceanic anoxic events" by the geologists S.O. Schlanger (1927–1990; Northwestern University, Illinois) and H.C. Jenkyns (University of Oxford, UK). In the original publication, one "OAE" is a time span within the Aptian, one is at the Cenomanian-Turonian boundary, and there is one in the Santonian (Fig. 13.8). The suggestion is that sea level was high at times of anoxia. According to Hugh Jenkyns (in a review published in 2010), there is a possibility that methane release

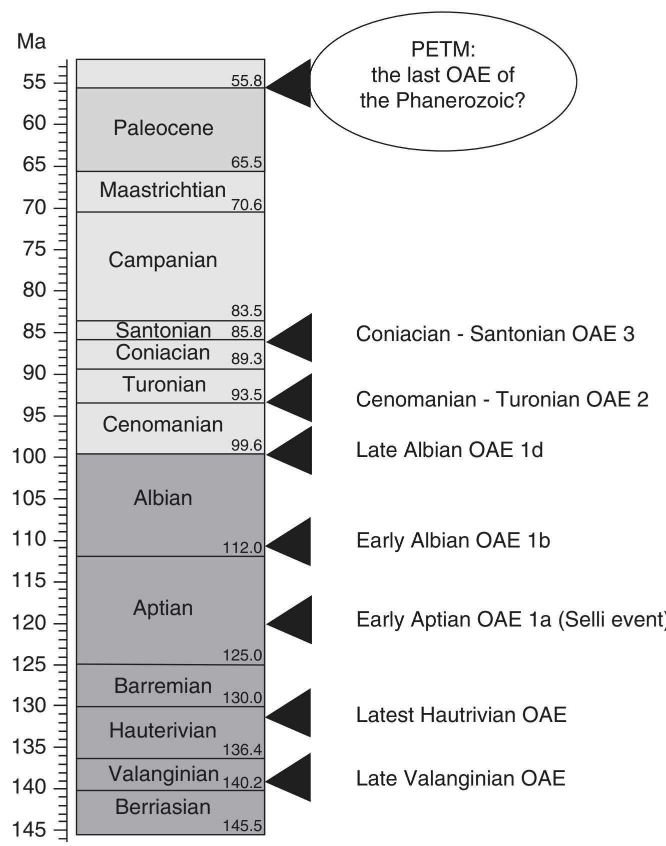

Fig. 13.8 Late Mesozoic periods with abundant organic-rich marine sediments ("oceanic anoxic events"), according to Hugh Jenkyns, 2010. The PETM would be an early Cenozoic equivalent event in the Jenkyns scheme (After a review by H. Jenkyns in Geochemistry, Geophysics, Geosystems v. 11 (3))

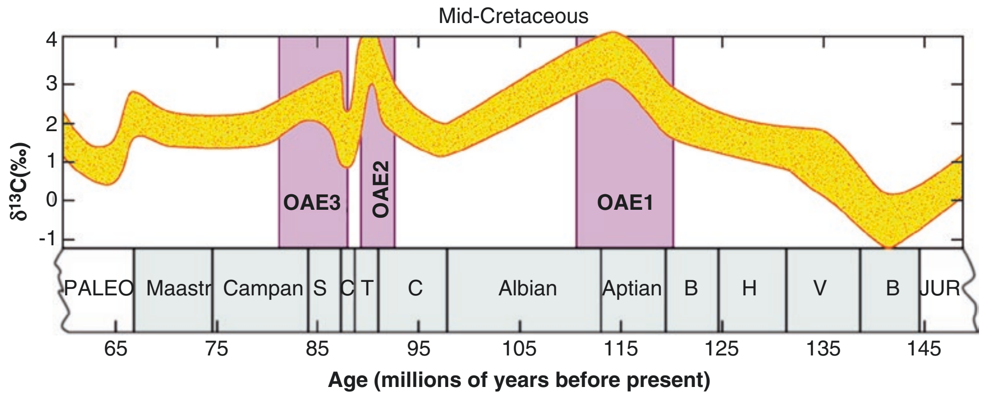

Fig. 13.9 "Oceanic anoxic events" of S.O. Schlanger and H.C. Jenkyns, as seen in the Cretaceous δ^{13}C signal of pelagic limestone. Stages at bottom from Berriasian to Maastrichtian simplified (for full names, see previous figure) (After M.A. Arthur, W.E. Dean, and S.O. Schlanger, 1985. Also see AGU Monogr. 32:504; figure here modified for clarity)

was involved in the origin of OAEs, which would make the PETM Event at the end of the Paleocene the last example of a series of similar events in the late Mesozoic.

A high sea level in the middle of the Cretaceous would call for increased generation of new and hot seafloor (standing shallow and displacing seawater therefore, as suggested by the Lamont geophysicist W. C. Pitman). The large number of Cretaceous-age basaltic edifices (plateaus, volcanoes) in the western Pacific basin would agree, in principle, with the proposition. Also, according to H. Jenkyns, large anaerobic events may have been typically accompanied by an increase in nutrient supply, enhancing productivity and oxygen demand. He thinks that the progressive evolution in redox conditions through phases of denitrification to sulfate reduction could have been accompanied by water column precipitation of pyrite framboids and may have resulted in fractionation within many isotope systems (e.g., N, S, Fe, Mo, and U), modifications that can be recognized in Mesozoic sediments today.

The Schlanger-Jenkyns label "oceanic anoxic event" while easily remembered may not describe the situation found quite precisely. Organic-rich layers may not define a single "event" but instead may just indicate higher propensity of occurrence of laminations. In other words, the technical term "oceanic anoxic event" (applicable to time intervals in periods marked by arrow heads in Fig. 13.8) often appears to refer to a time span of a high probability for a general shortage of oxygen, lasting for millions of years. In marine shelf deposits, pyrite nodules are common, indicating reduction of sulfate and reaction of the resulting sulfide with iron, presumably mostly within sediments (rather than in the water or on top of the seafloor, details being difficult to reconstruct). The laminations in the clay- and siltstones indicate the absence of large organisms on the seafloor from a lack of oxygen.

13.2.4 The δ13C Signal

A global signal for the propensity of making black claystone can be obtained from carbon isotopes, as pointed out by the US geologists P.A. Scholle and M.A. Arthur in 1980. The ratio of carbon isotopes in fossils and in limestone favors somewhat the rare ^{13}C whenever organic carbon is extracted from the sea in unusual amounts (Fig. 13.9). Indeed, the isotope stratigraphy in detailed sections suggests pulsed extraction in many cases. Such pulsation could result from crossing the same low-oxygen triggering condition numerous times, for example, by exceeding a balance between supply of organic matter and oxygen, a balance that presumably was unusually precarious during periods with OAEs.

13.2.5 "Anoxic Events" and Volcanism

Why were there OAEs at all? So it was warmer and perhaps the sea level stood higher. But why was that? We do not know for sure, of course. However, one scenario that is quite plausible is linked to the idea that extensive volcanism was responsible for the origin of the OAEs. There is evidence for enormous outpourings of basaltic lava in the western South Pacific in the Aptian, Albian, and Cenomanian (i.e., in the mid-Cretaceous). Many geologists believe that the associated release of carbon dioxide to the atmosphere was responsible, at least in large part, for the warming observed (and hence for some reduction of oxygen values in the sea).

Large amounts of the greenhouse gas carbon dioxide are given off by volcanic activity (this can be measured). The level of carbon dioxide in the present atmosphere is extremely low. An addition of the greenhouse gas can materially increase concentration. Indeed, repeated massive release of the gas can affect the temperature of the lower atmosphere for millions of years. (Cautionary remark considering occasionally encountered confusion about volcanogenic emissions of carbon dioxide: we are talking about long-term processes here, not about the human time scale. Also, the steady increase of carbon dioxide in the air since the Industrial Revolution demonstrably is *not* caused by volcanic activity.)

There are several solid witnesses of enhanced volcanic activity from that time in the mid-Cretaceous, some in the

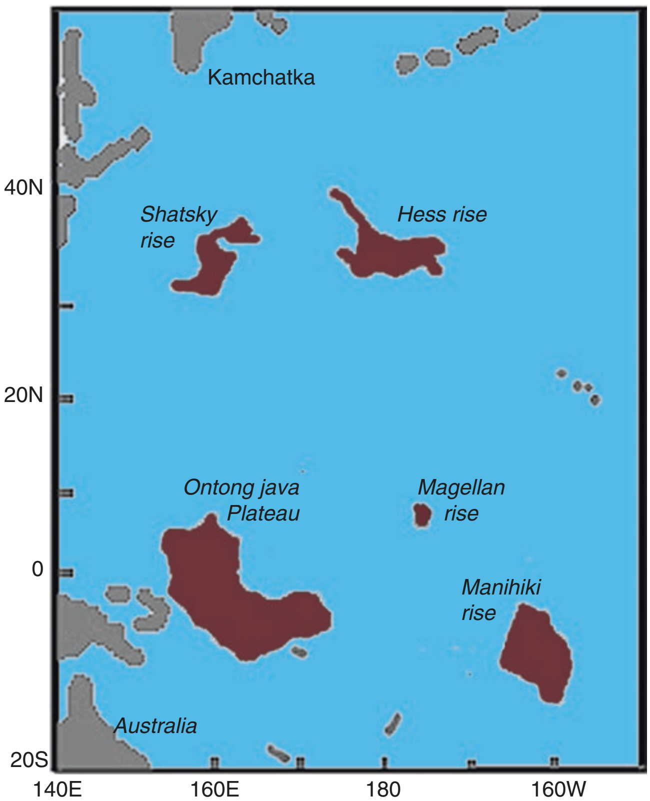

Fig. 13.10 Basaltic plateaus of Cretaceous age in the western Pacific. The largest of these is Ontong Java Plateau, more than once a target for drilling to reconstruct the history of Cenozoic and Cretaceous time from the calcareous sediments accumulating on elevated seafloor (Map from ODP)

shape of great basaltic mesas or plateaus (Fig. 13.10). One of these is the Ontong Java Plateau east of New Guinea. It is 40 km thick and has the size of Texas. A "superplume" event may be responsible for its origin; that is, the plateau may be derived from a great hot blob of magma that rose all the way through the mantle from the core-mantle boundary during the early Cretaceous in the manner of a "hot spot" and arrived at the seafloor sometimes in the Aptian. Being relatively warm, the basaltic crust of the plateau is less dense than the normal seafloor material and stands high above it. Being deep-seated, it presumably stays elevated for millions of years, thanks to slow diffusion of heat through rock, collecting carbonate ooze and thus providing an excellent record of Cretaceous and Cenozoic history.

13.2.6 Coincidence of Volcanism, Anaerobic Conditions, and Petroleum Abundance

There is a striking coincidence of major volcanism, anaerobism, petroleum formation, and high sea level in the timing of events in the Cretaceous (Fig. 13.11). As the geophysicist and marine geologist R.L. Larson (1943–2006; Rhode Island) pointed out, the frequency of magnetic reversals changed at the time of black shale formation. Perhaps the delivery of large plumes of magma extracted energy from the core-mantle boundary, halting the processes responsible for magnetic reversals. Larson's graph emphasizes the complexity of factors that are important in ocean and Earth history and the connection between seemingly unrelated elements of the Earth system. Especially the influence of mantle processes on climate evolution has caught the attention of ocean historians and of Earth scientists in general. So far, it is fair to say, the proposed explanations for the correlations, while interesting and worth considering, are quite hypothetical. The graph is suggestive, but correlations remain difficult to document in any detail.

Is the formation of black shales and of related petroleum source rocks really determined by volcanism and the making of oceanic crust? Presumably, volcanic activity and processes related to plate tectonics are indeed important (see timing of the origin of "ocean crust" in Fig. 13.11), but these factors may be joined by others, less obvious. We do not know for sure.

13.2.7 Milankovitch in the Cretaceous?

Many pelagic sediments of Cretaceous age show distinct cyclic deposition, suggesting that the balance for near-zero oxygen was global and that it was readily disturbed everywhere by relatively small but persistent changes in irradiation patterns in privileged regions (in analogy to Milankovitch Theory). Alternations of organic-rich and carbonate-rich rock types are typical in marine shelf sediments. On the deep seafloor, one finds variations that appear to be strictly cyclic (Fig. 13.12). They are readily captured as changes in shading in the recording of calcareous ooze, and they are seen in changes of carbonate content, as well. Oscillations can persist for millions of years. According to the marine geologists T.D. Herbert (Brown University) and S.L. d'Hondt (Rhode Island), the cycles found in DSDP Leg 39 are of precessional origin (i.e., they are orbital in nature) and extend over vast areas.

What might be the mechanism or mechanisms translating orbital information into sediment cycles during periods when albedo-based enhancement of climate change is absent or modest, without the support from albedo changes in snow and ice? We must consider several factors, some related to temperature and oxygen supply, others to nutrients and productivity, yet others to dissolution and resulting changes in darkness of sediment at the surface of the sea.

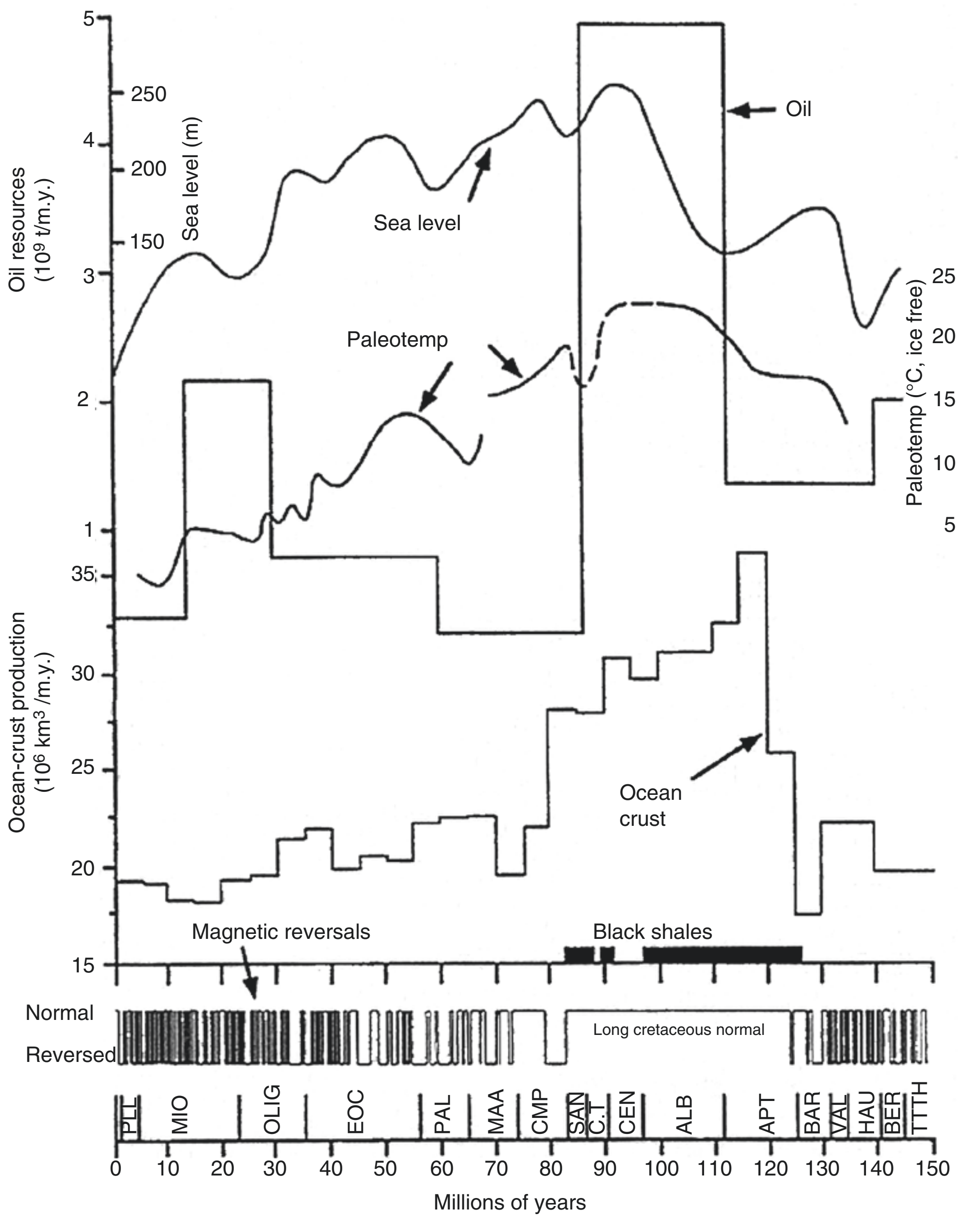

Fig. 13.11 Comparison of the production of basaltic ocean crust with sea level and temperature changes, black shale and oil formation, and magnetic reversal activity. Note the position of the three OAEs ("*black shales*") (Courtesy of R.L. Larson. See Geology 19:963 (1991))

Temperature may be crucially important in the production of the cycles. At present, much of the ocean below the thermocline has a temperature between 0 and 5 °C. Saturation values for oxygen in this cold water are near 7.5 ml/L. Typical actual values in the deep ocean are between 3 and 5 ml/L; that is, they are distinctly lower than saturation values because of oxygen consumption through decay ("AOU") on a millennial time scale (deep ocean mixing) or shorter. (Oceanographers refer to the difference between saturation at a given temperature and the actual value as "*apparent oxygen utilization*" or "AOU.")

For oceans in the mid-Cretaceous, isotopic measurements suggest deep water temperatures close to 15 °C. For water temperatures between 15 and 20 °C, the oxygen saturation is some 2 ml/L less than the present-day cold deep water. If productivity and organic decay effects (which control AOU) were roughly the same then as now (as suggested by the similarity of sedimentation rates), we can subtract that value from the expected saturation values of the warmer deep water (values that presumably indicate initial oxygen content of Cretaceous deep waters) and arrive at a typical value of 2 ml/L for average Cretaceous conditions. With such a modest starting value for oxygen content, it would hardly be surprising to end up in pulsed anaerobic conditions in many places, given some amount of variation about the average oxygen content in deep waters. Milankovitch forcing could have been working on

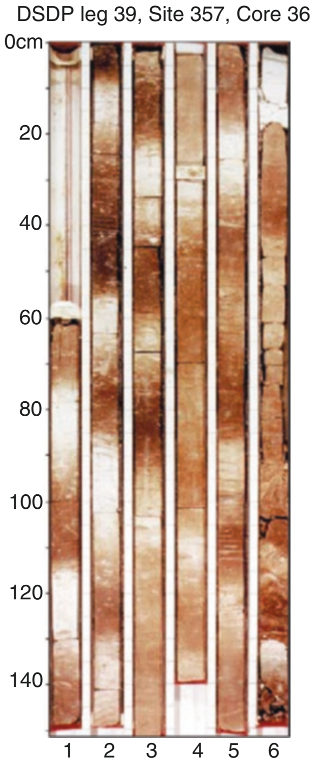

Fig. 13.12 Deep-sea carbonate cycles in Cretaceous sediments, identified as precessional (i.e., each cycle is between 20,000 and 25,000 years long). There are roughly 25 cycles in this core, for a deep-sea sedimentation rate of about 2 cm/ millennium (comparable to the rates of today's calcareous ooze) (T.D. Herbert, 1998. In: R.L. Larson (ed.) ODP's Greatest Hits. Ocean Drilling Program. College Station, Texas)

modifying conditions that were strongly predisposed to respond to small triggers with dark or white sediments. Evidently, the question about the mechanism of making rhythmic AOEs with Milankovitch periodicity may easily turn into a question about rhythmically changing controls of the AOU through time.

It is unlikely that we shall know the relevant mechanisms soon – after all, we do not understand the cycles of the ice ages very well – even though they govern the period we live in, and they have been studied in great detail for several decades. Unfortunately (for the geologists studying environmental history of the Phanerozoic), the entire Phanerozoic likely has conditions normally resembling Cretaceous ones much more than modern ones. Thus, while the Cretaceous opens worlds into the Phanerozoic (the last half billion years), it also raises a flag of caution, warning us about the lack of analogs and hence of understanding that accompanies the study of warm and ice-free worlds.

13.2.8 A Plethora of Unusual Minerals

If high salinity is a crucial property of moderately warm bottom water (making it heavy enough to sink to the seafloor), we might expect the development of *dead lakes* on the deep seafloor and also in basins of a deep shelf; that is, we might see evidence for stagnant water bodies with no free oxygen and with a plethora of nutrient-related elements and compounds collected over thousands of years. A modern analog (ice on land being salt-free) was postulated half a century ago by the late physical oceanographer L.V. Worthington of Woods Hole. Worthington proposed a "dead lake" (bottom water and deep water many millennia old) for the global deep sea after melting the ice masses of North America and Scandinavia, during the end of the last ice age (see Sect. 11.6). The time required to make any dead lake deposits thick enough to escape later mixing and oxidation would be of great interest. It would vary in length, of course, with no striking record left for many dead lake events. At this point, we do not know whether such bottom waters as postulated in the dead lake hypothesis ever existed.

What we do know is that an early DSDP leg (Leg 14) recovered many uncommon minerals from Cretaceous sediments, including rhodochrosite, siderite, barite, and plenty of clinoptilolite (a zeolite). Zeolites are commonly interpreted as alteration products of volcanic glass. However, there are ready-made alternative explanations for the origin of certain zeolites (and the occurrence of minerals suggesting an anoxic environment). Correct interpretation is likely to be very difficult to attain.

13.3 Remarks on Cretaceous Carbonate Reefs

13.3.1 Why Cretaceous Reefs Matter

Ancient reefs have long been the focus of a host of studies, in part for their academic interest regarding paleoecology and in part for their importance in economic geology in the context of hydrocarbon occurrence. Modern reefs are in the current news for various problems, presumably mainly problems related to human impact (bleaching, overfishing, acidification). As complicated and highly diverse ecological systems, coral reefs seem to have a special sensitivity to any disturbance. Their history holds special interest therefore, and the Cretaceous is no exception (except that there were no people).

Obviously, the effects of water properties on any marine organisms greatly depend on the nature of the water the organisms find themselves in, not just on one factor, arbitrarily chosen. Regarding the quality of the sunlit seawater during anoxic times, a number of questions arise: just how severe and how frequent were invasions from outside the shelf (and did the local shelf environment turn bad for coral at all) and how long did any postulated "bad" conditions last; just how acidic was the water (from the addition of volcanogenic carbon dioxide); just how much poisonous hydrogen sulfide was there (from the conversion of sulfate to sulfide in an anaerobic environment); and just how many usable nutrients were there. Loading of waters with substances typical of anoxia depended, one assumes, on where anoxic conditions prevailed in the water column: close to the bottom and in contact with sediment, or everywhere up into a halocline (boundary between salt-rich deep water and somewhat fresher upper waters) or thermocline, or in various water layers. The available data do not readily tell.

13.3.2 Rudists and "Bad" Conditions

The question arises whether anaerobic conditions in warm oceans could have been detrimental to the ancient reef-making organisms. If so, the coral was in trouble whenever anoxia dominated in basins of the sunlit zone. In an analogous question regarding high-nutrient conditions, the answer offered several decades ago by the US paleontologist P. Hallock (Miami) and the Austrian geologist W. Schlager (now Amsterdam) was that high-nutrient conditions must have interfered with reef building. (Many corals are desert specialists.) Besides water with low-nutrient content, stone corals may need plenty of oxygen, and water free of poisonous gases, conditions that may be interfered with in an anaerobic environment.

Pulsed anaerobism in an intermediate time scale would not have offered much of an improvement over a long-term anaerobic environment. Unacceptable is unacceptable, even if bad conditions prevail just for a short time. What is necessary is a way to escape the bad conditions, either by running or swimming away (not possible for sessile corals) or by withdrawing into a shell and closing it tight. The latter may have been a special ability of reef-making "rudists," that is, reef-building bivalves that flourished in the Cretaceous, the ones of the late Cretaceous being quite abundant (Fig. 13.13). The species became extinct at the end of the Cretaceous, along with many other marine organisms.

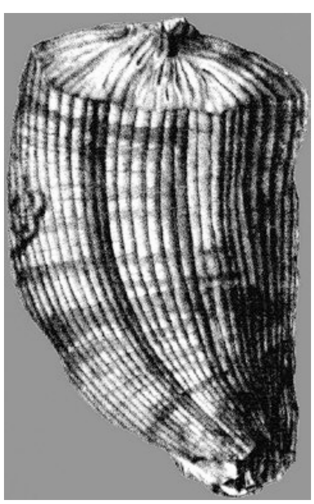

Fig. 13.13 A reef-building bivalve called a "rudist." Note the ability to close the shell container tightly. This type of organism flourished in the late Cretaceous, making large reefs (After M. Neumayr)

13.4 The End of the Mesozoic

13.4.1 On the Evidence for Sudden Termination

The end of the Mesozoic (like the end of the Paleozoic that preceded the Mesozoic) was catastrophic; that is, it demonstrably involved the extinction of a large number of animal types in a geologically short time. Exactly how the habitats and requirements of survivors differed from those species that succumbed to the deadly event has invited much study but still is not so well known. "Luck," with its suggestion of random choices, would readily provide an answer, but relying on chance may seem like a cop-out to many who are interested in the subject.

By convention and poetic inclination, we count the time since the end of the Cretaceous (shorthand term: the "K-T event"; K for Cretaceous, T for Tertiary) as the "Age of Mammals," replacing the "Age of Reptiles," the Mesozoic. However, the implied switch from reptiles to mammals took some time, of course, and implies overlap of the two time spans. While many reptiles (dinosaurs, ichthyosaurs, mosasaurs, and various other – saurs that happened to be around, as well as ammonites) may have gone extinct in a day or a month, the mammals must have had millions of years to evolve such disparate forms as bats and whales, whose fossils are found in the early Cenozoic, the "Paleogene." Presumably, some of the required diversification took place well before the great extinction. The environment may have been quite unfavorable for dinosaurs around the end of the Mesozoic, judging from the assertion of many paleontologists that a majority of species of the large reptiles were disappearing well before the mass extinction event. In any case, mass extinction apparently affected ammonites and belemnites (cephalopods), reef-making rudists and other clams, and many types of sea urchins, brachiopods, corals, and crinoids. Not only many large reptiles (including iconic dinosaurs) apparently went offstage at K-T time but well over one half of marine organisms including many taxa other than the ones most commonly mentioned.

In shallow marine sections on land, the evidence for profound change associated with the K-T event is obvious in many places. One such place, famous among geologists for the contrast between the fossil content of Cretaceous limestones and the Cenozoic ones, is in the cliff at the shores of Stevns Klint in Denmark, on the island of Zealand, south of Copenhagen. While the thin "fish clay layer" often cited as marking the boundary is somewhat elusive here, the sights are indeed

Fig. 13.14 The K-T boundary at Stevns Klint in Denmark. Shore erosion has kept the outcrop free of vegetation. The *arrow points* to the end of the Mesozoic sequence (Photo W.H.B.). The boundary is very accessible in a nearby quarry

spectacular (Fig. 13.14). A nearby museum offers geological details.

Pelagic sediments exposed on land for a long time have given the best evidence for a rapid and fundamental change in plankton biota at the end of the Mesozoic. Initial information on pelagic mass extinction emerged thanks to the studies carried out by a number of biostratigraphers in the 1960s (e.g., W.A. Berggren at Woods Hole, M.N. Bramlette at S.I.O, E. Martini in Frankfurt). Of these types of studies, none have attracted more attention than the one by the German and Swiss geologist H. Luterbacher and the Italian geologist I. Premoli Silva. These two stratigraphers took great care in measuring and describing the section near Gubbio (in the Apennine Mountains in Italy). The section (fully marine) nicely shows the end of the Mesozoic and the beginning of the Cenozoic in the plankton sequence, as a remarkable change from a large and flashy to an inconspicuous and smallish planktonic foraminifer fauna. The sequence described by Luterbacher and Premoli Silva was later used by the Alvarez team (father and son) and their associates to provide evidence for an impact from space as the cause for mass extinction at the end of the Cretaceous (Sect. 13.4.2).

Results from deep drilling in the seafloor fully confirmed the earlier results suggesting a mass extinction at the end of the Cretaceous. There is now no doubt whatsoever that marine plankton suffered major extinctions in a short time associated with the impact of an asteroid (or several) from space. Whether or not dinosaur extinction was already proceeding prior to the impact is an interesting question but is only a burning issue for those studying how much of dinosaur extinction was caused by impact. The reality that there was an impact and that it caused extinction is not made questionable by dinosaur evidence. That a sharp boundary marking the event is found at all is somewhat surprising because bioturbation and reworking can suggest a transition where there is none in reality. This problem is especially evident in silt- and clay-size fossils, for obvious reasons.

13.4.2 Evidence for an Impact

A gradual buildup of stress toward the end of the Cretaceous is unattractive as a cause for most of the extinctions at the end of the Mesozoic for a number of reasons. One of these is that warmwater plankton shows sudden change, while abyssal benthic forms do not. Apparently, we must look for a major unusual disturbance at the Earth's surface on a short time scale to produce the patterns seen. Such a disturbance, *an impact from space*, was proposed by Luis and Walter Alvarez (father and son, physicist and geologist, respectively, both at UC Berkeley) and their collaborators, the geochemists F. Asaro (Berkeley) and H.V. Michel (Sandia) in 1980, in an article in the journal *Science*. Evidence for an impact scenario also was presented at the same time by Jan Smit (then a doctoral student in Amsterdam) and by J. Hertogen (Gent University, Belgium) in the journal *Nature*. Both the "Science" and the "Nature" articles call on collision of our planet with a large mass of rock coming in from outer space and ending the Mesozoic. The Alvarez team estimated the likely diameter of a solid impacting asteroid near 10 km, with the evidence mainly resting on minute but unusually high concentrations of the rare element *iridium* within the K-T boundary layer.

Jan Smit focused on a record located in Spain rather than the one near Gubbio, Italy. Also, he turned to the presence and distribution of certain droplets of molten crustal rock, so-called *tektites*, within boundary-near sediments to document aspects of the collision. Tektites (once molten pieces of rock) are commonly found in connection with impacts of large meteorites. Additional evidence for the K-T impact was provided at several other places in part by other metals and mineral matter (including shocked quartz) and by tektite-rich sequences found in deep-sea sediments by drilling (Fig. 13.15).

The Alvarez team hypothesized that the proposed impact raised sufficient dust to block sunlight and thus interfered with photosynthesis for some time, in analogy to a scenario called "*nuclear winter*" (which became a subject of much discussion in the 1980s). A crater of K-T age (~65 Ma) was subsequently found in Mexico on the Yucatán Peninsula (the *Chicxulub crater*). It was dubbed the *Crater of Doom* by Waiter Alvarez, in a book dealing with the mass extinction.

There may have been other craters as well, of course. This would be not surprising if the observed impact of a series of comet-like bodies on Jupiter in July 1994 (Shoemaker-Levy 9)

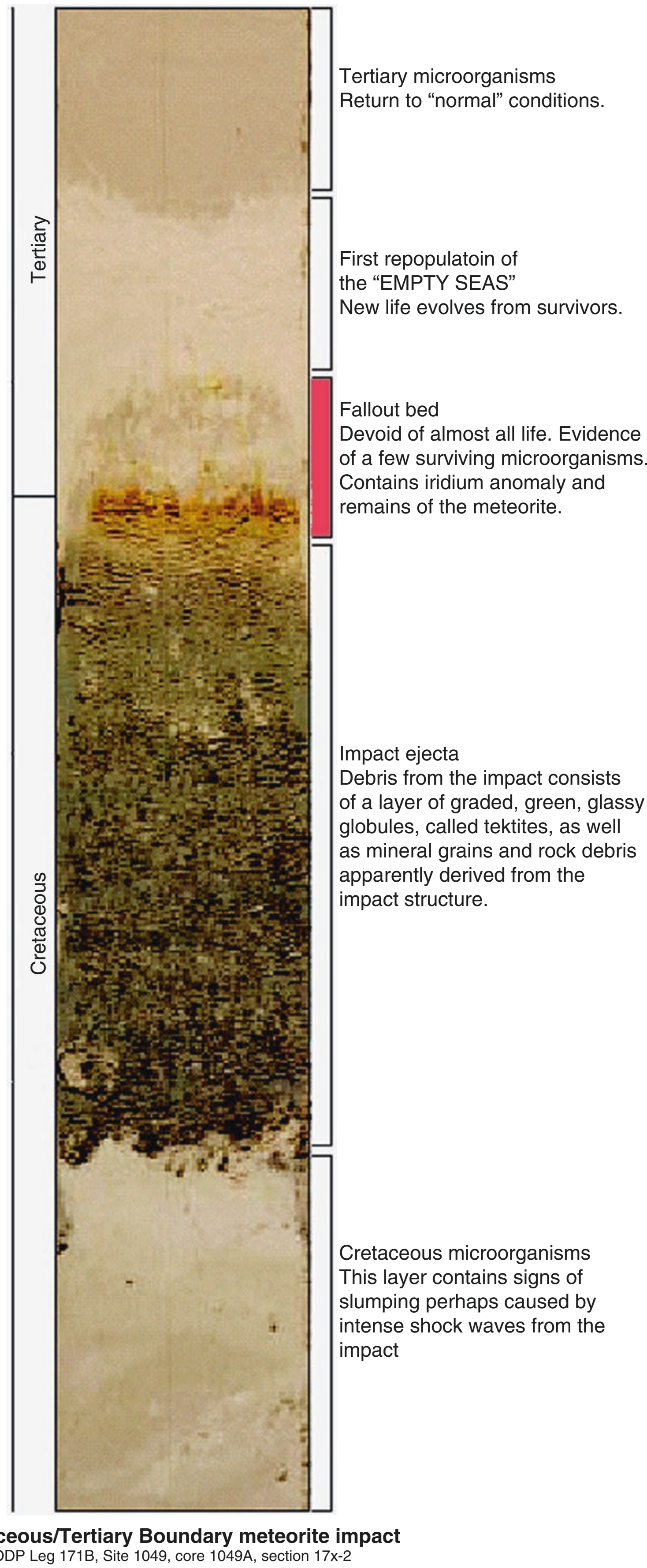

Fig. 13.15 The K-T boundary in a core of the Ocean Drilling Program. Leg 171 B, Site 1049, Core 1049A Section 17 X-2 (R. Norris, 1998. In: R.L. Larson (ed.) ODP's Greatest Hits. Ocean Drilling Program. College Station, Texas)

may be taken as an analog to the inferred K-T impact. One extremely large crater-like feature of the right age has recently been reported from off western India. If it is indeed an impact crater at the K-T boundary, the crater in Yucatán would have had a companion much larger than itself. The size of a crater may not be the only thing that matters, though. One would like to know what was hit and with what effect. In any case, the timing of the proposed impact in India coincides with a great outpouring of basalt (known as the "Deccan Traps" to geologists).

The mixture of processes responsible for dealing widespread death to extant organisms of the latest Cretaceous is not known. A prolonged darkening of the sun (from particles in the stratosphere), acid rain (from sulfurous impact sources), and drastically elevated temperatures (from the addition of greenhouse gases) are possible stress factors that have been discussed. Another one with some credentials in the timing is volcanism (the "Deccan Traps"). Regardless of the detailed mechanisms causing the great extinction, as far as life on the planet, the main effect of the event seems to have been to "clear the table" and let new organisms replace the types that dominated the scene, perhaps for a hundred million years or more. Diversity recovered in a few million years and resumed (surprisingly) a previously pursued upward trend, as the planet cooled.

Much has been written about the reluctance of established geologists to accept the new ideas emerging from the impact hypothesis. Surprise at such reluctance, of course, may expose ignorance about how regular science works rather than illuminating shortcomings of the people whose work is being discussed. The first task for any trained scientists is to sort crackpot ideas from acceptable ones. Once new ideas are well supported by evidence and appear in acceptable technical articles and textbooks (as the Alvarez hypothesis did), opposition wanes. What remains is a healthy skepticism toward confident assertions about what happened in the distant past which has to be reconstructed from indirect clues.

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14 Resources from the Ocean Floor

14.1 Background

14.1.1 Principles and Expectations

The one resource originating in marine deposits and also of prime importance economically at present is *petroleum*. Energy use is basic to a flourishing economy, so that the item has traditionally received much supportive attention both from elected politicians and from other decision makers in the industry and elsewhere. However, in recent decades, it has become obvious that the use of carbon-based fossil fuels involves uninsurable risks. Manageable and insurable risks are largely confined to activities concerning obtaining the resources. The risks with regard to energy waste disposal (mainly carbon dioxide), however, are much larger, involving unwanted climate change, as well as sea-level rise and other associated serious problems.

Generally, when geological services are in demand (including those for marine geology), they increasingly involve dealing with conflicts stemming from the danger of pollution and other topics concerning the public interest. Problems related to climate change have become especially prominent, since the burning of carbon for obtaining energy produces an important greenhouse gas that interferes with the natural radiation balance.

To be sure, a growing demand on carbon-based energy (regardless of any associated downsides) presumably will dominate the economic scene yet for some time. One might assume (based on what is happening) that oil and gas will remain the most sought-after resources from the seafloor, whatever our individual preferences or insights are concerning what the future might be. To emphasize the importance of less-discussed but similar issues, we note that the need for waste disposal likewise will keep growing. Disposal of waste at sea can represent considerable economic benefit to the dumper (and dispersed environmental risk to all others).

Conflicts are common. Geologists involved with conflict resolution are likely to be faced with extremely difficult problems arising from urgent requirements. For example, cities produce waste, and they need to deal with the pileup, preferably in the cheapest possible way. On the other hand, the tourist industry needs clean beaches and a clean ocean for its customers if it is to stay in business.

Getting an acceptable beach in the first place, while simple in principle, may turn out to be rather expensive. Who is to bear the cost? The search for cheap and suitable offshore sand to use for beach replenishment will acquire increasing urgency as sea level rises from the melting of ice on the planet, with considerable impacts on the tourist industry and the safety of nearshore structures.

From an economic perspective, a shallow resource is better than a deep one, because access is better (i.e., cheaper) in shallow than in deep places. Inasmuch as many important resources represent growing biological products, they are linked to shallow (sunlit) waters and hence tend to be rather accessible, that is, they are found largely in certain shelf basins and also on the upper continental slope. The fact that many resources share this property of shallowness increases the likelihood of conflict, of course.

For hydrocarbons, there is a requirement for maturation from the organic matter originally deposited, which implies an abundance of heat and geologic time, that is, special circumstances. Obviously, where tar or other petroleum products ooze out of the ground, such geological requirements are likely to have been met (Fig. 14.1), and the resource may safely be assumed to be abundant in the underground.

14.1.2 On the EEZ

The EEZ is a fact that requires paying much attention to whenever exploring the seafloor anywhere near the shore. The

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_14

Fig. 14.1 Petroleum in the seafloor of Southern California. *Left*: tar seeping from the beach cliffs at Carpenteria; *right*, offshore production rig off nearby Santa Barbara (Photos W.H.B.)

"exclusive economic zone" has its origin quite obviously in economics. The EEZ denotes a 200-nautical mile zone adjacent to national boundaries. Established by agreement between nations in 1982 (United Nations Convention on the Law of the Sea), it restricts the right of geological exploration and any resource extraction on areas next to national boundaries to the nation owning the EEZ. In essence, since hydrocarbons are the one resource that really matters, the convention assigns the rights to explore for or produce hydrocarbons from the seafloor adjacent to a country. To be sure, fisheries played a role in introducing the concept and are still important in places. But the focus of the EEZ is largely on hydrocarbons.

The EEZ is of interest in all marine geological work that occurs within or close to 200 miles of land, including islands. Permission must be obtained from the EEZ owner for any exploration within the EEZ (including seismic profiling, sampling, and drilling, whether or not for commercial purposes). Permits are obtained commonly well ahead of an expedition. The perception is that geological knowledge is an important economic factor, whatever the intent of obtaining it. Thus, there is a strong motivation for controlling the creation and dispersion of the type of knowledge produced by marine geologists.

For many barren islands, especially small ones, the EEZ seems to be the one item that matters to the owning nation. Presumably, small barren and surf-swept rocks defining the EEZ are of little or no interest by themselves in most circumstances. It is the control of the surrounding seafloor that counts.

14.2 Petroleum Beneath the Seafloor

14.2.1 A Focus on Petroleum

The chief reason for the great interest in petroleum from the seafloor is economic: it is possible to extract petroleum profitably in large amounts (as long as proper safety rules of extraction are followed). Besides, demand on fossil energy sources is rising markedly, decreasing chances that eventually excess supply will depress prices to unacceptably low levels. Also, the earnings by the world's oil production are measured in trillions of dollars. They are in the same category as national incomes even of large nations. Offshore production plays a significant role in this, with revenues in the USA reported in the tens of billions of dollars per year and job creation in the tens of thousands. Demand is generally increasing globally for petroleum, gas, and coal, regardless of expert warnings and of any expressed national preferences for non-carbon energy. The discrepancies between perceived economic necessity (decades) and planetary reality (which includes millennia and centuries and even millions of years) obviously result in confusion and also in conflicts. Commonly, both confusion and conflict interfere with the goals of international climate conferences. Avoidance of tough decisions is another problem. For the sake of clarity, confusion, conflict, and avoidance, so far, have prevented the adoption and execution of the type of economics acceptable to climate scientists (i.e., an economy that avoids carbon-based energy).

The "exclusive economic zone," with resources commonly in shallow waters and hence quite accessible, naturally has been an object of intense scrutiny for hydrocarbon resources over the last several decades in many nations. The payoff sought is measured in terms of supply of energy that is considered economically affordable, in terms of jobs created and of tax monies collected. Economy-based concerns about conflict of carbon-energy use with the tourist business and with regional fisheries have led, in some cases, to significant restrictions on offshore hydrocarbon development in the EEZ. Concerns about undesirable effects on global climate change of carbon-based energy use – although well recognized as an issue by relevant scientists and by many politicians – apparently have had much less recognizable impact on policy.

In 2015, world production was about 5 billion tons of petroleum (up from 3 billion near the end of the last century). Prices were highly variable (they are measured in US dollars per barrel, a unit seven times smaller than dollars per ton). The total value was well in excess of a trillion US dollars (i.e., more than a thousand billion). In addition, there were 3 trillion cubic meters of natural gas production (at a price of around $50 per thousand cubic meters). Around one half of the oil used and more than one fourth of the "natural gas" (mainly methane) came from offshore sources. The discovery that much oil and gas previously unavailable can be extracted from known (and previously exploited) hydrocarbon and carbon deposits by creating new pathways for the movement of fluids and gas underground has changed all assessments for hydrocarbon resources in the ground and their accessibility. For many instances, invoking the possibility of *fracking*, the availability of water may constitute a severe limitation, one that presumably would not apply in an offshore setting. "Fracking," even well offshore, may however result in conflicts with users of potable groundwater close to the shore, largely for similar reasons as apply on land.

14.2.2 Methane Ice and Hydrate

Conventional hydrocarbon energy resources have a new potential partner: methane ice, called methane *hydrate* by geochemists. "Hydrate" is the name for ice that holds large amounts of gas in its structure, which is much more accommodating than would be regular water ice (which contains some air, mainly as bubbles). The gas of interest commonly is methane (CH_4), but other types of gas can be at issue. Thus, hydrate has been discussed in the context of sequestration of carbon dioxide on the deep seafloor. Almost all natural gas hydrate resources in the sea are thought to occur below the seafloor close to continents. (Petroleum reservoirs on land, coal, and permafrost masses have large potential resources of this type, as well.) So far, commercial exploitation of marine methane is not much in evidence, presumably mainly because generating natural gas from hydrates and transporting it to market, while possible, are not sufficiently profitable at this time to spread widely and vigorously.

Finding a way to exploit hydrates for energy use is commercially attractive: many professional geologists concerned with the matter estimate that the abundance of methane in gas hydrates greatly exceeds the volume of known conventional gas resources in coal. Since we do not know how much oil, gas, and coal is in the ground, and how much clathrate is below the seafloor, such comparisons are difficult to make. What we do know is that the burning of natural gas, whether from clathrates or from elsewhere, produces carbon dioxide, which warms the planet and changes the climate. It is true, though, that the production of the greenhouse gas carbon dioxide from methane is considerably less for the same amount of energy than that produced from burning coal or oil (coal, including the so-called clean coal, has the highest output of carbon dioxide per unit of energy produced among the fossil carbon-based energy sources).

14.2.3 Risk Assessment

Exploitation of resources invariably implies economic risks other than the enormous ones linked to anticipated climatic effects of carbon dioxide release and the (largely unknown) consequences of *acidification*. Recovery may fail, or that what is recovered may not bring a profit, or costly damage may accompany recovery attempts or transportation. Such risks are taken very seriously: large amounts of money are potentially involved. The risks addressed (and insured against commercially) are not necessarily the largest ones, however. Insurable risks can be estimated from history and therefore can be handled reasonably well by the insurance industry. Future costs of climate change are not in that category.

Climate risks are widely ignored (although there is plenty of discussion about the subject). A general list of potential problems is daunting and extremely difficult to work with, both for political and for scientific reasons. A scientific focus on one obvious and ubiquitous type of damage (say, involving the rise of sea level) might help in that situation, putting emphasis on an item of great interest and hence useful in education. Educating the public about possible problems ahead is definitely a necessary and welcome activity. But confounding risk and fact in the process would be unhelpful. Such confounding may result in severe discounting of the most important problems faced by humanity.

One source of risk is accidents, especially those that result from failure in safety measures. Much attention has been paid to the danger of oil spills, based on experiences with a blowout off Santa Barbara (Union Oil Platform) in 1969, the Amoco Cadiz running aground off France in 1978, and the Pemex blowout event in Campeche Bay (Ixtoc I; Gulf of Mexico 1979) among others. The Exxon Valdez hits rocks off Alaska in 1989, causing much local damage. In 2010, the extremely expensive Gulf of Mexico Deepwater Horizon accident occurred, involving BP, a major oil company. Damages for the largest spills are commonly measured in hundreds of millions of dollars; for the Deepwater Horizon event, it was billions (many thousands of millions). As a result of that event, whose cost greatly exceeded all others, a focus on progress in technology for the reduction of risk from spillage may have lost much credibility. Instead of a focus on how to drill, a focus on where to drill may be called for. Such an emphasis has produced a perfect safety record for DSDP and ODP. The difficulty regarding general application of the types of safety measure used in scientific drilling to commercial drilling is obvious: scientific drilling explicitly avoids hydrocarbons.

Oil spills, while highly visible, are not the only commercial risk in the resource business, of course. There are many others, even when just concentrating on hydrocarbons. Some risks are related to various political developments, which can impact profitability. Whether such risks are insurable is another question, one that is certainly beyond the scope of marine geology.

14.2.4 Origin of Petroleum

In general, the conversion of organic matter to commercial hydrocarbon is rather inefficient. We have to ask why this is so. A number of items are at issue, beginning with the organic chemistry of hydrocarbons (Fig. 14.2).

The production of commercially important hydrocarbons is a complicated process, as is evident in the listing of requirements that follows. (1) Organic matter captured within marine sediments must be converted to fluid petroleum by thermochemical processes, commonly requiring a thick blanket of sediments (more than 1 km thick) and temperatures of 50–150° Celsius. To be sure, long reaction times can compensate for relatively low temperatures, and high temperatures can compensate to some degree for a thin sediment cover. This is demonstrated, for example, in the *Guaymas Basin* in the Gulf of California, where hydrothermal activity produced oil from sediments less than 5000 years old. At excessively high temperatures, however, some of the reactions do not proceed to make oil, and we get natural gas instead (Fig. 14.2). (2) Petroleum must migrate from organic-rich source rock sediments to porous and permeable reservoir rock such as sandstones or porous limestones if it is to be recovered in conventional ways (Fig. 14.3). (3) Reservoirs must be big enough to be of commercial interest, and they must trap the petroleum with impermeable cover rocks such as shales or evaporates. Otherwise, the more volatile hydrocarbons can escape to the surface and to the atmosphere, which results in pitch lakes and tar accumulations (familiar to Southern Californians from the *La Brea tar pits* in Los Angeles). Tar also occurs in some of the sea cliffs in Carpenteria near Santa Barbara (Fig. 14.1). (4) The several petroleum-forming processes must take place within the correct time

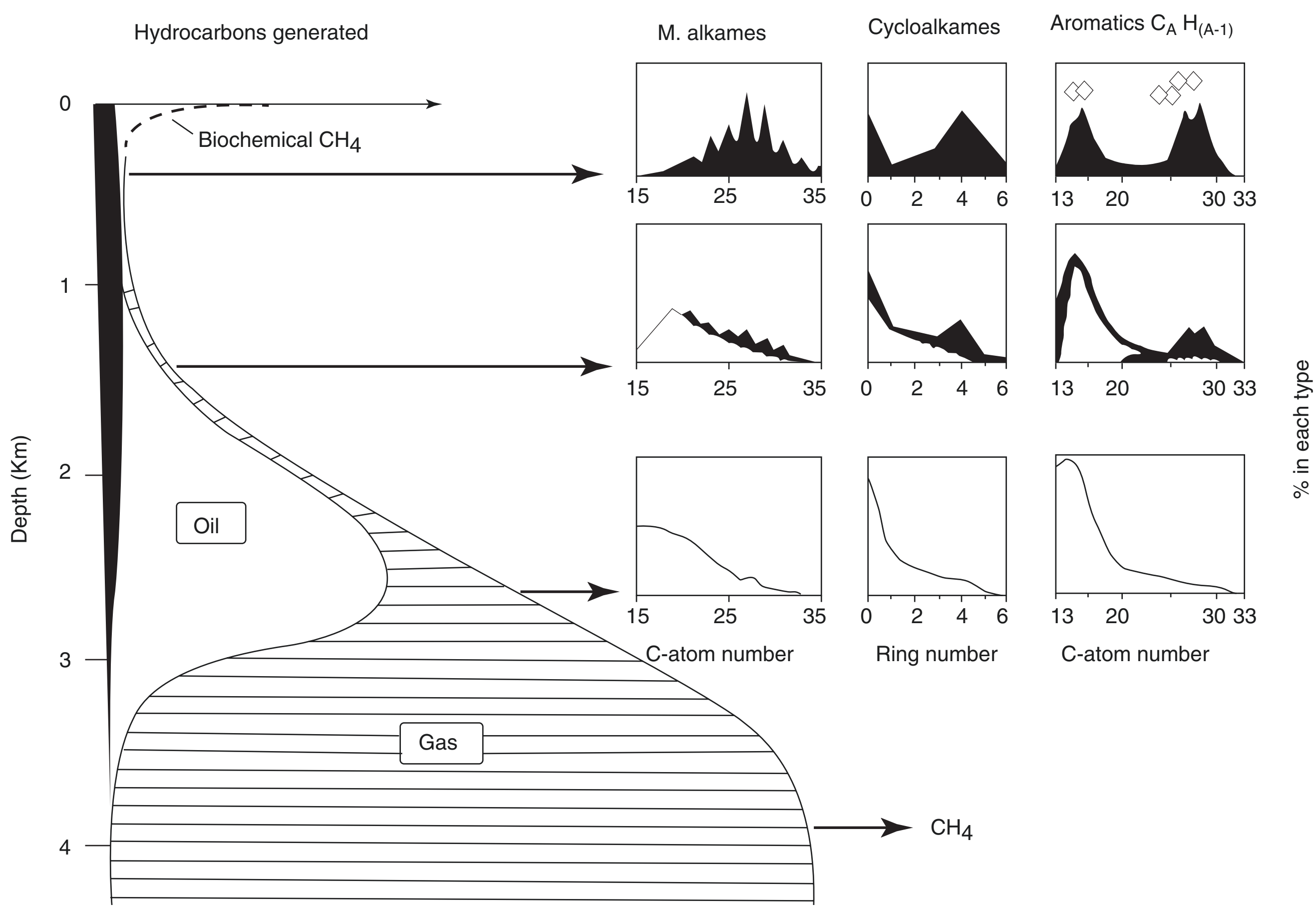

Fig. 14.2 Origin of petroleum. After burial of the source material (*black*), carbon chains and rings are produced. Gas is produced under a large overburden and can help push the oil out during production (After B.P. Tissot and D.H. Welte, 1978. *Petroleum Formation and Occurrence*. Springer, Heidelberg)

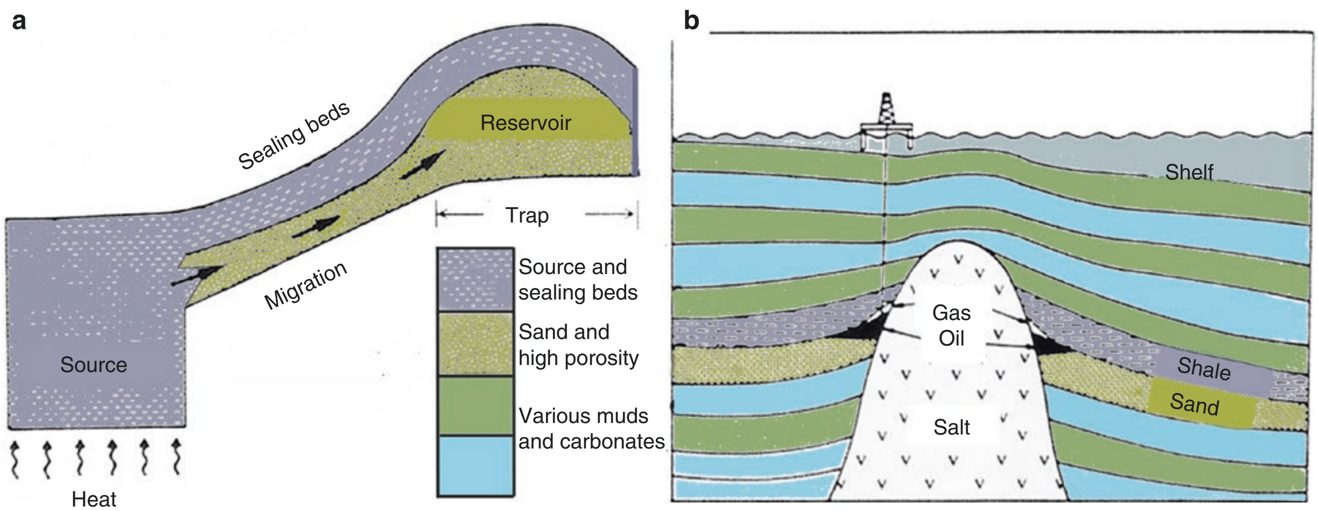

Fig. 14.3 Schematics on hydrocarbon accumulation. (**a**) Basic elements of a petroleum reservoir system, with migration from heated source rocks to a porous reservoir. (**b**) Offshore salt dome tectonics generating trapping conditions

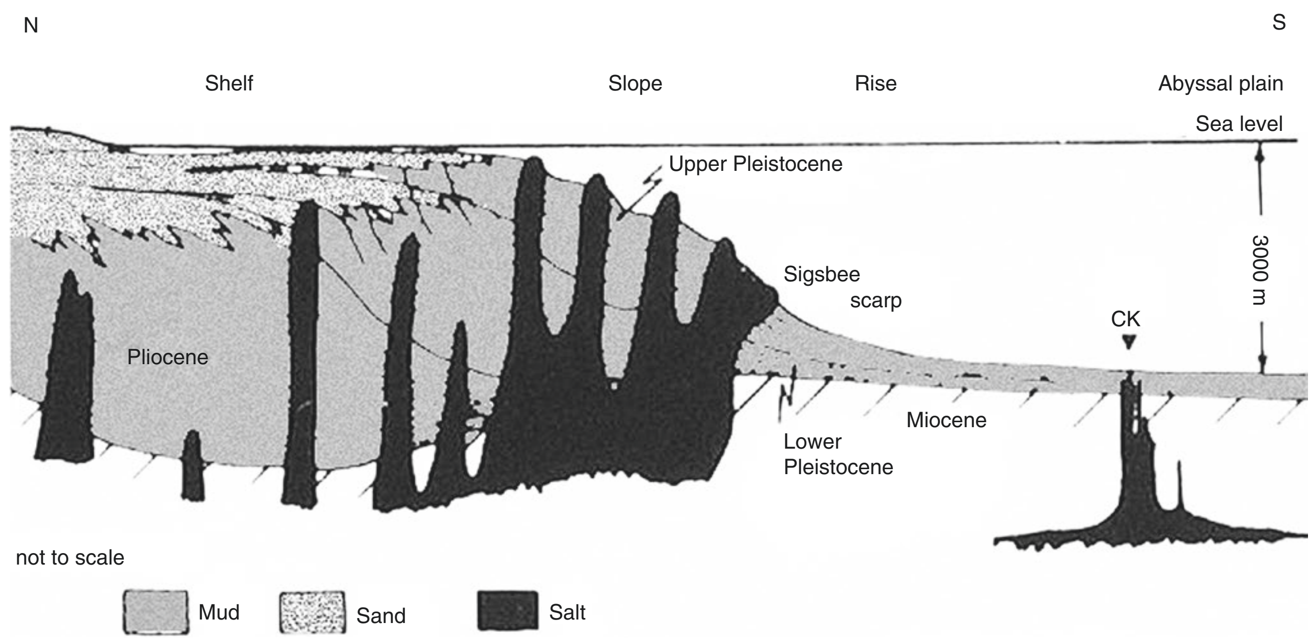

Fig. 14.4 Salt plumes in the Gulf of Mexico: Mesozoic salt intruding Neogene mud (After C.J. Stuart and C.A. Caughey, 1977, *AAPG Geol. Mem.*, 26, modif.; "CK", Challenger Knoll, a salt dome below the continental slope (position from DSDP Leg 1))

frame to result in a usable resource. Each process needs to complete its turn in sequence. A drill hole might be "dry" even though conditions seem right for successful exploitation, because the *timing* in the interplay of the natural processes was wrong.

14.2.5 Where Offshore Oil Is Found

The conditions of entrapment of oil in the marine realm are much like those on land (Fig. 14.3). Marine sediments can overlie salt deposits that are gravitationally unstable, being less dense than the overburden. The salt pushes up in plumes, making a so-called *salt domes*. Such domes are common at depth in the Gulf of Mexico (Fig. 14.4). Upturned sedimentary strata butting against the salt can provide traps for hydrocarbons (as shown schematically in Fig. 14.3b). The label "CK" in Fig. 14.4 denotes the "Challenger Knoll," which was drilled during the first leg of the Deep Sea Drilling Project (Hole 2). In subsequent drilling legs, such salt dome targets were avoided for safety reasons, as were any dome-like structures on the continental slope. In the well seaward of the continental slope, the sediment cover is generally too thin (and productivity too low) to cause serious concern in regard to hydrocarbons.

Offshore oil has been produced from the *Continental Borderland off Southern California*. There, the oil originates from Miocene organic-rich layers originally formed in upwelling areas and under conditions of oxygen deficiency. The oil is trapped in sandy layers ending at faults, which can promote considerable natural seepage into the sea (Fig. 14.1). On the *Arctic shore of Alaska*, there also are highly eligible areas for the recovery of offshore petroleum, and extraction is proceeding in Prudhoe Bay. Climatic conditions are very inclement, however, making exploration and production hazardous.

Thanks to recent technological developments' boosting recovery, the area of the North Sea (including portions of the Norwegian Sea), Europe's chief source of marine hydrocarbons, is still a large potential source, in spite of many years of vigorous hydrocarbon extraction. It has thick sediments, with source rocks common in the Mesozoic, and much storage in the early Tertiary sands. Permian Zechstein salts are involved in producing salt domes (Fig. 14.5). Thus, Phanerozoic sedimentary rocks of all ages combined forces here in generating the resources. Of the various western European nations involved in exploiting the resources, the UK and Norway have had the largest shares. In 1989, production of oil was some 175 million tons (53% from UK concessions, 43% from Norwegian ones) and 150 billion cubic meters of natural gas (30% UK, 21% Norway, 48% Netherlands, including onshore sources relatively close to sea level).

Production has waned since the turn of the millennium, when it may have attained maximum recovery for the oil (*peak oil*). Thanks to new technology, future recovery of hydrocarbons may be of the same order as recovery in past decades. To be sure, the *decommissioning* of existing oil fields has started (abandonment has to follow strict safety rules); nevertheless, projections regarding recovery are optimistic. However, the details of hydrocarbon occurrence in the North Sea region are complicated involving a host of sources and traps. In any case, future production is notoriously difficult to predict accurately.

In the south of the hydrocarbon-rich region, a broad belt of gas fields stretches east-west from Germany to southern England. Gas was once migrated from Carboniferous coal measures to porous Lower Permian sandstones and was sealed by overlying Zechstein (upper Permian) evaporates. Oil fields, in contrast, are concentrated around the north-south rift in the northern North Sea, which is reported to have originated in the first stages of opening of the northern North

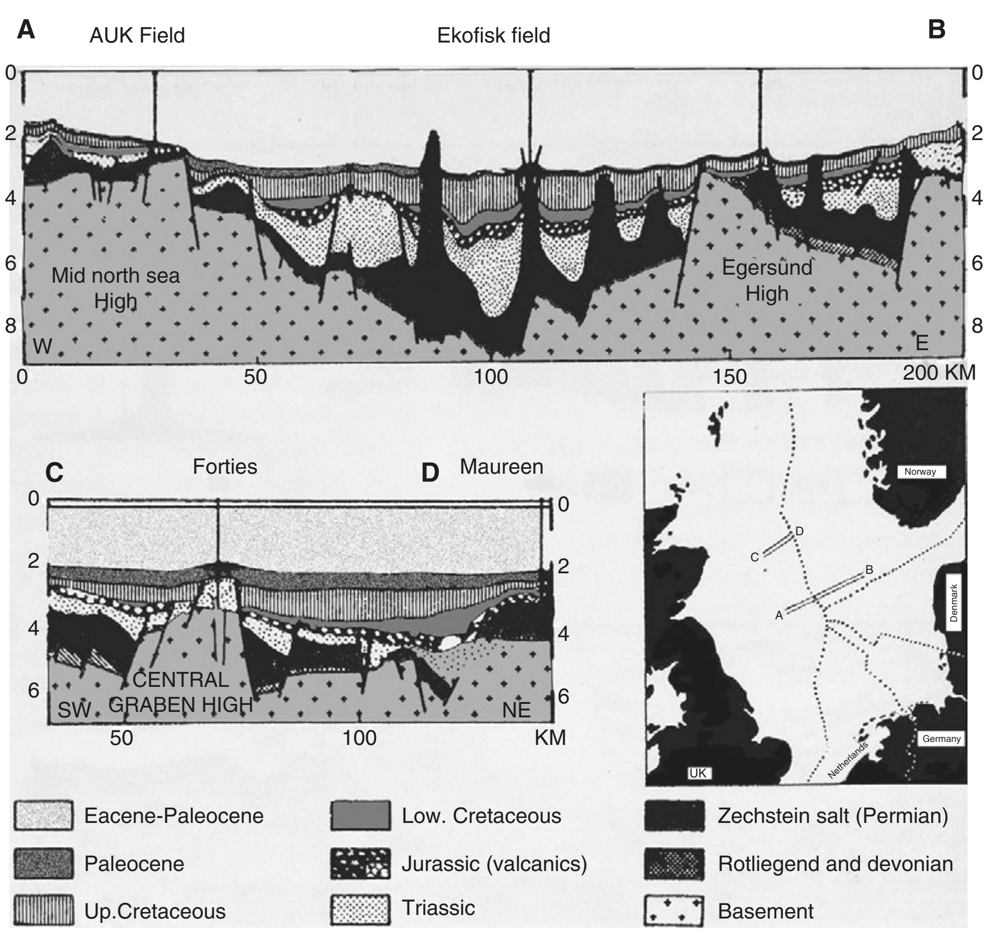

Fig. 14.5 Geological profiles under the North Sea, with hydrocarbon fields. Vertical exaggeration 6.7×. *Insert*: Map with national shelf boundaries (After P.A. Ziegler, 1977, modified for clarity. See *Geol. Journal* 1:7)

Atlantic. In the south, the oil occurs in Cretaceous chalk reservoirs, next to salt domes. Some fields are developed on Jurassic sandstones on the crests of horsts and tilted blocks. The Forties and other fields get oil from reservoirs in basal Tertiary sands.

We see that offshore oil, for reasons having to do with high organic production and thick sediment cover, is most likely found in shallow water close to continents. It is here that one finds thick sediments (Fig. 14.6). The locations are such that heavily populated regions can readily generate issues adverse to untrammeled oil extraction, starting with considerable objection to hydrocarbon exploration from fears centered on oil pollution. Experience shows that such fears are not without justification.

In addition to sediment thicknesses being unimpressive at great oceanic depths and organic matter content relatively low, both the risks associated with drilling and with production and the cost of recovery of hydrocarbons increase greatly with depth of water. In short, in deeper water, successful exploitation is increasingly less likely. Thus, shallow-water regions are and likely will be targets for commercial exploration. Minima of people aggregations (and associated conflicts) are located in icy polar areas. Thus, activities in shallow waters in high latitudes are likely going to define the future of petroleum recovery from the seafloor in the foreseeable future, despite the very palpable risks arising from inclement weather. Of course, concerning the projected damage to climate from hydrocarbon burning, the source regions for the hydrocarbon used do not matter.

14.2.6 Offshore Methane as a New Energy Source?

Much attention is being paid to methane as a new energy source with a reduced output of carbon dioxide upon burning relative to coal and to oil. So far, promises have been slow in being met. Methane is, however, an important risk factor in global warming. Judging from the stability relationships identified by geochemists (Fig. 4.5), a warming of the deep ocean along the slopes might set free much of the methane now safely shut into icy cages. The result of freeing the methane depends on the fraction converted to carbon dioxide by microbes and the rate of such conversion. Methane is much more powerful as a greenhouse gas than the oxidized gas (a factor of 25 has been suggested on the century scale). Presumably, fast release implies less conversion than slow release. Again, the time scale matters. A lack of information on rates of destruction at various temperatures and pressures within slope sediments may contribute to difficulties in predicting likely developments following the release of methane.

We have mentioned clathrates as a type of chemical sediment, gas accommodated in water ice with a very open structure. Much gas is trapped below the clathrate level where temperatures are low enough on the continental slope for clathrate to exist (Fig. 4.5). Thus, much of the gas may be at temperatures too high to make clathrate, that is, a large proportion of the methane in the system may be close to the stability zone, hard to recover but sensitive to disturbance.

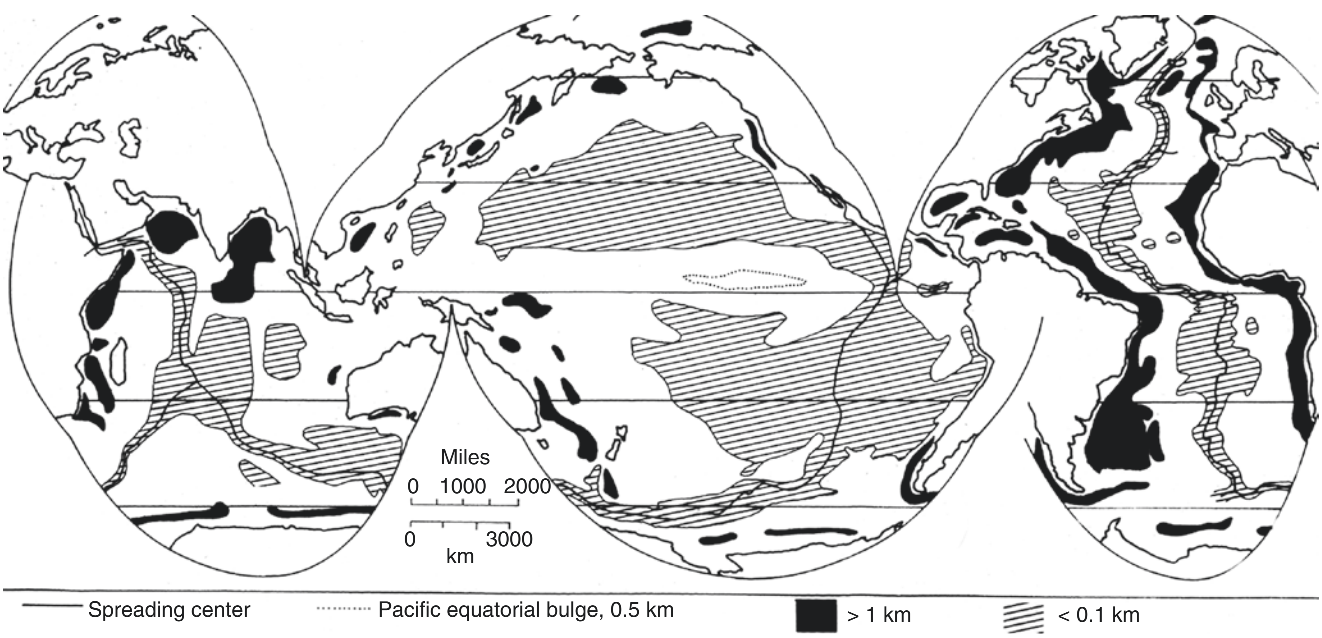

Fig. 14.6 Regions with thick and with thin sediments in the ocean (W.H.B., 1974. In: C.A. Burk and C.L. Drake (eds) *The Geology of Continental Margins*. Springer, Heidelberg Berlin)

Any planned removal of clathrate for commercial purposes has to be assessed with this problem in mind. On the seafloor, one can see places of presumed methane that exits the sediment in so-called mud volcanoes even without any provocation (Fig. 4.6). Presumably, one would not wish to increase the number of such "volcanoes."

14.3 Solid Raw Materials from Shelves

14.3.1 Phosphorites

Ground-up phosphatic rocks and products resulting from reactions with sulfuric acid are used as fertilizer in agriculture. Consumption is measured in millions of tons. Marine sources play a relatively minor part, although much of the phosphorite (*phosphate rock*) mined is marine in origin. Offshore (Neogene) phosphorites typically contain less than 25% of P_2O_5.

Modern marine phosphorites have a general formula containing the elements calcium and the bases phosphate and carbonate. They are, on the whole, fluoride- or hydroxide-bearing apatite rocks. The rocks at issue typically occur in areas of high productivity – off Southern and Baja California, for example, and in other upwelling areas (e.g., off Peru and Namibia, see Fig. 14.7). They occur as black or brown nodules forming pellets with various diameters (up to head size) and as irregularly shaped cakes. In places, they look like replacement products of fine-grained carbonate rock. In other places, they are interpreted as mineralization of preexisting organic matter, as precipitates from microbes filling cavities within carbonates or as direct precipitates from interstitial water.

Marine phosphorites on the present seafloor are typically Neogene in age (i.e., post-Paleogene), with the Miocene strongly represented. They are found in upwelling regions (Fig. 14.7). However, the early Tertiary phosphatic rocks and even Cretaceous ones are commonly associated, even though their origin may differ from the Neogene types. In many cases, it is not clear why phosphates of Neogene age tend to occur together with much more ancient ones. Possibly some of the conditions favorable for the making of Neogene phosphatic rocks (i.e., high productivity) also enhance the chances for preservation of preexisting older phosphates. The origin of phosphates on seamounts is not necessarily obvious. Some of the phosphates may stem from a time when the seamounts were islands and had bird-producing guano and were surrounded by a strong oxygen minimum zone. In any case, ancient phosphatic rocks on seamounts presumably did not form in the conditions in which they are found at present. In Fig. 14.7, only the Neogene and non-seamount phosphates are shown.

The common depth of deposition of phosphatic materials is on the shelf and upper slope. For reasons discussed in Chap. 7, productivity is high here, and oxygen may be in short supply (a condition of phosphorite formation according to a pers. Communication of the marine geochemist Y. Kolodny in Jerusalem). Also, there is a chance for considerable rework-

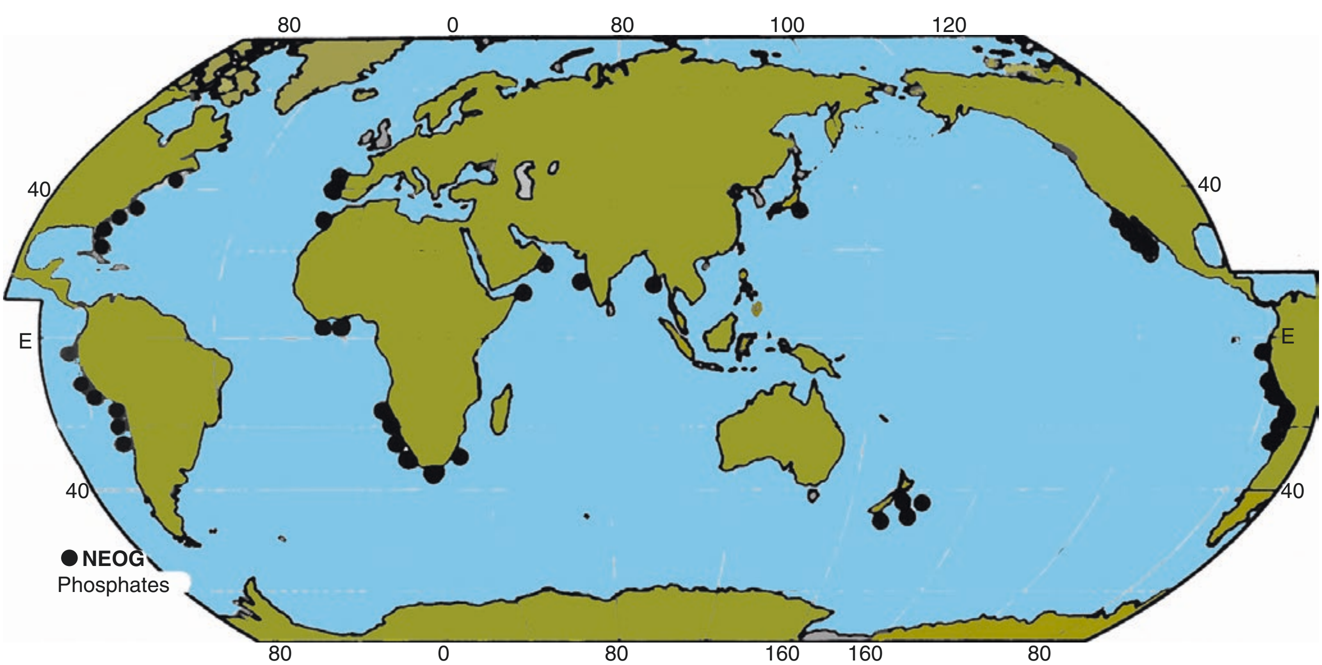

Fig. 14.7 Distribution of Neogene phosphorites according to G.N. Baturin and P.L. Bezrukov, 1979 (*solid circles*). Omitted compared with original graph: phosphatic rocks (regardless of age), phosphate on seamounts, and much of the longitude-latitude grid (Marine Geol. 31:317; modified) (note the importance of upwelling areas in focusing occurrences of Neogene phosphate rocks)

ing (and attendant concentration of the resource) from drastic changes in seawater cover owing to pulsed ice buildup and ice decay. In many places, shallow marine phosphate deposits are accessible on land (e.g., in Florida and in Morocco) where mining is economically inviting. About two thirds of the US production of phosphatic rocks is from Neogene sources in Florida, enriched by weathering processes. The mineral "francolite" (a carbonate-fluorapatite) is the chief carrier of the sought-after phosphorus here. Until recently, the USA mined more phosphate rock than any other country (almost 30 million tons in 2011, satisfying more than 90% of its own demand), but some other large nations have caught up. Morocco is the chief exporter of phosphorite. It has abundant weathering-enriched phosphate rocks of the early Tertiary and Mesozoic age from the ancient Tethys seaway.

The association of geologically young phosphorite deposits with present-day regions of upwelling suggests that the source of the phosphorus is organic matter (and that the relevant upwelling process is geologically young, which agrees with the 10 million-year age commonly quoted for the beginning of strong upwelling). During decomposition of organic debris on the seafloor, phosphate is released and becomes highly concentrated in interstitial waters. Apatite precipitation can then proceed, as well as diagenetic replacement of preexisting carbonate minerals. Off Peru, nodules with diameters of several centimeters grow within soft sediments with a rate of several millimeters per millennium, suggesting upward diffusion of dissolved phosphorus. The resulting phosphatic concretions are resistant to transport by currents; they can be mechanically exhumed and concentrated during periods of sediment reworking. Indeed, concretions are commonly associated within the sediment with hiatuses or other discontinuity surfaces with drastically reduced net deposition rates. Alternatively or in addition, particles can be sorted by size and density. Turbidity currents, in places, have concentrated phosphatic particles in layers in the (Miocene) Monterey Formation, for example.

14.3.2 Calcareous Shell Deposits

Calcareous shell deposits were or are dredged in places as raw material for calcium carbonate and for obtaining road-building material. For example, oyster shells have been mined in San Francisco Bay and in Galveston Bay in the Gulf of Mexico, as well as in Chesapeake Bay. Unfortunately, the large-scale removal of oyster shells apparently damages existing oyster reefs (by removing suitable sites for oyster spat). In addition, the dredging for shells may adversely affect the productivity of the seabed. In any case, conflicts commonly arise where shell dredging and fishery activities overlap.

Pretty shells have become collection items for tourists. With the advent of worldwide souvenir markets and of face mask diving, the gathering and sale of shells and of corals are now an important source of income for many islanders and other coastal peoples. Not surprisingly, attractive and rare species suffer considerable depredation from such collecting. There is safety in looking drab. Hence, according to Darwinian selection, the relative abundance of drab specimens is likely to increase in the years to come, while that of pretty shells is bound to decrease further.

14.3.3 Metals, Heavy Minerals, and Diamonds

Concentrations of heavy minerals and ore particles on beaches and in estuaries are locally mined for metals such as titanium, gold, platinum, thorium, zirconium, and tungsten and for valuable minerals such as diamonds. Some of the material is in uplifted beaches and in dunes. Seventy percent of the world production of zirconium is being extracted from placer deposits off eastern Australia (what is mined here is the Zr silicate). Diamonds are found in beach deposits of Southwest Africa and Namibia, as well as offshore. Magnetite is being mined from beach placers in Japan and New Zealand. Gold has been mined from beach deposits near Nome, Alaska. The big rush was at the very beginning of the twentieth century. Today, tourists wash gold there. Thousands of tons of ilmenite (an iron-rich titanium oxide) were extracted at one time from Redondo Beach, California. Also, titanium minerals were taken from beach sands along the northeastern Florida coast during World War II in "Mineral City" (now Ponte Vedra Beach, a golf resort). The biggest producers at present are Australia, with a million tons, and South Africa, with another million; Canada is third. Beaches in the western Oregon have chromite and other heavy minerals as well as gold and platinum.

14.3.4 Placer Deposits

As mentioned, magnetite is being mined from beach placers in Japan and New Zealand. (Magnetite is conspicuous in many places on beaches in Southern California but is not suited for mining here.)

The mining of black sands supports large operations in several places, including at several beaches in India, Indonesia, and the Philippines (Fig. 14.8). Unfortunately, the removal of sand from beaches, for whatever reason, can have serious consequences, including loss of protection from storm waves. In some places, therefore, mandatory restoration activities are linked to beach-mining permits.

How do placers originate? The process of concentrating heavy particles on the beach has much in common with panning for gold – water motion works on particles with different settling velocities and with different sizes, separating heavy from light and large from small (Fig. 14.9). Thick

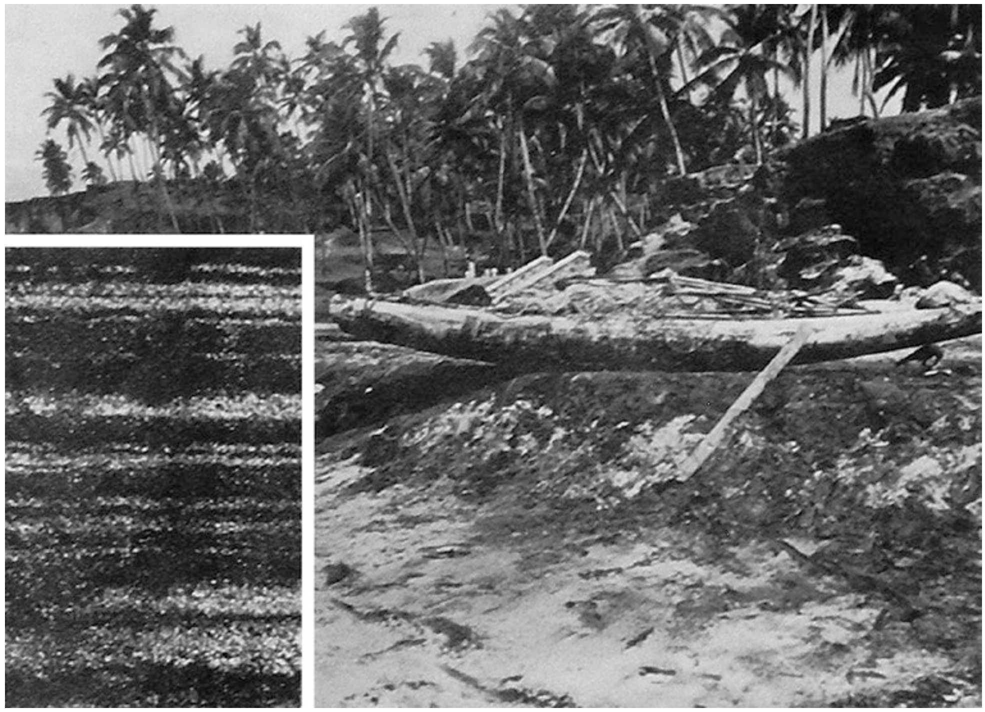

Fig. 14.8 Black sand in beach placers. The beach south of Quilon (southwest India) has a heavy mineral deposit on top of the beach (where the boat rests). During the southwest monsoon, high waves sort the sand into layers of dark and light mineral (*inset*, 20 cm high) (Photos E. S)

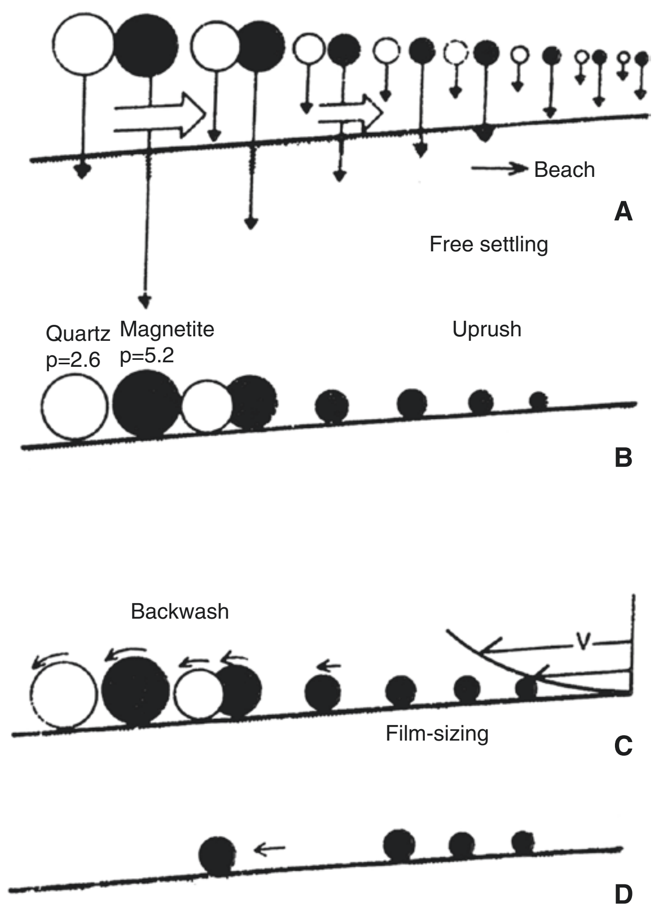

Fig. 14.9 Origin of heavy mineral placers, schematic. (**a**) Suspended sediment particles are washed onshore with incoming waves, large and heavy ones first (*black*). (**b**) Resulting grain association is enriched in grains that settle easily. (**c**) Backwash rolls away large grains preferentially, owing to increase of water velocity off but still close to the floor (note v in *inset*). (**d**) Resulting well-sorted and heavy mineral assemblage (E.S.,1970. *Chem. Ing. Tech*. 42: A 2081)

placers of heavy particles form in the beach zone only. Offshore placers do exist; presumably the ones observed formed when sea level was lower than now. River deposits may contain heavy mineral concentrations also, for example, the cassiterite deposits in Thailand, Malaysia, and Indonesia. In the relevant places, former river beds on the shelf, now submerged down to about 100-m water depth, are of economic interest, and some have been exploited offshore in shallow water for many decades.

14.3.5 Sand and Gravel

Locally considerable amounts of sand may be taken from the beach for building roads and houses, as well as for coastal protection (dams) and for fill. However, the main use of beach sand is indeed for building sand castles and for plain enjoyment. Recreation is one of the most important uses of beaches – by far exceeding mineral extraction in business value. Beaches are commonly eroded by winter storms (Sect. 4.2.2). In some areas, sand may be brought in from offshore to replace eroded material, at considerable expense.

Gravel rarely reaches the sea, unless high mountains are close to the shore or unless we deal with glacial debris. Such debris, when washed by waves or rivers on the exposed ice-age shelves, can readily yield gravel. Around the Baltic and in the North Sea region, for example, gravel is mined as filler for concrete. Taken altogether, the economic importance of sand and gravel and other raw materials from shelves is rather modest. For a high economic impact of marine resources, one must look to hydrocarbons, recreation, and waste disposal, perhaps fisheries.

14.4 Heavy Metals on the Deep Seafloor

14.4.1 Importance of Manganese Deposits

The metal deposits on the deep seafloor have been a popular topic of discussion among marine geologists ever since the *manganese nodules* (better *ferromanganese concretions*) were described by the *Challenger* Expedition more than a century ago (Fig. 14.10) and were shown to be rich in copper, cobalt, nickel, and other heavy metals. The abundances of these metals in seawater are extremely low, largely because their oxides and hydroxides have low solubility. The metals are nevertheless extracted by biological processes and sent to the seafloor with organic matter.

The nodules grow extremely slowly, rates being measured in millimeters per million years. They have growth rings that reflect Cenozoic history with widely spaced events. Both nodules and crusts have been studied. Nodules, some being perfectly spherical, may be subject to (geologically) frequent overturning motion. (Not exactly a dance of the objects on hot springs at depth as visualized by the Bavarian chief geologist W. V. Gümbel more than a century ago. Gümbel was sent some nodules from the *Challenger* Expedition, perhaps by mistake.)

The ferromanganese nodules are quite abundant in places, as seen on photographs of the seafloor (Fig. 14.11). The amount of nodules on the Pacific Ocean floor has been estimated as exceeding one hundred billion tons. However, the economic value of the deposits is not materially greater than zero. It is expensive to mine them and any profits (however unlikely) are at risk: the ownership of the deposits is not clear. While this is not true for certain deposits in the Baltic Sea that are located within the EEZ of various bordering countries, those deposits likewise have little economic value, being not large enough for commercial exploitation and commonly not particularly attractive as concerns associate metals. (In the Baltic region, the Mn and Fe presumably are derived from glacial deposits.)

14.4.2 Origin of Manganese-Bearing Deposits

Where do the ferromanganese deposits occur? How much of the valuable trace metals do they contain? And how did they originate?

Distribution and composition may have some clues to origin. The nodules are generally restricted to areas of low sedimentation rate: they would soon be covered up in regions of high sediment supply. Surprisingly, brown pelagic clay accumulates slowly enough (at 1–2.5 m per million years) to accommodate the nodules (which grow a thousand times

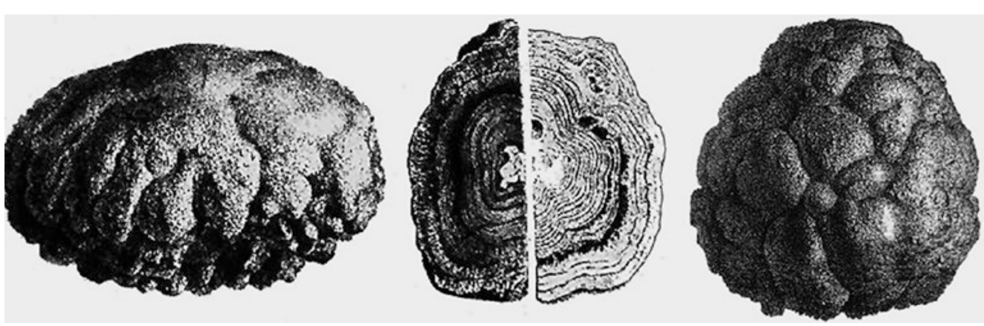

Fig. 14.10 Ferromanganese concretions from the Challenger Reports (The objects shown have diameters of ca. 5 cm). Sectioning of nodules yields growth rings containing information on Cenozoic history

more slowly). Apparently, they are moved about sufficiently by large benthic organisms to stay on top of the clay. In fact, such movement may account for the roundness of nodules and their failure, in many places, to make crust or pavement by coalescing (Fig. 14.11). In places, bottom currents prevent clay deposition or even cause erosion over large areas. It is there that ferromanganese nodules tend to be concentrated. The nodules are more common in the Pacific than in the Atlantic, possibly a result of additional sources in the Pacific and of low sedimentation rates there. Figure 14.11 suggests an important role for sediment cover, regarding the abundance of nodules. Perhaps a temporary covering by slowly migrating sediment waves and by burrowing mounds are more common than is generally realized.

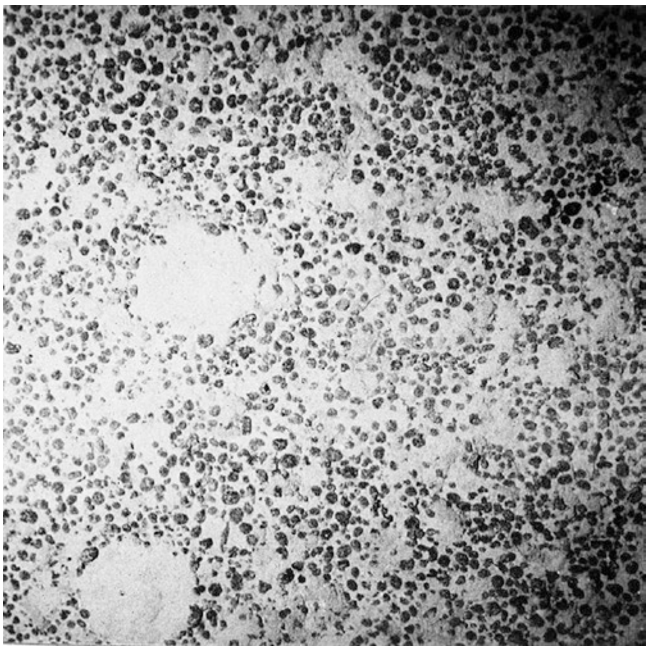

Fig. 14.11 Nodule-covered deep seafloor, central tropical Pacific (Photo courtesy of Metallgesellschaft Frankfurt)

In the various discussions about the origin of ferromanganese concretions (and about their commercial value), the trace element content and the *iron-to-manganese ratio* (its inverse) play an important role. Typical values are listed in Fig. 14.12 in a table. Note that the Mn/Fe ratio is greater in the Pacific than in the Atlantic. Also, the content of trace elements in Pacific sediments is typically considerably larger in Pacific concretions than in Atlantic ones, suggesting differences in origin and in value.

14.4.3 Ores from Spreading Axes

14.4.4 Hot Vents and Black Smokers

For some of the ferromanganese deposits on the deep seafloor, the origin is hardly in doubt: the concretions are located near the crest of active spreading ridges. At the hot vents (and beyond), seawater circulates through cracks in the newly formed crust, reacting with the hot basalt (Fig. 14.13). Precipitation of sulfides (Fe, Mn, Cu, Zn, etc.) originating there may become economically important some day; they are not at present, even though something like a hundred discoveries (among many more) have massive sulfides. Access is difficult and resource volumes are modest, compared with those on land. In recent years, increased interest has been shown by some developing nations in the mining of ridge crest sulfides, especially in back-arc basins where accessibility is enhanced by shallower depths and by the fact that the

%	*Ferromanganese nodules*					FerroMn Crusts	Hydrothermal Oxides	Hot vent sulfides Average	Max.	Min.
	Avg. Pacific	*Indian O.*	*Atlantic*	Max.	Min					
Mn	17.2	14.9	13.6	34.0	5.4	21.6	27.3	-	-	-
Fe	11.8	14.6	15.5	26.3	4.4	16.5	11.6	19.1	34.0	2.5
Ni	0.63	0.38	0.33	2.0	0.13	-	-	-	-	-
Co	0.36	0.31	0.24	2.6	0.045	0.63	0.0023	-	-	-
Cu	0.36	0.17	0.16	2.5	0.028	-	-	2.9	9.2	0.2
Pb	0.047	0.053	-	0.51	0.046	-	-	0.71	12.1	0.03
Zn								16.9	36.7	4.0
SiO2								10.2	28.1	1.2

Fig. 14.12 Metal contents of manganese nodules, crusts, and hot vent sulfides (Sources: J.S. Tooms, 1972. Endeavour 31:113; F.T. Manheim and C.M. Lane-Bostwick, 1989. *Nature* 335:59; S.D. Scott, 1991. In: K.J. Hsü and J. Thiede (eds.) *Use and Misuse of the Seafloor*. Wiley, Chichester, UK)

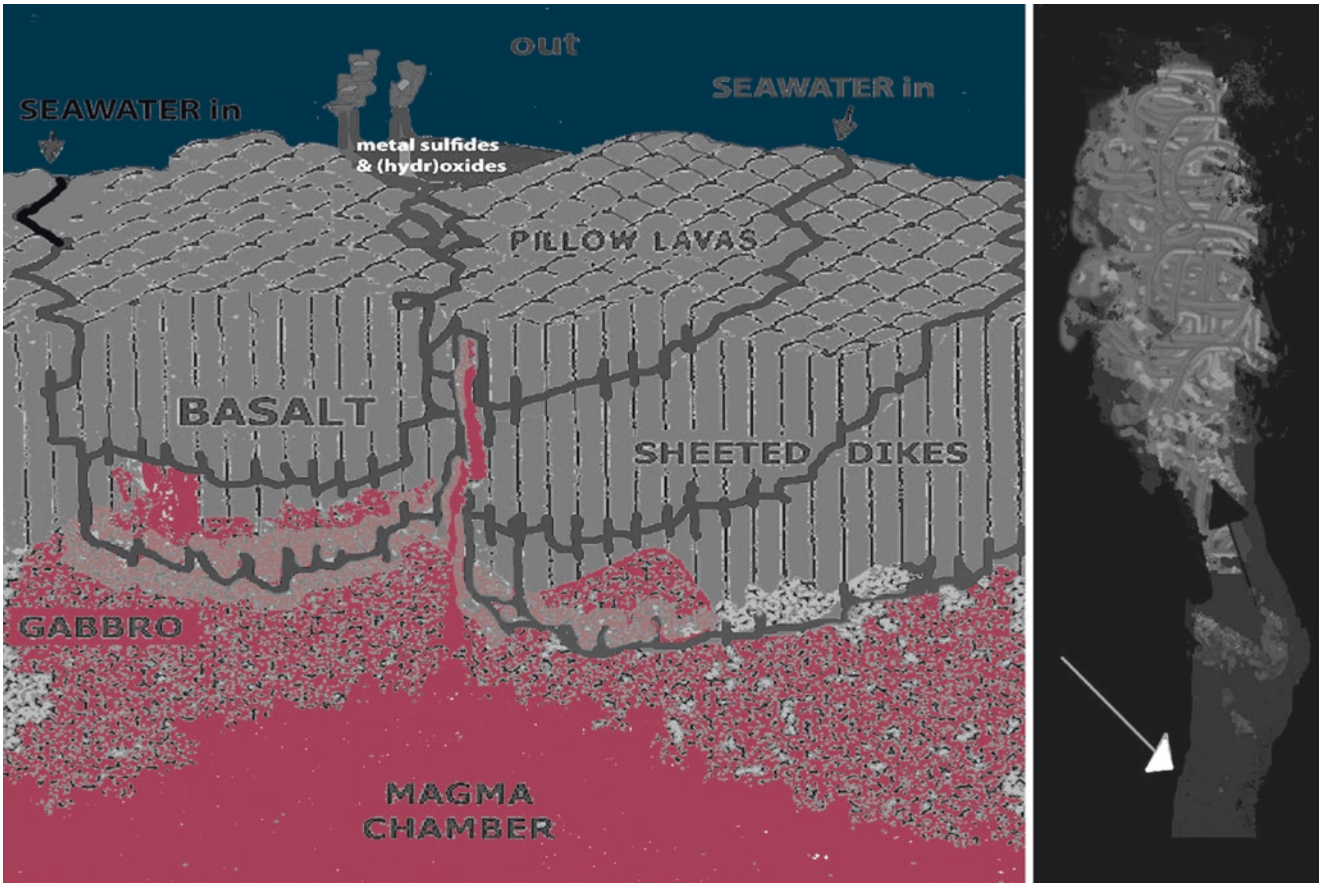

Fig. 14.13 Black smokers as a resource for metals. The smoker has reaction products of seawater with hot basalt exiting from chimneys (*black arrow*). Resources are largely sulfide chimneys (*white arrow*) (*Right*, schematic, using a photo taken from Woods Hole's scientific submarine ALVIN (here redrawn for simplicity); *left*: see Fig. 1.9 for source)

EEZ is involved in many cases. Most discoveries have been in the Pacific. The first notable efforts in mining hot vent deposits were made in 2006.

Many surficial land-based ores are in fact of marine origin and originated in ways similar to known ridge crest deposits, with massive sulfides resulting from stripping seawater sulfate of its oxygen by reactions of seawater with reduced iron within the hot basalt and with (hydr)oxides resulting from reaction of metal-bearing fluids with the oxygen in the cold deep seawater. Depending on how the metal-bearing fluids emerge and mix, mounds, miniature spires, and various baroque edifices can be built, some several meters high. The growth of modern spires may take but years, rather than decades or centuries.

A chimney made of anhydrite in the North Fiji basin (back-arc spreading) was labeled "La Dame Blanche," in reference to its ghostly appearance. The white structure stands in stark contrast to the more common "black smokers," vent fluids shooting from dark chimneys made of metal sulfides and oxides (Fig. 14.13). In a broad zone around the vents, iron- and manganese-(hydr)oxides are precipitated, as the reduced iron and manganese (Fe2+ and Mn2+) meets normal, oxygen-rich seawater and becomes insoluble.

As the various minerals precipitate within the cracks of the basalt, they tend to clog the system and shut it down. Thus, on the whole, the hydrothermal systems are short lived. However, where resources are voluminous in certain mining areas on land systems, there is evidence that the vent system was renewed in many pulses, each presumably ending in a shutdown.

14.4.5 Red Sea Ore Deposits

The "Red Sea" likely gets its name from certain plankton blooms, not from any deposits on the seafloor, which remained unknown till well into the last century, after the "Red Sea" acquired its colorful name. The rift represents a rather specialized case of ridge crest accumulation of heavy metal deposits. The deposits are of considerable economic interest, and the area has been visited by many research vessels, including drilling vessels. The various ships were sampling heavy metal deposits, commonly present as a type of mud in so-called deeps. The deeps were named after the vessels that discovered them.

In the central Red Sea – an active spreading center that opened only a few million years ago and whose axis does not form a ridge yet, therefore – there are several "deeps," that is, enclosed basins. The Atlantis II Deep is more than 2000-m deep and only 6 by 15 km in area. The bottom is filled with brine having a temperature of about 60 °C and a salinity of 25%, seven times that of seawater. The brine is rich in iron, zinc, and copper. The sediment below the brine is incredibly colorful, brick-red layers alternating with ocher, white, black, and greenish layers. A variety of minerals provides the coloring; economically of interest are the sulfides in the dark layers. Trace element contents are high. Zinc is abundant within the fine-grained mud, and copper is present, along with its chemical homologs silver and gold.

The metals are trapped within the brine, which forms water bodies below seawater within the relevant deeps. Brine, being very heavy, does not mix well. The salt presumably comes from older deposits, formed as the rift first opened. While the Atlantic II Deep may contain millions of tons of zinc and thousands of tons of copper and silver (and be quite rich compared with other marine deposits of interest), these metals are still more readily mined on land. For the countries adjacent to the Red Sea, however, mining within their own EEZ might be of interest eventually.

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15 Problems Ahead

15.1 The Central Problem

For thousands of years, we humans have found ways to control our natural environment – using fire against the cold, building shelters against bad weather, and extracting sustenance from the biosphere on land and in the sea. All the way through this history, these activities kept increasing, especially in the last two centuries. Now we have to learn to cope with a new geologic force of planetary proportions – our own activities disturbing the "regular" course of events.

Prime examples of the impact of the new forces are found in the overexploitation of biosphere resources, such as fish and forests. ("Overexploitation" is *unsustainable*; that is, the annual harvest is greater than the annual addition to the resource and thus cuts into the potential harvest of future generations.) To this must be added climate modification and waste disposal. (Both items are of geologic interest. They exceed nature's ability to accommodate disturbance.) Efforts are underway to meet the challenges arising. On the whole, success in dealing with negative effects of human activities in anticipatory fashion has been very modest.

15.2 A Role for Marine Geologists

As mentioned in the preface, geologists do have a responsibility to bring basic knowledge and professional experience to the ongoing discussions. Quite commonly, the task of a marine geologist is simple enough: to show (i.e., offer evidence) that human activities have geologic consequences. This includes the fact that short-term activities have long-term consequences for climate change and other environmental conditions.

There are a few things marine geologists can contribute to current discussions, even though geologists might not be more adept at finding answers to the central questions arising than others pondering the matter.

From the marine record, we can certainly learn how the climate system has responded to past disturbances and how fast. True, it is likely that the system will respond differently in the future, since the ongoing global human disturbance is fast and unique. There is the danger of offering geologic palliatives for serious threats (such as trotting out a long time scale, involving thousands of years and more, when a short one of decades and centuries is more appropriate for human concerns at issue). In any case, however, what we know for sure is that if something happened it can indeed happen: no ad hoc dismissal of the evidence is possible.

Relevant discussions about geologic history presumably are not about remote possibilities on a million-year scale (the geologic time scale of Earth history) but about applicability of observations on a human time scale of a century and less. There is no question that the deep-sea record has more information in terms of sea-level change than do most other records. But the best resolution in deep-sea records of our extant planetary natural climate system (ice age climate fluctuations) is about one millennium (see Chap. 11). Information for events taking place on shorter time scales than millennial ones must be looked for in special locations or in certain organisms. Examples of such records are found in varved (=finely layered) sediments or in ice cores or in growth rings of trees or in various long-lived marine organisms, including certain corals. Unfortunately, organisms can engage in biased reporting, discriminating against reporting "bad" conditions by ceasing to grow, seasonally or interannually. Also, reporting organisms can adapt to a changing environment, thus reducing the range of the physical information recorded. Ice core records obviously are limited by the location they report on.

Records on changes in the frequency of large storms and of large eruptions also may conceivably become available for examination or verification of relevant assertions. (The ones whose assertions are being examined might consider that the

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3_15

examiners, while skeptical, take the assertions seriously enough to check them against history).

15.3 The Seafloor as Waste Receptacle

15.3.1 Humans as a Sediment Source

One of the oldest signs of human impact, one may assume, is the increased delivery of sediments to estuaries and harbors within the Holocene. The increase is caused by intensification of erosion on land from agriculture and from deforestation. Basically, a natural growth of erosion has a similar effect: an acceleration of the filling in of estuaries and of affected lagoons. By far most of the increase in erosion observed for recent geologic time (on the million-year scale) is natural: the action of ice on land is ultimately responsible. But some of the rise in erosion (the one from agriculture and deforestation, beginning a few thousand years ago) is of human origin. Some of the new erosion provides "extra" sediment to the seafloor, beyond the natural supply. One might classify it as waste. Some of it definitely indeed *is* waste. Where waste increases nutrient input, it may result in "eutrophication" stimulating algal growth, including harmful algal blooms.

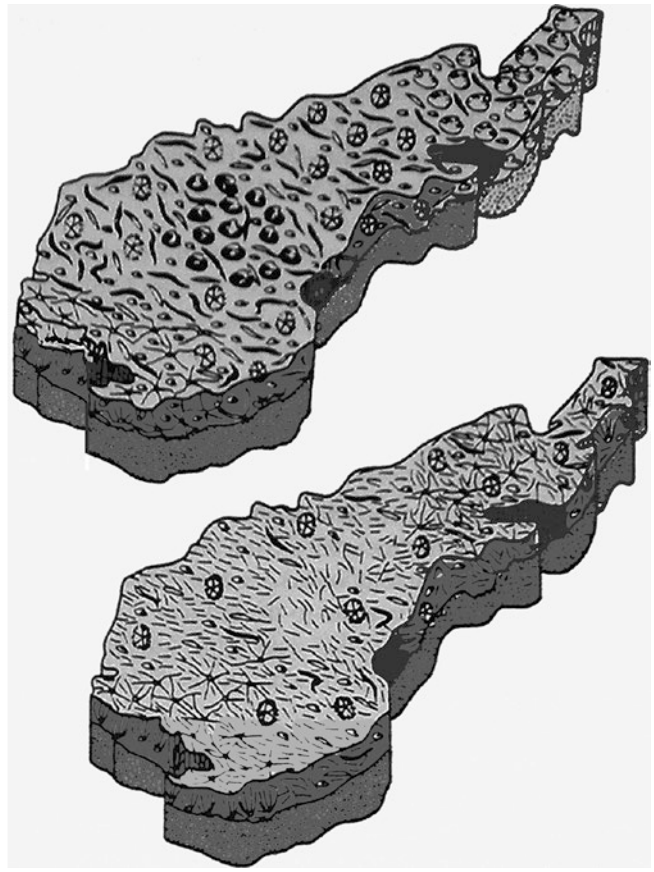

Fig. 15.1 Biota on the seafloor as indicators of environmental health in the North Sea (Dogger Bank). Mussels and sea urchins were replaced by brittle stars and worms in the time interval shown, suggesting a drop in oxygen levels and an increase in nutrient supply (Courtesy J. Lohse, modified for clarity. For the original drawing see J. Lohse et al., 1989, Die Geowissenschaften 7.6, 155)

15.3.2 Increasing Waste from Cities and from Agriculture

Coastal cities and agriculture have contributed nutrients to the sea for centuries, stimulating algal growth in places, some harmful. This process has recently greatly accelerated owing to population growth and the increased production of waste and fertilizer. In places where waste amounts are sufficiently large and water bodies are restricted, considerable damage can ensue from waste disposal and fertilizer runoff both in terms of impact on tourism and regarding biological resources. Some effects are readily seen on the seafloor, in the changes wrought in benthic biota (Fig. 15.1). Difficulties in separating natural from human causes do not constitute evidence, of course, that human causes are unimportant. That sort of conclusion (not uncommon) ignores the rules of logic.

In recent decades, the amount of waste delivered by cities to the adjacent sea has increased by leaps and bounds, and the sites of deposition have expanded correspondingly. Negative effects include dangerous levels of disease-causing microbes and blooms of harmful phytoplankton. Although many of the effects observed concern biology in the water, the seafloor is not exempt from the untoward consequences of the release of industrial pollution into the sea. The waste causing special concern does involve biology: it includes substances that tend to accumulate along the food chain, with organisms feeding at a high trophic level on the seafloor becoming poison-bearers. Thus, in many cases, the biological amplification affects the quality of marine food consumed by people.

In contrast to agricultural effects and deforestation, which increase sediment supply to coastal areas, the building of large dams to intercept rivers for water and for hydropower presumably has the consequence of trapping sediment before it gets to the sea, starving beaches, and thus promoting the erosion of the coast. For example, the construction of the Aswan Dam in 1964 in the uplands of Egypt apparently has helped increase shoreward erosion around the Nile delta region.

15.3.3 Sewage and Sludge

Industry consumes energy and raw materials and produces waste as a by-product of its target output. Much of the waste, such as sewage sludge, may be relatively innocuous provided it is free of obnoxious substances such as certain metals, for example, or biocides, or disease-causing microbes. Portions of the sea at ocean margins are commonly rather limited regarding the waste they can absorb and process. This is especially true for estuaries and marginal basins. (Examples are readily found within the Baltic Sea, which is both a restricted basin and surrounded by large cities.) *Eutrophication* (reflected in algal blooms from overfertilization) is a common phenomenon in such places. Resulting sediments tend to be organic-rich, black, oxygen-poor, laminated, and stinky. Eutrophication commonly results in a shortage of oxygen due to greatly increased oxygen demand at the organic-rich seafloor. Such a shortage, sporadically exacerbated by varying climatic conditions, can produce massive fish kills.

Even “deep” dumping in the open sea and away from shelf areas of cities can be problematic. Disposal of municipal sewage sludge in moderately deep water as far as 185 km off the coast of New Jersey from 1986 to 1992 at a rate of 8–9 million wet tons per year affected benthic organisms some 2.5 km below the dumping site. It had been hoped that dispersal and dilution of sewage particles in shallow waters would prevent waste from entering the food web at such depths. In many coastal cities, seaside recreation is big business. Thus, the pressure for “clean” disposal of sewage and sludge can become intense. “Beach closed” signs obviously are highly unwelcome, especially if it results in turning away economically important tourists and thus hurts hotels and merchants.

15.3.4 Garbage in the Deep Sea

Not even the remote deep seafloor is exempt from the effects of pollution and human deposition, as is demonstrated by the “garbage patch” phenomenon in northern central gyres, referring to floating debris. Some of the material is bound to reach the seafloor, as well. Much of the material involved (plastics) excludes an origin from natural sources. Unfortunately, apparently the plastic involved is mostly long-lived. Numbers mentioned for the typical life span of this type of trash include decades, centuries, and even millennia. Actual experience and observation covers decades only, of course, so one is likely to obtain estimates of varying quality. Items resembling food, but largely worthless in terms of nutrition, interfere with the proper workings of the food chain, touching many animals in the sea, including (one suspects) benthic ones on the seafloor.

15.3.5 Hydrocarbon Pollution

Hydrocarbon pollution comes in two versions: (1) sporadic and intense, and (2) chronic. The first type is referred to as a spill (Chap. 14). It gets much media coverage. The second is background and attracts not much attention, but is nevertheless an extremely important hydrocarbon pollution source, receiving a steady input from shipping and unregulated boating. Other sources exist in various kinds of hydrocarbon-based travel, in drains, in dumping, and in the operation of offshore rigs. It is sometimes claimed that background input is much larger than either natural seeps or catastrophic spills. Although supporting information is not very precise, there may be truth in such claims, especially when referring to conditions of decades ago. Apparently, much background input actually involves natural seeps, which obviously cluster in oil-rich regions. Thus, some tar balls on the beaches of southern California may not originate from industrial activities but may reflect natural processes. That many tar balls are in fact of industrial origin is suggested by the timing of occurrence and sometimes is proven by chemical analysis.

Major spills commonly involve either blowout from wells under pressure or else the demise of oil-carrying vessels that run aground or collide with others. Acts of war may also play a role, as in the enormous Kuwaiti spills resulting from hostile activities, in 1991. A recent large pollution event in southern Californian oil country occurred at Santa Barbara, where a pipe broke (a pipe of Plains All American Pipeline) and spilled thousands of gallons of petroleum, some of which apparently ended up as tar balls contaminating beaches in Los Angeles County.

In reported cases of catastrophic spills, oil dispersed by seawater demonstrably damaged the environment, killing seabirds and mammals and other marine organisms, according to both marine scientists and media observers. Negative impacts are especially large nearshore, in estuaries, and in tidal wetlands. It commonly takes several years in the case of the largest spills till the system comes back to what may be considered “normal” (a state not commonly well defined, however).

15.3.6 Radioactive Pollution

One legacy of the time of nuclear testing in the middle of the twentieth century is known to every marine geologist: the human-made radiocarbon spike around 1963 that is obvious in records with annual resolution. It is commonly noted in records of the last several decades (e.g., in laminated sediments, in coral, and in large mollusk shells).

The massive release of radioactive substances to the atmosphere by the *Chernobyl* power plant, in 1986, provided a unique opportunity to study the pathways in the marine environment of a number of dangerous radionuclides. Results from such studies (e.g., trapping and coring in the Black Sea) suggest that many, perhaps most, of the metals involved (e.g., the notorious cesium-137; beta decay, half-life 30 years) are rapidly removed from surface waters and transported to the seafloor within biogenic debris. Similar fortuitously benign removal has been observed off the US West Coast when studying the effluent from the Hanford reactor near the end of the Columbia River.

Problems of an entirely different order arise in connection with high-level radioactive waste from power plants (such as Cs-137 and Sr-90). Given that large amounts of such waste exist on land now in a few places, and given that they can pose a considerable hazard, the disposal of such materials through burial in the seafloor is an option that has received some discussion and study. One type of proposal envisions canisters of various shapes placed into deep-sea sediments (including spear-shaped ones to aid penetration). Geologic questions that must be considered are as follows: (1) How stable is the target area? Is there erosion or uplift or faulting going on in the area at issue? (2) What type of circulation will develop around the hot containers, and what will be the effect? (3) What types of reactions of the sequestered material might be expected and what are the implications for a lasting confinement? "Lasting confinement" may have to involve thousands of years (in the case of certain transuranium elements with long half-lives). Given that many civilizations and nations have a history measured in hundreds, not thousands of years, a task requiring monitoring of hazardous substances for time spans exceeding many centuries seems extraordinarily challenging.

15.3.7 Risk Assessment

Clearly, there are risks involved in using the seafloor as a waste receptacle, and it is desirable to assess such risks *before* starting the dumping. In any case, one would hope that existing (albeit unfortunate) instances of hazardous dumping are used to study the processes that determine the fate of the materials in question. These include *dispersion* of the substances and alteration through *chemical reactions.*

Dispersion is controlled by bioturbation, resuspension, and transport by currents. Chemical reactions on and within the sediment can change the toxicity of materials and the ease with which they migrate through the sediment or are taken up by organisms. A well-known example with dire consequences is the mobilization of mercury by *methylation* within anaerobic sediments. ("Methylation" here describes the chemical reaction binding mercury to methane.) Methylmercury is a potent neurotoxin. When present in the environment, it gets concentrated in marine food (fish, shellfish) up the food chain. Consuming the food causes *Minamata disease* if ingested in sufficient quantity. The disease (disorientation, loss of balance and of control of the body in general) is named after the bay in Japan where it struck in the middle of the twentieth century and up to around 1970, owing to release of mercury by local industry.

15.4 Climate Change and the Seafloor

15.4.1 Revelle's "Great Experiment"

Sometime in the middle of the last century, it was widely realized that the continued addition of carbon dioxide to the air from the burning of coal and petroleum would unavoidably affect climate in harmful ways. Global warming was anticipated even at the end of the nineteenth century already, but it was not necessarily seen as harmful at the time. Also, in the first half of the last century, it was still widely believed that the sea would take up the bulk of the gas, thus greatly slowing the greenhouse gas buildup in the atmosphere. The Californian marine geologist and oceanographer Roger R. D. Revelle (1909–1991; Scripps director from 1951 to 1964) was adamant in insisting that this is not necessarily so on the short time scale considered. The ocean is not an ideal agent mitigating effects of carbon dioxide release. For once it responds but slowly to any change in atmospheric concentration, owing to long mixing times. In addition, concerning carbon dioxide, the ocean's potential for chemical reaction (and hence uptake of carbon dioxide) is quite limited by the abundance and availability of carbonate ions. Also, the sea warms at the top, which slows vertical mixing and thus decreases the rate of uptake from the air of any gas during warming, not just the one (carbon dioxide) that promotes warming of the planet by absorbing heat radiation.

Revelle's 1957 pioneering article on the subject (written together with the Austrian-American physical chemist Hans Suess) emphasized the limiting chemistry of the relevant reactive species in the sea. His article contains the often-quoted phrase about the "great geophysical experiment" that is carried out by mankind by modifying the composition of the atmosphere in a fashion that could only be done once. Realizing a need for documentation, Revelle caused the CalTech geochemist Charles D. Keeling (1928–2005) to come to Scripps, where Keeling began his famous series, documenting the inexorable rise of carbon dioxide in the atmosphere from one year to the next.

The addition of carbon dioxide increased atmospheric content roughly 0.5 percent per year after 1980 (when the concentration was 340 ppm). The concentration of carbon dioxide in the air is small, hence easily altered ("ppm" is "parts per million"). In 2015 a level of 400 ppm of CO_2 was reached. In the coming decades, it is expected by many observers, non-carbon energy use might well keep increasing, but the rise of CO_2 will continue unabated. The presence of "positive feedback" suggests that there may be a point when the melting of ice cannot be readily stopped by ceasing to release greenhouse gases. When the horse has left the barn, thus the popular wisdom, closing the barn door is not the answer.

Incidentally, the observed rise in carbon dioxide, a greenhouse gas, *must* result in global warming; discussion of this fact is unnecessary since we are dealing with basic experimental physics. One *can* argue about delay and about feedback processes that affect the precise *amount and rate* of warming, but not about the basic principle. The warming presumably is enhanced by an increase in humidity in the lower atmosphere and by various albedo effects, largely "positive feedback" on the human time scale, that is, reinforcing the given change on the scale of decades and centuries (not actually a positive development, but one re-enforcing the initial disturbance, potentially rendering it dangerous). Milankovitch theory of the ice ages likewise invokes positive feedback, as we have seen (Chap. 11). In fact, his climate theory of the ice ages presumably works *because* it relies very heavily on albedo feedback in high latitudes.

Revelle's "experiment" implies global warming, since carbon dioxide is a greenhouse gas; that is, a gas hindering heat radiation back to space, a radiation that normally balances the energy received from the sun (Fig. 9.1). A warming has to occur if energy is suddenly prevented from leaving. A warming indeed has been observed within the last century, culminating in an unusually strong rise toward the end of it. The resulting shape of the temperature history resulted in the term "hockey-stick" graph. The significance of the "hockey stick" has been the subject of much heated discussion, not necessarily discussion aimed at improving understanding. Discussion has subsided since. Apparently the data base used by the professional climatologists who discovered the hockey-stick shape of recent temperature history is not a substantial problem. As expected, data centuries old and subject to interpretation are somewhat less trustworthy than more recent ones.

15.4.2 Time Scale Problems

Time scale problems are pervasive. For example, after the year 2000, apparently there was a puzzling occurrence of little or no rise in temperature for more than a decade despite a continuing rise of carbon dioxide. The concept of missing warming (if bolstered by evidence) suggests that background temperature variations (some conceivably not wholly natural but partly caused by global climate change) can be strong enough still to mask some of the climate change from excess carbon dioxide. What emerges most clearly from the various discussions is that the ongoing experiment noted by Revelle apparently has different messages for people with different backgrounds. The identification of trends is a difficult matter, especially when using data from short time scales. Regarding climate evidence from marine geology, most of its information applies to scales much longer than decades or even centuries. Geology is about Earth behavior and includes long time scales measured in millions of years. While the lessons learned from long time scales indeed are useful for many purposes, they are obviously difficult to apply to problems arising on short time scales.

15.4.3 Some Basic Considerations on Climate Change

The phenomenon of global warming, as mentioned, reflects an emerging imbalance between the energy coming to Earth from our central star and the energy leaving the planet. The two must balance if the temperature on Earth is to stay the same (Fig. 9.1). The planet must warm if the incoming solar heat increases too rapidly to be gotten rid of simultaneously, for example, by suddenly providing for additional obstacles to reradiation (such as excess greenhouse gases in the atmosphere). Certain sediments on the seafloor (laminated ones for recent changes) can be consulted for clues to the history of the planetary energy balance, as reflected in temperature variation. (Technically, the energy balance is usually referred to as "radiation balance.")

As far as is known, the incoming amount of heat from the sun is comparatively stable on a short time scale (a century or less), although satellite data do allow for small variations and there is an 11-year sunspot cycle that is relevant to discussions of solar variation. Solar input apparently varies on longer time scales as well, as inferred from anomalous radiocarbon fluctuations on a 1000-year scale and as implicit in Milankovitch theory, which addresses change on a multimillennial scale. Some have regarded a fuzzy 1.5 millennia cycle based on changing sedimentation as evidence for a longish solar cycle. However, the currently observed (fast) warming is ascribed by working climatologists mainly to the buildup of obstacles for reradiation to space, that is, to the rise in excess greenhouse gases, rather than a brightening of the comparatively steady sun. To repeat for clarity, according to many or most experts, the sun is *not* the main reason for currently observed warming. We should be able to measure a brightening in the sun and not have to rely on fuzzy sedimentary evidence for inferring change relevant to now occurring events.

Of course, the climate system of the planet, that is, the various interactions of the many factors involved in energy transformations on the surface of Earth, results in enormous complications (Fig. 15.2). One important complication, as mentioned, concerns transfer of information from one time scale to another. Also, plenty of difficulties arise when contemplating cooling and warming effects from changes in cloud cover, in any time scale. Thus, the underlying simplicity of the radiation balance tends to be obscured, and interpretations of changes in this balance are correspondingly difficult. Problems arising from uncertainties are the fundamental reason why some attack on the scientific consensus is unavoidable (and why the IPCC is reporting implications of scientific findings in terms of somewhat ill-defined probability rather than insisting on strict quantification). What we are quite sure of is that human-released excess carbon dioxide warms the planet – it would seem that everything else is open

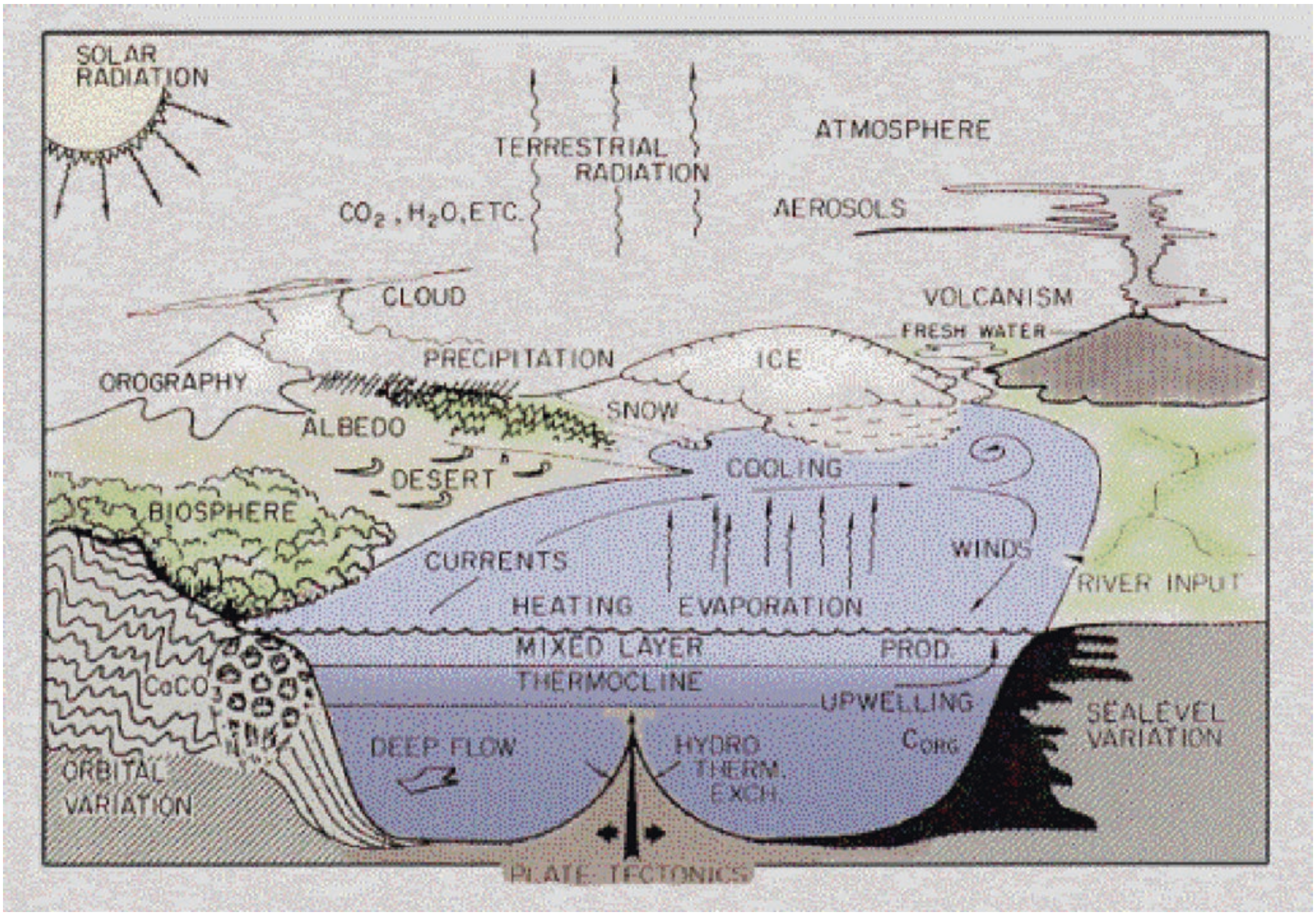

Fig. 15.2 Diagram depicting the major elements of the climate system in schematic fashion. Major sources of instability and uncertainty include clouds, seasonal variation, ice mass, volcanism, deep-water production, upwelling, volatile gases in the seafloor, and human activities (Graph from W.H.B. and L.D. Labeyrie (eds.) 1987. Abrupt Climatic Change: Evidence and Implications. Reidel, Dordrecht. Color here added)

to discussion, especially information involving the precise path of future developments.

On occasion questions have been raised about the quality of mathematical modelling of climate change. We offer several comments regarding this much argued-over topic: (1) Granted that the climate system generated by feedback mechanisms active on several time scales is extremely complex, a mathematical treatment seems much preferable for producing estimates of future conditions to a nonmathematical seat-of-the-pants approach. (2) It seems quite obvious that a mathematical model cannot be more trustworthy than the rules of climate behavior that it assumes, and (3) those assumptions cannot be better than current understanding of the behavior of the various elements of the climate system in the time scales at issue. Large uncertainties persist both for long and for short scales. For example, there is the treatment of clouds, which can both cool and heat the area below them (by shading the ground or by sheltering it from loss of heat). Obviously it is of great interest which process – heating or cooling from clouds – is the stronger in any given situation of a short time scale and also overall as a factor affecting latitudinal zonation, on a geological scale. The question is difficult to answer. (4) On a short time scale, when geologic hazards threaten, we must deal with risk and uncertainty, rather than assuming that a lack of certainty implies safety. As the late climatologist Steve Schneider said to one of us: "Uncertainty cuts both ways: the outcome of the (Revelle) experiment could be worse than anticipated." Schneider's comments are a reminder that this particular experiment (of releasing carbon dioxide in large amounts) is not suited for gambling. In other words, throwing doubt on expert opinion is dangerous in this case.

15.4.4 Abbreviating the List of Anticipated Calamities

The list of possible impacts from a general large warming is quite long and frightening. The rise of sea level is comparably easy to conceptualize and measure. Among other things, the conversion of ice on land to water in the sea raises sea level, with many dire consequences for people. To be sure, there are many other potentially very undesirable consequences of global warming. Thus, the planetary heating presumably interferes with ocean circulation and alters the wind system, causing a host of changes in productivity and other environmental processes. Potentially there also are grim corollaries for storm activity and rainfall distributions. Concerning the chemistry of the sea, there is the (observed) issue of an expansion of the anoxic zone, presumably mainly reflecting nutrient pollution, but exacerbated by decreased solubility of oxygen in warming seawater. Also, there is the *acidification* of seawater resulting from the uptake of carbon dioxide by the surface waters of the ocean. Marine geologists will increasingly be faced with such changes as seen within the composition of sediments on the seafloor and within the calcareous skeletal structures built by corals and by many mollusks.

Regarding biodiversity, any rapid climate change results in vast numbers of organisms poorly adapted to the new environment and thus subject to a culling that may result in mass extinction. Much of the recovery of biodiversity from a large disturbance seems to involve a geologic time scale (thousands and even millions of years, rather than decades or centuries). Concerning human well-being, anticipated changes of biodiversity on a human time scale have implications that are commonly quite unwelcome. The list is long and calamitous. Unfortunately, feedbacks in climate change seem to be largely positive on a human time scale over a wide range of change, while any negative feedback (system response favoring a return toward initial conditions) may reside in time scales too long for comfort. It is possible that both positive and negative feedbacks are always present but are operating at entirely different rates. If so, confusion of the pertinent issues would be expected.

15.4.5 Why Not Simply Stop Digging?

A well-known proverb advises to stop digging whenever you find yourself in a hole of your making. So, what prevents us from ceasing to add greenhouse gas to the atmosphere? The answer is rather straightforward: an overwhelming and urgent demand for energy and a traditional view of how economic issues are best approached. In a nutshell, the level of public worry about the potentially terrible effects of excess greenhouse gases in the climate system is as yet difficult to translate into political action. One may deplore the underlying weighting of arguments. However, perception can create a reality of sorts. It may be based on nonscientific reasoning, but it still must be dealt with. At this point, it is not clear where usable answers for meeting the conundrum will emerge. In the meantime, sea level is rising distinctly and observably. It is a relatively simple and measurable process.

15.4.6 Rates of Sea-Level Rise on a Short to Intermediate Time Scale

Sea level is bound to rise as a result of a general warming, no matter what the cause of the warming. A rise of between 150 and 200 mm characterized the twentieth century. It was detected despite all local complications, which caused much uncertainty before the arrival of satellites. The rise may have accelerated in recent years. One expert reconstruction, for example, suggested a global rise of 40 mm in the first decade of the twenty-first century. If this satellite-based value is correct, one can then expect a rise of nearly one half of a meter by the end of the century using straight extrapolation. Given the reality of continuing increase of input of carbon dioxide and other greenhouse gases, it seems more likely that sea-level rise will accelerate some more. The result will be, one assumes, a steering of the system toward a rate of ice melting similar to the one that prevailed at the end of the last ice age 10,000 years ago. At that time sea level rose at an average rate greater than 1 m per century (1000 mm per 100 years) for several thousand years.

Sea-level rise is likely to impact millions, since the coast is a popular site for human settlement, judging from regional conditions (Fig. 15.3). Already one fourth of the average rise during deglaciation (i.e., about 30 mm per decade) can result in enormous damage given enough time, including making traditional housing and agriculture obsolete near the coast.

At the present time, most of the sea-level rise apparently is from thermal expansion. However, the potential for the contribution from polar ice is very large indeed: melting Greenland ice can yield around 7 m of sea-level rise and melting the ice on Antarctica ten times that. The mass (and hence the potential contribution) of the rather vulnerable West Antarctic ice is roughly equal to that of Greenland ice. East Antarctica has ten times that amount. The total available

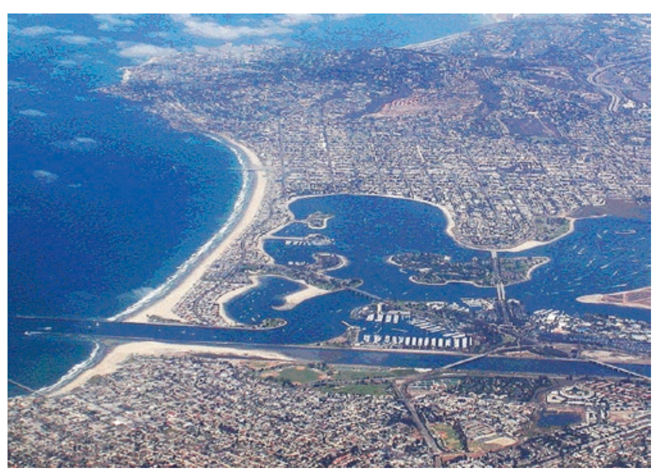

Fig. 15.3 Coastal settlement patterns, San Diego between La Jolla (N) and Ocean Beach(S) (Photo W.H.B)

ice mass that can be melted is roughly 80 m – approaching the amount of ice involved in the terminal melting event of the last ice age (ca. 125 m) and presumably repeating some of the ice and meltwater behavior at that time. The last melting event (and analogous earlier ones) may have lessons concerning the way polar ice melts. Ice responds to warming, but this is only one of the factors, there being a host of positive feedbacks on melting, including sea-level rise itself.

15.4.7 Rates of Sea-Level Rise: The SPECMAP Evidence

During the last major melting ending the last ice age, an average rise of more than 1 m per century was maintained for many tens of centuries (roughly 120 m in 10,000 years). This particular information has been available for several decades (Fig. 6.5). We are confident since the 1960s that sea level rose by more than 100 m during somewhat less than 10,000 years in the last deglaciation event. Since at least in the 1970s, it is thought that deglaciation was strongly pulsed and that the average rise rate was exceeded substantially on occasion. A rise of a meter per century (1 cm per year) or slightly more simply was nothing unusual during the time of major melting. It was close to average for thousands of years. Analysis of data in the literature produced in the 1980s readily confirms it. A prominent curve on sea-level positions suitable for this type of analysis is the "SPECMAP" standard of J. Imbrie et al., published in 1984.

The "SPECMAP" oxygen isotope series (purportedly depicting the history of the last third of the Pleistocene) can be converted to sea-level change abundances by assuming a total change in sea level of 120 m during the last deglaciation and matching this change to the corresponding one observed in the oxygen isotopes. One fundamental assumption in this procedure is somewhat questionable: It is that temperature effects do not matter in the conversion, because they tend to run parallel to ice mass. The SPECMAP data are listed back to 782,000 years ago in 2000-year steps in the data source. They are greatly smoothed. The series, it turns out, is usable to about 650,000 years B.P. Maximum rise rates in this exercise emerge at slightly higher than 1.5 m per century; that is, the "maximum" is fairly close to the average rise of sea level, for millennia.

It should be pointed out that the period before 450,000 years contains sea-level events that apparently have causes different from those of later events, making the entire set analyzed inhomogeneous. This may not be relevant for the comparison with the modern sea-level story, however, there being no analogs to modern events in any case. Other problems exist also. For example, there is a surprising lack of information on the last 8000 years or so in the SPECMAP data, perhaps an indication of loss of surface sediment during (piston-) coring. Any Holocene gap on recording historical changes during the last deglaciation would have an effect on calculated sea-level distributions: Closing a recording gap of several thousand years in the Holocene with available data (the series purportedly starts at zero age going backward) would introduce a spurious sea-level rise into the Holocene and decrease the overall rise during the last deglaciation correspondingly.

The data suggest, that a substantial buildup of ice is what is most commonly seen in this record on a millennial scale, but that the ice can melt sporadically unexpectedly fast (also see Fig. 6.5). Analysis of other isotopic ice age data, less smoothed, indicates that a rise close to 3 m per century, while definitely not common, can be seen on occasion on a millennium scale. Actual century-scale changes presumably occasionally substantially exceeded this rate, which only represents an average valid for a millennium that was especially prolific regarding the contribution of meltwater.

15.4.8 Neogene-Type Ice Masses

Of course, great caution is advisable when studying the ice ages for lessons applicable to the ongoing warming. We study a system that has ice at both poles and changes rather slowly (the ice age system) to better understand one that is on the way to become entirely lopsided, with very little ice on the northern hemisphere, a system much like the one in much of Neogene time (Fig. 15.4). According to the late climatologist H. Flohn (Bonn), the asymmetry is bound to increase in the future and possibly quite fast.

What then can we learn for the future from studying the ice-age past of the planet?

We can certainly learn much about the behavior of the climate system in general. But application of particular insights to prediction of future developments is heavily constrained. What emerges as a strong suggestion is that ice is likely to display unstable behavior (in melting unexpectedly fast and sporadically) and any regular (linear or nonlinear) extrapolation of past sea-level rise to the future is therefore

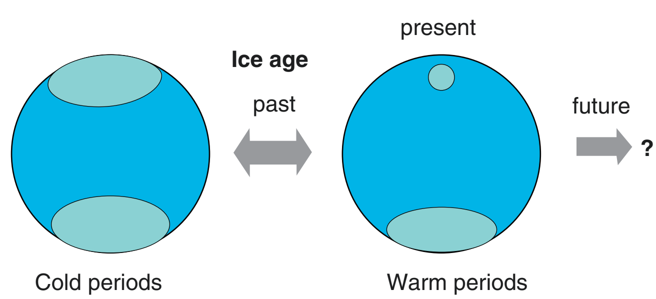

Fig. 15.4 Schematic concerning the relevancy of ice-age climate to global warming, according to concepts of the late H. Flohn (erstwhile meteorologist at the University of Bonn)

tagged with major uncertainty. Uncertainty need not entail lack of action (it does not when contemplating military threats). However, uncertainty encourages a reliance on personal opinion and unavoidably affects political perception and international negotiations. Asking a medical doctor for proof for his assertion that certain behavior may result in major problems is rather dangerous, of course. Proof may be too late to be useful.

What the marine record suggests is that sea level will keep rising thanks to positive feedback once the process has entered the cm-per-year magnitude (the typical rate of natural deglaciation). To what degree this insight applies to current events is uncertain. It is unlikely to be entirely irrelevant, however. We may quite literally approach sea-level rise rates last seen during the last deglaciation. The soaking up of meltwater by parched land (as suggested by some NASA scientists) perhaps brings some temporary relief, but is likely to end up as another irrelevant discussion topic as far as the millennial time scale.

15.5 The Carbon Cycle Modified

15.5.1 Relevance of the Carbon System

The science of the carbon cycle is central to discussions in the present text for several reasons. One is that the largest of the highly active carbon reservoirs is in the sea (total dissolved inorganic carbon is about 60 times atmospheric content). Also, some of the very large (and much less active) geological reservoirs are made up of marine carbonates, of marine petroleum-type substances, and of carbon-bearing gases of marine origin (as discussed in the previous chapters). The scientific discipline relevant to carbon cycling belongs to the discipline of biogeochemistry. Its chief ingredients are in the name: biology, geology, chemistry.

On the whole, the largest reservoirs are part of the multimillion-year cycle, while smaller but highly active ones help define the short-term cycles (including decadal ones). The largest marine reservoir by far is part of the crust of the planet: it is in the marine carbonates on land and in its crust and in the oozes and chalks of the seafloor. The reservoir has more than a hundred thousand times the atmospheric carbon mass or "ACM." Organic carbon mass in marine sediments (e.g., petroleum and other hydrocarbon) occupies second place, as far as abundance of carbon (less than half of the amount of carbon in the carbonates). The next largest reservoirs, and quite relevant on scales of centuries to millions of years, are the amount of dissolved inorganic carbon in the sea (ca. 60 times ACM) and the dissolved marine organic carbon (less than one tenth of that). Soil organic carbon, forests, and peat are next (probably in that order), each being in a class with the atmospheric reservoir and highly relevant to ongoing changes. The phytoplankton in the sea is a distinctly smaller reservoir, albeit a highly active one (i.e., it generates a large carbon flux compared with its size).

All reservoirs vary in tandem with changing environmental conditions, affecting geologic processes accordingly, for example, through deposition of carbonate and organic matter. Carbonate deposition affects, for instance, the building of continental margins and presumably can facilitate the motion of plates when entering the upper mantle. Organic matter deposition affects all aspects of the productivity of the sea, and hence the sea's biological and biogenic resources, and provides clues to marine evolution. Many carbon reservoir fluctuations are well known. Fluctuations of the atmospheric carbon dioxide gas content are particularly well studied for the ice ages thanks to the work begun in the laboratories of French and Swiss geophysicists (Jean-Claude Lorius, Grenoble, and the late Hans Oeschger, Bern). A substantial portion of Milankovitch-scale climate variation presumably is attributable to the well-tracked (but poorly explained) carbon dioxide variations that closely follow orbital forcing. Any attempt at such explanation brings us welcome additional familiarity with the controls on atmospheric carbon dioxide, thus teaching us important aspects about Revelle's great experiment.

15.5.2 "Sensitivity" to a Doubling of Carbon Dioxide and "Safe" Levels

It has been said, correctly, that in essence we humans are returning to the environment, on the time scale of a century, the carbon that was stored in coal and oil over hundreds of millions of years (a factor of several million in terms of applicable rates). This is the essence of the "great experiment" referred to by R. Revelle. We are now cognizant of the difficulties arising, an alarm having been sounded by a number of perceptive pioneers (including Revelle) and by the IPCC founder and his organization (the Intergovernmental Panel on Climate Change). The founder was the Swedish meteorologist and carbon-biogeochemist Bert Bolin (1925–2007) whose concern was to get reliable information on climate change to the public and to decision makers.

A doubling of background atmospheric carbon content is expected within the present century by many expert climatologists. Will marine geologists recognize the event in the record when it happens? At this time, we cannot possibly know.

Many or most climatologists put the "sensitivity" of the climate system (warming from a doubling of carbon dioxide) in the range between 2 and 3 °C (4–6 °F). However, large uncertainties exist, resulting in about a factor of two on the central estimate (2.5 °C for a doubling of CO_2). Because of the gaps in knowledge exemplified by the uncertainty regarding "sensitivity," it is futile to assign the word "safe" to the atmospheric content of carbon dioxide at any level above 300 ppm (the one that worked for thousands of years). Time scales matter, and a "sensitivity" established for one scale is not necessarily valid for another. Our life spans invite us to emphasize short time scales of a century or less, but the long scales are very real for ecology and evolution, that is, for life on the planet.

15.5.3 Possible Lessons from the Late Neogene Changes in Marine Productivity

It appears that anything that affects wind velocity will also affect the productivity of the sea. Wind lives off temperature gradients. Thus, it would seem, a general warming more intense in high latitudes than in the tropics, by reducing the planetary temperature gradient, will eventually slow surface circulation and the productivity-enhancing zonal processes around the gyre margins. (The well-behaved gyres presumably are replaced with much less steady and sporadically active currents.) The time scale of the hypothesized processes involved, however, is not well constrained from marine geological insights. Observations of the California Current suggest that its productivity has been relatively low for decades already. The mechanism or mechanisms responsible for productivity and its collapse are poorly known and may be creating but a temporary anomaly, although a link to global warming is quite possible, perhaps even likely.

From marine geological studies of drilled Neogene sediments, we know that coastal upwelling started in earnest at the end of the middle Miocene, about 10 million years ago. Presumably, general cooling at the time fostered strength and zonal character of the wind field and thus enhanced mixing and currents linked to marine production. As soon as snow formed, it was able to affect seasonal albedo contrast, which probably affected upwelling in ways that boosted productivity (by increasing seasonal change and by alteration of mixing and stratification sequences). If so, a reverse development (from a general warming) of a reduction of snow and seasonal contrast may result in a reduction of marine productivity.

15.5.4 "Engineering Fixes"

The likely outcome of doing nothing or very little now to address the growing concerns regarding the problem of planetary warming will be a call for "engineering fixes," if and when problems become hard to deal with. Commonly such "fixes" are planned to reduce incoming solar energy by shading Earth's surface in one way or another or by facilitating the rejection of incoming energy by increasing planetary albedo. One alternative to shading and albedo manipulation is the reduction of greenhouse gas in the atmosphere. In essence, this would be an attempt to remedy the missed opportunity to restrict input of the greenhouse gas carbon dioxide to a supportable level in the first place. One problem would be that past damage may be irreversible.

The process of mentally preparing for emergency action has begun, at least among many climate scientists. (Some "business-as-usual" people hardly foresee a problem). What can marine geologists contribute? Like other scientists, they can examine proposed "fixes" and point out possible unintended consequences for the manipulation of Earth systems including climate. Unintended consequences (by definition) are likely to result from poorly understood processes. Attention to factual history will be useful. Marine records documenting changes on short time scales do exist, as illustrated in the next section in this chapter. Their detailed study is only a few decades old, however, and they have to deal with the full complexity of the climate system. Insights are correspondingly limited and may not apply to the evaluation of any planned "engineering fixes." In a rapidly changing system such as ours, the challenge to identify those elements of the system studied that can be modified in satisfactory ways to provide "fixes" is likely to be enormous. *Caveat emptor.*

Again, a marine recording device appropriate for keeping track of millennial change is an unlikely candidate for tackling urgent problems arising on time scales with much higher resolution than ice ages. Ice as a recorder avoids many of the problems common in marine recording and may be more promising than marine sediments or records from marine biology. But ice may not turn out to be an ideal substitute for the marine record. On a warming planet, useful information from sediment stacks of ice comes from ever higher latitudes (or ever higher elevations). What will likely be missing is a reliable way of finding the implications of observed high-latitude recordings for low latitude systems (impacting millions) at a time when the systems are undergoing rapid and poorly understood changes. Again, to apply partial knowledge to extant problems (as may be proposed) may invite much-feared types of unforeseen consequences.

15.6 Short-Term Climate Change in the Marine Record

15.6.1 Varved Sediments

Finely layered marine sediments ("varves") deposited on the seafloor in anoxic environments and thus not stirred by burrowing organisms can deliver information on changes relevant for short time scales. Such sediments can be used to reconstruct historical trends on the scale of decades and centuries and to check for short-term periodicity in depositional sequences (e.g., such as the 11-year solar cycle). Also available for such checks are coral records (Fig. 8.4). As a general rule, one finds that short time scales tend to focus on local information; that is, the long-term records tend to have globally valid information, while short-scale records may mainly address local change.

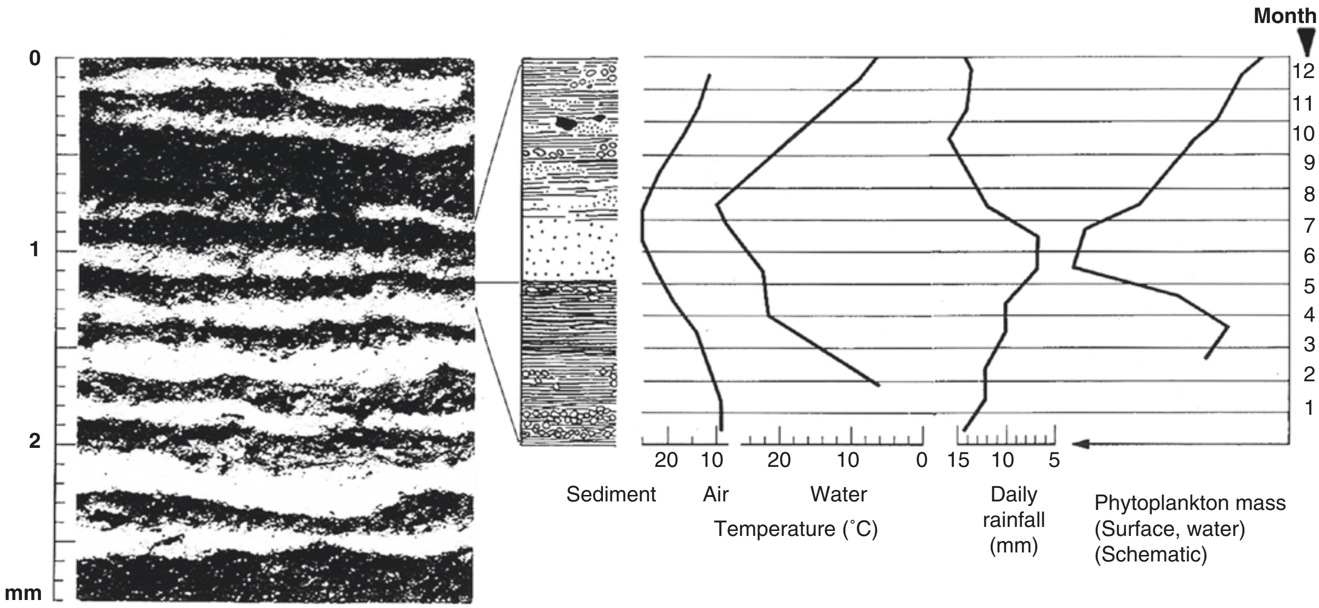

Fig. 15.5 Annual layers ("varves") in sediments in the Adriatic Sea. In the (simplified) microphotograph (*left*) one light-dark pair corresponds to 1 year. The light layers represent carbonate precipitation by phytoplankton blooming early in summer (*right panel*). In fall and winter, rains bring terrigenous matter, which together with organic detritus provides for dark colors (Photo E.S.)

Some of the earliest studies of marine "varved" deposits were done on the shelf in the Mediterranean Sea, in the bay next to Mljet Island, in the middle of the last century (Fig. 15.5). The nature of the varves as layers of alternating terrigenous and planktonic supply emerged at that time from a detailed study both of the laminae and the seasonally changing environment.

Detailed studies on marine varves in deep waters off California were carried out in the 1960s by the oceanographer and marine geochemist S. Calvert (then a graduate student and subsequently a postgraduate researcher at S.I.O.) with a focus on diatom deposition within the plankton layer (rather than carbonate as on the Mediterranean shelf in sediments studied earlier by E.S. and associates). Calvert found that diatom production, not volcanism, is responsible for high silica values within the varved sediments. The varves themselves represent a record of seasonal change in sedimentation.

15.6.2 Santa Barbara Basin: Recent Decades and Centuries

The basin off Santa Barbara is invaded by oxygen-poor water from the offshore oxygen minimum zone and made anaerobic by additional removal of oxygen from the overlying upwelling-type productivity in coastal waters. Its varved muds serve as a repository of detailed information about the history of the California Current in the region (Sect. 9.1.7). Within the time span of the last century, two items stand out (among several). One is a crash of diatom productivity in the early 1970s, discovered by C.B. Lange. It was followed several years later by partial recovery (Fig. 9.8). Another is the evidence for recent warming, from a change of foraminiferal content, beginning several decades ago and described by the marine biologist D.B. Field in his Ph.D. thesis (S.I.O., UCSD 2004).

The thickness of the varves cannot be determined precisely, because the necessary correction for water content in the uppermost portion of a core is arbitrary. Unless a precise value is desired, though, this is not a serious problem. That the recording sediment is changing on a short time scale is evident both in the thickness of varves and in their contents (such as the fish scale abundance studied by the late J.D. Isaacs and his colleague and erstwhile assistant, A. Soutar). Periodicity of deposition is evident in the various types of varve records. Periodicity is easily determined by applying Fourier analysis, but the cause or causes of cycles emerging remain poorly understood. One vexing problem is that the mix of cycles changes over the centuries, sometimes rather abruptly. The reason is not known. Another problem is that missing varves (e.g., from sheetlike slumping at the place of deposition, following earthquakes) apparently can provide for an incomplete record, difficult to recognize as such when counting varves, but emerging on detailed dating.

15.6.3 A Question Re Solar Cycles

A changing sun has been proposed by some as a source of variability of modern climate. The Swiss physicist Jürg Beer and some of his colleagues, for example, have ascertained that a large portion of the climate change of the last 150 years is owed to changes in solar brightness. An important question arises: Can a variation of the solar "constant" (the energy input from the sun) be recognized in the marine geology record? A related question addresses the fuzziness of the evidence: how well should the solar cycles be defined, and how well should they be matched in the record before a finding of such cycles is considered significant for understanding climate change?

As an aside, the late climatologist Stephen Schneider (NCAR in Boulder and later Stanford University) pointed out to one of us that documenting past solar effects on climate does not bear importantly on the sun's potential influence on present or future climate change. The questions to a geologist faced with the record are most commonly about what happened, of course, not about what is happening or what is likely to happen.

We can be sure that certain solar cycles are real: the sunspot observations of the last 300 years (collected, e.g., by the Royal Astronomical Observatory in Belgium) as well as the historical aurora record collected by the historian D.J. Schove (St. David's College, UK), and published in the Geologische Rundschau in 1964, with more than 1000 years of observations, yield a strong cycle near 11.1 years when analyzed by Fourier's method. In the aurora record, one also finds a cycle near 205 years, presumably the so-called "Suess" Cycle, named after the late Austro-American physical chemist Hans Suess (S.I.O.). He reported the cycle from radiocarbon records in certain tree-ring sequences. The ca.-200-year solar cycle also is known as "de Vries" cycle, named after a meritorious pioneer Dutch solar astronomer. In contrast, the much cited ca.-80-year "Gleissberg" cycle is not readily seen when using mathematical analysis on reliable solar data. The suggestion is that while the "Suess" cycle and the "11-year"cycle indeed describe solar activity, the "Gleissberg" cycle describes something else, possibly linked to solar activity in poorly understood ways.

Apparently the sediment supply from the mountains and drylands in southern California is not particularly responsive to solar variation on the decade to century scale. In any case, none of the dominant cycles below three decades long (near 29, near 23.4, and near 7 years) seen in sediments of Santa Barbara Basin are obviously of solar origin. They do suggest, however, that the varve thicknesses are capable of reflecting decadal-scale cycles if they exist. As concerns the ca.-200-year cycle, the available data series was not long enough for testing the Suess proposition. Thus, while we cannot categorically exclude the possibility that the sun affects climate (and hence marine sedimentation), the evidence available from thickness variation in varved sediments off Santa Barbara in southern California, for the last 500 years, is not demonstrably there.

Of course, just because we don't find convincing evidence in thickness variations of varves in one of the Santa Barbara cores does not mean that the proposed phenomenon does not exist. It just means that we cannot support the proposition of solar variation from these data, which are barren regarding the 11.1-year sunspot cycle and unconvincing with respect to the 205-year Suess cycle. Interestingly, analysis of a foraminiferal record in the Cariaco Basin published by D.E. Black of Stony Brook and associates yielded even less evidence for a solar effect (a Suess cycle was definitely not seen in their 825-year record). Thus, proven solar cycles do not show without fail in sediments.

Does the climate in fact vary with solar activity? Judging from ice-age information, it would seem that it can do so within a millennial time scale. The well-documented sun-forced Milankovitch fluctuations in deep-sea sediments, in polar ice, and in loess deposits involve tens of millennia, though. Perhaps what marine geology can contribute is to find locations where shorter periods are seen to have left a record also.

15.6.4 Bermuda Coral

Coral growth also may have information on variations in the environment. Presumably the building of a carbonate skeleton captures such information. However, the reaction of coral to solar variation, such as that of the massive *Montastraea cavernosa* studied by J. Pätzold and G. Wefer (then Kiel, now Bremen) in Bermuda, apparently is extremely complicated. One indeed finds cycles in the growth data, but they are not clearly linked to known solar forcing. The reason presumably has to do with the fact that supply of warm water and of nutrients are two fundamentally different growth requirements of the coral, implying at least two controlling factors (e.g., solar and tidal) that may interact. Interacting cycles of forcing may result in cyclic variations that are difficult to interpret.

15.7 Discussing the Future of Doing Marine Geology

Marine geologists (as is true for other scientists) are faced with the problem that we humans are on a path toward environmental catastrophe, according to a host of experts. (At least one well-known climatologist suggested in a popular book that we may generate a lifeless planet.) The chief concern is that we are ignoring unintended consequences of releasing carbon dioxide into the air. The greenhouse effect is a fact – it makes the planet habitable. Without the atmosphere and its greenhouse effects, our climate might be more like that of the moon, instead of what we have. There is no discussion or doubt about the benefit of greenhouse gas in the atmosphere. What is a matter of discussion among the experts is how much greenhouse gas is too much. As we humans increase the relevant gas content in the atmosphere at a rate and to a level unprecedented in the last several thousand millennia, we do get serious warnings from well-respected scientists about undesirable consequences of doing so. The most frightening warning is about the prospect of a runaway effect from a massive release of the gas methane, fast enough to escape prompt microbial conversion to carbon dioxide and harmless water, by oxidation.

The most reliable of the awful projections is that sea-level rise will continue to accompany global warming. We can be 100% sure of that because sea-level rise is already happening and can now be measured quite reliably using satellites, videos, and listening devices. On the one hand, we have admonitions from a great number of working scientists that we cannot continue with business as usual without facing unacceptable risks. On the other hand, we have stern warnings from certain sociologists and politically active persons (including representatives of industry) about troublesome economic and political consequences from abandoning or greatly reducing carbon-based energy use. There also is a third message. The message from many Dutch engineers is this: let's make the dams higher, just in case. The arguments for all three messages have changed very little for decades. What *has* changed is a sense of urgency, a perception that we are running out of time for discussions.

Incidentally, serious skepticism is not the issue when examining the work of scientists, including that of marine geologists. All scientists are trained to be skeptical; that is, to prefer evidence over just-so assertions. When discussing the future development of climate, of course, there *is* no good evidence. The future, by definition, has not happened yet. Thus, there cannot be any records to study. Unfortunately, when our future does happen, it is likely to bring an unfamiliar world, a world even more poorly understood than the one we live in. Increasingly, therefore, as we get closer to an unfamiliar world, our projections are likely to be out of sync with reality. The task for marine geologists is to gather real evidence and to ignore assertions based on personal preference or fear (attitudes sometimes confused with "skepticism"). The task is, obviously, to help generate acceptable risks. Among the many challenges arising in the context, helping to avoid unintended consequences of purported remedial action is well worth tackling, as is the prospect of sea-level rise and methane runaway.

Suggestions for Further Reading

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Appendix

1.1 Conversions Between Common US Units and Metric Units

One of the most common conversions seen is the one for temperature (Fig. A1.1):

$$C = (F - 32) \times 5/9$$

$$F = C \times 9/5 + 32$$

where C stands for degrees Celsius (centigrade) and F for Fahrenheit

$$K = C + 273.2$$

where K stands for degrees Kelvin (absolute temperature).

Figure A1.1 Many thermometers can be read in °C or in °F, demonstrating a simple linearity in the conversion (see equations)

Length	
1 cm = 0.394 inch	1 inch = 2.54 cm
1 m = 3.281 feet	1 foot = 0.305 m
1 km =0.621 statute miles	1 mile = 1.609 km
	1 nautical mile = 1.852 km
Volume	
1 mm = 1000 μm (micrometer or microns) (international)	
1 liter (l) = 1000 milliliter (ml) = 0.264 US gallons	1 US gallon = 3.781 l
Mass	
1 barrel (oil) = 42 gallons =159.1 liter	
1 kilogram (kg) = 1000 grams (g) = 2.205 US pounds	1 pound = 0.454 kg
1 metric ton =1000 kg = 1.102 US short tons	1 short ton = 907.4 kg

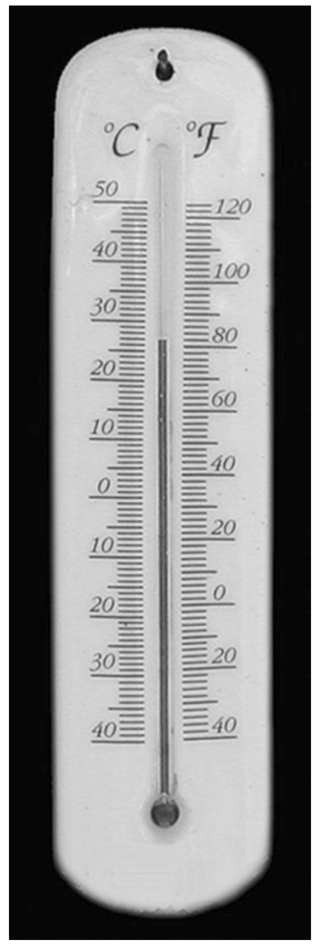

Fig. A1.1 Thermometer with two temperature scales (Photo W.H.B.)

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3

1.2 Topographic Statistics

Earth		
Equatorial radius = 6378 km	Polar radius = 6356 km	
Ocean		
Atlantic Ocean (incl. Arctic and marginal seas)		
Area = 106.5 million km^2	Volume = 354.7 million km^3	Mean depth = 3332 m
Indian Ocean (incl. Adjacent seas)		
Area = 74.9 million km^2	Volume = 291.9 million km^3	Mean depth = 3897 m
Pacific Ocean (incl. Adjacent seas)		
Area = 179.7 million km^2	Volume = 723.7 million km^3	Mean depth = 4028 m
Total Ocean (incl. Adjacent seas)		
Area = 361.1 million km^2	Volume = 1370.3 million km^3	Mean depth = 3795 m
(Source: E. Kossinna, 1921, as quoted in H.U. Sverdrup et al., 1942. The Oceans. Prentice-Hall, Englewood Cliffs, NJ)		
Shelf seas		
(Slightly over 3% of the total ocean area.) Collation of continental and oceanic areas in Kossinna's data prevents recognition of shelf sea properties. The data shown are from the compilation in Table 2.1 in the third edition of the present text		
Area = 11.4 million km^2	Volume = 2.28 million km^3	Depth (defined!) = 200 m

1.3 Geologic Time Scale

On the modern seafloor, ages greater than 100 million years are rare. Nevertheless, it is well to keep in mind that more ancient marine rocks are common on land. Shown: time scale for marine fossils for the entire Phanerozoic, beginning with the Cambrian (Figs. A3.1 and A3.2).

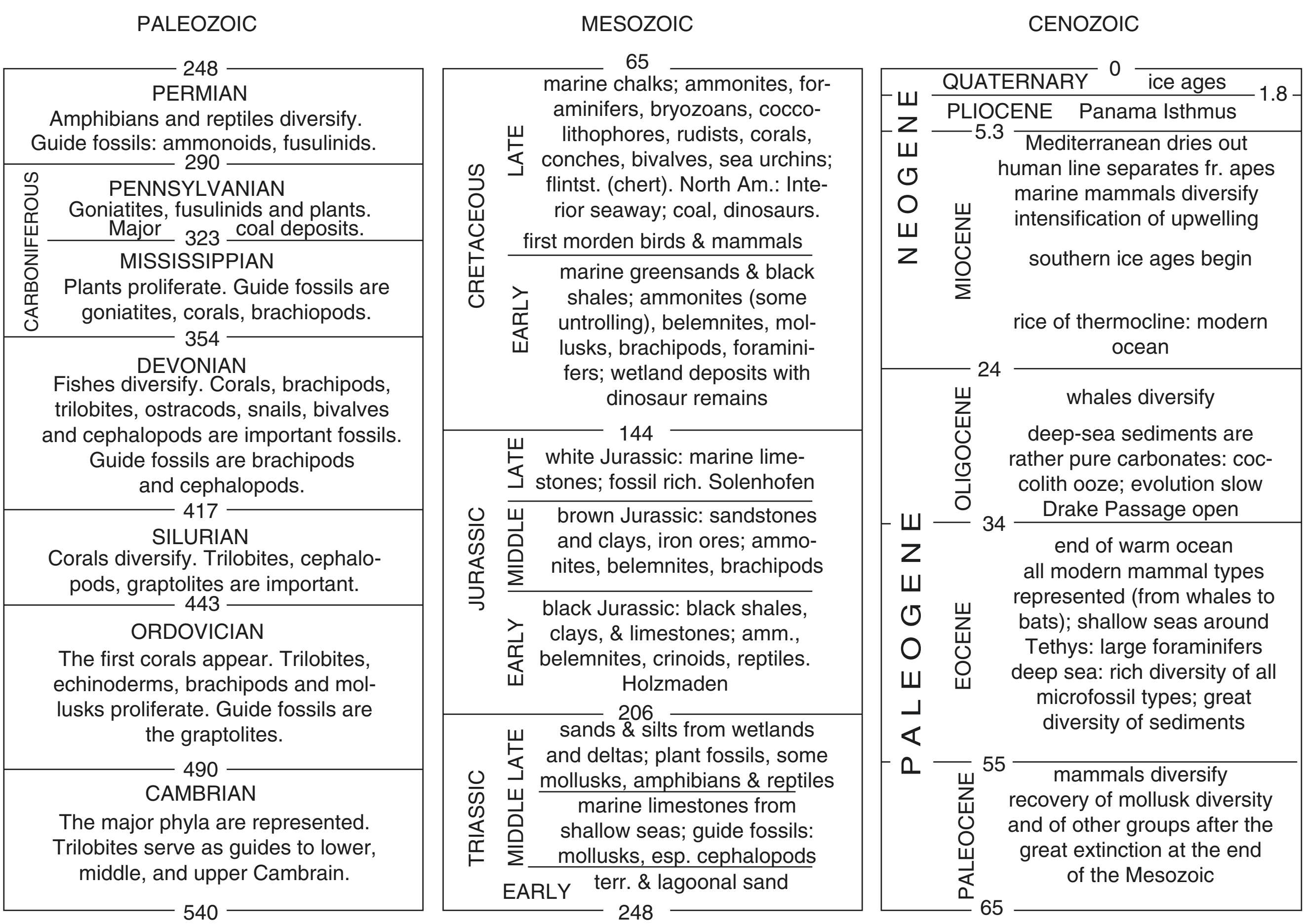

Fig. A3.1 Approximate time scale for the Phanerozoic (i.e., for rocks with marine fossils) in millions of years (Sources: The Geological Society of America, Boulder, Colorado and various geology textbooks)

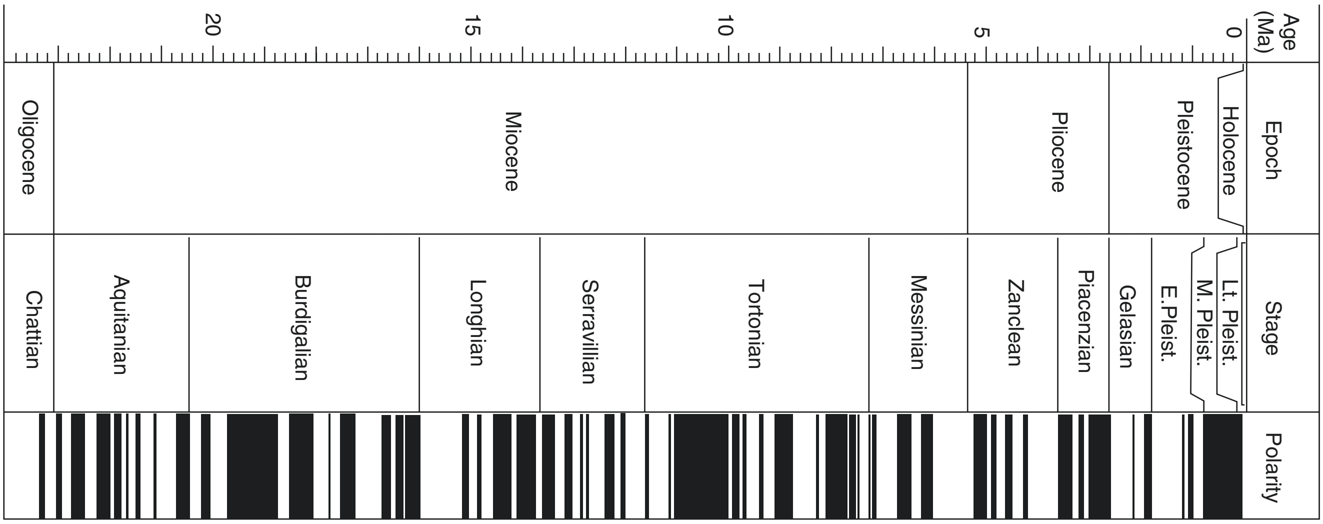

Fig. A3.2 Modern age assignments rely heavily on correlations using the geomagnetic time scale, especially in Neogene sediments (Sources: the Ocean Drilling Program after 1990)

1.4 Common Minerals

1.4.1 Silicate Minerals

Perhaps the most widely known mineral is the ubiquitous clear or white-colored glassy-looking *quartz,* with the formula SiO_2. It is common in granitic rocks (i.e., in continental crust) and is made of *silicon tetrahedra* (Fig. A4.1) such that each oxygen atom is shared by a neighboring tetrahedron (hence the formula, with two rather than four oxygen atoms for each silicon atom). "Opal" is the same but less well ordered and with plenty of water molecules accommodated between the silicate tetrahedra. By replacing some Si atoms with Al atoms, the tetrahedra structure retains a (negative) charge,

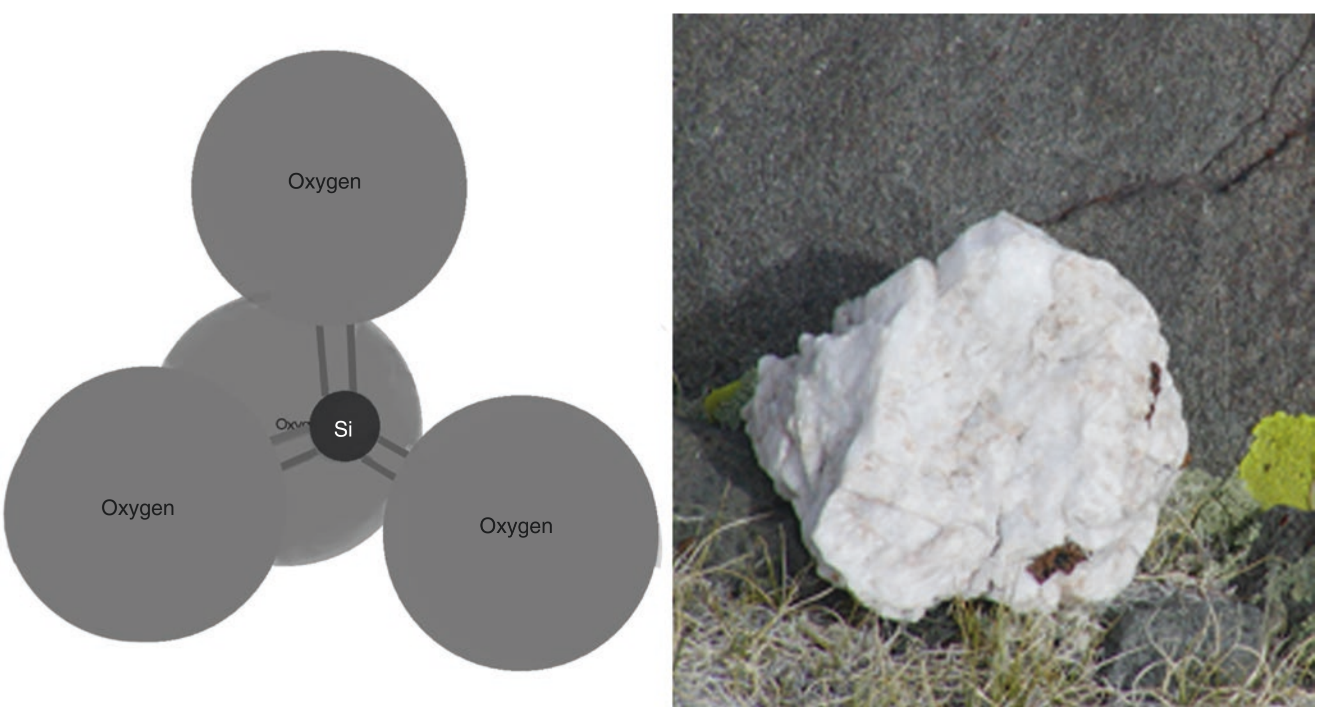

Fig. A4.1 Quartz, the most common silicate mineral on the surface of the planet and its structure. *Left*: silicate tetrahedron: a silicon atom surrounded by four oxygen atoms (fourth set of binding forces obscured). The spheres represent the atoms. [See any text book on minerals] *Right*: milky quartz crystal on White Mountain, near Bishop, Calif (Photo W.H.B.)

which is balanced by cations (Ca, Na, or K). This arrangement results in *feldspars* (e.g., albite, $NaAlSi_3O_8$, microcline and orthoclase, KAl Si_3O_8, anorthite, $CaAl_2Si_2O_8$). Feldspars are ubiquitous, with sodium and potassium feldspars common in granitic rocks and calcium feldspars in basaltic rocks.

The shimmering platy minerals commonly seen on sandy beaches are *mica*. Mica represents ubiquitous potassium-rich silicate minerals common in granitic rocks and in all marine sediments because of ease of transportation. Its sheet structure derives from the fact that its silicate tetrahedra are joined by three atoms in a plane, rather than by four oxygen atoms in a three-dimensional structure. Familiar examples of the mineral are the silvery "muscovite" (used, e.g., for windows into furnaces) and the dark magnesium- and iron-rich "biotite." Clay minerals are similar to "mica" but are deficient in cations. The most notable representatives are "smectite" (or "montmorillonite"), a weathering product of basaltic rocks; "illite" (somewhat similar, but commonly on a path toward "mica," with less water and more potassium than "smectite"); "chlorite," a weathering product of metamorphic continental crustal rock; and "kaolinite," a product of chemical weathering, stripped of cations.

Other minerals of note are the chain-like silicates in the "hornblende" group (double chains, "amphiboles") and those in the "augite" group (single chain, "pyroxenes"). "Pyroxenes" (note the reference to pyr = fire) are common in volcanic rocks, as are "olivines" (dark Mg- and Fe-rich silicates with tetrahedral in insular arrangement).

Zeolites (Fig. A4.2) are feldspar-like silicate minerals precipitated in places on the seafloor and within deep-sea clays, commonly in volcanogenic and other silica-rich environments. Unlike many other silicates, they are commonly fully marine in origin.

1.4.2 Nonsilicate Minerals

Carbonates. Bulk of biogenous sediments. Examples:
- Calcite ($CaCO_3$) calcareous shells and skeletons
- Aragonite ($CaCO_3$) ditto. Easily dissolved
- Dolomite (Ca Mg $(CaCO_3)_2$) product of diagenesis. Very resistant.

Evaporite minerals. Most common: calcium sulfate (anhydrite and gypsum)
- Commonly seen in drylands as a white cover of dried-up marine wetlands. Much less common: halite = sodium chloride Na Cl (kitchen salt)

Iron oxides and sulfides. Oxides and sulfides of iron are very common in marine sediments (here commonly hydroxides at the modern surface) and ubiquitous in all crustal rocks. The presence of sulfide minerals within sediments (mostly iron sulfides) indicates reduction of sulfate either on the surface or (mainly) after burial. Iron sulfide is abundant in organic-rich sediments, which also may have abundant diatoms (Fig. A4.3).

1.4.3 Heavy Minerals

The minerals listed here (excepting carbonates and evaporites) also are counted as heavy minerals if they are compact and make sand grains. Heavy minerals are useful as tracers for the sediment they are part of.

Details on minerals are in mineralogy textbooks and in textbooks on general geology.

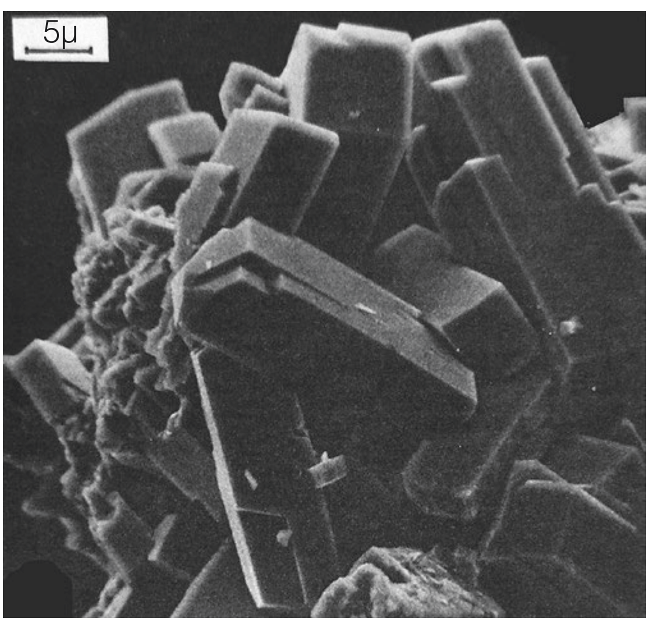

Fig. A4.2 Silt-size zeolite crystals from Paleocene sediments in the central Atlantic (clinoptilolite) (W.H.B. and U. von Rad, 1971. DSDP Leg 14; SEM by C. Samtleben, Kiel)

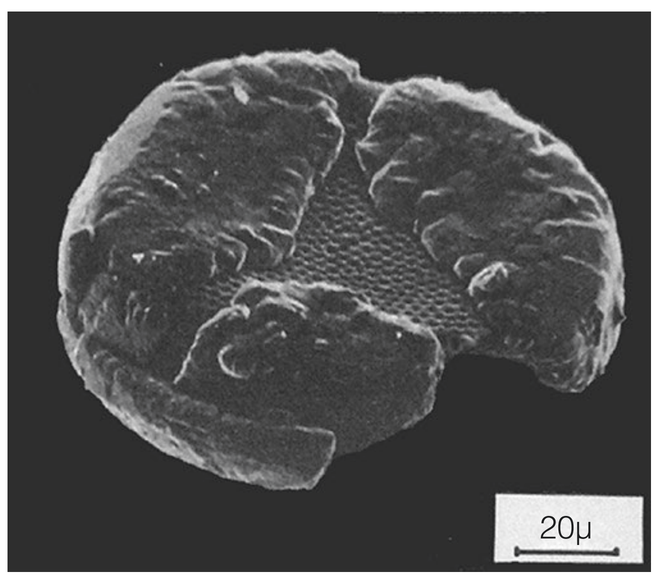

Fig. A4.3 Miocene diatom shell from sediments in the central Atlantic, with iron sulfide precipitation (W.H.B and U. von Rad, 1971. DSDP Leg 14. SEM by C. Samtleben, Kiel)

1.5 Grain Size Classification for Sediments

Boulder	Gravel	Sand	Silt	Clay
mm	256	2	0.063	0.004

In some classifications, 0.002 mm separates silt from clay. Also see Sect. 4.5.1 and Chap. 5.

1.6 Common Rock Types

1.6.1 Igneous Rocks

The Earth is hot inside (Fig. A6.1), in and below the lowermost crust, oceanic, and continental. It is here one finds igneous rocks, that is, rocks that are crystallized from hot melt (Engl. cognate: ignition). Igneous rocks make up the bulk of the Earth's crust and much of the uppermost mantle material (some of which is soft, being very hot). Continental crust: largely granitic, rich in silicate minerals (large ones typical for slow cooling). Granitic rock is commonly seen in uplifted roots of mountains, such as the Sierra Nevada and peninsular extension in Southern California (Fig. A6.2). In contrast, all of the oceanic crust is basaltic (rocks seen on the surface are in large part from ash, some welded, or from volcanic extrusions; see Figs. A6.3 and A6.4). The fundamental difference in rock types of continental and oceanic crust was used by A. Wegener in his arguments for continental drift, early in the twentieth century. The classification of igneous rocks is based on silicate content (high for granitic rocks, low for basalt) and on crystal size (coarse crystals for intrusive rocks and microscopic ones for extrusive rocks).

Fig. A6.1 Evidence that Earth is hot inside: geysirs in Iceland (Photo W.H.B.)

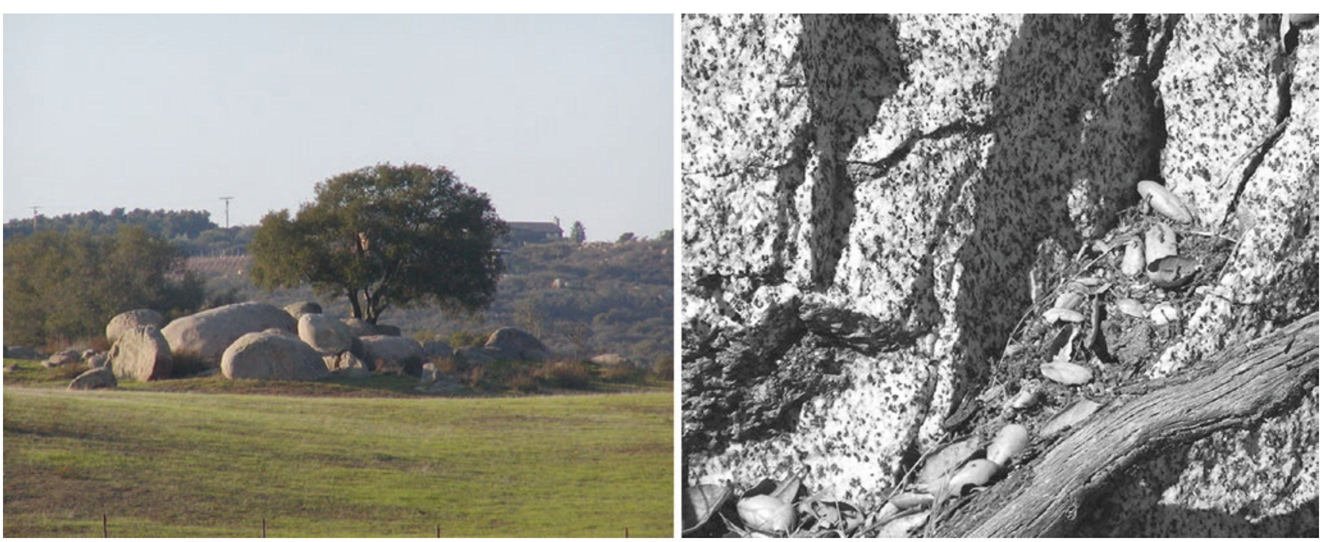

Fig. A6.2 Light-colored granitic rock, mountain roots exposed after removal of "overburden," San Diego County (Photos W.H.B.)

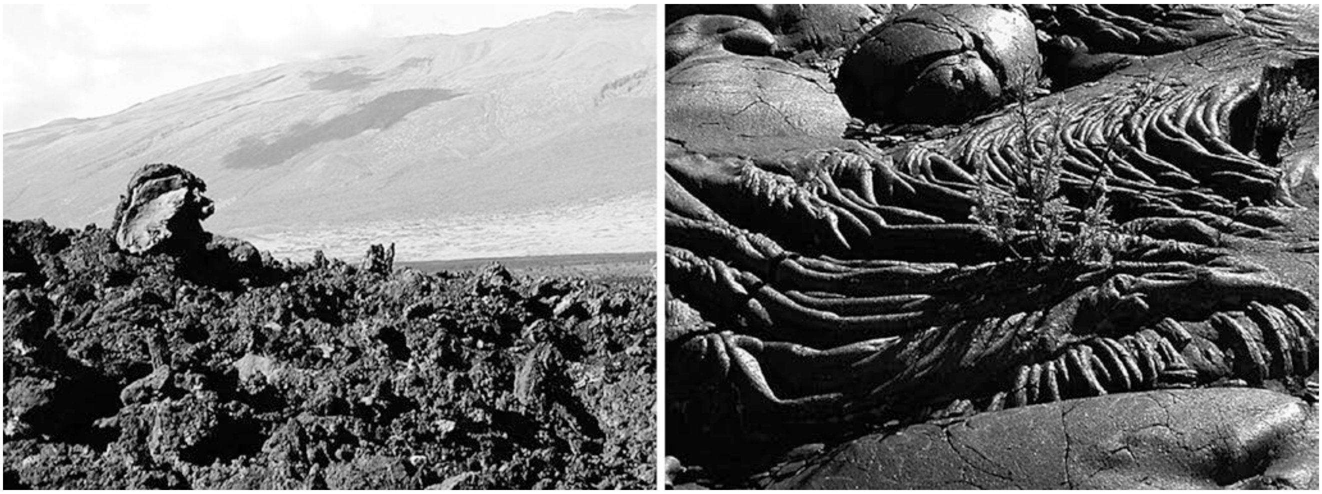

Fig. A6.3 Dark basaltic lava rocks, Hawaii (*Left*: rough aa lava; right: smooth pahoehoe flow) (Photos W.H.B.)

Fig. A6.4 Volcanogenic boulders (scoriaceous basalt; two light gray and white coral pieces, beach, Honolulu) (Photo W.H.B.) "scoria" = slag

Fig. A6.5 Basalt columns in Iceland. The entire island is made of volcanic rock (Photo W.H.B.)

Much of the seafloor basement is on oceanic crust and made of basaltic volcanic rock, some apparently derived from eruptions and intrusions (as in Hawaii and in Iceland. Much deep-sea sand, especially where influenced by the volcanic "Ring of Fire," has a strong showing of volcanic debris (Figs. A6.5 and A6.6).

On land, one can find ash deposits surrounding explosive volcanoes (Fig. A6.7). Examples are abundant in the Andes (here: Chile, Villarica). Layers of ash (seen on the flanks of the volcano) mark eruptions.

"Andesites" are intermediate in composition between granitic and basaltic rocks. The intermediate composition presumably is the result of mixing materials from the two major types of crust in Andes-type mixing machines (as in Peruvian mountains) next to subduction zones. The "andesite" line separates andesitic volcanoes and their rocks from (oceanic) basaltic ones. It is parallel to the Pacific "Ring of Fire" for much of its course. It was first mentioned (and named) in the nineteenth century.

1.6.2 Sedimentary Rocks

The most abundant types of sedimentary rock are "shale," "sandstone," and "limestone." "Limestone" refers to rocks made of calcium carbonate. The rock can be massive (Fig. 13.1) or layered (Fig. A6.8). Shale derives from siltstone, claystone, or mudstone. Commonly, it parts along very thin bedding planes. Much of it presumably originates from once widespread anaerobic marine sediments – sequences of ancient seafloor horizons deficient in oxygen. Examples are especially common in the Paleozoic and Mesozoic, but they also are seen in the Cenozoic. In the Paleozoic, some ancient shales have delicate marine fossils (presumably once drifting organisms) preserved as pyrite structures. The pieces show in X-ray graphs. Geologically young black shales may be well lithified or else still contain considerable water. An outstanding Mesozoic example is provided

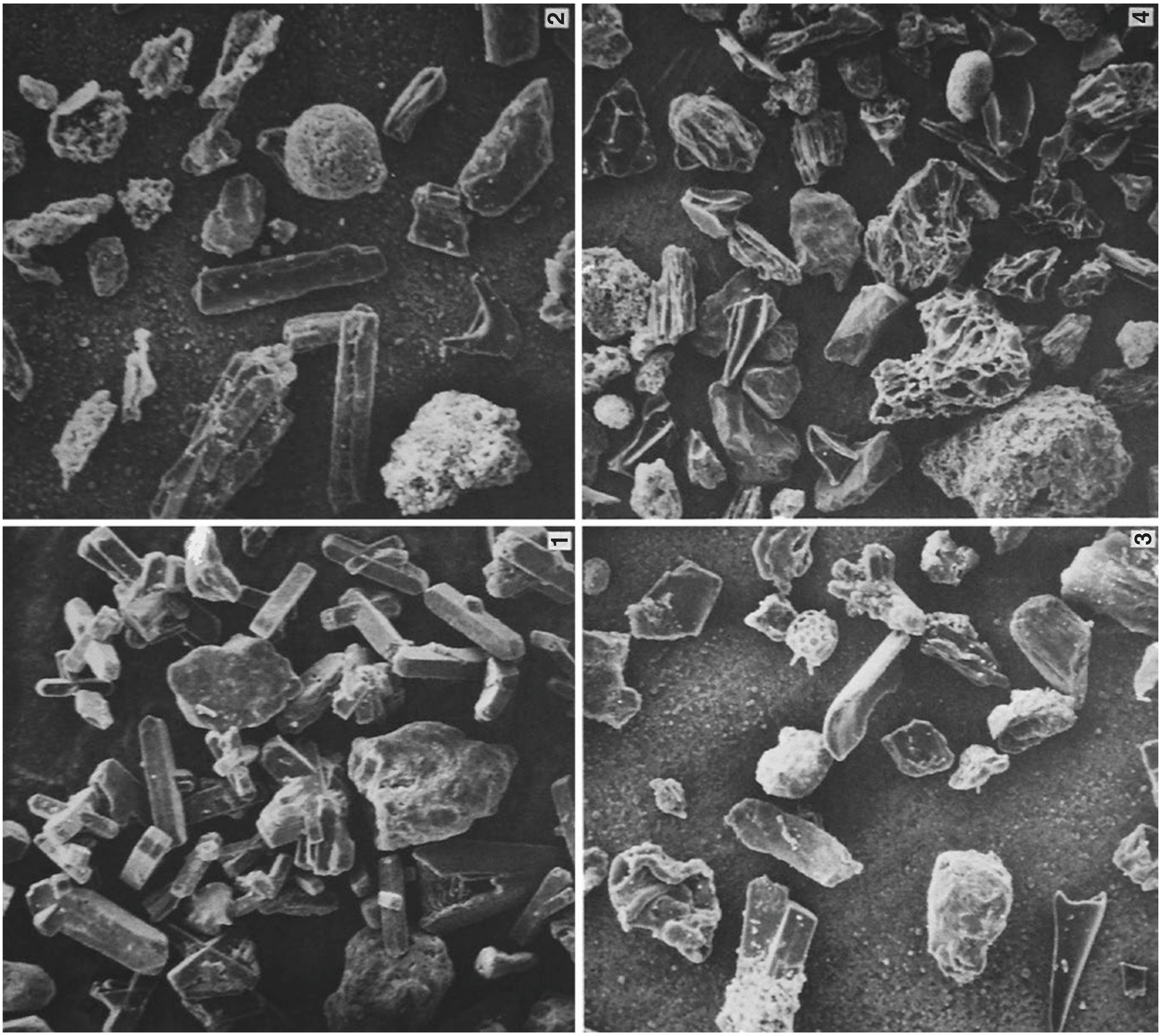

Fig. A6.6 Deep-sea sand with strong volcanogenic affinity (From A.C. Pimm, DSDP Leg 6. Northwestern Pacific. Images provided by S.A. Kling (then S.I.O.))

Fig. A6.7 Volcanic ash layers on the road to Villarica Volcano, Chile (Photo W.H.B.)

Fig. A6.8 Jurassic bedded limestone. Quarry near Crailsheim in Southern Germany (Photo W.H.B.)

by the Black Jurassic. Many fine examples of various types of Cretaceous marine sedimentary rock are now on land (Fig. 13.1).

Bedded (layered) limestones are familiar from Mesozoic sequences (Fig. A6.8) but occur abundantly all through the Phanerozoic and in earlier sequences. The bulk of this rock type is of marine origin, although other types do occur also, notably varieties derived from lake deposits ("limnic" sediments).

Among common but not dominant sedimentary rocks are cherts, coal, and phosphorites. Chert is typified by microcrystalline quartz, the latter presumably largely of biogenous origin (although this is difficult to document, especially in ancient rocks). The rocks range from silicified mudstones to very fine-grained quartzite. Coal usually contains terrestrial plant fossils. Phosphorite, a rock originating from phosphatic deposits, apparently is largely of marine origin: many phosphorites have marine fossils.

A widespread sedimentary rock (albeit not necessarily recognized as such) is ice, both in glaciers and at sea. On land, it originates from snow, that is, from sedimentary particles (frozen water) settling through the air. When close to its melting point, the ice is mobile and flows downhill. Being less dense than water, it floats at sea. For marine geologists, floating ice incorporating rock particles is very important as a transportation agent for IRD (*ice-rafted debris*). Sea ice results from freezing seawater at the surface. On and just below the seafloor, there can be methane ice, which is a type of water ice accommodating large amounts of methane. Upon melting, methane ice (*clathrate*) releases large amounts of combustible gas with strong greenhouse properties.

1.6.3 Metamorphic Rocks

Metamorphic rocks are mineral assemblages produced by deformation and recrystallization of igneous or sedimentary rocks, through elevated temperature and pressure, with or without addition of material through injection of solutions. Metamorphic rocks occur preferentially at collision margins, such as the ones in California, Alaska, or Japan, along the rim of the Pacific. Classification by origin (the most useful kind) is commonly very difficult, and thus classification tends to be descriptive in many cases (e.g., based on properties such as types of foliation, minerals present, grain size). Ages of metamorphic rocks on land have a large range, with a strong link to mountain building. Uplift and removal of overburden (some by polar ice masses) exposes ancient mountain roots with abundant metamorphic rocks (Figs. A6.9, A6.10, and A6.11). The resulting labels include the terms hornfels, amphibolite, granulite, slate, schist, and gneiss. In contrast, the labels marble and quartzite point to an origin from carbonate rocks or quartz sandstone, respectively.

Fig. A6.9 Metamorphic rock in Southern California, presumably of igneous origin. San Diego County (Photo W.H.B.) (US five-cent piece for scale.)

Fig. A6.10 Metamorphic rock, Norway (Photo W.H.B.)

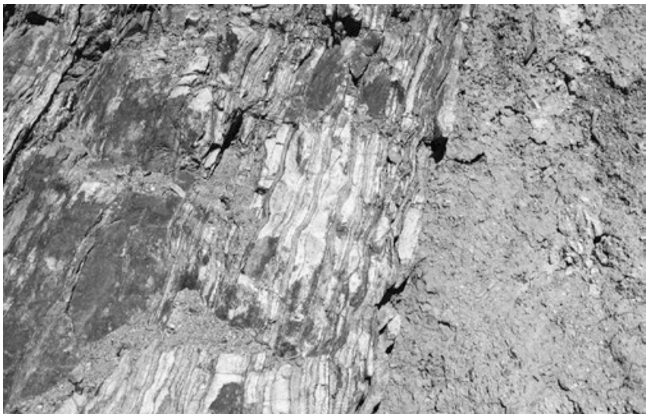

Fig. A6.11 Metamorphic sedimentary rocks from the mountain root zone in San Diego County, Southern California (Photo W.H.B.)

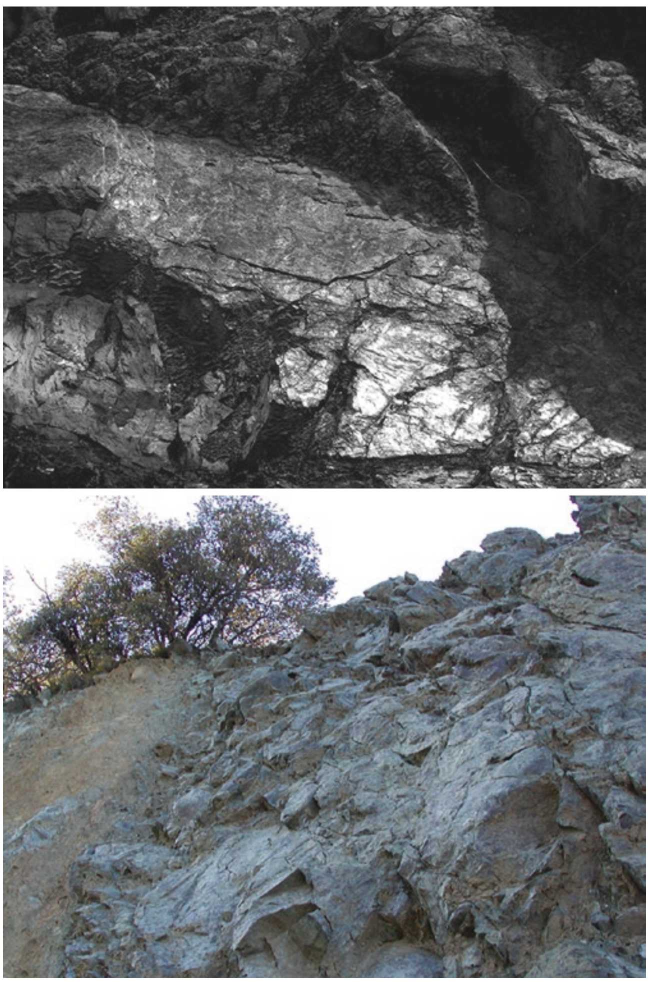

Fig. A6.12 Ophiolite (serpentinized basalt). *Top*: in a road cut near Santa Barbara, Southern California (*gray*). *Bottom*: road cut, Big Sur, California (Photos W.H.B.)

For marine geologists, metamorphosed and chemically altered basalt ("serpentinite," "ophiolite") is of special interest. It represents pieces of basaltic seafloor that ended up within active margin mountains (Fig. A6.12).

(Consult any general geology textbook or petrology text for details on rock types and their distribution.)

1.7 Geochemical Statistics

The Earth's crust mainly consists of aluminum silicates (oxygen, silicon, aluminum), with alkaline and earth alkaline elements for cations (Na, K, Ca, Mg) and with iron (Fe) ubiquitous within silicate or as sulfides and oxides mostly (see Table A7.1). Continental crust has a composition resembling that of granodiorite (a granitic rock with a reduced content of quartz, compared with classic granite, a feldspathic rock rich in mica and quartz). Ocean crust is mainly basalt (no quartz, rich in Fe and Mg). Seawater is largely a watery solution of kitchen salt (Na, Cl) with a bit of Epsom salt mixed in ($Mg\ SO_4$). The atmosphere is a mixture of nitrogen and oxygen gas (N_2, O_2), both intimately tied to biological cycling (i.e., to the carbon cycle) (see Table A7.1).

On the whole, igneous rocks and the sediments derived from them (mainly shale-type sediments) have very similar composition (Table A7.2). Sodium tends to be depleted in the sediment; much of it ends up in seawater. The high content in carbon in sediments is owing to the presence of carbonates and also of organic carbon (most of it finely dispersed). The sulfur content is largely linked to evaporites (e.g., gypsum). Compared with marine sediments on land, "Red Clay" is enriched in highly oxidized iron and manganese. Also it is unusually low in carbon (carbonate is dissolved; organic carbon is oxidized) but relatively high in water content**.** Differences in crustal composition (used by

Table A7.1 Chemical ingredients in crust, ocean, and atmosphere (weight %)

	Cont. crust	Ocean crust	Seawater	Atmosphere
O	46.3	43.6	85.8	21.0
Si	28.1	23.9	–	–
Al	8.2	8.8	–	–
Fe	5.6	8.6	–	–
Ca	4.2	6.7	0.04	–
Na	2.4	1.9	1.1	–
Mg	2.3	4.5	0.14	–
K	2.1	0.8	0.04	–
Ti	0.6	0.9	–	
H	0.14	0.2	10.7	–
P	0.10	0.14	–	–
Cl	–	–	2.0	–
N	–	–	–	78.1
Ar	–	–	–	0.9
CO_2	–	–	tr	0.04

Data of geochemistry. Sources for the tables: mainly R.W. Fairbridge (ed) 1972. The Encyclopedia of Geochemistry and Environmental Sciences. Van Nostrand Reinhold, New York.; and A.E.J. Engel and C.G. Engel, 1971. In A. Maxwell (ed) The Sea vol. 4, pt. 1. Wiley, New York

Table A7.2 Composition of igneous and sedimentary rocks and of Red Clay (percent)

	Igneous (continent)	Sedimentary (continent)	Tholeiite (deep-sea basalt)	Alkali basalt	Red Clay (deep seafloor)
SiO_2	59.1	57.9	50.2	48.2	53.7
Al_2O_3	15.3	13.3	16.2	16.5	17.4
FeO	3.8	2.1	7.1	7.6	0.5
Fe_2O_3	3.1	3.5	2.6	4.2	8.5
CaO	5.1	5.9	11.4	9.1	1.6
Na_2O	3.8	1.1	2.8	3.7	1.3
MgO	3.5	2.7	7.7	5.3	4.6
K_2O	3.1	2.9	0.2	1.9	3.7
H_2O	1.1	3.2	<1	<1	6.3
TiO_2	1.1	0.6	1.5	2.9	1.0
MnO	–	0.1	0.2	0.2	0.8
P_2O_5	0.3	0.1	0.1	0.5	0.1
CO_2	0.1	5.4	–	–	0.4
C(org)	–	0.7	–	–	0.1
SO_3	–	0.5	–	–	–

For details on the chemical composition of Earth and its crust, see any textbook on general geology or a text on geochemistry

A. Wegener in his arguments concerning continental drift) are quite obvious (see Table A7.1, columns 1 and 2).

1.8 Radioisotopes and Dating

Atoms of the same kind are called *isotopes* (a label referring to the fact that they take up the same place in the "periodic table"). The atoms that emit radiation are *radioactive isotopes* or r*adioisotopes.* Such isotopes may emit alpha, beta, or gamma radiation. An alpha particle is a helium nucleus (i.e., it consists of two protons and two neutrons) and is commonly ejected from the nucleus of a large element in the uranium series. A beta particle is a high-speed electron, commonly ejected from radioactive carbon. Gamma rays are electromagnetic waves similar to X-rays, except much more energetic.

Over the ranges of temperature and pressure investigated, radioisotopes were found to emit radiation independently of conditions, including chemical configuration. Thus, emissions from an element are considered to be only dependent on the number of radioactive atoms present. From this, it follows that the number of atoms remaining after time t from an initial number of radioisotopes is given by a simple decay formula, which may be written

$$N / N_0 = e^{-\lambda t} \qquad \text{(A8.1)}$$

Solving for "t" yields the time of decay that reduced N_0 to N. The decay constant λ obviously must be known before the solution can be found. For radiocarbon, it is near 0.00012.

$$\ln\left(N_0 / N\right) = \lambda \times t; \qquad \text{(A8.2)}$$

$$t = \ln\left(N_0 / N\right) / \lambda. \qquad \text{(A8.3)}$$

Setting N/N_0 equal to 1/2 and then solving for t yields the *half-life* of a radioactive isotope. Half-lives vary greatly. For radiocarbon (^{14}C), it is close to 6000 years. (Radiocarbon decays back to ^{14}N, from which it arose through cosmic ray bombardment of the air.) The half-life for the most common uranium (^{238}U) is near 4.5 billion years (close to the age of the Earth, it is thought). It decays to lead which is at the end of a long decay series involving several different types of radioactive elements.

After ten half-lives, only about one tenth of a percent is still present of the original number of radioisotopes. Thus, some radioisotopes become unsuitable for dating at some point. For the "short-lived" radiocarbon, with its 6000-year half-life, the point is reached near 40,000 years. Beyond about seven half-lives, the decay signal is very vulnerable to contamination. For the dating of older products (e.g., oceanic crust and also sediments), there are long-lived radioisotopes. Examples are potassium-40 (which decays to several elements, notably its daughter element argon-40). Dating by potassium-40 decay delivered the time scale for seafloor ages. (Transfer of such ages from one place to another is by correlation, including correlation using magnetic reversal sequences.) Other commonly used long-lived radioisotopes, besides potassium-40 (^{40}K), are rubidium-87, thorium-232, uranium-235, and uranium-238 (^{87}Rb, ^{232}Th, ^{235}U, ^{238}U). Their decay (in some cases involving decay series – a cascade of radioisotopes ending up in stable elements) has been studied in some detail and applied to the dating of rocks and sediments.

Man-made radioisotopes (such as plutonium and other bomb-related products) carry information not normally an important part of routine dating: their appearance is geologically extremely young.

Details on radioactivity in the sea and on the seafloor and its use in the study of seafloor age and sedimentation processes are available in any geochemistry text and in many textbooks in general geology.

1.9 Marine Organisms Involved in Seafloor Processes

1.9.1 General Remarks

There are many more different species on land than at sea, mainly owing to the proliferation of insects (phylum arthropods) on land. The number of *noninsect* species in the sea outnumbers those on land in the ratio of roughly 2 is to 1 (as does area of the sea to area of land). Marine benthic species greatly outnumber planktonic ones in practically all categories by various factors. It is a factor of more than 100 for foraminifers (not subtle). In the phylum echinoderms, the ratio approaches infinity for certain crinoids, adult forms being extinct in the plankton (they are still seen as plankton fossils in rocks of Mesozoic age, having been reported from quarries in Solnhofen and from DSDP Leg 11). Infinity is avoided if one counts larvae as establishing presence. Similarly poorly defined ratios obtain in some holothurians floating close to the bottom and of questionable affiliation, benthic or planktonic or both, like many a flatfish (phylum chordates).

On the basis of similarity of appearance (and most recently of chemistry), organisms are classified into species, which are grouped into *genera* (plural of *genus*), which in turn are grouped into *families,* then *orders*, then *classes*, and finally *phyla* (plural of *phylum*). The classification is a legacy of the Swedish naturalist Carolus Linnaeus (Carl von Linné), working in the eighteenth century.

Phyla that are similar to each other make up a *kingdom*. To establish similarity between different phyla can be a bit of a challenge: the groupings at this high level can be arbitrary. Commonly cell properties are being compared. Presumably, a kingdom has implications for common ancestry deep in the Precambrian, while classes imply ancestry in the early Paleozoic, with successively younger geologic ancestors for the groupings' orders, families, and genera, at least in large multicellular organisms. The discovery of a new kingdom in recent decades (the "archaea," by the biologist C. Woese, 1928–2012, working in Urbana, Illinois) suggests that the task of proper classification of organisms has by no means ended, especially when contemplating bacteria and archaea. The life-forms in question (prokaryotic microbes) are greatly involved in many geochemical processes, including those on the seafloor.

At this highest level of classification (kingdoms), there is a choice between fundamentally different ways of viewing life-forms. Some scientists are still aware of the ancient divisions of "animals," "plants," "fungi," "protists," and "monera," a classification that goes back to the early nineteenth century and has no biochemistry in it. The category "monera," once used for prokaryotic microbes such as bacteria, is now obsolete. Many biologists prefer to consider only prokaryotes (cells without a membrane-enclosed nucleus) and eukaryotes (cells with a well-defined nucleus) as of fundamental importance. In such a simplified scheme, the eukaryotes include animals, plants, fungi, and microscopic "protists" such as foraminifers, coccolithophores, and radiolarians, while the prokaryotes contain the bacteria and the archaea. The latter apparently are ancient life-forms now found abundantly in hot springs and salty environments and involved in anaerobic reactions. They may superficially resemble bacteria but are quite different from them biochemically.

In the modern threefold fundamental classification, there are then three "domains" (kingdoms) – bacteria, archaea, and eukaryotes. All are abundant on the seafloor. Whether eukaryotes are derived from prokaryotes by complexification or prokaryotes originate (at least in part) from eukaryotes by simplification is a matter of research and discussion. In the marine rock record, eukaryotes apparently are the younger group of organisms, arising hundreds of millions of years later than prokaryotes. Also, eukaryotes are organisms with plenty of multicellular representatives (i.e., with animals, plants, and fungi), which are not much in evidence before the Phanerozoic (although multicellular remains, presumably marine, are indeed present as fossils in certain late Precambrian rocks prior to the Cambrian).

When we discuss life on the seafloor, we tend to emphasize the types of organisms one finds at the shore and in the fish markets, that is, multicellular eukaryotes. But it is microbes that dominate the marine environment. Their remains cover most of the seafloor. Also, members of the group "microbes" perform the photosynthesis in corals. The sedimentary reactions on and within the seafloor, including decay of organic matter and recycling of nutrients, are largely performed by bacteria and archaea.

1.9.2 Microbes and Marine Sediments

Archaea and bacteria largely appear in their products (e.g., methane ice, iron sulfides, ferromanganese nodules) and more generally in chemical reactions in seawater and on the seafloor (e.g., photosynthesis, nitrogen reduction, many types of mineral precipitation, early diagenesis). Their activities dominate the marine environment. Nevertheless, the chief targets of microbial studies by marine geologists classically have been eukaryotes, especially foraminifers. All of the eukaryotic microbes are useful in stratigraphy, as well as in assessment of environmental conditions, since they normally leave recognizable skeletal remains in marine sediments (Fig. A9.1).

Much of micropaleontology is in fact focused on the contents of the fine sand fraction of marine sediments. Study targets include the ubiquitous foraminifers (Fig. A9.2) and other microbes. Nannofossils, the one group of microbes studied roughly equally intensely as foraminifers, are mainly in the fine silt, however. Occasionally, one sees fish teeth in marine sediments. Such remains, while difficult to identify, can be of some use for stratigraphic purposes, especially where other fossils are lacking.

1.9.3 Dominant Marine Organisms

Multicellular marine creatures are eukaryotic, as mentioned. The ones of relevance in marine geology make fossils or produce bioturbation. The best-known examples of the former are in coral reefs and include calcareous

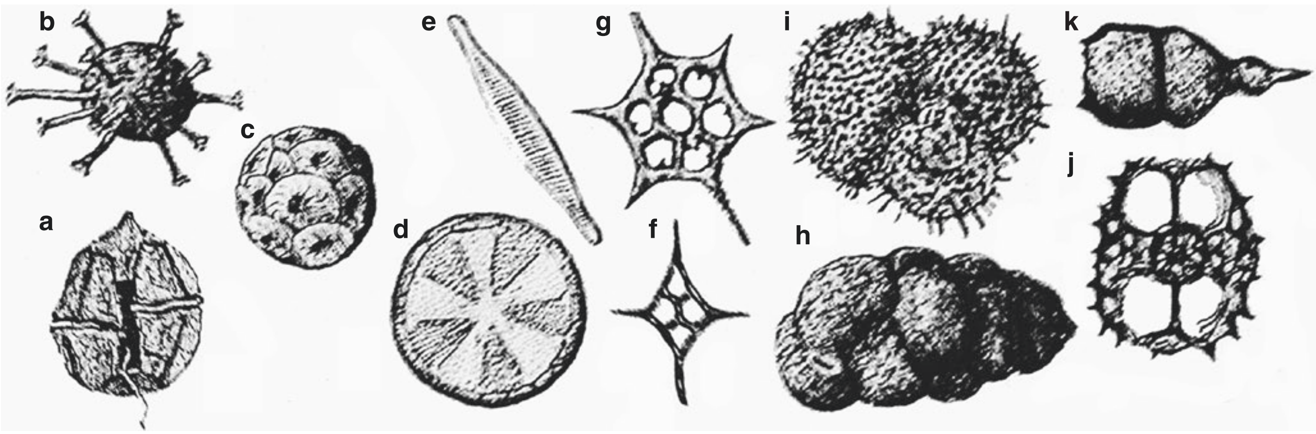

Fig. A9.1 Eukaryotic microbes found in marine sediments: fossils of micropaleontology. (**a**) and (**b**) dinoflagellates. Photosynthesizing. Common as symbionts in various organisms, including certain other microbes. (**c**) Coccolithophore (nannofossil). Photosynthesizing. Remains are ubiquitous in calcareous sediments; rock forming in many places. (**d**, **e**) Diatoms. Photosynthesizing. Siliceous skeletons. Colonial forms abundant in marine plankton. Can be rock forming. (**f**, **g**) Silicoflagellates. Similar to diatoms but with an internal skeleton. (**h**, **i**) Foraminifers. Heterotrophic plankton and benthos. Remains are common in "*Globigerina* ooze." Can be rock forming. (**j**, **k**) Radiolarians. Heterotrophic. Internal skeleton of silica. "Radiolarian ooze" in places. Ancient fossils (Drawings from the third edition, after various sources)

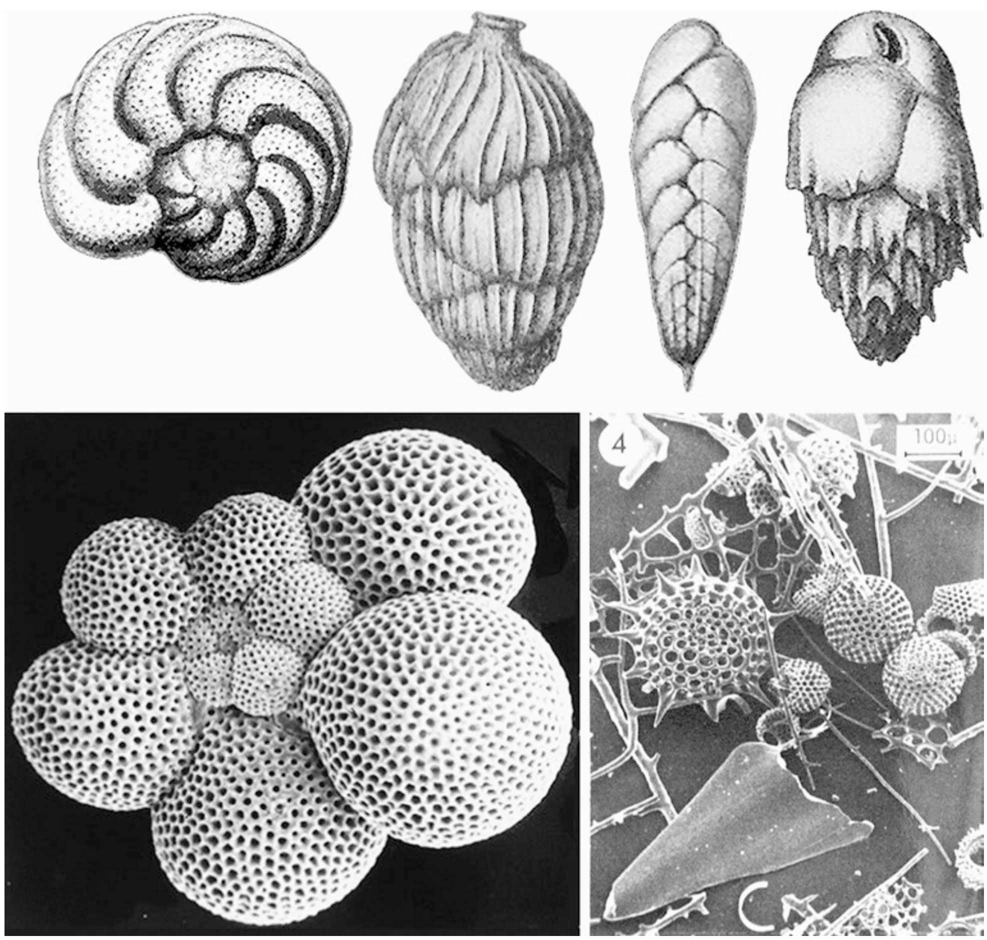

Fig. A9.2 Favorite objects of marine micropaleontology. *Top row*, benthic foraminifers from the Challenger Expedition; leftmost: dominant information carrier for stable isotopes; rightmost three: indicators of low-oxygen conditions. *Bottom row*, SEM photos (leftmost: planktonic foraminifer species, deep tow off California, S.I.O.; right: Neogene sediment sampled during DSDP Leg 14, SEM Kiel University) algae, sponges, corals, lamp shells, bryozoans, and a host of mollusks and arthropods, as well as echinoderms. In addition, there is a great variety of fishes. A similar assemblage (on the level of class and phylum) is seen at any shore, again with mollusks, arthropods, and echinoderms dominant, and with coral relatives and large algae much in evidence. At the shorelines of Southern California, we also have plenty of barnacles (a shell-bound sessile arthropod and in the same class as the crabs scurrying underfoot and the somewhat annoying kelp flies nearby). Mollusks, arthropods, and vertebrates, of course, dominate the biosphere not only at sea but on land as well. Echinoderms and cnidarians are found mainly in seawater, as is true for the phyla of many other creatures.

We might well ask whether the more restricted phyla are the more ancient ones or vice versa.

The question is difficult to tackle. With few exceptions, only fossil makers can be shown to have existed in the distant past. That is, for many organisms, we cannot even think of documenting the first appearance (not to mention the time of origin which is a near-impossible challenge even for organisms with hard parts). This is true not just for phyla but for any type of organism, actually, right down to the species level. (Neither do we know the time of last occurrence on the planet. What is known is what was found in what rock.) Another problem is that so many animals first appear in the Cambrian (hence the label "Phanerozoic," which is the time since the beginning of the Cambrian). The coincidence of first appearances suggests that environmental forces were at work in forcing geologic history at the time; the relevant changes are not known but much discussed. A third problem is that evolution changes the phyla. Just what we are talking about when discussing the origin of a phylum is not necessarily clear.

1.9.4 A Note on Large Marine Plants

Not all marine photosynthesizing algae are minute and reside in the marine plankton. Some live on the seafloor and some are large, like the calcareous algae, or very large, like certain species indicative of upwelling conditions ("kelp").

Details on marine organisms are available in any marine biology text and in many general biology texts.

Glossary

Marine geology, like many related disciplines of science, has experienced major changes within the twentieth century, and also in the twenty-first. One result is an index in flux, that is, one that reflects arbitrary preferences of authors regarding entries more strongly than desirable. [→ means "see entry for"]

Abyssal hill morphology: Sect. 2.3.6

Abyssal plains, deep seafloor usually covered by erosion products (→ Turbidites), commonly from extension of certain types of "continental rise." Sects. 1.3.3, 2.2.2, and 3.8; Figs. 2.2, 2.7, and 3.18.

Abyssal storms, pulsed deep-sea currents (transporting silt): Sect. 5.3.4.

Accelerated sinking (biological pump): Sects. 7.2.4 and 10.2.5; Figs. 7.6 and 10.7.

Acidification of seawater: Sects. 4.7.1., 13.1.1, 13.3.1, and 14.2.3.

Acoustic impedance (Eocene rocks, → chert), commonly associated with a change in density of the medium carrying sound: Sect. 10.4.3, Fig. 10.12.

Acropora, bushy calcareous coral genus, common in Caribbean, apparently selected for fast growth: Figs. 8.9 and 9.11; Sect.9.3.2.

Active margin (Pacific-type margin, collision margin): mountain-building region, factory of continental crust, → Trench, → Ring of Fire, Sect. 3.3, Figs.1.2, 2.9, 2.10, 2.11, 3.2, 3.9, and 3.10.

AFS (Auversian Facies Shift) also see → Auversian Facies Change, Eocene → CCD drop, → Plate stratigraphy: Sects. 12.2.2 and 12.2.3; Figs. 3, 1.7, 12.4, and 12.15.

Air gun records (sub-seafloor → echo sounding, also see → seismic profiling → side scan): Figs. 2.7, 3.6, and 3.14.

Albatross Expedition: Sects. 1.3.4 and 11.1.1; Fig. 11.6.

Alfred-Wegener-Institut, marine institute in Bremerhaven, Germany, focused on polar studies.

Algal mats, desiccation cracks, tidal flat deposits: Fig. 6.3.

Alvarez, L., pioneer, Calif. physicist, proposed bolide impact for end of Mesozoic (→ K-T): Sect. 1.1.1.

Alvarez, W., pioneer, Calif. geologist (son of L.), co-proposed → K-T impact: Sects. 1.1.1 and 1.1.5.

ALVIN, scientific submarine of Woods Hole (also see → hot vents): Sect. 1.2.1; Figs. 1.8 and 1.9.

Anaerobic (anoxic) and dysaerobic (low oxygen) conditions, humid sedimentation, and human impact: Sects. 8.2.3 and 9.5.2; Figs. 8.5, 8.6, 9.15, and 9.16.

Anaerobic conditions: result of complete → oxygen loss.

Anaerobic feedback in ocean productivity: Sect. 7.1.2.

Ancient sea-level fluctuations (reconstruction): Sects. 6.5 and 15.4.7; Figs. 6.10, 6.11, 6.12, 6.15, 11.9, 12.3, and 15.4.

Andrusov, N.I.: Sect. 1.3.6.

Anhydrite, $CaSO_4$, soft mineral, ("La Dame Blanche"): Sect. 14.4.3.

Anomalies (temperature, surface water): Figs. 7.10, 7.13, and 9.7; Sect. 9.1.5.

Anoxic conditions: Sects. 11.6.3, 13.2.2, 13.2.3, and 13.2.4; Figs. 13.8 and 13.9 (also see → Oceanic Anoxic Events).

Antarctic bottom water (AABW): Sect. 5.3.4; Figs. 5.15 and 7.12.

Antarctic deposition patterns in the Cenozoic: Sect. 12.2.3 (also see → Drake Passage opening).

Antarctic opal deposition: Sect. 10.4.2; Fig. 10.11 (also see → Siliceous ooze and mud).

Antarctic upwelling: Sect. 7.5.4; Fig. 7.12.

Anthropocene, Carbon Cycle Modified: Sect. 15.5.

Anthropocene, the Central Problem: Sect. 15.1.

Anti-estuarine circulation (margins): Sect. 5.3.5.

Aragonite vs. calcite: Sect. 4.6.2.

Archaea: Sect.1.2.2 (also see → methane and → Woese).

Arid and humid sedimentation: Preface, Fig. 2; Sect. 9.5.1; Figs.9.14, 9.15, 9.16, and 9.17.

Arrhenius, G.O.S, chemist on the *Albatross* discovered deep-sea carbonate cycles: Sect. 10.1; Fig. 11.5.

Asaro, F., co-proposer bolide impact at end of Cretaceous: Sect. 1.1.1.

Asymmetries in ocean circulation: Sect. 12.2.6; Fig.12.10 (also see → Basin-basin fractionation).

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
DOI 10.1007/978-3-319-51412-3

Atlantic-type margins (passive margins, trailing edge, sinking edge of continent): Sect. 3.2.

Atolls: Sect. 9.3.3.

Auversian Facies Shift (→ AFS, → Eocene-Oligocene silica sedimentation).

Back-arc spreading (spreading in collision margins): Sect. 3.3 (→ Active margin), Fig. 3.9.

Bacterial mat: Sect. 4.5.4 (also see → Bioturbation).

Bahamas (carbonate sedimentation): Sect. 4.7.2; Figs. 4.14 and 4.15. (Also see → Carbonate).

Baleen whale evolution (length of food chain argument, link to diatoms): Sect. 12.2.4.

Baltic Sea (humid circulation example): Sect. 9.5.2.

Baltic Sea (depth stratification of benthic foraminifers): Sect. 9.5.2; Fig. 9.16.

Barrier-type coast: Sect. 6.3.3; Fig. 6.9 (Galveston).

Barron, J.A., USGS, diatom expert (coauthor in reporting the mid-Miocene silica switch → Monterey).

Basin-basin exchange, (between Atlantic and Pacific, → Monterey): Sect. 10.2.3.

Basin-basin fractionation (corollaries of exchange circulation between Pacific and Atlantic): Sect. 10.2.3.

Basin-shelf fractionation (and AFS): Sect. 12.4.2.

Beach (morphology, California): Sect. 3.1.2; Fig. 3.3.

Bed load (vs. → suspension load): Sect. 5.1.3; Figs. 5.4 and 5.5 (Also see → Hjulström Diagram).

Bengal Fan (largest sediment wedge on the planet): Sect. 3.1.3.

Benthic boundary layer, water layer in contact with the seafloor: Sect. 5.3.4.

Benthic foraminifers (examples by E. Haeckel): Fig. 4.13.

Benthic organisms on and in the seafloor: for all major groups, benthic organisms are more diverse than planktonic ones, by far.

Benthic-organism-derived sediments: Sect. 4.6.2; Fig. 9.11.

Benthic life styles (infauna, epifauna, sessile, vagile, suspension feeder, etc.): Sect. 8.3.

Berger, A., Belgian astronomer and climatologist, recalculated Milankovitch input to high precision:

Sects. 1.1.1 and 11.1.2.

Berger, W.H., S.I.O./UCSD; junior coauthor of this textbook, worked on → CCD fluctuations, on ice-age history, and on selective preservation of deep-sea fossils (calcareous foraminifers).

Berggren, W.A., pioneer. Woods Hole geologist, worked on Cenozoic biostratigraphy: Preface, Fig. 5.

Bermuda coral (coral growth and solar cycles): Sect. 15.6.4.

Biodiversity (long time to recover from destruction): Sect. 15.4.4.

Biogenic sedimentary structures (trace fossils, bioturbation): Figs. 8.6, 8.7, 8.13, 8.14, 8.15, and 8.16.

Biogenous sediment: Sects. 4.4, 4.6, and 10.2.2; Figs. 10.5 (plankton), 10.2, 10.11, 4.7, 4.12, 4.13, 4.14, 4.15, 7.9, 8.1, 8.2, 13.3, and 13.4.

Biological pump: Sect. 7.2.4; Figs. 7.5, and 7.6.

Bioturbation (disturbance): Sects. 8.4.1 and 8.5.

Bioturbation (and the low abundance of thin layers in oxygenated sediments): Sect. 8.5.2.

Black sediment (modern seafloor): Sect. 13.2.2; Fig. 13.7.

Black shale, Cretaceous: Sect. 13.2.1; Fig. 13.6. (Also see → “Oceanic Anoxic Events”).

Bleaching of coral reefs: Sect. 9.3.5.

“Blue” vs. “green” ocean: Figs. 7.4 and 7.7.

Bohrmann, G., marine geochemist, Bemen University: Sect. 1.2.2; Fig. 1.10.

Bolin, B., Swedish meteorologist and biogeochemist, founder of the IPCC: Sect. 15.5.2.

Bottom shear stress: Sect. 5.1.3; Fig. 5.4.

Bottom-water circulation (global ocean): Sects. 5.3.3 and 5.3.4; Figs.5.15 and 5.16.

Bottom-water formation on Tethys shelf: Sects. 5.3.3 and 13.2.6.

Bottom-water production (NADW): Sect. 5.3.3; Fig. 5.15.

Bouma sequence = graded layer (turbidite): Sect. 3.7.

Bourcart, J., French marine geologist: Sect. 1.3.6.

Box-core (artificial outcrop of deep-sea ooze): Fig. 10.8.

Brittle stars: abundant in some muddy places offshore: Fig. 8.11.

Broecker, W.S., Lamont geologist and ocean chemist, involved in a great number of concepts and activities of interest to marine geology (radiocarbon dating, radioisotope dating of last warm time, terminations, differential dissolution, carbon isotopes, conveyor-belt circulation, Dansgaard-Oeschger Oscillations, Heinrich Events, and others).

BSR (“bottom-simulating reflector”): Figs. 3.10 and 4.5; (also see → methane clathrate).

Bullard, E.C.: Sect. 1.3.8; Fig. 1.14.

Burning coal and oil (recovery by carbonate dissolution): Sect. 10.3.4.

Burrows and trails: Sects. 8.4 and 8.5.

Burrows in “Red Clay”: Fig. 8.13 (X-ray shadow graph).

Calcareous ooze: Sect. 10.3; Figs. 4.7, 10.2, 10.3, 10.4, 10.5, 10.6, 10.8, and 10.9.

Calcareous shell (resource): Sect. 14.3.2.

Calcium sulfate deposit (resource, vent chimney): Sect. 14.4.3.

Calibration (distribution patterns vs. environmental parameters}: Sect. 9.2.1.

Californian margin (active, uplift, Big Sur): Fig. 3.2.

Carbon dioxide (effects of rise, basic considerations): Sect. 15.4.3.

Carbon dioxide (uptake by the sea); Revelle and Suess, 1957: Sect. 15.4.1.

Carbon dioxide (warming effect; discussions): Sect. 15.4.1; Fig. 15.2.

Carbon dioxide consumption (during dissolution of carbonates): Sect. 4.7.1.

Carbon dioxide release (during precipitation of carbonates): Sect. 4.7.1.

Carbon isotope signal, Cretaceous OAEs: Sect.13.2.4; Figs. 13.8 and 13.9.

Carbon isotopes, indicators of carbon cycle: Sect. 9.4.2.

Carbon system state and changes: Sect. 15.5.1.

Carbon transfer, sea surface to seafloor, schematic, estimated (blue vs. green ocean contrast): Fig. 7.7.

Carbonate crash (short-lived CCD rise, end of middle Miocene): Sect. 12.4.2; Fig. 12.15.

Carbonate cycles: Sect. 11.4.2; Fig. 11.5.

Carbonate rate of production: Sect. 8.3.3; Figs. 8.8, 8.9, and 8.10.

Carbonate saturation and precipitation (focus on geochemical definition): Sect. 4.7.1.

Carbonate (types of calcareous matter): Sects. 4.6, 4.7, and 4.8; Figs. 4.7, 4.11, 4.12, 4.13, 4.14, and 4.15.

Carbonate deposits with no recognizable biological structure: Sect. 4.7.

CCD and "Red Clay": Sects. 10.1, 10.2, and 10.3; Fig. 10.2.

CCD drop at end of Eocene (→ AFS, → Drake Passage opening): Sect. 12.2.3.

CCD, Cenozoic fluctuations: Sects. 12.4.1 and 12.4.2; Fig. 12.15.

CCD map: Fig. 10.10.

Cementation during early diagenesis: Sect. 4.3.2.

Cenozoic history from deep-ocean drilling: Sects. 12.1 and 12.2; Figs. 12.3 and 12.5.

Cenozoic stepwise cooling: Fig. 1.7.

Cenozoic time scale: Preface, Fig. 5.

Central gyres: Sect. 5.3.2; Fig.5.14. (Also see "Gulf Stream").

Challenger Expedition (1872–1876): Sects. 1.3.2 and 10.1; Figs. 10.1 and 10.2.

Charles Darwin (1809–1882), geologist, naturalist, and polymath known in marine geology for his amazing contributions to paleontology and biology, for his speculations about atolls, and his observations on dust at sea (among other things).

Chemistry of seawater (geochemistry of sediments): Sect. 4.3.

Chert (microcrystalline quartz): Sects. 4.6.2 and 10.4.3; Fig. 10.13.

Classification of sediments on the seafloor: Box 4.1.

Clathrate (→ methane ice): Sect. 1.2.2; Figs. 1.10 and 4.5.

Clay (clay size): Sects. 4.5.1 and 4.5.4.

Clay (mineral composition): Sect. 10.5.2; Fig. 10.14; Appendix A4, A5.

Cliff erosion: Fig. 5.9; Sects. 5.2.1, 5.2.2, and 5.2.3.

CLIMAP reconstruction, last glacial maximum (LGM): Sect. 9.2.2; Fig. 9.9.

Climate change (ongoing) and the seafloor: Sect. 15.4.

Climate change, short-term marine record: Sect. 15.6 (also see → Santa Barbara).

Climate indicators other than coral reefs, physical and biological indicators: Sect. 9.4.

Climate system (complexity): Sect. 15.4.2; Fig. 15.2.

Climate zonation (marine sediments and fossils): Sects. 9.1 and 9.1.2; Figs. 9.2, 9.3, 9.4, 9.5, and 9.6.

Climatic change vs. the economy (prevailing argument): Sect. 15.4.5.

Climatic clues from restricted seas: arid and humid zones on shelves: Sect. 9.5; Figs. 9.14 and 9.15. Climatic transgression (pattern of temperature): Sect. 9.2.2.

Coastal erosion: Figs. 5.8 and 5.9.

Coastal morphology and sea level: Sect. 6.3; Figs. 6.2, 6.3, 6.4, 6.5, 6.6, 6.7, 6.8, and 6.9.

Coastal upwelling and high production off California: Sect. 7.2.1; Figs. 7.2 and 7.10.

Coccolithophores (temperature reconstruction): Sect. 9.2.1.

Coccoliths, parts of → coccolithophores, the most abundant fossils on the planet (also called nannofossils because of their minute size): Sect. 4.4; Figs. 4.7 and 4.11; Table 10.1.

Cold seeps: Sect. 1.2.2.

Collision margin: see → Active margin.

Comet mark: Fig. 5.13.

Continental drift: Sect. 1.3.7; Figs. 1.13, 1.14, and 13.5 (Also see → Wegener).

Continental margins, general features: Sect. 3.1. (Also see → Cont. rise and → Cont. slope, → Shelf)

Continental rise and slope, coalescing fan deposits off continents. Differ by rates of sedimentation, steepness (both have gentle slopes) and type of underlying crust (continental or oceanic): Sect. 3.5.

Continental rise: Sect. 3.5; Figs. 3.13, 3.14, and 3.15.

Continental slope: Sects. 3.5 and 3.1.1; Figs. 3.13, 3.14, and 3.15.

Continental uplift: Sect. 2.1.2; Figs. 2.1 and 3.2 (Western Big Sur vs. East Coast).

Contour currents: Sect. 5.3.4.

Contourites: sediment layers formed by → contour currents.

Conversions for temperature, length, volume, mass: Appendix A1.

Conveyor belt circulation: Sects. 5.3.3 and 9.1.5; Fig. 5.15. (Also see → Bottom water and → Silica Switch).

Cooling and mountain building: Sects. 2.1.2 and 12.2.2.

Cooling trends (evidence from oxygen isotopes): Fig. 12.3; Sect. 12.1.3.

Cooling in the Cenozoic: Preface, Fig. 3; Figs. 12.1, 12.3, 12.4, 12.9, and 12.13; Sect. 12.1.1.

Coquina deposits: Sect. 8.3.5.

Coral reef stress: Sect. 9.3.5.

Coral reef, associated organisms: Fig. 9.11; Sects. 2.2.2 and 9.3.1 (Great Barrier Reef: Sects. 2.2.2 and 9.3.4).

Coral reefs (general distribution): Sect. 9.3.1; Fig. 9.10.

Coral reefs, markers of tropical climate: the response of reefs to a warm climate is not obvious because a warm ocean differs from a cold one not just in temperature: Sect. 9.3.

Coral symbiosis: Sects. 2.2.2 and 9.3.1.

Coring (the first collection of long cores was from the *Albatross* Expedition, 1940s): Sect. 10.1.

Coring (the first cores from the seafloor were from the *Meteor* Expedition, 1920s): Sect. 10.1.

Coring: widespread use of piston cores taken by Lamont's RS *Vema* (see → *Vema*).

Coring and drilling: drilling vessels *Glomar Challenger* and *JOIDES Resolution*. (See → Drilling, deep sea).

Coriolis force (discussed with upwelling): Sects. 7.5.1 and 7.5.3; Figs. 7.10 and 7.11.

Corliss, J., pioneer, S.I.O.-trained marine geologist, Oregon, discovered the first → hot vent: Sect. 1.2.1.

Cox, A., pioneer, USGS geologist who worked on → magnetic reversals: Sect. 2.3.3.

Crest (MOR): Sect. 2.3.5; Figs. 2.4 and 2.5.

Cretaceous carbon energy (background): Sect. 13.1.1.

Cretaceous carbonate reefs (why they matter): Sect. 13.3.1.

Cretaceous environments (evidence from land and from deep-ocean drilling): Sects. 13.1 and 13.2.

Cretaceous leitmotif: Sect. 13.2.

Cretaceous marine fossils (background): Sect. 13.1.2.

Cretaceous marine rocks (background): Sect. 13.1.1.

Cretaceous Milankovitch cycles: Fig. 13.12; Sect. 13.2.6.

Cretaceous rudist reefs: Sect. 13.3.2; Fig. 13.13.

Croll vs. Milankovitch (emphasis on ice buildup vs. melting): Sect. 11.1.4.

Croll, J., Scottish pioneer geologist and climatologist of the nineteenth century (a brilliant but failed) Proposed orbital theory of ice ages: Sect. 11.4.1. Also see Sect. 9.1.5 (Gulf Stream heat transport).

Crust (continental and oceanic): Sect. 3.1.1; Fig. 3.1.

Cryptic predation: Sect. 8.5.1.

Currents and sediments (the faster the current, the larger the particles moved): Sects. 5.1, 5.2, and 5.3.

Cyanobacteria (blue-green algae, stromatolites): Sect. 6.2.2.

Cycles of the Pleistocene: Sect. 11.4.

Dalrymple, G.B., pioneer, USGS geologist who worked on magnetic reversals: Sect. 2.3.3.

Daly, R.A., pioneer, Harvard geologist, examined reef growth and ice-age sea level: Sects. 6.4.3 and 1.3.3.

Dansgaard-Qeschger oscillations (found in ice and in lake sediments) are named after physicists in Copenhagen and in Berne, who discovered long-term (millennia-long) regular climate variations in the Pleistocene: Sect. 11.6.2 (Also see → Heinrich Events).

Dead lakes on the Cretaceous seafloor: Sect. 13.2.7.

Dead zones (→ oxygen loss): Sect. 7.2.4, final paragraph.

Décollement, major fault system in collision systems, nearly horizontal; commonly greased by methane-bearing fluids: Fig. 3.10 (Also see → Active margin).

Deep biosphere; prokaryotic microbes are everywhere, even deep within sediments: Fig. 1.11; Sect. 8.6.

Deep-ocean drilling and the ice-age target: Sect. 11.5.

Deep-sea fans and abyssal plains: fans are found off valleys carved into → shelf and slope; abyssal plains are, in essence, extensions of rises and fans: Sect. 3.8; Figs. 3.13, 3.14, 3.15, 3.16, 3.17, and 3.18 (Also see → Turbidites).

Deep seafloor morphology: Sect. 2.3.6; Figs. 2.7 and 2.8.

Deep-sea sediment cycles (ice ages): Sects. 1.3.4 and 11.4.

Deep-sea sediments – patterns and processes. Three major facies (types of rock or sediment) exist – calcareous ooze and "Red Clay" (separated by the → CCD) and the mud (commonly siliceous) of the coastal ocean (mainly on cont. slope): Sects. 10.1 and 10.2.

Deglaciation: sea-level rise and coastal morphology: Sects. 11.6; Figs. 6.5, 6.6, and 6.7.

Delta deposit structure: Fig. 6.7.

Deposit feeders, interference with suspension feeders: Sect. 8.3.5, fourth paragraph.

Depth of the sea: Sect. 2.1.

Desert at gyre center: Sect. 7.2.2; Figs. 7.3 and 7.4.

Desiccation cracks, algal mats, tidal flat deposits: Sect. 6.2.1; Fig. 6.3.

Diagenesis in coral reefs and other carbonate deposits: Sects. 9.3.2, 4.3.2, and 4.7.3.

Diagenesis, chemical modification of deposits: Sects. 4.3.2, 7.6, and 4.7.3; Fig. 7.15.

Diatoms, shelled siliceous plankton – weakly silicified in some species, strongly in other instances.

Diatoms (marine): Sects. 4.4 and 7.4.2; Figs. 4.7 and 7.9.

Diatoms (size and length of food chain): Sect. 7.4.2.

Diatom distribution pattern: Fig.7.4.

Diatom crash, Santa Barbara Basin: Sect. 9.1.7; Fig. 9.8.

Diester-Haass, L., Kiel-trained marine geologist (then at Saarbruecken) discovered the Pleistocene silica paradox (more production, less silica during glacials, off Namibia): Sects. 11.4.4 and 7.4.2.

Dietz, R.S., pioneer, US Navy, San Diego, proposed seafloor spreading: Sect. 1.3.9.

Discussing the future (a task for marine geology in the modern world) Sect. 15.7.

Dissolution of calcareous shells on the seafloor: Sect. 10.3.2; Figs. 10.9 and 10.10.

Dissolution of siliceous fossils: Sects. 10.4.2 (first paragraph) and 10.4.3; Fig. 7.15.

Diversity of marine organisms (benthos far more diverse than plankton): Sect. 8.1; Fig. 4.13.

Diversity (benthic foraminifers, different ways of building a shell): Sect. 8.1.2; Fig. 8.10.

Diversity of calcareous shells (benthic mollusks and other macrofauna): Sect. 8.1.2.

Diversity (link to differences in habitat): Sect. 8.2.

Diversity (link to food supply, may be counterintuitive): Sect. 8.2.1, first and last paragraph.

Diversity of microplankton (Cenozoic changes): Fig. 12.2.

Doell, R.R., pioneer, USGS geologist who worked on magnetic reversals, Sect. 2.3.3.

Dolomite, a product of diagenesis: Sect. 4.7.3.

Douglas, R., geologist, USC Los Angeles, codiscovered stepped cooling: Sect. 1.1.1.

Drake Passage opening: Sect. 12.2.3; Fig. 12.9.

Drilling, deep sea: Sects. 1.1.4, 1.3.5, and 11.5; Figs. 3, 1.6, 1.7, 6.16, 11.10, 12.2, 12.3, 12.5, 12.9, 12.10, 12.11, 12.13, 12.15, 13.4, 13.6, 13.8, 13.9, 13.11, 13.12, and 13.15.

Earthquakes (mantle motion): Sect. 2.4.3; Fig. 2.12. Also see → MOR (shallow quakes).

Earthquakes in continental margins: Sect. 3.5; Fig. 3.14.

Eccentricity, frequency of variation: Sect. 1.1.3, second paragraph; Fig. 11.10.

Echo-sounding records: Sect. 2.3.1; Figs. 2.2, 2.7, 3.6, 3.10, 3.14, 10.6, 10.12 (side scan), 4.6, 5.13, and 5.16.

EEZ (exclusive economic zone): Sect. 14.1.2.

Elevations: Sect. 2.1.1 (first paragraph); Fig. 2.1.

Emiliani, C, (isotope record principle): Fig. 1.5; Sect. 11.4.5.

Emiliani, C. (isotopic stages in the Pleistocene): Figs. 1.4 and 11.10.

Emiliani, C., pioneer paleoceanographer, introduced oxygen isotopes to reading the record of ocean history: Sects. 11.1.1 and 1.1.3; Fig. 1.5.

End of the Mesozoic: Sect. 13.4; Figs. 13.14 and13.15.

Endogenic and exogenic processes: Sect. 2.2; Figs. 1.2, 1.14, and 4.9.

Endogenic forcing (tectonic forcing largely controlled by processes in Earth's mantle): Sect. 2.2.1.

"Engineering Fixes": Sect. 15.5.4.

ENSO (El Niño-Southern Oscillation): Sect. 9.1.6 (Also see → Warm Pool).

Environments at sea (main depth habitat labels): Sect. 8.2.1; Fig. 8.3.

Eocene-Oligocene silica drop: Sect. 10.4.3; Fig. 10.12. Also see "AFS" and "chert."

Epsom salt, magnesium sulfate, is made of the second most abundant metal ion and the second most abundant base (sulfate) in seawater; yet, seawater evaporation is not the foremost source of the salt.

Epstein, S., pioneer Canadian-US isotope geochemist and paleontologist: Sects. 8.2.2; 11.4.5.

Equatorial upwelling (high production and silica deposition): Sects. 7.2.3 and 7.5.3; Fig. 7.11.

Ericson, D.B., twentieth-century Lamont geologist: last use (1960s) of Penck scheme of ice ages by a prominent marine geologist.

Erosion, coastal: Figs. 5.8, 5.9, 5.10, and 5.11.

Erosion (mechanical vs. chemical): Sect. 4.2.1.

Erratics, boulders important in the discovery of ice ages, rare on the global seafloor: Sect. 4.5.1 (last paragraph).

Error discussion (estimating past temperatures): Sect. 9.2.1.

Estuarine and anti-estuarine circulation: Sects. 5.3.5 and 9.5.1; Preface, Fig. 2; Figs. 9.14, and 9.15.

Eukaryotes: organisms with one or more cells with a well-defined nucleus. Appendix A9.1, A9.2, and A9.3. Others are "prokaryotes," i.e., bacteria and archaea.

Euler's theorem: Sect. 1.4.2 (Also see → Bullard and → Morgan).

Eustatic sea-level change: Sect. 6.1.2 (first paragraph).

Evaporites (marine): Sect. 4.8.1.

Events (forced suspension in sediment transport): Sect. 5.1.4.

Evidence for an impact at the K-T boundary: Sect. 13.4.2; Fig. 13.15.

Ewing, M., pioneer US geophysicist: Sects. 1.3.8 and 1.3.3; also see → Lamont (Ewing was the founder).

Exogenic forcing: Forcing of processes on Earth that is ascribed to solar energy (including erosion, but which obviously depends on tectonic uplift as well; i.e., on tectonic forcing): Sect. 2.2.

Facies, type of rock formation, one of any number of types at several levels of classification.

Facies pattern as function of depth of deposition in the sea: Sect. 6.1.1.

Fairbanks, R.G., pioneer Lamont geologist worked on → deglaciation (Fig. 6.5) and Cenozoic evolution.

Fecal pellet transport (→ accelerated settling in the water column): Sects. 10.2.5 and 4.5.4; Figs. 7.6 and 10.7.

Feedback, negative, stabilizing: Sect. 1.1.1. (Also see → Le Chatelier, → Vernadsky.)

Feedback, positive, run-away: Sect. 1.1.1. (Also see → Milankovitch.)

Felspars = certain common Ca−, Na−, and K-silicates (Ca− in basaltic rock; Na− and K− in granitic rocks (e.g., continental crust).

Ferromanganese nodules: Sect. 14.4; Figs. 14.10 and 14.11 (table).

Feynman, R., famous Calif. physicist, erstwhile CalTech and Berkeley: Preface.

Fiddler crab (benthic, California wetland): Fig. 8.6.

Fischer, A., pioneer Alpine geologist (Wisconsin-trained), worked on ancient marine shelf sediments.

Flank sinking, MOR: Fig. 2.6.

Flooded shelves vs. exposed shelves (albedo): Sect. 6.1.2 (second paragraph).

Flysch = Alpine turbidite sequences (local Swiss name): Sect. 3.7.

Food chain (or food web): Sect. 7.4.1.

Food chain: the order of being eaten, top consumer at the top. Also web and pyramid. Since there is much loss from one eating level (→ trophic level) to the next, a long food chain (starting with extremely small organisms) produces less for top consumers than a short one: Sect. 7.4.

Fool's gold: golden-yellow iron sulfide. Can conceivably be confused with real gold.

Foraminiferal ooze, calcareous ooze with planktonic foraminifers: Fig. 10.1; Table 10.1.

Fossil information (biased recording): Sect. 8.2.2 (fourth paragraph).

Fourier analysis (statistics): Sects. 15.6.3 and 15.6.2.

Fracture zones: Sects. 1.4.2 and 2.3.4.

Framework and fill (coral reef): Sect. 9.3.2.

Fusulinids, grain-shaped Paleozoic benthic foraminifers, presumably bore light-processing symbionts.

Future problems: Sect. 15.7.

Gabbro, basaltic dark rock above the → "Moho," made of large crystals: Sect. 2.3.4; Fig. 2.5.

Gaia hypothesis, so named by the British engineer and chemist J. Lovelock, negative feedback: Sect. 1.1.1 (last paragraph).

Galveston Bay: Fig. 6.9.

Garbage in the deep sea: Sect. 15.3.4.

General cooling and productivity: Sect. 7.1.3.

Geochemical climate indicators: Sect. 9.4.1.

Geochemical statistics (element distributions): A7 (table).

Geologic time: not available before the end of the nineteenth century: Sect. 4.9.

Glacial sea level (also see → shelf break): Fig. 6.5 (range of variation during last deglaciation).

Glauconite: greenish iron-bearing silicate mineral, common in certain continental margins: Sect. 4.8.3.

Global conveyor (deep-basin exchange circulation): Sects. 5.3.3 and 9.1.5.

Global warming, methane release (concern, positive feedback): Sect. 1.2.2.

GLOMAR Challenger, first ship employed for scientific drilling, 1968: Sects. 1.1.4; Figs. 1.6 and 12.5.

Graded layers: → turbidites; Sect. 3.7.

Grain size and sediment transport: Sects. 4.5.1 (last paragraph), 5.1.2, A5; Box 4.1; Figs. 5.4 and 14.9.

Great Barrier Reef: Sects. 9.3.4 (specific) and 4.6.2 (general).

Great Man Drowning: Storm flooding A.D.1362, North Sea: Sect. 6.2.1 (fourth paragraph).

Greenhouse effect: Sects. 9.1.1 (last paragraph) and 15.7.

Greenland ice today: Fig. 11.4 (air photo, 10 km up).

Gubbio: Sect. 1.1.5.

Gulf Stream: Sect. 5.3.2; Fig. 5.14.

Guyot = flat-topped seamount (so named by H. Hess, whose office was in Guyot Hall, Princeton University): Sects. 2.6.1 and 2.6.2.

Halite = kitchen salt mineral, NaCl.

Hawaiian island chain: Sect. 2.6.2; Figs. 2.14 and 2.15.

Heat anomaly (ocean): Sect. 5.3.3.

Heat piracy (ocean, N Atlantic): Sects. 5.3.3 and 9.1.6; Fig. 9.7.

Heat transfer, link to → NADW production: Sect. 9.1.5.

Heavy metals on the deep seafloor: Sect. 14.4.

Heavy minerals: A4.3.

Heezen, B.C., pioneer marine geologist (Lamont), morphology of the seafloor, seafloor spreading: Sects. 1.1.1, 1.3.3, 1.3.6, and 1.3.9; Fig. 1.12.

Heinrich events (geologically short-lived cooling events somewhat similar to the Younger Dryas cold spell, seen in deep-sea cores, named after the geologist who is credited with seeing them first and with dating the relevant layers): Sect. 11.6.2.

Heirtzler, J.R., Lamont geophysicist who documented the symmetry of magnetic anomalies across the mid-ocean ridge south of Iceland: Sect. 2.3.3; Fig. 2.4.

Hemipelagic mud ("hemi-"= Greek for "semi-"; "pelagic" = pertaining to the sea well away from land: Sect. 10.6.

Hertogen, J., Belgian physicist (Gent University); proposed a bolide impact for the end of the Mesozoic, offering evidence from exposed marine sediments in Spain.

Hess, H.H., geologist (Princeton) and US Navy officer, discovered numerous flat-topped seamounts in the Pacific (he named them → "guyots"), proposed seafloor spreading in 1960 (unpublished ms), subsequently proposed SFS in a publication (1962): Sect. 1.3.9.

Hessler, R., marine biologist, helped explore hot vents, Woods Hole, and S.I.O.: Sect. 1.2.1; Fig. 1.20.

Hiatus development: Sect. 10.2.4.

Hilgen, F.J., Dutch geologist (Utrecht), extended orbital stratigraphy back into the Miocene: Sect. 11.6.3.

Hjulström diagram: the standard textbook link between transport of particles of different sizes and current speed. Not entirely clear: Sects. 5.1, 5.2, and 5.3; Fig. 5.4.

"Hockey stick" (description of the graph containing an unusual temperature excursion upward near the end of the last century.): Sect. 15.4.1.

Hot spots, island chains, seamounts: Sect. 2.6.

Hot vents: Sect. 1.2.1; Figs. 1.8 and 1.9.

How fast did sea level rise? Illustration of an ice-age data analysis based on oxygen isotope ratios in foraminifer shells: Sects. 6.7 and 15.4.7; Fig. 6.16.

Human time scale vs. geologic time scale: Sects. 15.4.5 and 11.2.2.

Humid and arid deposits in shelf basins: Sects. 9.5.1 and 5.3.5; Figs. 9.14 and 9.15.

Hydrate (methane): Sect. 1.2.2; Fig. 1.11.

Hydrocarbon pollution: Sect. 15.3.5.

Hydrogen sulfide produced in early diagenesis: Sect. 4.3.2.

Hydrogenous sediment, made in in situ, in water or below the seafloor (not including → biogenous): Sects. 4.4 and 4.8; Fig. A4.2; Box 4.1.

Hydrological cycle: Sect. 4.5; Fig. 4.9.

Ice-age fluctuations: Sects. 6.4.2 and 11.4.6; Figs. 6.10, 11.9, 11.10, 11.11, and 11.12.

Ice age – a confusing term used in more than one way: Sect. 11.2.2 – cycles (multiple) and Sect. 11.4.1 (background).

Ice-age carbon cycles: Sect. 11.4.7; Figs. 11.6, 11.7, and 11.8.

Ice-age carbonate productivity cycles: Sects. 11.4.2 and 11.4.3.

Ice-age lessons re-climate change: Sects. 11.2.1, 11.2.2, and 11.2.3.

Ice-age lessons: Lessons other than broad principles are difficult to extract from the geologic record, largely owing to timescale problems. The melting of large ice masses is of considerable interest, however. Crucially important clues to climate-relevant processes can emerge from Pleistocene sea-level studies: Sects. 11.2 and 11.4; Figs. 11.10, 6.16, and 15.4.

Ice-age oceans: upper waters were colder in an ice age, and currents were altered: Sect. 11.1.

Ice-age silica productivity cycles (Walvis Paradox): Sect. 11.4.4.

Ice-ages onset: Sect. 12.3.1; Fig. 12.13.

Ice input (→ IRD, ice-rafted debris) Sect. 4.2.2; Figs. 4.2 and 9.13.

Ice-age studies by deep-ocean drilling: Sect. 11.5.1.

Ice-driven sea-level fluctuations: Only 20,000 years ago, ice forming on land had bound some 200 m worth of seawater; much shelf area was exposed or had but a thin water cover. Even now (during an interglacial time period) there is still about 80 m worth of sea-level rise bound up in ice. Notable sea-level fluctuations have been recorded for the last several million years using oxygen isotopes. Suggested corresponding variations in ice mass: Sect. 6.4; all of Chap. 11.

Iceland low: Sect. 9.1.6.

Igneous rocks: crystalline rocks made inside the Earth: Sect. A6.1.

Imbrie, J., Brown Univ. geologist, pioneer of a system approach to marine geology and to ice-age temperature reconstruction, Preface, Fig. 4, Sect. 9.2.1 (leader, CLIMAP group). Also see Chap. 11.

Impact (end of Cretaceous, Cretaceous-Tertiary boundary): Sects. 1.1.1 (first paragraph) and 13.4; Figs. 13.14 and 13.15.

Imprinting of zonal (temperature) pattern on planktonic foraminifers: Sect. 9.1.3; Figs. 9.3 and 9.5.

Inman, D.L. (1921–2016), S.I.O., marine geologist who discovered beach sand transport cells in Southern California, in the 1950s. Cells commonly end in a → submarine canyon: Figs. 3.16 and 5.1.

Interstitial water, chemical modification of deposits: Sect. 4.3.2.

Intertidal zone: Sect. 6.2.1; Figs. 6.3, 8.6, and 8.7.

IRD (ice-rafted debris), indicator of polar conditions: Sect. 9.4.3; Fig. 9.13.

Iron sulfide: Sect. 4.3.2.

Iron, abundance linked to that of oxygen: Sect. 4.8.3.

Island chain: a label applied to islands in a row. Within the central ocean islands are of volcanic origin, without exception. The iconic example in the Pacific is Hawaii: Sects. 2.6.1 and 2.6.2; Figs. 2.14 and 2.15.

Island chains, seamounts, hot spots: Sect. 2.6.

Isostasy (vertical motion of crust from loading or unloading): Sect. 3.4.

Jannasch, H., Woods Hole biologist, studied hot vent organisms and related forms: Sect. 1.2.1.

Japan Trench: trench system off Japan, parts of which serve to illustrate important processes related to subduction (volcanism, earthquakes, mélange formation, submarine landslides, tectonic erosion, décollement, expulsion of fluids, and others): Sect. 3.3.

Jenkyns, H.C., British geologist, Oxford, studied "ocean anaerobic events:" Sect. 13.2.3; Fig. 13.8.

JOIDES Resolution: Sect. 1.1.4; Fig. 1.6.

Karig, D.E., Prof. em. Cornell, erstwhile grad student at S.I.O., UCSD, discovered → back-arc spreading.

Kastner, M., Israeli-US geochemist and general marine geologist, UCSD, studied reactions at hot vents, changing productivity of the oceans, methane formation, and other fundamental processes recorded on the seafloor: Sects. 1.2.1 and 1.2.2; Figs. 1.8 and 1.10.

Keller, G., Swiss-US marine geologist, Princeton, codiscovered the mid-Miocene → silica switch in deep-sea sediments. Also, she questions the end-of-Cretaceous → impact scenario as the final word on the matter of ending the Mesozoic.

Kennett, J.P., pioneer of Cenozoic ocean history, N-Z trained (Wellington), Prof. em. UCSB, studied Cenozoic cooling, deglaciation, methane, and the role of impact in the origin of the → Younger Dryas: Sects. 1.1.1 and 12.1.3.

Klenova, M.B.: Russian pioneer marine geologist, wrote one of the earliest textbooks in the field of marine geology (1948), documenting abundant Russian contributions: Sect. 1.3.6.

Köppen, W., German pioneer climatologist born in St. Petersburg, used vegetation to classify climate types. Earliest supporter of Milankovitch Theory.

K-T event (bolide impact, massive extinctions): Sect. 1.1.5.

K-T event (end of Mesozoic): Sect. 13.4.

K-T event boundary Denmark (Stevn's Clint): Fig. 13.14.

K-T event boundary Gubbio: a section described by H. Luterbacher and I. Premoli-Silva, (marine pelagic record) and subsequently analyzed by → L. Alvarez and associates.

Kuenen, Ph.H., pioneer marine geologist, studied → turbidites: Sects. 1.3.3 and 3.7.

Kullenberg, B., Swedish oceanographer on the *Albatross* Expedition invented the piston corer: Sect. 11.1.1.

Lagoon: Sects. 6.3.3 and 3.1.1; Figs. 3.2, 3.4, and 6.8.

Lamont Geological Observatory; later Lamont-Doherty Earth Observatory: Marine geological institute that played a major role in the evolution of plate tectonics. Founded 1949 by the US geophysicist Maurice → Ewing, erstwhile professor at Columbia University in New York. Many marine geologists refer to the institute as "Lamont."

Lancelot, Y. (1938–2016), French marine geologist, studied deep-sea corollaries of plate tectonics, anoxic deep-sea sedimentation, and other fundamental issues of deep-sea geology: Sect. 12.4.1; Fig. 13.6.

Laplace, P.-S., famous French mathematician and astronomer: Sect. 1.1.3.

Larson, R., Scripps-trained marine geophysicist, URI, discovered various parallelisms in endogenic processes, sedimentation, and environment: Fig. 13.11.

Last Glacial Maximum (LGM): Sect. 11.3; Fig. 11.3.

Le Chatelier, H.L., French chemist and engineer, gave his name to negative feedback in chemistry, asserting that reactions minimize any change: Sect. 1.1.1.

Lebensspuren (trace fossils): Fig. 8.14.

Leitmotif of the Cenozoic (cooling): Sect. 12.1.

Le Pichon, X., French marine geophycisist who determined plate sizes and motions, chiefly at Lamont (Columbia University, New York).

Levin, L., S.I.O.-trained biogeochemist, studies the biology and chemistry of cold seeps and of the dysaerobic and anaerobic seafloor: Sects.1.2.1 and 1.2.2.

Limiting nutrients (phosphate, nitrate, silicate, iron, and various other compounds and trace elements): nutrients whose abundance is sufficiently restricted to limit growth: Sect. 7.1.1.

Lister, C.R.B., marine geophysicist, (U. Washington), early proposed hydrothermal processes at spreading centers: Sect. 1.2.1; Fig. 1.9.

Lithogenous sediment, continent-derived or volcanogenic particles: Sects. 4.4 and 4.5; Figs. 4.8, 4.9, and 4.10.

Living on the seafloor: requirements differ for mobile and sessile organisms and for those that live exposed to the environment or that make burrows in sediments or in rocks: Sects. 8.2 and 8.3.

Lonsdale, P., marine geologist, studied morphology of deep seafloor, using deep-towed, side-ways looking echo-sounding equipment designed by physicist and engineer Fred N. Spiess (erstwhile short-term multiple director of S.I.O.): Sect. 1.2.1; Fig. 5.16.

Lovelock, J., British environmental chemist, creator of "Gaia hypothesis" and "daisy world;" see → feedback, negative.

Lugworm (benthic, wadden): Sect. 8.3.2; Fig. 8.7.

Luterbacher, H., German-Swiss geologist (paleontology), co-described (with Isabella Premoli-Silva) in some detail the pelagic K-T section near Gubbio, Italy: Sects. 1.1.1 and 1.1.5; Fig. 13.15.

Lyell, C., uniformitarianism: Sect. 1.1.1 (third paragraph).

Lysocline (differential dissolution, selective preservation): Sects. 10.3.2 and 10.3.3; Fig. 10.10.

Magma chambers (MOR): Sect. 2.3.4; Fig.2.6.

Magnesium, abundant ionic metal in seawater, homologous to calcium: Sect. 4.1.2.

Magnetic anomalies: widely greeted as proof of seafloor spreading; notably the symmetry of anomalies across certain spreading ridges. The ultimate cause of the pattern of magnetic anomalies on the seafloor are reversals of the Earth's magnetic field: Sect. 2.3.3; Figs. A3.2, 13.11, and 2.4.

Magnetic reversal scale, Neogene: Sect. A3; Figs. A3.2, 7.14 (back to Gilbert, side bar), 2.4 (J.R.Heirtzler).

Mammoth extinction during deglaciation (along with many other large mammals): Sect. 11.6.2.

Manganese deposits (composition, origin): Sect. 14.4; Figs. 14.10, 14.11, and 14.12.

Mangrove thicket, brush and forest at the beach: Fig. 3.4; Sects. 3.1.2 and 6.3.4.

Mantle of Earth: Sects. 1.1.2 and 2.5.2; Fig. 1.1.

Margin (continental): Sects. 3.1.1 and 3.2; Figs. 3.1, 3.2, and 3.5.

Margins (organic matter): Sect. 3.1.3.

Margins (origin of passive margins): Sect. 3.2; Figs. 3.5, 3.6, and 3.8.

Marine organisms: A9.

Marine record of short-term climate change: Sect. 15.6.

Martin, J., erstwhile Calif. ocean chemist, proposed a strong role for iron in diatom growth: Sect. 10.4.2 (fourth paragraph).

Mass extinction (at → K-T boundary): Sect. 13.4.

Matuyama Diatom Maximum (ODP Leg 175): Sect. 7.5.5 (fourth paragraph); Fig. 7.14.

Mediterranean drying out (end of Miocene): Sect. 12.2.5.

Mediterranean sapropels (recovered by drilling): Sect. 11.6.3.

Mélange (mixture; applied to landslide rock mixtures in active margins): Sect. 3.3; Figs. 1.2, 3.9, and 3.10.

Mesozoic fossils and rocks: Figs. 13.1, 13.3, 13.4, 13.6, 13.12, 13.13, 13.14, and A6.8.

Mesozoic seafloor abundance: Sect. 13.1.3.

Meteor Expedition (1925–1927): Sects. 1.3.4 and 5.3.4; Figs. 2.2 and 5.15.

Methane clathrates, methane-bearing ice-like substance: Figs. 4.5, 12.16, and 11.2.2.

Methane ice and deglaciation (also see → permafrost): Sect. 11.6.2 (third paragraph).

Methane ice in the Cenozoic: Sect. 12.5; Fig. 12.16.

Methane offshore: Sect. 14.2.6.

Methane runaway feedback potential: Sect. 15.7.

Methane, produced in early diagenesis and by fermentation: Sects. 4.3.2 and 8.6.

Methane (burning ice): Sects. 1.2.2 and 11.2.2; Fig. 1.10.

Michel, H., was on the Alvarez team, involved in the K-T revolution: Sect. 1.1.1.

Micrite = lime mud and silt, commonly made in carbonate reefs and platforms.

Microbes (archaea and bacteria) deep below seafloor: Fig. 1.11.

Microfossils (eukaryotic microbes): A9.2.

Mid-ocean ridge morphology: Sect. 2.3.

Mid-Pleistocene climate shift: Sect. 11.5.2; Fig. 11.10.

Milankovitch theory (principles): Sects. 1.1.3 and 11.1.3; Fig. 11.2.

Milankovitch theory of the ice ages: support from deep-sea sediments: Sects. 11.1.2; Fig. 1.4.

Milankovitch, M., Serbian civil engineer, pioneer ice-age student, orbital theory: Sects. 1.1.1 and 11.1.2.

Milankovitch cycles in the Cretaceous: Sect. 13.2.6; Fig. 13.12.

Miller, K.G., pioneer marine geologist, Rutgers U., plotted isotopic core data from the Atlantic for an early view of Cenozoic cooling and ice buildup: Preface, Fig. 3; Fig. 12.3.

Minamata disease: Sect. 15.3.7.

Minerals, silicates: A4.1 and nonsilicates A.4.2.

Minette ores, oolitic iron (hydr)oxide pellets, abundant in certain Jurassic shelf deposits but apparently not forming at present: Sect. 4.8.3.

Mining the seafloor: Sect. 14.1.1.

Mixing intensity and upwelling, bringing nutrients: Sect. 7.2.4 (last two paragraphs).

Mljet Island (varved sediments): Sect. 15.6.1; Fig. 15.6.

"Moho," Mohorovičić discontinuity, referring to a sudden change in speed of sound within the Earth. The sound speed is relatively low in the crust and high in the mantle, the boundary being near 10 km below the deep-ocean floor (and deeper below the continents): Sect. 2.3.4; Fig. 2.5.

Mollusk diversity (large species): Sect. 8.1.1; Fig. 8.1.

Mollusk diversity (snails, benthic vs. pelagic): Sect. 8.1.2, Fig. 8.2.

Monterey Canyon: extremely large Grand Canyon size → submarine canyon: Fig. 3.16.

Monterey climate event (Monterey C isotope shift; also "silica switch"): Figs. 12.10, 12.11, and 12.12.

MOR (mid-ocean ridge): Sect. 1.3.9; Figs. 1.12, 2.2, 2.3, 2.4, 2.5, and 2.6.

Moraines (and other inherited shelf sediments): Sect. 3.4.

Morgan, W.J., pioneer geophysicist, Princeton U., contributor to plate tectonics: Sects. 1.4.2 and 2.6.2.

Morphology, deep seafloor, MOR to abyssal plain: Sect. 2.3.6; Fig. 2.7.

Mountain making: → subduction.

Mountain, G.S., Rutgers Univ., marine geologist, codiscovered general Cenozoic cooling: Preface, Fig. 3.

Mud volcanoes, holes in the seafloor, presumably made from strong gas sources: Sect. 4.3.2; Fig. 4.6.

Mud, hemipelagic ("continental slope" sediments, mainly): Sect. 10.6.

Murray, W., Scotsman, naturalist on R/V *Challenger* (1872–1876), discovered the chief patterns of sediment distribution on the deep seafloor: Sects. 1.3.2 and 10.1.

Natland, M.L., pioneer geologist, Los Angeles, proposed Pliocene turbidites in Ventura Basin: Sect. 1.3.3.

Namibia upwelling (ODP Leg 175): Sect. 7.5.5, Fig. 7.13.

Nannofossils, geologic label for → coccoliths and → coccolithophores.

Neogene time scale: Fig. 5.

Neogene-type seasonal productivity effects (guess): Sect. 15.5.3.

New York margin: Fig. 3.2.

Nonsilicate minerals: A4.2.

Non-skeletal carbonate: calcareous deposit with no recognizable biological structure: Sect. 4.7.

North Sea oil (occurrence): Sect. 14.2.5; Fig. 14.5.

North-Atlantic Deepwater production (NADW production): Sect. 9.1.5.

Nutrients and waste: Sects. 15.3.1 and 15.3.2; Fig. 15.1.

Obliquity, tilt of Earth axis, frequency of variation: Sect. 1.1.3.

Oceanic anoxic events (OAEs): Sects. 13.2.3 and 13.2.4; Figs. 13.8 and 13.9.

Oil production from the seafloor (ballpark estimates): Sect. 14.2.1.

Olausson diagram, deep-sea sediments, example tropical eastern Pacific: Fig. 10.4.

Onset of the ice ages: Sect. 12.3.

Ontong Java Plateau (Cretaceous sequence, link to volcanism): Sect. 13.2.5; Fig. 13.11.

Oolites, Bahamas, precipitation aided by microbes: Sect. 4.7.2; Fig. 4.15.

Ooze → biogenous sediment made of microscopic shells, commonly seen on and below the deep seafloor: Fig. 4.7.

Ophiolite ("snake rock"): Sects. 2.4.4 and 3.3 (third paragraph); Figs. 2.12 and A6.11.

Orbital ice-age revolution: Sect. 1.1.1.

Organic matter distribution pattern: Fig. 7.4.

Organisms, marine (dominant forms, high-level taxonomy, common-language terms): A9.3.

Organisms associated with coral reefs, Fig. 9.11.

Oscillations (and anomalies): Sect. 9.1.6.

Overwash fan: Fig. 6.8.

Oxygen and carbon isotopes (as climate indicators): Sect. 9.4.2.

Oxygen deficiency (various localities): Sect. 8.2.3.

Oxygen isotope cycles (Urey, Emiliani): Sect. 11.4.5.

Oxygen loss from sediment (early diagenesis): Sect. 7.6; Fig. 7.15.

Oxygen minimum: Sect. 7.3.1.

Oxygen reconstruction in the environment: Sect. 8.2.3; Fig. 8.5.

Oxygen supply vs. marine production in the Cretaceous: Sect. 13.2.1.

Oxygen utilization (AOU): 10.2.5.

Pacific-type (active) margin: mountain-building, factory of continental crust: Sect. 3.3; Fig. 3.2.

Paleoceanography and piston coring: Sect. 11.1.1; Fig. 11.1.

Paleogene ocean environments (guesses): Sect. 12.2.1.

Paleomagnetism on the seafloor: Sect. 2.3.3; Figs. 2.4, 7.14, 13.11, and A3.2.

Paleotemperature and climate zonation: the reconstruction of temperature patterns is fraught with difficulties stemming from partial preservation, from continuing adaptation, and from the applicability of greatly differing time scales depending on record resolution, among other problems: Sect. 9.2.

Panama Paradox: Sect. 12.3.2 (also see → onset).

Parker, F.L., pioneer paleontologist and interpreter of foraminiferal distributions: Figs. 8.5 and A9.2.

Partial preservation of the record and bioturbation: Sect. 8.5.

Passive margin: sinking edge of continent: Sect. 3.2.

Pause (in temperature rise): Sect. 15.4.2.

Pebble beaches: Sect. 5.2.1 (fourth paragraph).

Penck, A., pioneer Alpine ice-age expert, proposed multiple ice ages (4 or 5): Sect. 11.4.1 (third paragraph).

Peridotite, black pyroxene-rich rock of the upper mantle: Sect. 2.3.4 (last paragraph).

Permafrost (on land and in shallow water, in high latitudes) shrinks during deglaciation: Sect. 11.6.2.

Persian Gulf: shelf basin in the arid realm, effects on incoming plankton: Sect. 9.5.3; Fig. 9.17.

Peterson, M.N.A., S.I.O. geochemist in charge of the first deep-sea drilling program: Sect. 10.3.2.

PETM (Paleocene-Eocene Temperature Maximum): Sect. 12.5; Fig. 12.17.

Petroleum (origin and Gulf of Mexico): Sects. 14.2.4 and 14.2.5; Figs. 14.2 and 14.4.

Petroleum (off Southern California): Sect. 3.1.3; Fig. 14.1.

Pettersson, H., Swedish marine geophysicist, led the *Albatross* Expedition: Sect. 1.3.4.

Phanerozoic time scale: Fig. A3.1.

Phosphorites (upwelling, resource): Sect. 14.3.1; Fig. 14.7.

Photosynthesis as sea-level indicator: Sect. 6.2.2.

Phylum, phyla (high-level grouping of orgs., e.g., mollusks, arthropods): Sects. A9.1 and A9.3.

Pillow basalt, lava frozen in water in the shape of pillows: Sect. 2.3.4.

Pioneers of marine geology: Sect. 1.3.

Pioneers of paleoceanography: Sect. 11.1.1.

Piston coring (Kullenberg machine and subsequent modifications): Sect. 11.1.1; Figs. 11.1.

Pitman, W., Lamont geophysicist, proposed flooded shelves from increased seafloor spreading.

Pitman hypothesis (seafloor heat content determines depth of shelf): Sect. 6.5.2.

Placer deposits (beach, resource): Sect. 14.3.3; Figs. 14.8 and 14.9.

Plagioclase, Ca felspar, mineral common in → tholeiite: Sect. 2.4.4.

Plankton fossils: → coccoliths, → planktonic foraminifers, → diatoms, → radiolarians. Also Sect. 4.6.3.

Planktonic foraminifers (temperature reconstruction): Sect. 9.2.1.

Planktonic foraminifers, important part of calcareous → ooze, besides → coccoliths: Fig. 4.7.

Plate stratigraphy and CCD fluctuations: Sect. 12.4.

Plate tectonics revolution: Sect. 1.1.1.

Plate tectonics, forcing: Sect. 1.1.2.

Plate tectonics (pioneers): Sect. 1.3.8.

Plate tectonics (summary): Sects.1.4.1, 2.5.1, 2.5.2, and 2.6; Figs.1.2, 2.3, 2.11, and 2.13.

Plate tectonics (open questions): Sect. 2.6.3.

Plate tectonics (seismology): Fig. 2.11.

Pock marks, gas-spewing holes in the seafloor: Sect. 4.3.2 (last paragraph); Fig. 4.6.

Pollution of the seafloor: Sect. 15.3.

Postglacial sea-level rise and coastal morphology: Sect. 6.3; Figs. 6.5 and 6.6.

Premoli-Silva, I., Italian geologist, co-described (with Hans → Luterbacher) the Gubbio K-T transition.

Preservation effects (planktonic foraminifers}: Sect. 9.2.1; Fig. 10.10.

Problems ahead: Sect. 15.1.

Production patterns in the sea: Sect. 7.2.

Productivity and temperature: Sect. 9.1.3; Fig. 9.4.

Productivity and thermocline: Sect. 7.3.2.

Productivity controlling factors: wind stirring and nutrient abundance: Sect. 7.1.3.

Productivity indicators: Sect. 7.2.3; Fig. 7.4.

Productivity of the ocean: a misnomer; what is meant is *production.* It is controlled by nutrients and amount of sunlight. An alternation of mixing and stratification favors production. Thus, the structure of the water column and circulation are important, as well as seasonal change between stirring the uppermost waters and stable stratification.

Productivity, blue-green dichotomy: Sect. 7.2.2.

Prokaryotes: Sects. A9.1 (third paragraph) and A9.2.

Prokaryotic microbes deep below the seafloor: Sect. 8.6.

Protective structures (coastal): Sect. 5.2.2; Fig. 5.10.

Pteropod ooze, small pelagic snails, aragonitic, shallow parts of the seafloor, easily dissolved: Fig. 10.1.

Pumice, vuggy volcanic rock (gas-filled floats): Sect. 4.2.4 (second paragraph).

Radiation balance: Sect. 9.1.1; Fig. 9.1.

Radiocarbon and bioturbation: Sect. 8.5.2; Fig. 8.16.

Radioisotopes and age assignment: A8.

Radiolarians, geologically the oldest of planktonic microfossils: Sect. 4.4; Fig. 4.7.

Radiolarian (siliceous) distribution pattern (esp. equatorial upwelling): Fig. 7.4.

Rapid burrowing: Sect. 8.3.5 (second paragraph).

Rate of sea-level rise (evidence from ice-age data): Sects. 6.7, 15.4.6, 15.4.7, and 15.7; Figs. 6.16, 11.9, and 15.4.

Rates of sedimentation: indicator of global erosion rates. The range of rates of deposition (more than a factor of ten) reflects intensity (decreases with distance from continent): Sects. 4.2.1 and 4.9; Fig. 4.16.

Raw material, solids from shelves: Sect. 14.3.

Reconstruction of ancient sea-level fluctuations from physical and biological indicators: Sect. 6.5.

Reconstruction of the tertiary: Sect. 12.2.

"Red Clay" and "clay minerals": Sect. 10.5.

"Red Clay" vs. off-margin mud: Sect. 10.2.1; Fig. 10.3.

Red Sea (ore-bearing mud): Sect. 14.4.4 (rifting, NASA satellite photo): Fig. 3.5.

Redox reactions during early diagenesis: Sect. 4.3.2 (second paragraph).

Reef structure (Mg, chert): Sects. 4.6.2 (reef talus and reservoir rock) and 8.3.4 (last paragraph).

Regression (retreat of sea from shelf): Sect. 6.1.2.

Relict sediments, legacy sediments, offer clues to past environments: Sect. 5.1.3 (last paragraph).

Renard, A.F., Belgian naturalist, authored (with John → Murray) the *Challenger* Report on geology.

Residence time: Sect. 4.3.3.

Resource, shallow vs. deep (shallow is cheaper to obtain): Sect. 14.1.1 (fourth paragraph).

Resources from the ocean floor: Sect. 14.1 (introduction to resource chapter).

Revelle, R.R. (The *"Great Experiment"*): Sect. 15.4.1.

Revelle, R.R., Californian geologist and oceanographer, flagged differences in the carbon systems of Pacific and Atlantic (Sect. 10.2.3). Chiefly known as a pioneer of early warnings concerning input into atmosphere and ocean of excess carbon dioxide after the Industrial Revolution. Erstwhile Director of S.I.O.; i.e., Vice Chancellor for Marine Sciences at UCSD.

Revolutions in geobiology and geochemistry: Sect. 1.2.

Revolutions in geology of the twentieth century: Sect. 1.1; also see Chap. 2 for plate tectonics.

Ring of Fire morphology: Sects. 2.4 and 3.3; Figs. 1.3, 2.9, 2.10, and 2.11.

Rip rap, → protective rubble (against wave attack): Sect. 5.2.2 (last paragraph), Fig. 5.10.

Ripple marks: Figs. 5.2, 5.3, and 5.11.

Risk assessment (and human impacts on the seafloor): Sect. 15.3.7.

Risk assessment (and oil spillage): Sect. 14.2.3 (second paragraph).

River input (sediment cycle): Sect. 4.2.1.

Rocky seafloor as habitat: Sect. 8.3.4.

Salinity and temperature reconstruction: Sect. 8.2.2; Fig. 8.4.

Salt age of the ocean: → residence time.

Salt deposits (rift deposits): Fig. 3.7.

Sand and pebbles (beach, resource): Sect. 14.3.4.

Sand, a familiar term, strictly denoting grain size in marine geology: Sect. 4.4.2; Fig. 4.8.

Santa Barbara Basin (varves): Sects. 15.6.2; 15.6.3.

Satellite remote sensing by camera from space is heavily focused on everything that is visible from space, including ocean production. The processes seen belong to a short-term time scale.

Savin, S., pioneer geochemist, Ohio, codiscovered stepwise Cenozoic cooling: Sect. 1.1.1; Fig. 1.7.

Schlager, W., Austrian-Dutch geologist, made important contributions to carbonate platform knowledge.

Schlanger, S.O., pioneer geologist, expert on Pacific islands, codiscovered (with H. → Jenkyns) evidence for anaerobic conditions in the Cretaceous Pacific.

Schott, W., German geologist on the original *Meteor* Expedition: Sects. 1.3.4, 9.2.1, and 10.1.

Seafloor exploration, deep-sea history: Sect. 1.3.4; Fig. 10.1.

Sea level in geology, rise and fall define sustained "transgression" and "regression," which provide the calendar of geologic history: Sect. 6.1; Fig. 6.1.

Sea level and coastal morphology: Sect. 6.3.

Sea level and reef growth: Sect. 6.4.3.

Sea level and the fate of Venice: Sect. 6.6.

Sea-level fluctuations, ice driven: Only 20,000 years ago, ice forming on land had bound some 200 m worth of water from the sea. Even now there is still some 80 m worth of sea-level rise bound up in ice.

Notable fluctuations have been recorded for several million years: Sect. 6.4.

Sea-level indicators: Tidal wetlands occur where the land runs (gently) into the sea: Sect. 6.2.

Sea-level reconstruction, lacking ice: proceeds from physical and biological indicators. The result may have large uncertainties measured in tens of meters: Sect. 6.5.

Sea-level rise (asymmetric ice, Neogene type): Fig. 15.5.

Sea-level rise (ice-age data from the seafloor): Sect. 15.4.7; Figs. 6.15 and 15.4. Also see Sects. 15.4.6 and 15.7; Fig. 15.3.

Sea-level fall in the Cenozoic, albedo change, and other corollaries of cooling: Sect. 12.2.2.

Sea-level rise, postglacial: Fig. 6.5.

Sea level, how fast did it rise? Illustration of a sediment analysis based on oxygen isotope ratios in foraminifer shells from the late Pleistocene: Sects. 6.7 and 15.4.7; Fig. 6.15.

Seafloor as habitat: requirements differ for mobile and sessile organisms, and for those that live exposed to the environment or that make burrows in sediments or in rocks: Sects. 8.2 and 8.3.

Seafloor as waste receptacle: Sect. 15.3.

Seafloor spreading: Sects. 1.3.9 (second paragraph), 1.4, 2.3, and 3.2; Fig. 1.2.

Seamounts, island chains, and hot spots: Sect. 2.6.

Seasonal flux, trapping: Sects. 7.2.4 and 10.2.5; Figs. 7.6 and 10.7.

Seawater chemistry and sediment: reactions of rocks and deposits with seawater: Sect. 4.3.

Sediment cycle: pathways of sediments, some recycled: Sect. 4.1; Fig. 3.1.

Sediment sources: (erosion, weathering vs. production, productivity): Sect. 4.2.

Sediment thickness distribution and offshore oil: Fig. 14.6; Sect. 14.2.5.

Sediment transport and grain size: Sect. 5.1.2; Fig. 5.4.

Sediment transport offshore, Southern California: Fig. 5.1.

Sediment transport, shallow water, ripples, and subsea dunes: Fig. 5.2.

Sediment types (on the seafloor): Sect. 4.4; Box 4.1.

Sediment wedges (continental margins): Sect. 3.1.3, Figs. 3.6.

Sedimentary rocks: A6.2.

Sedimentation (and erosion) rates: Sect. 4.2.1 and 4.9; Fig. 4.16.

Sediments and seawater chemistry: reactions of rocks and deposits with seawater: Sect. 4.3.

Sediments, hydrogenous: produced by mineral precipitation: Sect. 4.8.

Sediments, lithogenous: (rock produced, from erosion): Sect. 4.5.

Seibold, E., pioneer of marine geology, Kiel and Freiburg, senior coauthor of the present book: Preface, Fig. 1, post-Preface essay (obituary).

Seismic profiling: Figs. 3.6, 3.7, 3.10, 3.14, 3.18, 6.11, 10.6, 10.12, 13.6, 14.4, and 14.5.

Sensitivity (variability of conditions, ease of forcing): Sects. 9.1.4 and 9.1.5.

Sensitivity of climate system and safe levels of carbon dioxide: Sect. 15.5.2.

Settling a matter of particle size: Sect. 5.3. (Also see → accelerated sinking.)

Sewage and sludge: Sect. 15.3.3.

Shackleton, N.J., British geophysicist and pioneer isotope geochemist (Cambridge), applied Milankovitch Theory to the ice ages and extended Milankovitch dating deep into the Cenozoic, using oxygen isotope stratigraphy: Sect. 1.1; Fig. 6.10.

Shear (downhill soil creep): Sect. 8.4.2 (second paragraph), Fig. 10.8.

Shelf (approximate area in percent of seafloor): Sect. 3.1.1 (third paragraph). (Continental shelf): Sect. 3.4; Figs. 3.1, 3.10, 3.11, 3.12, and 3.13.

Shelf currents: Sect. 5.3.1.

Shelf sediments as zonal climate indicators (distribution): Sect. 9.4.3; Fig. 9.12.

Shelves and shelf break: continent covered by the sea and edge: Sect. 3.4.

Shepard, F.P., pioneer marine geologist, S.I.O., focused attention on marine sedimentation off California and on submarine canyons worldwide: Sects. 1.3.3 and 1.3.6.

Shoemaker, E., pioneer NASA geologist, contributions to understanding impact events: Sect. 1.1.5.

Short-term climate change in the marine record: Sect. 15.6.

Side-scan echo sounding: Figs. 4.6, 5.12, 5.13, and 5.16; Sect. 3.8 (second paragraph).

Silica loss from sediment (early diagenesis): Sect. 7.6; Fig. 7.15.

Silica Paradox see → Diester-Haass.

Silica Switch: a major event in the Middle Miocene involving basin-basin exchange and the silica cycle. Discovered by the geologists G. Keller of Princeton and J. Barron of the USGS. (See Monterey.)

Silicate minerals: A4.1.

Siliceous ooze and mud: Sects. 4.4, 10.4; Figs. 4.7 and 10.11.

Silt (beyond the shelf break): Fig. 4.11.

Sink, technical term in geochemistry, process that balances “source:” Sect. 4.3.3.

Sinking of MOR (mid-ocean ridge) flanks: Sect. 2.3.5; Fig. 2.7.

Siphon (clams), adaptation (obtaining oxygen and food while hiding): Sect. 8.3.2; Fig. 8.7.

Size classification, sediments: A5.

Slope → continental slope.

Slump: incoherent landslide, also see → submarine landslides.

Smit, J., geologist and paleontologist (Amsterdam), made fundamental contributions to K-T knowledge.

Soft bottom (mud, sand) as habitat: Sect. 8.3.5.

Solar cycles (in the marine record): Sects. 15.6.4 and 15.6.5.

Solid raw material from shelves: Sect. 14.3.

Southern California Borderland: Sect. 3.5; Fig. 3.13.

Spiess, F.N., pioneer oceanographer, geophysicist and engineer (S.I.O.), made important contributions to identifying the location of hot vents: Sect. 1.2.1.

Stability field (T & P, methane hydrate): Fig. 4.5.

Stacking (combining of data points, commonly averaging them, limiting resolution): Sect. 8.5.3.

Stepwise Cenozoic cooling (revolution): Sects. 1.1.1 and 12.2; Fig. 1.7.

Storm damage (coastal): Sect. 5.2.2.

Stott, L., marine geologist (USC, Los Angeles) codiscovered (with Jim → Kennett) the Paleocene-Eocene temperature maximum event (PETM): Sect. 12.5 (fourth paragraph).

Strakhov, N.M., Russian pioneer geologist, studied (with A.D. Arkhangel'sky) the geological history of the Black Sea: Sect. 1.3.6 (first paragraph).

Stratigraphy, nannofossils: Sect. 12.2.1; Fig. 12.6.

Stromatolites, organic structures, commonly finely laminated mounds, football size. Among oldest marine fossils, the presence indicates shallow water cover. Sect. 6.2.2; Fig. 6.4 (Precambrian).

Strontium isotopes. Fig. 12.4; Sect. 12.1.4.

Subcrustal erosion (collision margin): Sect. 3.5 (fifth paragraph).

Subduction: Fig. 1.2, Sect. 3.3.

Submarine canyons: Sects. 3.6 and 5.2.1; Figs. 3.16 and 3.17.

Submarine dunes (giant): Sect. 5.3.4; Fig. 5.16.

Submarine landslides: Sect. 3.5; sixth paragraph and following; Figs. 3.14 and 3.15.

Suess, E., pioneer German-US American geochemist (Kiel and Oregon), studied export of organic matter from the sea surface by trapping, and methane release from the seafloor: Figs. 7.6 and 4.5.

Suess, H., solar chemist (S.I.O.). A 205-y solar cycle has his name: Sect. 15.6.3.

Suspension feeders (common at continental margin): Figs. 8.11 and 8.12.

Suspension load, settling velocity: Fig. 5.5.

Swedish Deep-Sea Expedition (*Albatross*): Sect. 1.3.4.

Symbiosis in benthic foraminifers: Fig. 8.10.

System concept: Preface, Fig. 4.

Taira, A., marine geophysicist, Japan, plate collision studies: Fig. 3.10.

Tektites, glassy objects, usually sand size, commonly classified as objects from space. Occasionally abundant: Sect. 4.9 (penultimate and last paragraphs).

Tectonic erosion, removal of material of a growing active margin: Sect. 3.3 (fourth paragraph), Fig. 3.10.

Temperature and productivity: Sect. 9.1.3; Fig. 9.4.

Temperature and salinity reconstruction: Sect. 8.2.2; Figs. 8.4 and 9.14.

Temperature history reconstruction: Sect. 9.2.1; Fig. 8.4.

Tempestites, storm-produced layers on the seafloor: Sect. 5.1.4 (second paragraph).

Tephra aprons (volcanogenic eruption deposits): Sect. 4.2.4; Fig. 4.4.

Terminations (geologically fast, major melting seen in the ice-age record): Fig. 11.11.

Terraces (California): Sect. 3.4.

"Terrigenous" = from continents: Sects. 2.2.2, esp. second paragraph, 4.1.1, and 4.1.2. Also see Sect. 4.2.

Tethys, an ancient tropical seaway: Sect. 4.6.2, second paragraph, Figs. 12.8 and 13.5.

Tethys closing toward end of Eocene: Sects. 12.2.3 and 12.2.5.

Tetrapods = certain protective structures against breaking waves: Sect. 5.2.2; Fig. 5.10.

Tharp, M., Lamont geomorphologist who worked with Bruce → Heezen to prepare the famous "physiographic map" of the seafloor: Sects. 1.3.6 and 1.3.9 (second paragraph); Fig. 1.12.

Thera eruption (Santorini volcano) in the Aegean Sea may have ended the Minoan culture about 3600 y ago: Fig. 4.4.

Thermocline: the transition layer between shallow (light) water and deep (heavy) water. In the modern ocean, the transition is between warm and cold water and commonly comprises several hundred m of water below 100 m of depth: Sect. 7.3; Fig. 7.8. The thermocline contains the oxygen minimum.

Thickness of deep-sea sediments: Sect. 10.2.4.

Thierstein, H.R., Swiss marine geologist and nannofossil expert, prof. em.ETH, Zurich, erstwhile prof. at S.I.O.: Figs. 1.7 and 12.2.

Tholeiite, type of basaltic rock, common on the mid-ocean ridge: Sect. 2.4.4.

Tidal flats: Sect. 6.2.1; Fig. 5.11 (wadden, North Sea).

Tides, large waves produced by gravitational interaction of moon and sun with a rotating Earth and its complicated morphology: Sect. 5.2.3.

Time scale: there are (at least) three: a human-relevant scale (decades to a few centuries: varves etc., Sect. 15.4.2), the ice-age scale (millennia, Chap. 11), and the geological scale (millions of years, Sects. 12.1.4, 8.1.1, and A3; Fig. 12.4).

Toba volcanism: perhaps the largest witnessed by humans, eruption 73,500 years ago: Sect. 4.2.4.

Topographic statistics, Earth and ocean: A2.

Trace fossils, trails and burrows, bioturbation: Sects. 8.4, 8.5. (Also see → lebensspuren).

Tragedy of the Commons hypothesis (G. Hardin): Sect. 14.1.1.

Transform fault: Sect. 2.5.1

Transgression: Sect. 6.1.2.

Trask, P.D., editor of a famous book with classic articles in early marine geology: Sect. 1.3.1.

Trenches: Sects. 2.4.1., 2.4.2, and 3.3; Figs. 1.2 (lower panel), 2.9, and 2.10.

Trophic level: habitat level in the → food chain.

Tsunami: a wave or a series of waves generated in the deep sea (e.g., by an earthquake), traveling at very high speed and growing to great size upon hitting the shelf: Sect. 3.4.

Turbidity currents, turbidites: muddy flows traveling downslope and deposits of same. Sects. 1.3.3, 3.5, 3.6, 3.7, and 3.8; Figs. 3.16 and 3.18. Also see → Kuenen, → submarine canyons, and → abyssal plain.

Upwelling (coastal): Sects. 3.1.3 and 7.5.1; Fig. 7.10. "Upwelling" pertains to vertical circulation of the upper 300 m or so, in certain coastal waters (e.g., off California, off Portugal, off Peru, off Namibia) and also along the equator. Around Antarctica, deeper waters are involved. Upwelling waters bring nutrients into sunlit depths, thus enabling high production: Sect. 7.5. Upwelling sediments commonly contain much gas (owing to oxygen shortage and fermentation): methane may be made here, resulting in → methane ice and a bottom-simulating reflector or "BDR" (→"clathrate").

Urey, H.C., outstanding physical chemist, Chicago, invented the use of oxygen isotopes for reconstruction of temperature: Sect. 9.4.2. Geologists use the method on calcareous fossils.

Uyeda, S., Japanese marine geophysicist, 1978 textbook on the new tectonics: Fig. 3.1.

Vail sea-level reconstruction: Fig, 6.11.

Vail, Hardenbol, and Haq sea-level curves for the Cenozoic: Fig. 6.12.

Van Dover, C.L., marine biologist, authored a book on the ecology of hydrothermal vents: Sect. 1.2.1.

Varves, annual layers deposited at a high rate with relatively little disturbance from bioturbation (lack of oxygen for large benthic organisms): Sects. 4.9, 9.1.7, and 15.6.1.

Vema, Lamont ship, took a comprehensive early collection of cores: Sects. 1.3.4 and 10.1 (second paragraph).

Venice (sea-level problems): Sect. 6.6.

Vening-Meinesz, F.A., Dutch geophysicist, pioneered gravity measurements in trenches: Sect. 1.3.9.

Vernadsky, V., pioneer Ukrainian geochemist, founder of biogeochemistry, identified negative feedback in geological processes: Sect. 1.1.1.

Vincent, E., French-US micropaleontologist, USC, Los Angeles and S.I.O, expert in Neogene stratigraphy, proposed (with W.H.B.) the → Monterey Event: Figs. 12.11 and 12.12.

Vine, F.J., British geophysicist, proposed (in 1963, with D.H. Matthews, his advisor) magnetic lineations on the seafloor as a corollary of seafloor spreading and magnetic reversals: Sects. 1.1.1 and 2.3.3.

Volcanism, → Ring of Fire, → subduction: Sects. 3.3 and 4.2.4; Figs.1.3 and 4.4. Also see → Trench, → Tephra.

Volcanism and Cretaceous sediments, western Pacific: Sect. 13.2.5; Figs. 13.10 and 13.11.

Volcanogenic beach sand, common on and around oceanic islands, largely black: Sect. 4.5.2; Fig. 4.10.

Volcanogenic margins: Sects. 3.2 and 3.3.

Von Herzen, R.P., Woods Hole geophysicist, studied heat flow patterns on the seafloor: Sect. 1.2.1.

von Humboldt, A., eighteenth- to nineteenth-century naturalist, noted similarity of opposing Atlantic margins: Sect. 1.3.7.

Walther, J., German pioneer geologist, studied environments of marine sedimentation: Sect. 1.3.1.

Walvis Paradox: Silica Paradox (Pleistocene silica cycles off Namibia, higher production during glacial periods resulting in lower sedimentation of siliceous fossils (see → Diester- Haass).

Warm ocean and dearth of oxygen (Cretaceous): Sect. 13.2.

Warm Pool: mass of warm surface water in the west of the equatorial Pacific, created by trade winds, collapses whenever trades become weak or cease (El Niño): Fig. 9.4. (Also see → ENSO.)

Waste (seafloor as waste receptacle, human-derived sediment): Sect. 15.3.1.

Wave base: Sect. 5.2.1 (last paragraph).

Wave-cut terrace: Fig. 6.2.

Waves and sediment transport: Sect. 5.2; Figs. 5.6 and 5.7.

Wegener, A., German-Austrian geophysicist, proposed the hypothesis of continental drift (rejected for many decades, now widely seen as corollary of plate tectonics): Sects. 1.1.1, 1.3.7, and 1.4.1; Fig.1.14.

Wilson Cycle, opening and closing of Atlantic (switching from rifting to collision, proposed by the Canadian geologist J.T. Wilson): Sect. 3.2, last paragraph.

Wilson, J.T. (Toronto), geologist who proposed the ruling hypothesis regarding the origin of the Hawaiian island chain and other chains like it: Sect. 2.6.2; Figs. 2.8 and 2.15. Also see → Wilson Cycle.

Wind input, eolian deposits: Sect. 4.2.3; Fig. 4.3.

Winogradsky, S., Russian microbiologist, early flagged the role of microbes in soil chemistry: Sect. 1.2.3.

Winterer, E.L., (S.I.O.), expert on Alpine and Pacific geology, co-proponent of → plate stratigraphy: Sects. 12.4 and 6.4.3; Figs. 12.15 and 12.14.

Woese, C., (Illinois) microbiologist who realized that prokaryotic microbes come in two very different versions, archaea and bacteria: Sect. 1.2.3, A9.1 (third paragraph).

Wuerm (last glacial period): Sect. 6.4.1.

Younger Dryas: Sect. 11.6.1 (third paragraph); Fig. 11.12.

Suggestions for Further Reading

Sundquist, E.T., and W.S. Broecker (eds.) 1985. The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present. Geophys. Monogr. Ser., 32.

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Mann, M.E., and L.R. Kump, 2009. Dire Predictions – Understanding Global Warming. Dorling Kindersley, New York.

IPCC, 2014: Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Core Writing Team, R.K. Pachauri and L.A. Meyer (eds.)]. IPCC, Geneva.

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