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# Paleoceanography: A review for the GSA Centennial

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## ABSTRACT

**The central problem of paleoceanography is the history of the circulation of the ocean. Although speculation about ancient oceanic circulation goes back to the past century, the field of paleoceanography was founded in the 1950s as oxygen-isotope studies suggested that oceanic deep waters were warmer in the past than they are today. Extensive coring of deep-sea sediments by numerous expeditions after World War II was followed by the ocean drilling programs, providing a rich data base. Paleoceanographic interpretations have tried to explain the most obvious changes in sea-floor sediments and their contained fossils: changing paleotemperatures indicated by oxygen isotopes, fluctuations in the calcium carbonate compensation depth, accumulations of organic carbon-rich sediments, and the unexpected abundance of hiatuses in a setting which had been thought to be the ultimate sedimentary sink. The result has been the intriguing discovery that although the positions and circulation of the major surface gyres is generally stable, the deep circulation of the ocean may reverse on a variety of time scales. It has been suggested that formation of North Atlantic Deep Water, which causes the uneven distribution of nutrients, alkalinity, and oxygen in the deep sea today, may have been replaced by formation of North Pacific Deep Water during the last deglaciation, reversing the concentration gradients of nutrients, alkalinity, and oxygen. On a longer time scale, the present general circulation, which is dominated by production of oxygen-rich cold deep water in the subpolar regions today, may have replaced a pre-Oligocene general circulation in which warm, saline, oxygen-poor deep waters were formed in warm seas in the arid zones. Paleoceanography is still in its infancy; many new clues to the history of the ocean are being discovered, and many new ideas about conditions in the past are being developed. The beginning of the next century should see continuing rapid growth and maturation in this exciting new field.**

## INTRODUCTION

Although systematic geologic investigation of the Earth has been under way since the 18th century, until 25 years ago it was essentially restricted to the study of sediments and rocks exposed on land or penetrated by mining or drilling activity. The 70% of the Earth's surface covered by water was virtually inaccessible to geologic investigation and contributed very little to our understanding of the development of the planet. It now seems strange that geologists of the 19th and the first half of the 20th century should have been so confident that they knew the Earth in spite of this obvious gap in knowledge. It has generally been assumed by those dealing primarily with rocks exposed on land that the history of the oceans can be deduced from observation of the deposits of shallow seas and geosynclines. Thus, Schopf's *Paleoceanography* (1980) discusses what is known of Precambrian. Paleozoic, and Mesozoic oceans from indirect knowledge gleaned from the continents. Paleooceanography is the study of the oceans of the past, and ancient oceans have been at least as different from marginal seas as are the modern oceans. As Berger (1979, 1981) has observed, the central problem of paleoceanography is the history of circulation of the ocean. Although marginal seas may offer important clues, real information that can be used to deduce the history of ocean circulation must come from the ocean basins proper.

## EARLY IDEAS ABOUT THE DEEP SEA

Deep-sea sediments had been sampled in the 19th century first as part of the effort to lay the trans-Atlantic telegraph cables, then as a routine function of the great voyages of scientific exploration of the sea. As recently as 1970, the deep sea was regarded by the majority of geologists as an unchanging primordial environment which had been the most constant feature on the surface of the Earth throughout most of geologic time. Paleontologists considered the cold lightless deep sea to be the final refuge for such forms of life as the hyalosponges and coelacanths which had been unsuccessful in the competition for Lebensraum in shallower waters. The climate of the continents and the shallow epeiric seas had obviously changed with time, but the open ocean was regarded as the great stabilizing feature of the Earth's surface. So ingrained was this idea in geologic thinking that the pioneer micropaleontologists J. J. Galloway and J. A. Cushman studiously avoided examining the biostratigraphic potential of the planktonic foraminifera and radiolaria because it was well known that oceanic plankton did not evolve. C. G. Wallich (1861) had given all coccoliths a single name, *Coccolithus oceanicus*, for the same reason. Only T. C. Chamberlin (1906) bothered to consider the possibility of a radically different ocean, suggesting that salinity rather than temperature differences may have driven oceanic circulation in the past.

Geology is unique among the sciences in having a historical aspect. Probably because of the vast amount of historical geological evidence which had been accumulated and investigated, geologists tended to be very conservative and slow to accept radical new ideas. The classical maxim

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attributed to James Hutton, that "the present is the key to the past," was so thoroughly ingrained in the reasoning of geologists that very little attention was given to the consideration of a world that would have been very different from that observed today. Recall that Alfred Wegener and the idea of "continental drift" were rejected by the great majority of the world's geologists for more than 50 years.

## DEVELOPMENT OF SAMPLING TECHNIQUES AND EARLY RESULTS

No unequivocal ancient oceanic deposits were known until the invention of the piston corer by Kullenberg (1947) and its subsequent extensive use by the Swedish Deep-Sea Expedition. The long piston cores recovered by the Swedish Deep-Sea Expedition in 1947–1948 added historical perspective to the sedimentary record of the ocean basins, albeit only the Pleistocene and late Pliocene were penetrated. Arrhenius (1952) noted that the cores displayed significant variations in carbonate content, which could be correlated with glacial and interglacial ages.

Although sedimentological studies were important in demonstrating that the oceanic environment has in fact changed with time, the new branch of science, "Paleooceanography," can be said to have been founded by Cesare Emiliani, who in three classic papers presented an outline of ocean thermal history and demolished the notion of constancy of the oceanic environment. Emiliani and Epstein (1953) presented the results of analysis of $^{18}O/^{16}O$ ratios (now commonly expressed as enrichment or depletion of $^{18}O$, $\delta^{18}O$, relative to the PDB standard, a belemnite rostrum from the PeeDee Formation in South Carolina) in planktonic foraminifera in samples from the lower Pleistocene of southern California indicating substantial glacial to interglacial temperature fluctuations. Emiliani and Edwards (1953) and Emiliani (1954) reported the results of oxygen isotopic analysis of benthonic foraminifera from pre-Pleistocene samples from the Pacific Ocean basin. These suggested bottom-water temperatures of 2.2 °C for the late Pliocene, 7.0 °C for the early middle Miocene, and 10.5 °C for the middle Oligocene (the supposed Oligocene sample has subsequently been determined to be Eocene; K. G. Miller, 1988, personal commun.). In 1955 Emiliani published the record of Pleistocene variation of the $^{18}O/^{16}O$ ratio in surface-dwelling planktonic foraminifera; the possibility of phyletic influence on isotopic fractionation was excluded by using a single species. He stated that the record of change in the isotopic composition of the tests must be due to (1) change in the over-all isotopic composition of the ocean, which would be a function of the mass of ice existing as glaciers, and (2) the temperature of the surface waters at the time of shell formation. These papers opened a new perspective on the history of the oceans and introduced the prospect of quantification of ocean temperatures of the past, a major descriptor of global climate. In 1958 Emiliani discussed the results of these studies of ancient ocean temperatures in what may be considered the first summary article on paleoceanography.

The Kullenberg piston corer is limited in its depth of penetration by the strength of the materials involved and by the mechanical properties of the sediment being cored. In effect, these limitations restricted the coring efforts in the deep sea to Pleistocene sediments except in those few areas where unlithified older sediments are exposed or are very close to the sea floor; consequently, knowledge of the older history of the ocean basins was virtually non-existent, and discussions from the middle of this century now seem quite fanciful. Kuenen (1950) acknowledged the remote possibility of continental drift, but preferred to assume that the ocean basins were ancient features, and that deposition in them had been going on continuously since the early Precambrian; he estimated that the average sediment thickness should be about 3 km. He mentioned the adaptation of echo sounding to measurement of sediment thickness, described by Weibull (1947), but very few data were available at the time. Poldervaart (1955) noted that Oliver and others (1953) had found that average sediment thicknesses in the ocean were about 600 m and remarked that such thicknesses could be accounted for by deposition during the past 200 m.y. He suggested that "with perhaps lower relief of the remaining land, absence of *Globigerina* and virtual absence of turbidity currents, practically all deposition could have occurred in shallow waters, and most of the ocean floor could have remained barren of a sedimentary cover until modern conditions prevailed."

The 1950s and 1960s were the golden age of piston coring. Ericson and others (1961) reported on studies of 221 Atlantic and Caribbean cores, 41 of which contained pre-Pleistocene sediments, the oldest being Upper Cretaceous. It was evident that unconformities are common in deep-sea sediments, that turbidity-current deposits are widespread, that rates of accumulation in the deep sea are highly variable, and that "a drastic reorganization of . . . the Earth's crust now covered by the oceans took place at some time during the latter part of the Mesozoic era" (p. 282). It had become evident that the ocean had a complex history, but it was not at all clear what that history was.

## THE OCEAN DRILLING PROGRAMS

Because of the great interest aroused by the Pleistocene sequences recovered from the deep sea, it was not surprising that Emiliani championed the cause of deeper coring of the sediment on the ocean floor. It was obvious that the only way in which cores could be recovered from the more deeply buried part of the sedimentary section was through the application of industrial drilling and coring techniques being developed for offshore exploration for oil and gas. In the late 1950s, it had been proposed that there be a national program to drill through the Mohorovičić Discontinuity in the Pacific Ocean basin where it was thought that samples of the mantle could be recovered by penetration to a depth of only about 12 km beneath the sea surface (Bascom, 1958, 1961; Hess, 1959), but these plans called for bypassing the sediment column. Emiliani (1981) has presented a personal recounting of the events of this period, which eventually led to establishment of the National Sediment Coring Program and its operational offspring, the Deep Sea Drilling Project (DSDP), and its successor, the Ocean Drilling Program (ODP).

Scientific drilling operations in the deep sea began on March 6, 1961, using the drilling barge *Cuss I* of Global Marine Exploration Company of Los Angeles in a water depth of 945 m off La Jolla, California; it ended 9 days later after a penetration of 1,315 m into the sea floor. Immediately following this, a second site was drilled 40 nautical miles east of Guadalupe Island, off Baja California. Known as the Experimental Mohole, this effort was conducted in 3,558 m of water and penetrated 183 m of sediment and 13 m of basalt. It demonstrated the feasibility of recovering scientifically valuable cores from depths well below the limit of penetration of the Kullenberg corer. The Mohole effort was conceived in a fixist context; the consensus among geologists was that the ocean basins were permanent features. Project Mohole was abandoned for a variety of reasons, but interest in scientific ocean drilling to recover the sedimentary record continued to develop.

The next effort to recover older oceanic materials was called "Project LOCO" (LOng COres). With funding from the National Science Foundation (NSF), the Global Marine vessel *Submarex* carried out drilling and coring operations on the Nicaragua Rise, from November 27 to December 17, 1963. Weather conditions were unfavorable, but drilling and coring operations penetrated the sediments to a depth of 56.4 m in a water depth of 610 m, with better than 30% recovery (Bolli and others, 1968).

In 1964, four United States oceanographic institutions, The Institute of Marine Sciences (now the Rosenstiel School of Marine and Atmospheric Sciences) of the University of Miami, Lamont (now Lamont-Doherty) Geological Observatory of Columbia University, the Woods Hole Oceanographic Institution, and the Scripps Institution of Oceanography of the University of California, joined together as a consortium which took the name JOIDES (Joint Oceanographic Institutions for Deep Earth Sampling). They proposed and NSF supported a six-hole drilling effort off Jacksonville, Florida, using Global Marine's vessel *Caldrill* in April and May of 1965; these later became known as the JOIDES holes. The six sites were continuously cored to sub-bottom depths of more than 1 km. Examination of the cores revealed that significant changes in the oceanographic conditions on the east Florida shelf and Blake Plateau had occurred since the Late Cretaceous. The samples provided very well preserved assemblages of planktonic and benthonic microfossils; this material was of major importance in developing the biostratigraphic schemes in use today.

The Experimental Mohole, *Submarex* hole and JOIDES drilling off Florida had convinced the scientists involved of two things. (1) Deep-sea sediments contained long sequences of extraordinarily well preserved microfossil assemblages which could be used for global stratigraphic correlation. Before samples from these sites were available, the basis for global stratigraphic correlation had been a fragmentary record preserved on land, often in structurally complex areas such as Trinidad (Bolli, 1957; Bramlette and Wilcoxon, 1967; Hay and others, 1967) where structural complexity made determination of stratigraphic relations difficult. For refinement of biostratigraphy alone, a campaign of further drilling was justified. (2) Although Emiliani's pioneering work of the 1950s had suggested that oceans of the past might have been significantly different from those of today, the stratigraphic sections recovered by these early efforts (less than 5 months of operations at sea over a 4-yr period) confirmed that the history of the ocean basins was anything but simple and unchanging.

JOIDES next proposed an 18-month campaign of drilling in the Atlantic and Pacific Oceans, to be known as the Deep Sea Drilling Project, selecting Scripps to be the operator. The NSF approved and funded the project. Global Marine was able to modify a vessel already under construction to become a research tool specifically designed for scientific ocean drilling, core recovery, and analysis. The new vessel, christened *Glomar Challenger*, set sea from Orange, Texas, on July 20, 1968. This was the culmination of a decade of dreams of marine geologists, and van Andel (1968) summarized the expectations at the time. No one anticipated the full impact of the discoveries which would be made, nor that the at-sea operations of the *Glomar Challenger* would continue for 15 years. By the early 1970s, the project had been so successful (Hammond, 1970) that other United States and foreign institutions joined JOIDES in support of the program (the Department of Oceanography of the University of Washington; the Academy of Sciences of the USSR; the Bundesanstalt für Geowissenschaften und Rohstoffe of the Federal Republic of Germany; the Centre National pour l'Exploitation des Océans (now the Institut français de Recherche pour l'Exploitation de la Mer) of France; the Natural Environment Research Council of the United Kingdom; the Ocean Research Institute of the University of Tokyo, Japan; the Hawaii Institute of Geophysics of the University of Hawaii; the School of Oceanography of Oregon State University; the Graduate School of Oceanography of the University of Rhode Island; and the Department of Oceanography of Texas A&M University). On November 8, 1983, the *Glomar Challenger* returned to port in Mobile, Alabama, for the last time as JOIDES scientific drilling vessel. JOIDES has continued to grow, adding the Institute of Geophysics of the University of Texas at Austin; the Department of Energy, Mines and Resources of Canada; and the European Science Foundation in the 1980s. A larger and more sophisticated drillship was modified to continue the program of scientific drilling in the oceans. Operated by Texas A&M University, the drillship, officially registered as the *Sedco*/BP 471, was named the *JOIDES Resolution* by the scientific community. She set sail on the first of a 10-yr series of cruises on March 20, 1985; this new program carries the experience and ideas developed since 1967 to a new level of sophistication (Hay, 1987).

The impetus for deep-sea drilling was that the recovery of sediment and rocks from the ocean basins was expected to provide a new perspective on the history of the planet. The drive for recovering materials from the ocean floor, however, happened to coincide with the revolution in geological thinking introduced by the theory of sea-floor spreading and plate tectonics. It is one of the most fortuitous events in the history of science that the revolutionary theory of plate tectonics and the geological tools and methods essential for its investigation developed simultaneously but almost entirely independently, and that an unforeseen offspring, the burgeoning field of paleoceanography, would result from this lucky marriage of theory and technology.

## THE AT-SEA AND ON-SHORE ANALYSES

To understand the development of ideas in paleoceanography, it is important to know about the kinds of information which have become available, largely as a result of the Deep Sea Drilling Project and its successor, the Ocean Drilling Program. As each core is recovered, a shipboard scientific party performs certain routine tasks. Before the core is cut longitudinally, bulk density and gamma-ray attenuation are measured to determine the volume of pore space and mass of the solid phase; this is essential for later determinations of sediment mass accumulation rates. The core is then cut and described; smear slides are made at frequent intervals to aid in the descriptions. Although many investigators have made highly detailed descriptions of smear slides, including visual estimates of the abundances of different sorts of mineral grains, little of this detail appears in the Initial Reports of the DSDP or the ODP because the information is presented in a synthesized form in the core descriptions. [All of the original observations have been preserved, however, and are incorporated into the data banks for the projects. Copies of the complete data files for the DSDP will soon be available from the National Geophysical Solar-Terrestrial Data Center (NGSDC) of the National Oceanographic and Atmospheric Administration (NOAA) in Boulder, Colorado, in CD ROM format.] The split core is then sampled for a variety of other studies to be undertaken by the shipboard scientific party and by shore-based laboratories. The shipboard party includes micropaleontologists who study the planktonic foraminifera, calcareous nannofossils, radiolaria, diatoms, and silicoflagellates to provide a biostratigraphic and paleoenvironmental framework for the lithologic studies. Shore-based laboratories provide additional biostratigraphic determinations, grain-size distribution in terms of three fractions, $CaCO_3$ content, organic carbon content, and X-ray mineralogy. In the later phases of the DSDP and in the ODP, paleomagnetic determinations, Rockeval analysis, and more chemical analyses have become routine. The descriptive data and results of analyses of the cores have been published in the *Initial Reports of the DSDP* and *Proceedings Part A: Initial Reports of the ODP*, with volume numbers corresponding to the cruise legs, which were ordinarily two months long. It is in large part due to the homogeneity of this vast body of data that regional and global analyses and paleoceanographic interpretation are possible.

## DEVELOPMENT OF THE STRATIGRAPHIC BASE

Paleoceanographic interpretation requires that stratigraphic resolution be as fine as possible. Because of the inertia inherent in the system, the ocean's response to short-term climatic change is on the order of hundreds to thousands of years; unfortunately, this is the same order of magnitude as the blurring of stratigraphic resolution resulting from bioturbation of the sediment. Hence it requires unusual preservation of the sedimentary record in order to resolve the short-term changes in the ocean which would relate directly to the observations of physical oceanographers. The record of longer-term climate change, such as that from glacial to interglacial and on the order of thousands to tens of thousands of years, is preserved in most oceanic sediments and can be determined from study of stable isotopes. At the time the ocean-drilling studies began, it was expected that the study of materials from the deep sea would result in significant improvements in biostratigraphy, magnetic stratigraphy, and isotope stratigraphy and would improve the level of stratigraphic resolution to that required to relate paleoceanography directly to modern physical oceanography.

For the most part, oceanic oozes are made up of microfossils, and it was widely assumed that it would be possible to greatly refine existing biostratigraphic schemes and achieve much finer stratigraphic resolution. The biostratigraphic zonation schemes for the major calcareous oceanic microfossils, the planktonic foraminifera, and calcareous nannoplankton fossils had been developed just prior to the DSDP using mostly samples from land and the JOIDES holes. The stratigraphic zonation schemes for the calcareous groups were little affected by the new information from the deep-sea samples because it was found that differential dissolution of species confused their apparent ranges and because hiatuses were much more common than had been expected. For the development of biostratigraphic zonations for the siliceous microfossil groups, the diatoms, silicoflagellates, and particularly the radiolarians, the deep-sea material was critically important. The use of fish debris ("ichthyoliths") for stratigraphic work was a direct outgrowth of the need of scientists involved with the DSDP to develop a technique for correlating red-clay sequences. Instead of documenting and refining a global biostratigraphic zonation scheme, the new materials have documented differing species distributions in different areas. The contrasts between tropical, temperate, subpolar, and polar assemblages have become well known, but much remains to be done in correlating species distributions in different environments. The enormous abundance of the microfossils and the global coverage of sampling has permitted statistical studies of oceanic assemblages which have been of great importance to the development of paleoceanographic ideas.

Measurement of natural remanent magnetism in samples from deep-sea cores can determine whether the Earth's magnetic field had normal or reversed polarity at the time of deposition of the sediments. The sequence of reversals observed in the sediments should match that known from areal geophysical surveys of the magnetic-anomaly patterns on the sea floor. The stratigraphic reversal sequence had been worked out during the 1960s and 1970s using piston cores from the deep sea and sections exposed on land, but some parts of the record were uncertain or incomplete. Unfortunately, during the earlier part of the DSDP, disturbance of soft sediment by the drilling and coring process rendered cores unsuitable for paleomagnetic investigation. The DSDP cores became much more useful for paleomagnetic studies after it was decided that all sections should be continuously cored and after the development of the hydraulic piston corer (HPC). The HPC is a hydraulic ram coring device which is used with the drill pipe and which can recover relatively undisturbed cores at sub-bottom depths to as much as 300 m. Since Leg 64 of the DSDP in 1978, it has been standard practice to continuously core the upper part of the section at each drill site with the HPC and to continuously core the deeper parts of the section with the standard coring device. During Legs 73 and 74 of the DSDP, a special effort was made to correlate magnetic stratigraphy with the calcareous plankton zonations and to document the magnetostratigraphic position of important paleontologic datum levels (Moore, Rabinowitz, and others, 1984; Hsü, LaBrecque, and others, 1984; Hsü and Weissert, 1985). The horizons of magnetic reversal are thought to be essentially instantaneous and global, and they can serve as precise references for paleoceanographic studies, but correlation in the intervening intervals is often limited by the uncertainties inherent in biostratigraphy.

Emiliani had envisioned the use of varying ratios of stable isotopes as a routine means of stratigraphic correlation, and although this was proposed at the inception of the DSDP, it required such large numbers of mass spectrometric measurements that it was considered impractical. By the middle 1970s, improvements in mass spectrometry reduced the sample size and time required for such measurements, and more recently stable-isotope stratigraphy has become almost routine. Studies of stable isotopic variation in pre-Pleistocene sediments have shown that there are significant signals which should be useful for stratigraphic correlation; although this technique has to date rarely been utilized, it has in some instances achieved spectacular results, permitting resolutions of 10,000–20,000 yr in the late Pleistocene, 40,000–50,000 yr in the early Pleistocene, and 100,000 yr in the Pliocene and Miocene (Williams and Trainor, 1986; Trainor and Williams, 1987). Integrated biostratigraphic, magnetostratigraphic, and isotopic stratigraphies for part of the Cenozoic have been prepared by Miller and others (1985a, 1988) and Keigwin (1987). At present a number of studies of both the stable isotopes and the variations in $CaCO_3$ (for example, Dean and Gardner, 1985) are underway, seeking a breakthrough that will permit Milankovich cycles to be recognized throughout the Mesozoic and Cenozoic, at last providing the temporal resolution required to relate paleoceanographic changes to modern processes.

## THE DEVELOPMENT OF PALEOCEANOGRAPHY

Emiliani's pioneering work with stable isotopes of oxygen opened the field of paleoceanography in the 1950s. He had discovered that the deep waters of the Pacific had been much warmer during the Paleogene than they are at present. The DSDP confirmed the changes in the thermal structure of the ocean and added a number of new facets to paleoceanography, including clues to the circulation of the past oceans, such as fluctuations of the carbonate compensation depth; deposition of organic carbon-rich sediments over extensive areas; changes in biogeographic provinces; changes in oceanic productivity; and changes in ocean chemistry; as well as direct evidence bearing on the surface, mid-depth, and deep circulation of the ocean and the modes of vertical mixing.

## CHANGES IN THE THERMAL STRUCTURE OF THE OCEAN

The ocean today is characterized by a thin (200–500 m), warm layer of tropical-subtropical water lying between the Antarctic Convergence in the south and the Arctic Convergence in the north. This warm-water "sphere," characterized by waters of relatively low density ($<1.0270$), with temperatures ranging from 12 to 30 °C and salinities from 34 to 37.5 per mil, overlies denser ($>1.0270$) intermediate and deep waters, most of which have a much narrower range of potential temperature (–1 to 5 °C) and salinity (34.4 to 35.0 per mil). Numerical paleotemperatures can be interpreted from ratios of the stable isotopes of oxygen (Emiliani, 1971) and indicate that the present thermal structure of the oceans is characteristic only of the late Cenozoic and Pleistocene-Holocene.

The long-term changes in the thermal structure of the ocean since the middle Cretaceous have been documented in a number of works and summarized by Savin and others (1975, 1985), Shackleton and Kennett (1975), Hecht (1976), Savin (1977, 1983), Keigwin (1979, 1980, 1982a, 1982b), Woodruff and others (1981), Savin and Yeh (1981), Shackleton and Boersma (1981), Saltzman and Barron (1982), Hodell and others (1983, 1985, 1986), Muza and others (1983), Shackleton (1983), Haq (1984), Hsü and others (1984), Loutit and others (1984), Shackleton and others (1984), Johnson (1985), Kennett (1985), McKenzie and Oberhänsli (1985), Murphy and Kennett (1985), Oberhänsli and Toumarkine (1985), Savin and Douglas (1985), Weissert and Oberhänsli (1985), Williams and others (1985), Frakes (1986), Keigwin and Corliss (1986), Miller and others (1987a), Barrera and others (1987), and Mix (1987). The temperature of oceanic surface and bottom waters is interpreted from the $^{18}O/^{16}O$ ratios in the calcite tests of selected species of planktonic and benthonic foraminifera and in the prismatic layer of shells of the Mesozoic pelecypod *Inoceramus* (Barron and others, 1982). General trends for the surface and deep ocean are illustrated in Figure 1. The heavy line shows the long-term average trend; for the Cenozoic, this was established by averaging a number of measurements from different sites over intervals of sub-epoch length; but for the Cretaceous, it represents a small number of single data points from deep-sea benthonic foraminifera, *Inoceramus* (triangles; data from Saltzman and Barron, 1982), and Antarctic shelf benthonic foraminifera (Barrera and others, 1987). More-detailed records from surface-dwelling planktonic foraminifera and abyssal benthonic foraminifera from the Angola Basin are also shown (after Shackleton, 1983). Three major states of thermal structure can be recognized: a state which prevailed during Cretaceous-Eocene time, with Cretaceous tropical surface water temperatures about the same as or warmer than today (27–32 °C), as suggested by Barron (1983a, 1983b, 1984, 1986), and slightly cooler tropical surface-water temperatures in the Eocene (23 °C), as indicated from the studies of Shackleton and Boersma (1981). High-latitude surface-water temperatures may have been as high as 17 °C or as low as 0 °C in the Cretaceous (Barron, 1986), but bottom-water temperatures ranged from as high as 23 °C in the Angola Basin during the Coniacian to 13 °C in the later Cretaceous (Saltzman and Barron, 1982; Barron and others, 1982) and were 10–12 °C in the Eocene (Shackleton and Boersma, 1981). According to Barrera and others (1987), high-latitude Antarctic (Seymour Island) shelf waters had temperatures between 4 and 10.5 °C during late Campanian through Paleocene time.

A very different state began to develop in the early Oligocene. Frakes and Kemp (1972) suggested that ocean waters were warmer during the Eocene and cooler during the Oligocene. Kennett and Shackleton (1976) found oxygen isotopic evidence that bottom-water temperatures dropped sharply 38 Ma. Miller and Fairbanks (1983, 1985) confirmed that bottom-water temperatures decreased markedly near the Eocene-Oligocene boundary and noted that there were shorter periods when bottom-water temperatures became as cold as or colder than they are today (2 °C). They reported that the five coldest episodes were 36–35 Ma (early Oligocene), 31–28 Ma (middle Oligocene), 25–24 Ma (Oligocene/Miocene boundary), and starting at 15 Ma (middle Miocene), corresponding to episodes of glaciation in Antarctica. They suggested that these cold excursions coincide with 30- to 50-m glacio-isostatic falls of sea level. Keigwin and Corliss (1986) have analyzed $\delta^{18}O$ data from a global suite of tropical and mid-latitude locations. They found that the deep-sea benthonic foraminifera record major cooling steps at the middle/late Eocene boundary and near the Eocene/Oligocene boundary, but the oxygen isotopic record of planktonic foraminifera does not indicate an increase in planetary temperature gradient from the Eocene into the Oligocene. They concluded that the deep sea cooled in the early Oligocene by 2° at some localities and by 1° overall; they suggested that the Eocene/Oligocene cooling was associated with the development of continental glaciation. During most of Oligocene–early Miocene time, tropical surface-water temperatures were

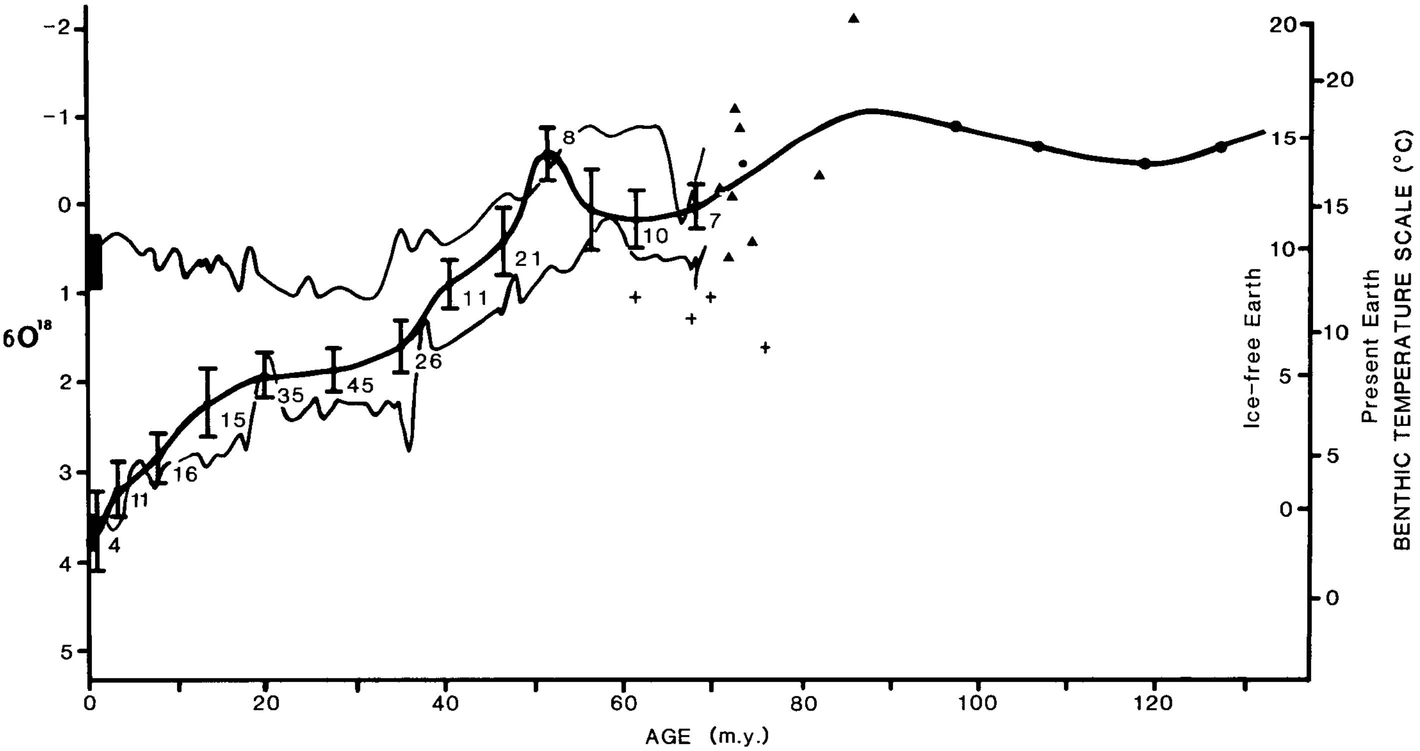



**Figure 1. Cenozoic and Cretaceous ocean-surface and bottom-water temperatures indicated by oxygen-isotope ratios. Heavy line is an average of multiple measurements of benthonic foraminifera over subepoch or stage time intervals with error bars (two standard deviations) shown. Upper thin line is for surface planktonic foraminifera from the Angola Basin; lower thin line is from benthonic foraminifera from the Angola Basin (both after Shackleton, 1983). Filled circles are single data points based on benthonic foraminifera. Triangles are single data points based on *Inoceramus* (Saltzman and others, 1982; Barron and others, 1982). Crosses are data points for Antarctic (Seymour Island) shelf benthonic foraminifera (Barrera and others, 1987).**

about 20 °C and deep-water temperatures about 3–7 °C. A third state has existed since the middle Miocene, with tropical surface-water temperatures about 25–29 °C and deep-water temperatures of 0–3 °C. The thermal contrast between tropical surface and deep waters during the Cretaceous was 10°–15 ° (Hay, 1983); in the Eocene, the thermal contrast was similar, about 13° (Shackleton and Boersma, 1981); in the Oligocene, the thermal contrast remained 13°–15° as both surface and bottom waters cooled (Brass and others, 1982); but from the middle Miocene to Holocene, the thermal contrast between tropical surface and deep waters has increased from about 20 °C to its present value of 25 °C as the deep waters remain cold and the surface waters warm. The significant drop in bottom-water temperatures that began near the end of the Eocene, between 40 and 35 Ma, marked the origin of the psychrosphere, the cold, constant deep-sea environment (Benson, 1975; Kennett and Shackleton, 1976; Benson and others, 1985).

Paleotemperatures have been interpreted from stable oxygen isotopes, but the conversion from the isotope ratio to temperature is not straightforward; an excellent review of the state of the art of interpretation has been presented by Mix (1987). Fractionation by vital effects between different species is eliminated insofar as possible by making the measurements on material from a single or closely related species. The interpretation of temperature, however, must also make assumptions about the isotopic composition of ocean water. At present, the ice on Greenland and Antarctica has a composition very different from that of sea water, being highly depleted in $^{18}O$; Greenland ice ranging from −25 to −35 per mil, and Antarctic ice ranging to −60 per mil, whereas the range in the ocean is only from +1 to −0.5 per mil (Garlick, 1972). Even if the isotopic composition of the ice caps were exactly known, it would still be impossible to determine the exact isotopic composition of the ocean on an ice-free Earth because the isotopes are also fractionated in the carbonate-rock output from the ocean. Nevertheless, it is possible to estimate the over-all isotopic composition of glacial ice and establish an approximate temperature scale for an ice-free world. There is, however, no general agreement as to when (if ever) the world was ice free. It is widely assumed that the Earth was ice free during the Cretaceous. Savin and Douglas (1985) have interpreted the oxygen isotopic data as indicating that the Earth was essentially ice free during the Paleocene, Eocene, parts of the Oligocene, and during the early Miocene. Shackleton and others (1984), Savin and Douglas (1985), and Miller and others (1987) wrote that there have been significant ice caps on the Earth during parts of the Oligocene and from middle Miocene to present. Matthews and Poore (1980) have presented an alternative interpretation of the oxygen isotopic data, suggesting that the Earth has usually had significant polar ice.

The intermediate and deep waters of the ocean are formed by modification of surface waters to increase their density and cause them to sink as downwelling plumes. The formation of denser water can be the result of chilling by cold air, or of increase in salinity as a result of evaporation or sea-ice formation. The ites of formation of such dense water are geographical accidents; the surface water must be quasi-isolated from the main body of ocean water so that it can be strongly modified by local air-sea interaction. The differing thermal structure of the oceans prior to the middle Miocene suggests that intermediate- and bottom-water sources then may have been different from those of today.

Deeper water temperatures have generally been assumed to correspond to high-latitude surface temperatures, following the early ideas of Emiliani (1954, 1961) and assuming that the sites of deep-water formation have always been polar or subpolar. Peterson (1979), however, noted that the deep water of the ocean will be produced by the dense-water source or downwelling plume which has the highest buoyancy flux. The buoyancy flux is the product of the density difference and the volume flow of the plume per unit time. The downwelling plume having the highest buoyancy flux need not be polar or subpolar, driven by the density increase from chilling of the water or increased salinity from sea-ice formation, but may form in the subtropics from density increases as a result of increased salinity caused by evaporation. This is likely when mediterranean seas or shallow shelf seas are developed in the arid subtropics and are able to feed the world ocean. The polar or subpolar origin of deep water in the ocean is unequivocal only for the Oligocene-Holocene. Prior to the Oligocene, the sources of deep water may have been at lower latitudes, and the deep-water temperatures cannot be equated with high-latitude surface temperatures. Furthermore, for the pre–middle Miocene, it is questionable how much of the difference in oxygen-isotope ratios between near-surface–dwelling planktonic microfossils and benthonic microfossils and *Inoceramus* is due to temperature and how much is due to salinity differences between the surface and deep waters. Water evaporating from the sea surface is depleted in $^{18}O$, so that as the salinity increases by 1 per mil, the oxygen isotopes shift by 0.1 to 0.6 per mil, depending on whether the isotopic composition of the evaporated vapor and the precipitation are the same, as in the tropics, or significantly different, as at higher latitudes. Unless taken into account, the shift in oxgyen isotopic composition as the salinity increases by 1 per mil would be interpreted as though the temperature had decreased by .5° to 2.7 °C, depending on the assumptions made about the isotopic composition of evaporated vapor and precipitation during the warm geological times. The question as to whether bottom waters are formed at high latitudes even during the times when the polar regions were much warmer than today, as suggested by Saltzman and Barron (1982) and by Barrera and others (1987), or were formed in the subtropics, as suggested by Brass and others (1982), remains to be resolved.

Shorter-term fluctuations of the oxygen-isotope ratios, in the order of thousands to hundreds of thousands of years, have been known since Emiliani's pioneering work of 1955. Lidz (1966) demonstrated that the variations in the oxygen-isotope record could be precisely matched by variations in assemblages of planktonic foraminifera, but the interpretation of these variations became a matter of controversy. Emiliani (1955, 1971) concluded that in the Caribbean and tropical Atlantic the change in oxygen-isotope ratio from a glacial to an interglacial was about 1.8 per mil, that the average composition of sea water during glacial ages was +0.5 per mil, and that the composition of glacial ice was −15 per mil. This would mean that about 30% of the isotopic signal is due to ice-volume changes from glacial to interglacial, and 70% due to changes in the water temperature. Dansgaard and Tauber (1969), on the basis of study of a Greenland ice core, suggested that the average composition of Pleistocene ice was probably −30 per mil; this would mean a glacial-interglacial change of 1.2 per mil and that 70% of the isotopic signal would be due to ice-volume changes. Shackleton (1967), assuming that bottom-water temperatures remained constant between glacials and interglacials, estimated the isotopic change to be 1.4 to 1.6 per mil; in this case, virtually all of the isotopic signal in tropical planktonic foraminifera is due to ice-volume changes. Shackleton and Opdyke (1973) noted that the isotopic variations in *Globigerinoides sacculifer*, the shallowest-dwelling planktonic foraminifer, in the equatorial Pacific are about 1.2 per mil between glacials and interglacials. They concluded that this is entirely a glacial ice-volume effect because of the similarity of their isotopic record to those established by Emiliani in the tropical Atlantic and Caribbean, because of the similarity of the curve with that for benthonic foraminifera and because of the good correlation of the curve with known sea-level variations. This estimate of the ice-volume effect (0.1 per mil for 10 m of sea-level change) was independent of the assumption that interglacial-glacial bottom-water temperatures did not change. Fairbanks and Matthews

(1978), from study of the isotopic composition of coral specimens from levels with known vertical distances from uplifted interglacial terraces on Barbados, were able to make an independent estimate that the oxygen isotopic composition of sea water varies between 0.10 and 0.12 per mil for 10 m of sea-level change. Shackleton (1986) has presented a recent review of interpretation of the history of the ocean from Pliocene-Pleistocene stable-isotope data.

Birchfield (1987) has made new determinations of the deep-water oxygen isotopic values for the Atlantic, Indian, and Pacific Oceans; he has concluded that the water isotopic change related to ice volume during the last glacial maximum is 74% of the total signal in the Atlantic, 78% in the Indian Ocean, and 84% in the Pacific Ocean. He also concluded that the temperature of deep water in the Atlantic was about 2° cooler during the glacial maximum than it is today; in the Pacific, the deep water was about 1.1° cooler; in the Indian Ocean, it was about 1.4° cooler. Labeyrie and others (1987), assuming that the temperature of bottom waters in the deep Pacific probably remained constant during glaciation, derived a mean global $\delta^{18}O$ record by piecing together the interglacial $\delta^{18}O$ record from a core from the Norwegian Sea with the glacial record of core V19-30 from the Pacific (corrected by the known difference in $\delta^{18}O$ for the part of the cores that overlap). The result was that they estimated that Atlantic deep waters were about +4 °C at 6 Ka and −1.8 °C at 25 Ka; the southern Indian Ocean deep waters were about +1.6 °C at 6 Ka and −1.8 °C at 25 Ka; the equatorial Pacific has been about +2 °C during the Holocene but was −1 °C at 8.4 Ka. They found that the interglacial/glacial shift is smaller (1.1‰) than earlier estimates (1.6‰).

## FLUCTUATION OF THE CARBONATE COMPENSATION DEPTH

The CCD is the boundary between carbonate-rich sediments (for example, calcareous oozes) and carbonate-free sediments (for example, red clay) and reflects an excess of dissolution demand over supply of pelagic carbonate to the deep sea. The CCD in the modern ocean was first described in detail by Schott (1935), on the basis of studies of transects across the South Atlantic by the Meteor Expedition of 1925–1927. The CCD is such an obvious boundary on the sea floor that it is inevitable that it should be considered one of the fundamental clues to the manner in which the ocean processes material.

Bramlette (1958) had noted that in the central Pacific calcareous oozes underlie red clays in depths well below the present CCD. Riedel and Funnell (1964) documented additional cores in which carbonate oozes lie beneath red clay. Heath (1969) suggested that the CCD had descended to depths greater than 5,000 m during the Oligocene and early Miocene, because cores with calcareous oozes of that age had been recovered from abyssal depths in the equatorial Pacific. Both Bramlette and Heath cited the higher abyssal temperatures of the middle Cenozoic as a possible cause for the apparently deeper CCD.

It was discovered on DSDP Leg 2 (Peterson and others, 1970) that there had been significant fluctuations of the calcite compensation depth in the ocean. It was expected that red clays might overlie calcareous oozes as was known from the Pacific, but on Leg 2 it was found that in the North Atlantic Pleistocene and Pliocene calcareous foraminifer-nannoplankton oozes overlie Miocene red clay, which in turn overlies older calcareous oozes, documenting that the CCD had moved both up and down. In the Initial Report for DSDP Leg 3, Hsü and Andrews (1970) followed the common knowledge of the day in assuming that the dissolution of carbonate and the level of compensation depth were determined by pressure. They concluded that the >1-km fluctuations of the CCD documented in the South Atlantic by DSDP Leg 3 were a direct reflection of changes in the depth of the South Atlantic ocean. Because these apparent depth changes did not correlate with known sea-level changes, they suggested that there must have been widespread vertical motions of the ocean floor.

The first compilations of the fluctuations of the CCD through time were by Peterson and others (1970) and Hay (1970), but their work antedated the publication of age-depth curves for the subsidence of ocean crust (Sclater and others, 1971), so that they were unable to determine the true magnitude of fluctuations of the CCD. More sophisticated compilations, taking sea-floor subsidence into account, were published by Berger (1972, 1973), van Andel and Moore (1974), Berger and Winterer (1974), van Andel and others (1975, 1977), Berger and Roth (1975), van Andel (1975), Heath and others (1977b), Le Pichon and others (1978), Melguen (1978), Thierstein (1979), Tucholke and Vogt (1979), Hsü and others (1984), Dean and others (1984), Moore and others (1985), and Hsü and Wright (1985). The effects of the fluctuations on sedimentary lithofacies are shown by the maps of McCoy and Zimmermann (1977). Figure 2 is a recent compilation of data on fluctuations of the CCD in different parts of the ocean. It shows that the CCD has fluctuated by more than 2 km in some parts of the ocean. It is also clear that the CCD behaved differently in different ocean basins during the Cretaceous but began to change in the same way during the Eocene, although the magnitude of the fluctuations differs from basin to basin. Individual basins, notably the Angola Basin, shown in Figure 2, and the semi-isolated Venezuelan Basin (Hay, 1985b) may show CCD fluctuations significantly different from the major ocean basins proper. Although fluctuation of the CCD has become a well-documented phenomenon, its causes are still not well understood.

The term "lysocline" was originally coined by Berger (1970) to indicate the depth zone through which the tests of planktonic foraminifera become dissolved. Broecker and Peng (1982) have noted that this is the depth at which the effects of dissolution become obvious in the sediment. According to Berger and others (1982), the lysocline occurs where about 20% of the carbonate has been dissolved; it is only slightly less obvious than the CCD where dissolution reaches 100%. The lysocline usually marks a transition from *Globigerina* ooze to nannofossil ("coccolith") ooze. Although most writers continue to use the term lysocline to refer to the zone of dissolution of planktonic foraminifera (for example, Berger, 1985), Kennett (1982) applied it in a general sense for any zone in which dissolution of microfossils occurs, that is, foraminiferal lysocline, pteropod lysocline, and coccolith lysocline. Because coccoliths and some other calcareous nannofossils are more resistant to dissolution than most other calcareous bioliths in the sea, the coccolith lysocline immediately overlies the CCD. Although the foraminiferal lysocline tracks the temporal fluctuations of the CCD, the vertical distance between the two does not remain constant, indicating that the slope of the dissolution gradient also changes with time, as was noted by van Andel and others (1975).

Peterson (1966) presented results of an experiment on dissolution of calcite spheres suspended in the Pacific, noting that the rate of dissolution increased rapidly below the top of the cold, $CO_2$-rich Antarctic Bottom Water mass (AABW). Berger (1967) related this to his studies of dissolution of planktonic foraminifera on the sea floor, demonstrating that the level of their rapid dissolution (the lysocline) also coincides with the top of the AABW. Considering the implications for the circulation of the ocean in the past, Berger (1972) attributed the high CCD prior to the late Miocene to more intense production of Antarctic bottom water. During the late Miocene and Pliocene, there was increasingly more significant production of North Atlantic Deep Water (NADW). He speculated that NADW replaced and mixed with the AABW to lower the CCD to its present level. Although this is attractive as an explanation for the Miocene-Holocene behavior of the CCD, it cannot be extrapolated into the Paleogene because the drop in the CCD from the Eocene into the Oligocene is

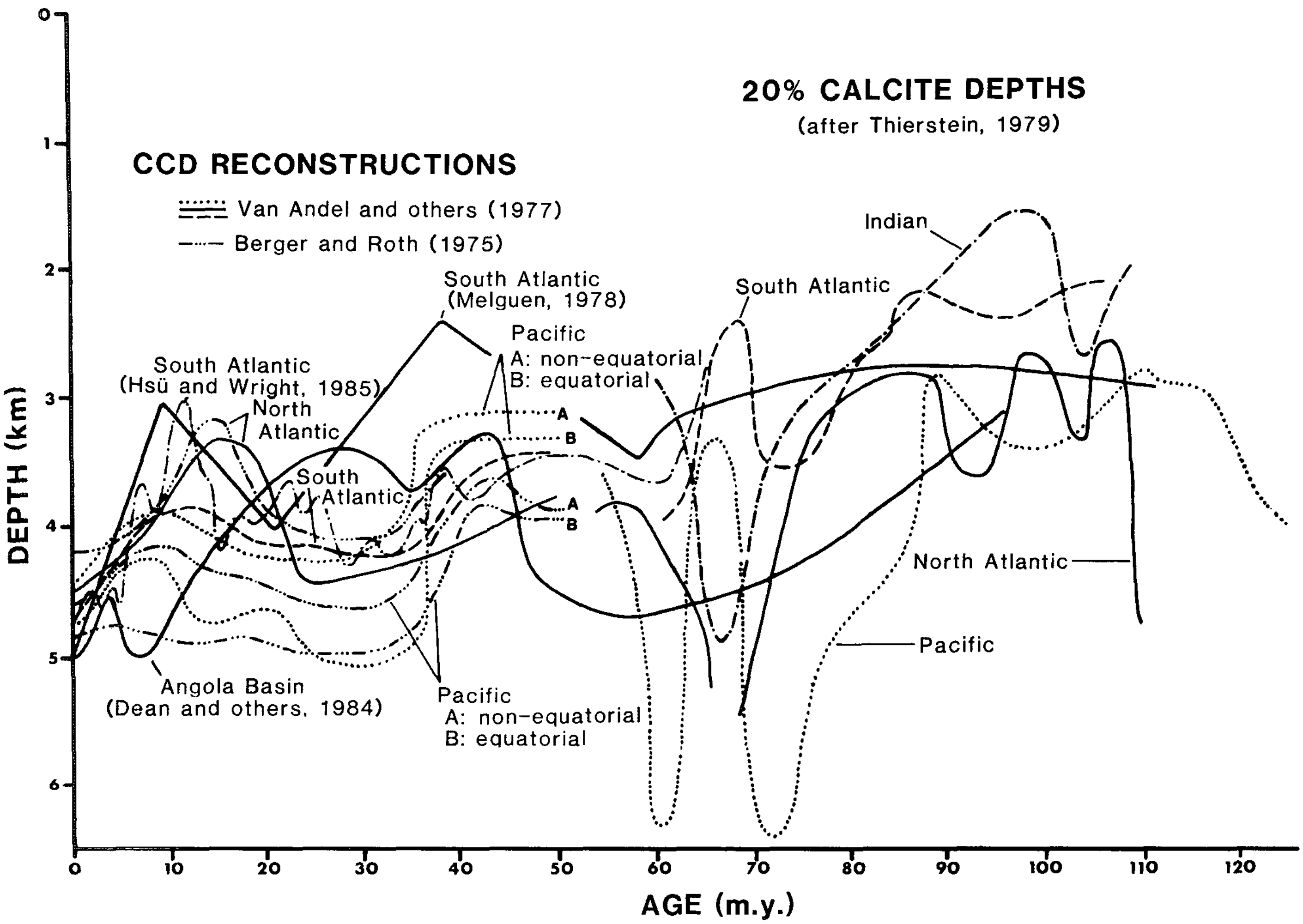



**Figure 2. Fluctuations of the calcite compensation depth (CCD; 0% carbonate) in different ocean basins, showing complications by several investigators. For the Cretaceous and Paleocene, the lines represent 20% carbonate content.**

accompanied by a marked lowering of bottom-water temperatures, as interpreted from the stable-oxygen-isotope ratios in benthonic foraminifera (Kennett and Shackleton, 1976). Re-evaluating the evidence, Edmond (1974) suggested that the lysocline in the Pacific may be related to increased deep-water current velocities. Adding to the uncertainty of the underlying cause of the development of the lysocline, recent studies of calcite dissolution in the Atlantic by Thunell (1982) have shown that in the western Atlantic the NADW/AABW boundary, as defined by a deep thermocline, coincides with the transition from saturation to undersaturation and with the lysocline; in the eastern basins of the Atlantic, however, no such deep thermocline occurs. Although there is no well-defined boundary between NADW and AABW, the lysocline is sharply developed but is significantly deeper than the saturation/undersaturation transition.

Milliman (1974), Berger and Winterer (1974), and Hay and Southam (1977) discussed why fluctuations of the CCD should be expected to follow sea level: higher sea levels would flood a larger shelf area and increase rates of sedimentation of shallow-water carbonates on the shelves; this would decrease the supply of carbonate to the deep sea, and without any concomitant decrease in dissolution demand, the CCD must rise. Verification of the sea-level control hypothesis was possible after the publication of the sea-level curves of Vail and others (1977). The first- and second-order changes in sea level described by Vail and others do seem to be well correlated with the over-all trends for the CCD, at least for most of the Cenozoic (Davies and Worsley, 1981). Hay and others (1982, 1984) have even suggested that Eocene and younger third-order cycles of Vail and others (1977) may be recorded as CCD fluctuations in the sediments of the Angola Basin.

The study of the distribution of carbonates in the Mesozoic and early Cenozoic has demonstrated that, until the Eocene, fluctuations of the CCD in different ocean basins were not synchronous and equal (see Fig. 2). The major ocean basins behaved differently, suggesting that there were local competing deep-water sources with different chemical characteristics; in particular, at the end of the Mesozoic, the CCD sank in the Atlantic and rose in the Pacific (see Fig. 2). This has been attributed to fractionation of carbonate between ocean basins, in much the same way that a balance between equatorial and extra-equatorial compensation has been postulated for the Pacific.

It is now clear that dissolution of different forms of calcite and aragonite may occur throughout the entire water column. Between the lysocline and the CCD, the sediments consist of calcareous nannoplankton which are relatively more resistant to dissolution than are the tests of planktonic foraminifera. As Broecker and Peng (1982) have noted, the percentage of $CaCO_3$ in the sediment is a deceptive index of dissolution because the rain rate of noncarbonate sediment is usually a small fraction of the carbonate rain rate. As a result, the proportion of carbonate in the sediment does not change appreciably until most of the carbonate has been dissolved.

In retrospect, it has become evident that calcium-carbonate deposition in the ocean is a highly complex phenomenon which is influenced by a number of variable factors (Broecker and Peng, 1982). The supply of calcium carbonate to the ocean may vary with time (Budyko and Ronov, 1979; Southam and Hay, 1977, 1981; Hay, 1985a; Budyko and others, 1985); the supply from subaerial erosion, the supply of $Ca^{++}$ from submarine weathering of mid-ocean–ridge basalts (Honnorez, 1983; Mottl, 1983), and the supply from submarine dissolution of older sediments (Moore and Heath, 1977; Moore and others, 1978; Thierstein, 1979; Thiede, 1979, 1981; Thiede and Ehrmann, 1986) are all subject to change with time and may even be interrelated in complex ways. If the model of Berner and others (1983) is correct, then an increase in the rate of sea-floor spreading would result in increased submarine weathering of mid-ocean–ridge basalts, releasing more $Ca^{++}$; it would also result in more volcanic activity and increased input of $CO_2$ into the atmosphere and ocean, as had been suggested by Budyko and Ronov (1979) and discussed more thoroughly by Budyko and others (1985). Because the $CO_2$ content of soils is already high with respect to that of the atmosphere, the increased $CO_2$ content of the atmosphere would not itself directly affect the rate of weathering of rocks on land, as suggested by Budyko and Ronov (1979) and Budyko and others (1985), but would result in general warming of the Earth which would in turn increase the rates of chemical reactions and cause subaerial weathering rates to increase (Berner and others, 1983; Lasaga and others, 1985). Evidence that the $CO_2$ content of the atmosphere has indeed changed has become available in recent years, first from ice cores which demonstrated that during the last glacial atmospheric $CO_2$ content was one-half to one-third that of the present (Delmas and others, 1980; Neftel and others, 1982), and subsequently from deep-sea cores (Shackleton and others, 1983). It has also been suggested that the changing $CO_2$ content of the atmosphere affects dissolution directly by increasing the total $CO_2$ in the ocean and by causing changes in oceanic circulation and productivity (Vincent and Berger, 1985; Berger, 1985; Boyle, 1986b; Berger and Mayer, 1987).

The partitioning of carbonate deposition between the continental shelves, marginal seas, and the ocean basins may be highly significant and is probably the first-order control on the CCD (Milliman, 1974; Berger and Winterer, 1974; Hay and Southam, 1977). The biological fixation processes for $CaCO_3$ may vary with time; fixation of calcium carbonate as calcite by marine plankton appears to have been significant only during the past 100–150 m.y. (Poldervaart, 1955; Hay and Southam, 1975; Sibley and Vogel, 1976). The apportionment between aragonite and calcite, two phases of $CaCO_3$ with different solubilities (Sandberg, 1983, 1985) may have varied with time. The apportionment between benthonic and planktonic fixers and even between biologic and abiologic fixation may also have varied with time. Calcite and aragonite are secreted by marine organisms in different ways, and as a result, the mineral phases produced by different taxa may have significantly different solubilities. The calcite spicules secreted by soft corals (gorgonians, plexaurids, and so on) dissolve in waters which are supersaturated with respect to calcite; at the other extreme, minute calcite coccoliths are more resistant to dissolution than are the much larger tests of planktonic foraminifera. The rate of fixation by any of these processes may vary with changes in the composition of sea water and because biologic agents are involved with the availability of nutrients. Changes in the rate of biological fixation in the surface waters has been invoked to explain the rapid fluctuations of the CCD between the last glacial and the Holocene. The mechanism of transport of the calcareous particles to the deep-sea floor (Honjo, 1980), whether incorporated in fecal pellets or as discrete particles, may vary with time. The flux of carbonate produced in shallow water and carried to the deep sea after winnowing and transport by turbidity currents may be significant at some times. The apportionment of C between the organic carbon and $CaCO_3$ reservoirs may be a significant variable. At present about one-fifth of the carbon leaves the ocean as organic carbon and four-fifths as $CaCO_3$ (Broecker and Peng, 1982), but variations have occurred in the past. Hay's (1985) analysis of Ronov's (1980) and DSDP data indicates that the ratio of $C_{org}$ to $CaCO_3$ is one-seventh for the Cenozoic and one-sixth for the Mesozoic; for the Paleozoic rocks still preserved on the continents, however, the ratio is less than one-tenth $C_{org}$ to nine-tenths $CaCO_3$. The preservation of $CaCO_3$ depends on its not being dissolved or subsequently eroded. Factors which affect dissolution of carbonate significantly as it descends through the water column or at the sediment surface have been discussed by Li and others (1969), Heath and Culberson (1970), Takahashi and Broecker (1977), Sclater and others (1979), and Broecker and Peng (1982); these factors include alkalinity and $CO_2$ content of the waters, temperature, pressure, and the velocity and turbulence of the water over the sediment-water interface (Edmond, 1974). Dissolution may occur within the sediment due to internal production of $CO_2$ as a result of oxidation of organic matter (Emerson and Bender, 1981), and even the rate at which burrowing organisms rework the sediment may be significant (Schink and Guinasso, 1977). Recently Delaney and Boyle (1988) have shown that information on the fluctuations of the CCD can be combined with information on the mean carbon isotopic ratio, the strontium isotopic ratio, the lithium-to-calcium ratio and the strontium-to-calcium ratio to assess possible changes in continental weathering rates and fluxes to the ocean, changes in the isotopic composition of materials being weathered on the continents, amount of carbonate deposited on shelves as opposed to in the deep sea, changes in the deposition or weathering rates of sediments, and changes in hydrothermal circulation. They concluded that, during the Tertiary, hydrothermal fluxes have been constant or have decreased, continental weathering and continent-to-ocean fluxes may have increased, and the proportion of carbonate deposited on the shelves has decreased. In spite of this great complex of factors which may affect $CaCO_3$ deposition in the deep sea, I believe the primary long-term control to be sea-level change and that the secondary controls are related to water-mass formation.

## SEDIMENTS WITH HIGH ORGANIC CARBON CONTENT

Prior to the DSDP, young deep-sea sediments with high organic carbon ($C_{org}$) content were known from the Black Sea (Ross and Degens, 1974) and eastern Mediterranean, and ancient sediments with high $C_{org}$ were known from the western North Atlantic (Windisch and others, 1968; Habib, 1968). Although more sediments with high $C_{org}$ content were recovered in the western Atlantic during Leg 1 of the DSDP (Ewing and others, 1969), the notion that any significant area of the deep sea could become anaerobic was rejected by ocean geochemists (Broecker, 1969). It was not until extensive drilling had been carried out that a pattern of widespread deposition of organic carbon in oceanic basins began to emerge (Fig. 3). "Black shales" have been found in the Cretaceous of all ocean basins; however, the term is used in a manner somewhat different from the term "black shales" on land. In the deep sea, the black shales are thin, $C_{org}$-enriched layers (millimeters to several tens of centimeters) incorporated in $C_{org}$-poor sequences of green, brown, or red clays or chalks. The black shales of the deep sea never exceed tens of centimeters in thickness, whereas the Chattanooga Shale, the black shales of the Pennsylvanian cyclothems, and the contemporaneous black shales of the

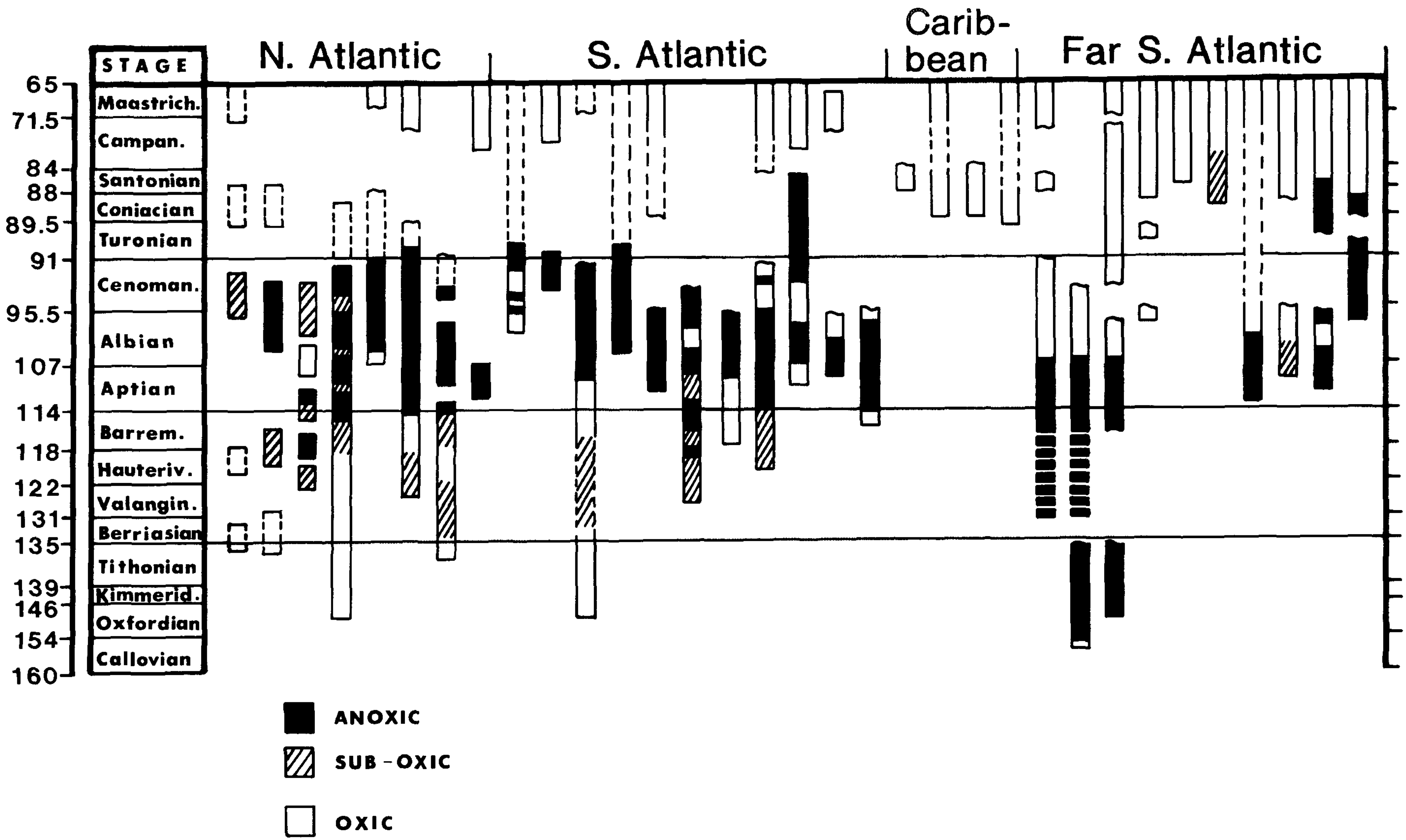



Figure 3. Distribution of anoxic, sub-oxic, and oxic conditions in bottom waters of the Atlantic and Caribbean interpreted from organic carbon content and sedimentary features of the sediments (after Krashenninikov and Basov, 1985).

Cretaceous western interior seaway of North America are usually meters to tens of meters thick.

As shown in Figure 3, the most extensive and intensive episode of deposition of $C_{org}$-rich sediments in the ocean basins occurred during the Aptian and early Albian (110–100 Ma) in the South Atlantic, which at that time was a long narrow sea connected with the world ocean only at its southern end, through the Cape and Argentine Basins. High organic carbon contents (up to 21%) have been found in sediments recovered from the Angola and Cape Basins (Bolli and others, 1978; Hay and others, 1984), and black shale layers extend south to the Falkland Plateau (Barker, Dalziel, and others, 1977; Ludwig, Krasheninnikov, and others, 1983). In the Angola Basin, the $C_{org}$-enriched sediments overlying an evaporite sequence of early Aptian age that was deposited as ocean crust appeared during the opening of the South Atlantic. The paleodepth range through which this Aptian–early Albian organic carbon enrichment occurs is 2.2 km, from the basin bottom at 3 km to about 0.8 km where the highest $C_{org}$ content occurred. It is thought that this episode involved development of an extensive oxygen minimum throughout the entire Angola-Brazil basin complex up to the level of the sills (Walvis Ridge and Rio Grande Rise) connecting it with the Cape-Argentine Basins (Arthur and Natland, 1979). On the Maurice Ewing Bank of the Falkland Plateau, the black shales are more massive than are those from the basins proper and are of Late Jurassic, possibly Neocomian and Aptian–early Albian age. They were formed in waters of outer-shelf, upper-rise depths (Basov and Krasheninnikov, 1983). In the later Cretaceous, another extensive episode of deposition of $C_{org}$-rich sediments occurred during the Turonian–early Santonian in mid-depth waters of the Angola Basin, extending west to the Sao Paulo Plateau between the Brazil and Argentine Basins (Perch-Nielsen and others, 1977). Again, sediments from the deeper parts of the Angola Basin have a lower $C_{org}$ content than do those at shallower depths, suggesting mid-depth intensification of the oxygen-minimum zone. Detailed examination of the middle Cretaceous organic-enriched sequence on DSDP Leg 75 (Hay and others, 1984) showed that the age range for all of the black-shale layers extends from late Albian to early Santonian, with the greatest abundance of $C_{org}$-enriched layers straddling the Cenomanian-Turonian boundary. At Site 530, where the stratigraphic section and occurrence of black shales is typical of both the South and North Atlantic, there are 262 beds of black shale in the middle Cretaceous, with $C_{org}$ content up to 19%. The black-shale layers range from 1 to 62 cm in thickness, averaging 4.3 cm. They are separated by green or red shales which are virtually devoid of organic carbon. The $C_{org}$ content of the black shales was found to be inversely related to the thickness of the enriched layer. Calculations show that the $C_{org}$ content of each layer is approximately that which would be expected if (1) the deep basin waters of that time had had a nutrient concentration comparable to that of the modern deep sea, (2) the entire water column with its nutrients had been cycled through the photic zone, and (3) all of the nutrients had been fixed with organic carbon in the Redfield ratio characteristic of modern plankton. The average $C_{org}$ content of the entire sequence is less than 0.2%, so that there is no over-all enrichment of $C_{org}$ over average marine sediment, but there is instead a sharp separation into enriched and depleted layers. The black layers, with high $C_{org}$ contents, are often sandwiched between green shales, and green-shale layers are often sandwiched between red-shale layers. Dean and others (1984) concluded that it was impossible to be certain that the bottom waters were anoxic at the time of deposition of all of the black-shale beds. About 80% of the black-shale layers were laminated, but about 40% of the layers were also bioturbated to some extent (Stow and Dean, 1984). Although most of the organic matter in the black shales was of marine origin (Deroo and others, 1984), the entire sequence containing the black shales included a number of fine-grained turbidites (Degens and others, 1986). It was not possible, therefore, to determine an exclusive mode of origin for the organic matter, nor was it possible to exclude the possibility that anaerobic conditions existed only in the bottom sediments themselves and not in the overlying water column (Dean and others, 1984).

The eastern and western North Atlantic behaved differently. In the North Atlantic, black shales occur in Aptian and early Albian strata throughout the depth of the basins, but even the layers most enriched in $C_{org}$ have relatively low contents (up to 4%). The greatest $C_{org}$ enrichments occur in late Albian–Turonian sediments and are restricted to the deepest part of the basins. In the western North Atlantic, moderate $C_{org}$ contents occur in sediments of Hauterivian to Aptian age and $C_{org}$ contents are greatest in mid-depths. Samples with up to 11% have been recovered from Cenomanian-Turonian strata. In the North Atlantic, however, organic matter in sediments from the western basins is mostly terrigenous, with woody fragments a common component, whereas organic matter in the black shales of the eastern basins is dominated by material of marine origin.

Organic carbon–rich sediments also occur in the Venezuelan Basin of the Caribbean, although except at Site 146/149 the $C_{org}$-enriched layers were overlooked by the DSDP shipboard scientific party (Hay, 1985b). At Sites 146/149, 150, and 153, high organic carbon content occurs in beds of Coniacian/Santonian age presently at depths of about 4,700 m; The $C_{org}$ contents range up to 11.2%. The location and paleodepth of the floor of the Venezuelan Basin in the Cretaceous are uncertain, but because sites 146/149, 150, and 153 are presently in the deeper part of the basin, it seems likely that they are due to brief pulses of warm saline bottom water which produced anoxia beneath a deep pycnocline. Organic carbon–rich layers also occur at Site 151, which is shallower, but are late Santonian; here they may be due to intensification of a mid-depth oxygen minimum.

No deposits containing significantly more than 1% $C_{org}$ have been found in the Indian Ocean except for Aptian-Albian samples which contain contents up to nearly 4%.

In the Pacific, significant $C_{org}$ contents have been found in isolated samples from middle to shallow depths in Barremian strata (9%, 1.8-km paleodepth, Site 306) and in Barremian-Aptian strata (28%, near surface paleodepth, Site 317); these also appear to represent intensification of the oxygen minimum.

Although the correlation of individual layers of black shale is almost always questionable, there were times at which the deposition of $C_{org}$-rich layers was favored on a regional or global basis. The Aptian–early Albian was a time of global tendency toward black-shale deposition, and the interval of the Cenomanian-Turonian boundary is exceptional both because it was a time when there was again a global tendency to black-shale deposition and because the boundary itself records a moment when black-shale deposition occurred simultaneously over a very wide area (Schlanger and Jenkyns, 1976; Graciansky and others, 1984; Arthur and others, 1985; Herbin and others, 1986). Known as the "Bonarelli bed" in the southern Alps, the Cenomanian-Turonian–boundary bed is a black-shale layer which can be specifically identified by a carbon-isotope signature (Scholle and Arthur, 1980; Schlanger and others, 1986), suggesting a global episode of carbon burial (Arthur and others, 1985).

Interpretation of the meaning of $C_{org}$ enrichment of sediments is more complex than merely the result of salinity stratification (by analogy with the modern-day Black Sea) or expansion and intensification of the oxygen minimum, as had been assumed (Schlanger and Jenkyns, 1976; Ryan and Cita, 1977; Fischer and Arthur, 1977; Thiede and van Andel, 1977; Arthur and Schlanger, 1979). In the simplest scenario, it was assumed that the rate of fixation of carbon in the surface waters, and hence the supply of carbon to the depths, remained constant, but the supply of oxygen is retarded by either salinity stratification or reduced rates of vertical mixing in the ocean, resulting in anoxic conditions. Alternatively, the supply of oxygen to the deeper waters could remain constant and the productivity of the surface waters might increase, again resulting in the development of anoxic conditions. The dissolved-oxygen content of intermediate and deep waters of the ocean reflects a dynamic balance between supply of oxygen to the surface waters by diffusion from the atmosphere, oxygen production within the surface waters as a result of photosynthesis, oxygen demand in the decomposition of settling organic matter, and advection of cold and/or saline former surface waters. Excellent discussions of these factors have been presented by Wyrtki (1962) and Broecker and Peng (1982).

Tissot and others (1979), Herbin and Deroo (1982), and Meyers and others (1986), however, have noted that much of the organic material in Cretaceous strata of the Atlantic, especially in the western basins, is of terrigenous origin and consists of forms of kerogen and charcoal, which are less readily oxidized than organic matter of marine origin and upon maturation is likely to produce gaseous hydrocarbons. The organic matter of the eastern Atlantic basins is mostly from marine plankton, and upon maturation, it is more likely to produce liquid hydrocarbons. Because the sources of the $C_{org}$ which becomes incorporated into the sediment are so different, it is difficult to give credence to a scenario which requires a constant supply of $C_{org}$. Rather, the accumulation of $C_{org}$ in sediments on the sea floor is a function of many of the same factors which affect the accumulation of $CaCO_3$: (1) variations in the supply of C from older rocks or volcanoes; (2) partitioning of C between the $C_{org}$ and $CaCO_3$ reservoirs; (3) variations in the rate of fixation of $C_{org}$, which is complex because fixation in significant quantities may occur on land as well as in shallow seas, marginal basins, and the deep sea; (4) the oxidation profile in the water column; (5) the oxidation capacity of the bottom and interstitial waters; and (6) the degree to which the sediments are worked by burrowing organisms. In spite of this complex of factors which may affect burial of $C_{org}$, its ultimate preservation in the sediment appears to be a nonlinear function of the sedimentation rate (Heath and others, 1977a; Mueller and Suess, 1979; Southam and Hay, 1981) such that as the sedimentation rate increases, the rate of $C_{org}$ burial and preservation increases rapidly. Although it had been considered that the overriding control on preservation might simply be the rapidity of burial, more recent studies (Emerson, 1985) suggest that the increase in amount of $C_{org}$ buried with increasing sedimentation rate is a result of a concomitant increase in the rain rates of both the $C_{org}$ and other sedimentary material. Lyle and others (1988) consider the rain rate of $C_{org}$ to be the most significant factor controlling the $C_{org}$ content of the sediments, and they noted that fluctuations are related to productivity on a regional scale.

It had been thought that anaerobic conditions in the bottom waters were indicated by the finely stratified or laminated sediments which are evidence of the absence of benthic burrowing organisms, but recently it has been discovered that in the basins off southern California mats of sulfide-oxidizing bacteria may occur at the sediment water interface, so that anaerobic sediments can be deposited beneath moving aerobic waters (Williams and Reimers, 1982).

Evidence that anaerobic conditions are due to stagnation induced by hypersaline conditions has been suggested, but it is less easy to document. Fischer and Arthur (1977) proposed that the accumulations of $C_{org}$ were the result of general oceanic stagnation resulting from sluggish oceanic circulation during "polytaxic" times, that is, episodes when the diversity of oceanic species was high. Arthur and Natland (1979) favored stable stratification as a result of sinking of dense saline waters produced in shelf seas or restricted low-latitude ocean basins. Drawing analogies with studies of saline lakes by Hay (1966), Natland (1978) had argued that the presence of phillipsite with authigenic kaolinite and illite at Site 361 in the Cape Basin and Site 364 in the Angola Basin strongly suggested elevated salinities. Brass and others (1982a, 1982b) suggested that during the Cretaceous the bottom waters of the ocean may have been warm saline waters, rather than cold waters as at present. Warm saline bottom (or intermediate) waters would be formed in marginal seas in the arid zones through evaporation. Because the temperature of the water at the site of formation is likely to be very warm, and because the solubility of oxygen decreases both as the temperature and as the salinity of the water increases (Weiss, 1970), dense warm saline waters would carry only about half as much oxygen to intermediate or deep levels in the ocean. Further, the basins of young opening oceans are often relatively isolated by shallow intervening ridges, and during the Cretaceous, there may have been many local sources of intermediate and deep waters.

$C_{org}$-rich sediments are not confined only to the Cretaceous, but are also especially common in the Pleistocene-Holocene deposits of several marginal seas, notably the Black Sea (Ryan, Hsü, and others, 1973; Degens and Ross, 1974), the eastern Mediterranean (Thunell, 1969; Thunell and others, 1977, 1983, 1984; Cita and Grignani, 1982; Thunell and Williams, 1983; Ganssen and Troelstra, 1987; van Hinte and others, 1987), Sulu Sea (Linsley and others, 1986), Sea of Japan (to be investigated on a future Leg of the Ocean Drilling Program), Cariaco Trench (Edgar, Saunders, and others, 1973), in the basins off southern California and in small basins on the northern slope of the Gulf of Mexico (Shokes and others, 1977; Trabant and Presley, 1978; Leventer and others, 1983). In these cases, anoxic conditions in the bottom waters have been caused by stratification of the water column, most often associated with the effects of lowered sea level, large volumes of glacial runoff during the deglaciations, and with strengthened monsoons supplying large quantities of fresh water to the region.

## PRODUCTIVITY

A fundamental question which has not been satisfactorily answered is whether organic productivity in the ocean waters during the Cretaceous was significantly different from that of today. Bralower and Thierstein (1984), making use of the relationships between productivity, sedimentation rate, and $C_{org}$ preservation determined for young sediments by Mueller and Suess (1979) and other data on Holocene sediments, have estimated that primary productivity in the ocean during the middle Cretaceous was an order of magnitude lower than it is today. De Boer (1986) also concluded that black-shale deposition was a result of decreased circulation but emphasized the importance of nutrients from land. In contrast, modeling exercises by Southam and others (1982) have indicated that because the oxygen content of the intermediate and deep waters and the supply of nutrients to the surface ocean are both linked to the vertical circulation of the ocean, development of extensive anoxia should occur when the thermohaline (or halothermal) circulation is rapid. This is not in conflict with the conclusion of Bralower and Thierstein that Cretaceous productivity was much lower than that at present, but it would indicate that turnover in the Cretaceous ocean may have been more rapid than at present.

The most obvious indicator of productivity in the ocean is the occurrence of biogenic silica (Lisitzin, 1972, 1974; Berger, 1976; Molina-Cruz and Price, 1977; Barron and Whitman, 1981; Broecker and Peng, 1982; Monin and Lisitzin, 1983; Leinen and others, 1986). At present biogenic silica accumulates in a belt around the Antarctic continent, in the equatorial Pacific, in the western Bering Sea, and in the Gulf of California (Calverg, 1966, 1968) and Sea of Okhotsk (Bezrukov, 1955), as well as in other upwelling regions. In each case, upward mixing of nutrient-rich waters containing dissolved silica results in high productivity in the surface waters, and a high proportion of silica-secreting plankton, such as diatoms and radiolarians, in the upwelling communities. The Gulf of California is a site of very high output of biogenic silica because of the year-around upwelling which shifts from one side of the Gulf to the other with the seasonal changes in wind direction. The Gulf of California and Sea of Okhotsk draw on the silica supply of Pacific intermediate waters and are, effectively, global sinks, with output rates per unit area two orders of magnitude greater than in the circum-Antarctic (Heath, 1974). The analysis of opal deposition by Miskell and others (1985) suggests that accumulation rates were much lower in the past than DeMaster's (1981) estimate of present-day inputs and output. Even taking the loss of sea floor through subduction into account, the opal accumulation on ocean floor investigated by the DSDP is much less than expected. It seems likely that in the geologic past a significant part of the opal output of the ocean was sequestered in small regions with intense upwelling, and that these remain unknown.

Recently (for example, Pokras, 1987) the recognition of particular species of diatom, such as *Ethmodiscus rex*, as productivity indicators has shown great promise toward development of an index of productivity, but a means of quantifying productivity with diatoms remains to be defined.

Ramsay (1974) had suggested that fluctuations of the $CaCO_3$ compensation depth might reflect changes in oceanic productivity, but it now seems more likely that the overriding control on calcite dissolution is changes in sea level; in any case, there are so many factors which can influence carbonate dissolution in the deep sea that the possibility of using the accumulation rates of carbonate sediments as indices of productivity has not been further explored as yet.

A recent study by Lyle and others (1988), comparing sedimentation of different biogenic components, confirmed the idea that the distribution of $CaCO_3$ is controlled by dissolution processes but found that the distribution of $C_{org}$ and opal reflect changes in productivity. They noted, however, that the distribution of fluctuations in $C_{org}$ can be traced over wide areas, whereas the fluctuations in opal-accumulation rates are highly variable.

Moody and others (1981, 1988) have pioneered the study of output of phosphorus into deep-sea sediments as an index for productivity. Although there are complications with partitioning of the output of phosphorus on shelves versus in the deep sea and with the addition of phosphorus from eolian dust transport, their studies indicate that significant increases in the intensity of upwelling and productivity are associated with the build-up of the Antarctic ice sheet and with northern hemisphere glaciation. Upwelling increased as a result of higher wind stress on the sea surface as the latitudinal thermal gradient developed.

The use of shifts in $\delta^{13}C$ as an index of global productivity is being developed, but it suffers from the fact that there is no unique cause for such shifts (Berger and Vincent, 1986; Berger and Mayer, 1987). In specific instances, such as at the Cenomanian/Turonian boundary (Scholle and Arthur, 1980; Schlanger and others, 1986) and in the middle Miocene (Vincent and Berger, 1985), the shift of $\delta^{13}C$ toward heavier values is clearly related to significantly increased organic carbon burial in the ocean in response to episodes of high productivity.

Recently Sarnthein and others (1987) have proposed a formula for estimating paleoproductivity based on the relationships between carbon-accumulation rate, water depth, and carbon-free bulk sedimentation rates calibrated with $\delta^{13}C$ in a benthic foraminifer. Using this method, they concluded that glacial productivity off northwestern Africa was three times greater than interglacial productivity.

The occurrence of barium in deep-sea sediments has been cited as having great potential as an index of paleoproductivity (Bostrom and others, 1979). It is thought that it may be a better measure of surface-water productivity than is silica, because barite is relatively insoluble, and the amount buried in the sediment is a close approximation of the supply to the sea floor. There are complications because of the introduction of particulate barite into the marine environment from terrigenous sources. Its occurrence in sediments in the Indian Ocean has been used by Schmitz (1987) to locate the belt of equatorial high productivity and to determine rates of plate motion, but a quantitative formulation of barium content as a measure of surface-water productivity has not yet been established.

A new line of investigation of ocean productivity has opened with the combined use of Cd/Ca and $\delta^{13}C$ ratios (Boyle, 1986a, 1986b) which may lead to applications in evaluating productivity; this technique can be used to evaluate (1) the relative amount of phosphorus entering the deep sea as organic particulates as a result of fixation in the surface waters and (2) the amount of phosphorus entering as preformed nutrient, that is, as dissolved phosphorus present in a water mass at the time it sinks to the deep ocean. This work has particular implications for ocean circulation and atmospheric climate and $CO_2$ content during the Pleistocene.

Recently, models of atmospheric circulation have been applied to ancient ocean configurations to attempt to identify ancient upwelling sites. Parrish (1982) and Parrish and Curtis (1982) assumed ancient atmospheric circulation systems to be analogous to that of today, so that predicted sites of upwelling in the past are located mostly on the western margins of landmasses in the subtropics. Barron (1985) has found that numerical climate models produce significantly different results for regions outside the subtropics; his "most realistic" Cretaceous simulation suggested upwelling around the margins of the Arctic Ocean. Parrish and Barron (1986) have recently summarized the state of knowledge on paleoclimates and economic deposits, including an extensive discussion of ancient upwelling systems. Unfortunately, many of the continental margin regions which would be critically important to understanding coastal upwelling are poorly known at present.

## HIATUSES

Saito and others (1974) reported that the Lamont-Doherty core library contained more than 900 pre-Quaternary cores and dredgings, attesting to the fact that hiatuses are widespread in the deep sea. After the DSDP had been operational for a number of years and a large number of stratigraphic sections sampled, it became apparent that there was paleoceanographic information content in the distribution of hiatuses in the deep sea. Moore and Heath (1977) noted a tendency for a greater proportion of older sections to be represented by hiatuses, so that for sediments older than the Cenozoic, only about 30% of the section is still present. Moore and others (1978) interpreted the increasing abundance of hiatuses in older sediments to be the result of dissolution of biogenic carbonate and opaline silica and erosion by bottom currents. It is also apparent that at certain times (middle and late Miocene, late Eocene–early Eocene, early Paleocene), hiatus generation was much more common than at other times. During the late Oligocene–early Miocene and late Paleocene–middle Eocene, hiatus abundance is less than a simple decay curve predicts. Subsequent studies have focused on the interpretation of hiatuses in terms of rates of production of bottom waters. The interpretation of hiatuses is not straightforward because (1) it is difficult to date the time of their formation, (2) dissolution alone may create hiatuses in carbonate and opaline sediments, and (3) the effect of formation and movement of ocean bottom waters on the sediments is controlled to a large extent by the local topography. Nevertheless, occurrence of hiatuses in particular locations have offered important clues to sites of bottom-water formation and flow paths in the past (Kennett and others, 1975; Kennett and Watkins, 1976;

Kennett, 1977; Ledbetter and others, 1978; Kaneps, 1979; Ledbetter, 1979, 1981; Ledbetter and Ciesielski, 1982; Johnson, 1982, 1983; Barron and Keller, 1983; Keller and Barron, 1983, 1987; Miller and Tucholke, 1983; Miller and others, 1985b, 1987a). Recently it has become possible to interpret seismic stratigraphic sections on continental margins and in the deep sea in terms of paleoceanographic events; Schlager and others (1984) suggested that hiatuses in the Late Cretaceous of the southeastern Gulf of Mexico may be the result of contour currents which entered the Gulf from the Atlantic before the Cuban arc arrived south of Florida. Pinet and Popenoe (1985), interpreting seismic profiles of the Blake Plateau, suggested that the initiation of the flow of the Gulf Stream over the Blake Plateau occurred in the early Tertiary. They were able to recognize two flow paths, one across the inner plateau along the Florida-Hatteras slope and the other across the central plateau; they attributed the shift from one path to the other to the effects of sea-level changes. Mullins and others (1987), from examination of a seismic stratigraphic succession off western Florida, suggested that an oceanographic event, that is, intensification of the Loope Current/Gulf Stream, occurred in the middle Miocene (12–15 Ma); they attributed this event to closure of the Isthmus of Panama to major ocean currents, intensifying the northward water transport of the western Atlantic.

## CIRCULATION OF THE SURFACE OCEAN

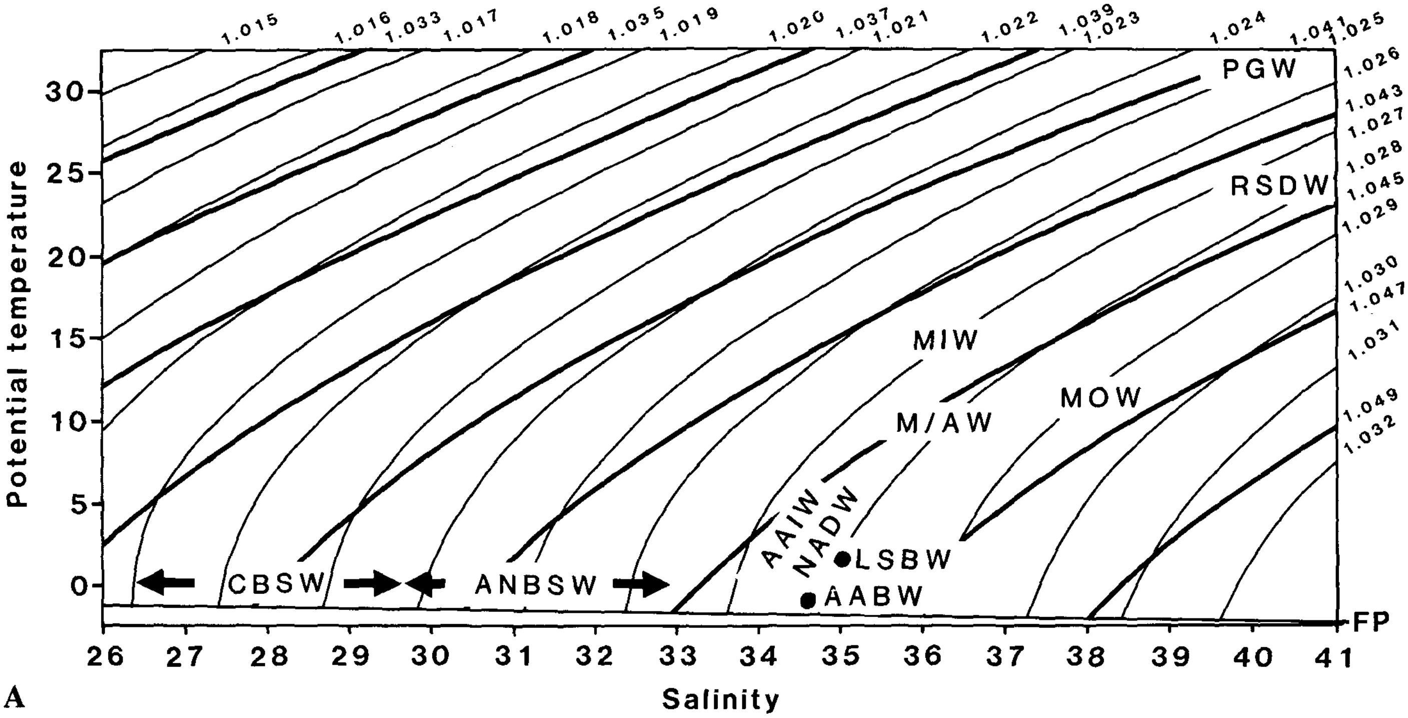
The first attempt to reconstruct ancient oceanic circulation patterns was the study of Berggren and Hollister (1974) who used the theory of wind-driven surface currents and paleobiogeographic data to postulate current flow directions in the Atlantic, western Tethys, and Caribbean from middle Triassic to present. A major flaw in their paleogeographic reconstructions was that they used modern shorelines, even for the Cretaceous when the Gulf of Mexico and Arctic Ocean were connected through the western interior of North America. Biostratigraphic schemes for the zonation of deep-sea sediments had become standardized by the early 1970s, and a number of micropaleontologists took up the new approach of using biogeographic distributions of oceanic microfossils to outline ocean-surface currents (Berggren and Hollister, 1977; Haq, 1980, 1981, 1984; Reyment, 1980; Thunell and Belyea, 1982; Boersma and Premoli-Silva, 1983; Poag and Miller, 1986; Berggren and Olsson, 1986; Boersma and others, 1987; Morley and others, 1987). Starting from the premise that assemblages of planktonic microfossils record surface-water temperatures (Schott, 1935), Imbrie and Kipp (1971) developed a quantitative method to relate a particular assemblage of planktonic foraminifera to oceanographic parameters such as sea-surface temperature through a "transfer function." A refinement of this technique (Kipp, 1976) has been used extensively to develop maps of distribution of temperature and salinity at the sea surface and at 100-m depth for summer and winter year at 18,000 Ma (McIntyre and others, 1976; CLIMAP Project Members, 1981). Ruddiman and McIntyre (1976) have shown that in the North Atlantic, the polar front migrates south to about 45°N during glacials, retreating very rapidly to the far north during interglacials.

## THE DEVELOPING PALEOCEANOGRAPHIC SYNTHESIS: THREE-DIMENSIONAL CIRCULATION OF THE OCEAN

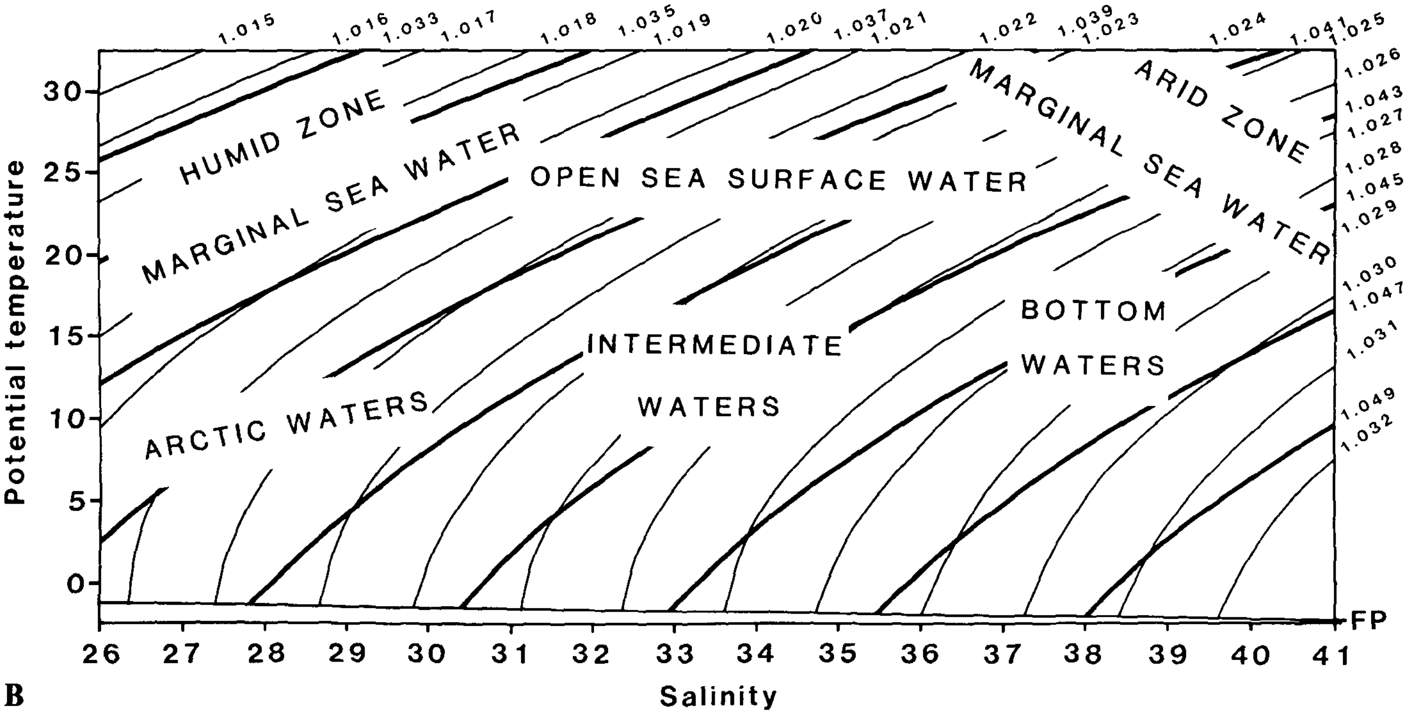
The information from oxygen isotopes, distribution of carbonate sediments reflecting fluctuations of the CCD, occurrence of $C_{org}$-rich sediments, occurrence of productivity indicators, location of hiatuses, and the paleobiogeography of oceanic plankton and benthos is currently being coupled with theoretical and model studies to provide a synthesis of the past behavior of the ocean. Rather than location of surface currents *per se*, it is the location of oceanic fronts and sites of intermediate- and deep-water production which are critical to the geologic history of the surface of the planet. Fronts mark significant climatic boundaries, and oceanic fronts, where waters of different temperature and salinity converge, separate major circulation systems which are, in effect, biogeographic provinces. Oceanic fronts are also marked by major changes in sediment type on the sea floor. The Antarctic Divergence, which closely surrounds the Antarctic continent, mixes nutrient-rich water upward, resulting in the high productivity belt of the Southern Ocean. South of the divergence, the sea floor is covered with ice-rafted debris; north of the divergence lies the circum-Antarctic band of diatomaceous sediment. Convergences mark absolute limits of transport of floating material. The Antarctic Convergence, which is about 50°S except off South America, where it lies more than 60°S, marks the northerly limit of drift ice and hence of the occurrence of ice-rafted debris on the ocean floor. Because of the complex distribution of land and sea in the Northern Hemisphere, the Antarctic Divergence and Convergence have no well-defined counterparts in the north. Using the occurrence of ice-rafted debris in Quaternary sediments, however, Lisitzin (1974) was able to estimate that during the glacials of the Pleistocene, the southern polar front extended to 40°S, and the northern polar front reached 40°N in the North Atlantic and Pacific. Subsequent studies by CLIMAP (1981), based on the analysis of fossil plankton, have confirmed Lisitzin's work. The subtropical convergences, which lie at about 30°N and 30°S, separate the westerly drifts of the higher latitudes from the easterly currents of the equatorial region, and mark the regions of lowest productivity in the ocean. The Equatorial Convergence, which is well developed in the Pacific and less well in the Atlantic, lies close to the thermal equator of the ocean; the Equatorial divergences on either side of it delineate the zone of open-ocean equatorial upwelling, with high productivity.

One of the most effective methods of learning about changes in atmospheric and oceanic circulation in the past is to trace the position of oceanic fronts back in time. This has been done specifically by Ciesielski and Grinstead (1986) and Hodell and Kennett (1985). Wind-driven upwelling systems can offer valuable insights into the strength and direction of winds in the past (Sarnthein and others, 1981, 1982; Rea and others, 1985; Pisias and Rea, 1988).

Studies of production of intermediate and deep waters in the present-day ocean have shown that they result from a complex interaction of factors. For water to sink and become intermediate or deep water in the ocean, it must be modified by interaction with the atmosphere so that its density is increased; this can be either (1) through evaporation, which increases the salinity of the surface water and hence its density; (2) through the formation of sea ice, which, because sea ice is less saline than the sea water from which it forms, expels salt into the surrounding surface water, increasing its density; and (3) through chilling of the water surface by conduction of heat to the atmosphere or by loss of heat due to evaporation. The relations between salinity, temperature, pressure, and density are defined by the equation of state of sea water and are illustrated in Figure 4. As dense water forms at the surface, it sinks, and in doing so, it entrains the surrounding water, thereby decreasing its density contrast. Peterson (1979) found that the source of dense water which produces a downwelling plume having the highest buoyancy flux will overcome all other sources and fill the depths of the ocean. The buoyancy flux is the product of the volume flux of the plume and its density excess over the surrounding water. As it descends, the density of the water is increased by the pressure of the overlying water; cold water, especially colder than 7 °C, is more compressible than warm water, and thus gains density faster. Figure 4 shows density at the surface and at a pressure of 400 bars, corresponding



A



B

Figure 4. A. Relation of potential temperature, salinity, and density, with fields occupied by some important modern water masses. Potential temperature is the temperature the water would have if raised adiabatically to the surface, that is, it would eliminate the increase in temperature of the water due to compression. Thin lines are surface densities (1.105–1.032) calculated for the sea surface beneath one atmosphere of pressure; thick lines are ocean-bottom densities (1.033–1.049) calculated for a pressure of 400 bars (approximately equal to 4,000 m depth. CBSW = Canadian Basin (Arctic) Surface Water; ANBSW = Amundsen-Nansen Basin (Arctic) Surface Water; AAIW = Antarctic Intermediate Water; NADW = North Atlantic Deep Water; AABW = Antarctic Bottom Water; LSBW = Labrador Sea Bottom Water; M/AW = Mediterranean/Atlantic Mixed Water; MIW = Mediterranean Inflow; MOW = Mediterranean Outflow; RSDW = Red Sea Deep Water; PGW = Persian Gulf Water.

B. Relation of potential temperature, salinity, and density, with approximate fields for some Cretaceous water masses.

approximately to a depth of 4,000 m, near the average depth of the ocean (3,730 m); the densities were calculated from the recent high-pressure equation of state of sea water and the new International Equation of State of Sea Water (Millero and others, 1980; Millero and Poisson, 1981). The deep circulation of today's ocean is driven by thermohaline circulation, that is, the major sources of bottom water obtain their excess density by virtue of their coldness. As T. C. Chamberlin (1906) speculated, however, the ocean may in the past have had a halothermal circulation in which the deep-water sources obtained their excess density by increase of salinity.

The formation of deep waters in the present ocean illustrates the complex interplay of factors. The salinity, temperature, and density of many of the world's major water masses are indicated in Figure 4. The major water masses of the Atlantic are shown in Figure 5A. The Mediterranean Sea plays a special role in global circulation today; almost isolated from the world ocean and located in a region where the annual evaporation exceeds precipitation by about 1 m, the eastern Mediterranean is the site of formation of the densest water in the ocean. The salinity of water flowing into the Mediterranean through Gibraltar is increased to over 39.2 by the high evaporation in the eastern Mediterranean (Miller and others, 1970). As a result, the density increases to as much as 1.0292 g/cc (Worthington and Wright, 1970) in the winter. This dense water sinks in the eastern Mediterranean to flow back into the western basins and then into the Atlantic over the Gibraltar sill. Enroute it mixes with a larger volume flux of less saline but cooler deep water formed in the winter in the northwestern Mediterranean (Medoc Group, 1970; Bryden and Stommel, 1982, 1984). Although the Mediterranean outflow, with a density as high as 1.0291 g/cc, is still one of the densest waters found in the ocean (see Fig. 4), the volume of outflow is not sufficient to give it a buoyancy flux adequate to make it become the bottom water of the North Atlantic. Instead, the Mediterranean outflow becomes less dense through entrainment of the surrounding waters so that at a depth of about 1 km it begins to spread on the 1.02751 g/cc (calculated from Fuglister, 1960), isopycnal (equal-density) surface. Advection of water in the ocean takes place predominantly on the isopycnal surfaces, and the 1.02751 isopycnal reaches the surface of the North Atlantic near the entrance to the Norwegian-Greenland Sea (Reid, 1979, 1981; see also Gorshkov, 1977). As the North Atlantic water, which is anomalously warm and saline because of its admixture of Mediterranean outflow water, enters the Norwegian-Greenland Sea near the pycnocline depth between Greenland and Scotland, it mixes with water of similar density and sinks to become the major source of North Atlantic Deep Water (NADW) (Peterson and Rooth, 1976; Swift and others, 1980). As can be seen in Figure 4, when two oceanic waters of equal density but different salinity and temperature mix, the resulting mixture is more dense and sinks. Other sources, notably the Labrador Sea, supplement the Norwegian-Greenland sea source, and the NADW fills the intermediate levels of the Atlantic with water which is nutrient-poor but oxygen-rich (Bainbridge, no date). NADW, diluted by entrainment of other waters it has encountered on the way south, returns to the surface at the Antarctic Divergence, where it is again chilled. The water which flows south from the divergence becomes involved in sea-ice formation in the Weddell Sea; its salinity is increased and it sinks as nutrient-rich, silica-rich, but oxygen-poor Antarctic Bottom Water, which underflows NADW to the north. The Coriolis effect, deflecting the flow to the left in the Southern Hemisphere, and the barrier of Walvis Ridge in the eastern South Atlantic cause AABW to flow northward through the western basins of the South Atlantic. Water which flows north from the Antarctic divergence escapes the sea-ice area, but it is both chilled and diluted by mixing. It sinks beneath the surface but is relatively light, and it overrides the NADW as it flows north as nutrient-rich but silica- and oxygen-poor Antarctic Intermediate Water (AAIW). Thus, each of the

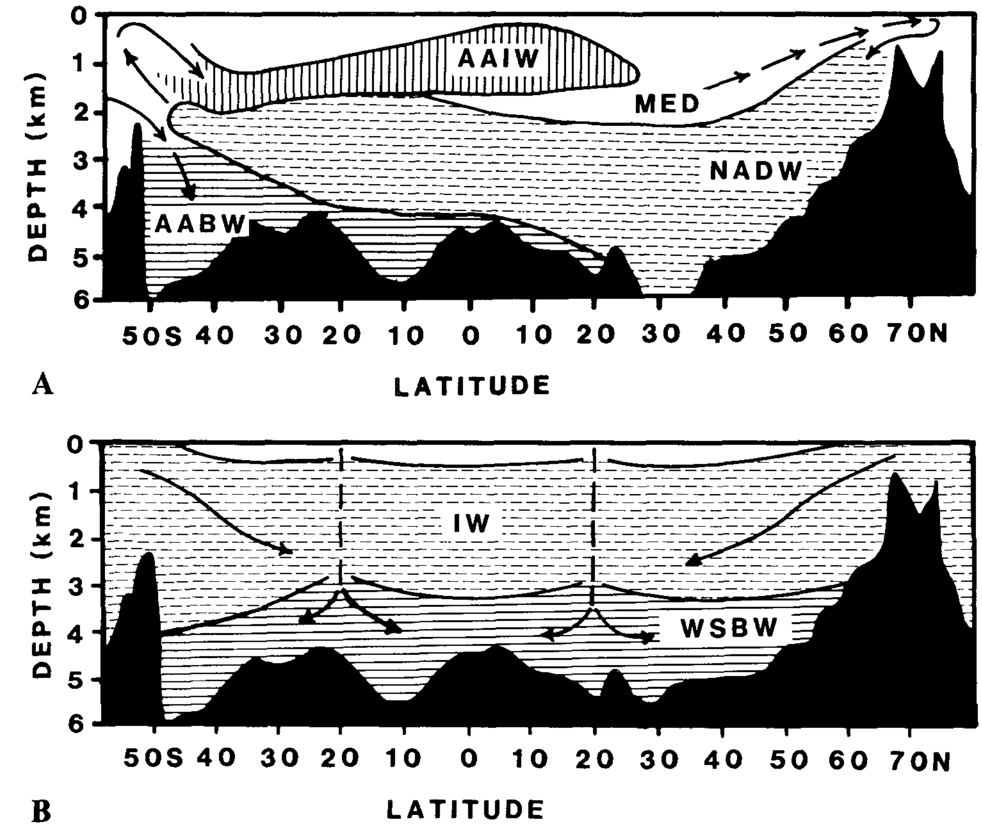



**Figure 5. A. Schematic diagram of the relation of some important water masses in the modern Atlantic. AABW = Antarctic Bottom Water; AAIW = Antarctic Intermediate Water; MED = Mediterranean Outflow mixing into Atlantic water at 1.5 km depth; NADW = North Atlantic Deep Water.**

**B. Schematic diagram of possible water masses in the Cretaceous. IW = Intermediate Water; WSBW = Warm Saline Bottom Water.**

major water masses has a different chemistry. The upwelling systems of the South Atlantic margin draw mostly from AAIW. AABW with its incorporated component of NADW also flows into the basins of the Indian and Pacific Oceans (Reid and Lynn, 1971; see also Gorshkov, 1974). Concentrations of nutrients, bicarbonate ion, and dissolved oxygen all indicate that the flow of bottom waters must be from south to north in the Indian and Pacific Oceans. There is no possibility of a northerly deep-water source for the Indian Ocean, but the warm saline outflows from the Red Sea–Gulf of Oman and Persian Gulf form a major intermediate water mass in the Arabian Sea, as is evident in Gorshkov (1977). The Bay of Bengal is diluted by the large fresh-water outflow from the Ganges-Brahmaputra river complex, and under these conditions, it could never become a deep- or intermediate-water source. There is presently no source of deep water in the North Pacific, but the potential of a source area certainly exists in the complex geography of the Bering Sea and Sea of Okhotsk. Sea-level changes could readily increase the isolation of the marginal seas and make the waters in them more susceptible to modification by the atmosphere.

Streeter (1973) and Schnitker (1974), in the pioneering studies of the differences in interglacial and glacial benthonic foraminifera, had noted that the distribution of certain modern deep-sea benthonic foraminifera in the North Atlantic coincides with the present distribution of bottom waters from different sources. Documentation of the changes in glacial-interglacial distribution of the benthonic foraminifera initially led to the conclusion that deep waters were warmer in glacial times and cooler in the interglacials, although the absolute magnitude of the changes could only be estimated. Schnitker (1974) assumed that this was the result of more limited distribution of Antarctic bottom water. The situation was clarified when Duplessy and others (1975) compared the oxygen-isotope record from a core from the Norwegian Sea with the record from Shackleton and Opdyke's (1973) west equatorial Pacific core. The amplitude of glacial-interglacial changes in the west equatorial Pacific core was 1.1 per mil greater than in the core from the Norwegian Sea. It seemed unreasonable to suggest that the Pacific bottom waters would have been 4.5° cooler during the glacial than they are today (1°), and more reasonable to assume that Norwegian Sea bottom waters were 4.5° warmer during the glacial than they are today. Streeter and Shackleton (1979) suggested that this effect was a result of cessation of North Atlantic Deep Water (NADW) production as a result of ice cover of the Norwegian-Greenland sea. Duplessy and others (1980) then compared the oxygen-isotope record from a core in the northeast Atlantic with that from a core taken beneath Circumpolar Deep Water in the southern Indian Ocean, where they assumed that the bottom-water temperatures had remained constant. The difference in glacial-interglacial amplitudes was 0.3 per mil, and from this they concluded that the bottom waters of the North Atlantic were 1.3° cooler during the glacial than they are at present; they also concluded that the North Atlantic has been a source of deep water for the past 75,000 yr.

In 1982 Hester and Boyle announced the discovery that the shells of benthic foraminifera deposited in oxygen-depleted water are enriched in cadmium. Boyle and Keigwin (1982) applied this discovery to paleoceanographic interpretation; they examined the Cd/Ca ratio, $\delta^{18}O$, and $\delta^{13}C$ in benthic foraminifera from a Pleistocene core taken northwest of the Azores at 42°N, 32°W at a depth of 3.2 km, the level of the NADW core. They found that the Cd/Ca ratio is always lower than it is in the Pacific, indicating that at no time had the flux of low-nutrient water ceased. The highest Cd/Ca ratios occur during the extreme glacials, and the lowest ratios, in the interglacials. The Cd concentration is inversely proportional to the flow rate; they concluded that during glacials the flux from the nutrient-depleted NADW sources was diminished in importance relative to that from southern deep and intermediate sources, but that the flux of NADW never ceased. Boyle and Keigwin (1987) interpreted Cd/Ca and $^{13}C/^{12}C$ data from shells of benthic foraminifera in the North Atlantic as indicating that there was a marked depletion of nutrients in intermediate water masses in association with the reduced NADW flux during the last glacial. They suggested that cold high-latitude sea-surface temperatures enhanced intermediate-water formation at the expense of deep-water formation. The complex history of changing surface conditions and bottom-water sources during the Pleistocene has been recently reviewed by Duplessy and Shackleton (1985), Corliss and others (1986), Williams and Fillon (1986), Ruddiman and others (1986), Ruddiman (1987), and Kellogg (1987). An important driving factor in altering thermohaline circulation in the Atlantic may be the nature of the Mediterranean outflow water (MOW); Zahn and Sarnthein (1987) concluded from a study of stable isotopes in benthic foraminifera near the Strait of Gibraltar that the MOW outflow decreased significantly during the deglaciations, and that its temperature was decreased by as much as 5° during glacial time. They suggest that this indicates that the importance of saline MOW to the formation of NADW may be less than has been assumed, but the time of reduced MOW flux cited by Zahn and Sarnthein corresponds to a time glacial meltwater flux to the North Atlantic was a maximum and when production of NADW was also reduced (Berger and Vincent, 1986; Broecker and others, 1988). The longer-term possible change in Mediterranean/Atlantic relations has been discussed by Thunell and others (1987).

Berger (1968), Moore and others (1977), and Broecker and others (1985) have suggested that oceanic circulation during glacials may be fundamentally different from that of today. It has been speculated that the Norwegian-Greenland Sea may become covered by permanent sea ice during the glacials (Broecker, 1975; Hughes and others, 1977; Lindstrom and MacAyeal, 1986), and the formation of NADW may be reduced or cease. Greatly increased sea-ice formation in the north Pacific may transform that region into a source area for oceanic deep waters. This would result in a complete reversal of gradients of nutrients, dissolved oxygen, and alkalinity in the deep sea. At present nutrients and bicarbonate have their lowest concentrations in the North Atlantic and highest concentrations in the North Pacific, but with a transfer of the Northern Hemisphere sites of deep-water formation to the Pacific and cessation of deep-water formation in the North Atlantic, the highest concentrations of nutrients and bicarbonate would be in the North Atlantic, and the lowest concentrations, in the North Pacific. Similarly, the related oxygen concentration would reverse, with the North Atlantic waters becoming depleted and the North Pacific waters oxygen-rich. Berger and Vincent (1986) have suggested that production of NADW may have been shut off during the glacial-Holocene transition (Dryas), which would again cause redistribution of biolimited materials in the oceanic deep waters, and Berger (1987) suggested that there was indeed deep-water production in the North Pacific during the glacial and deglaciation.

Cenozoic paleoceanography of the North Atlantic has been reviewed by Tucholke and Mountain (1986), of the South Atlantic by Hsü and others (1984) and Williams and others (1985), and of the Southern Ocean by Wise and others (1985).

The most significant changes in surface circulation during the Pliocene occurred with the final closure of the Central American Isthmus and resulting complete separation of tropical and subtropical circulation between the Atlantic and Pacific (Emiliani and others, 1972; Malfait and Dinkleman, 1972; van Andel, 1976; Keigwin, 1982a; Gartner and others, 1983; Haq, 1984). The final separation of Atlantic and Pacific appears to have occurred about 2.5 Ma, and the last interchange of water was probably between the Pacific and the Gulf of Mexico, across Isthmus of Tehuantepec, bypassing the Caribbean (Gartner and others, 1987). The salinity contrasts between the Arctic Ocean, where the surface waters of the Canadian Basin are diluted below a salinity of 30 to as little as 27.5 by the plume of the Mackenzie River (Gorshkov, 1980), and the salty (34.92) Atlantic and less salty (34.60) Pacific (Worthington, 1981) resulting from large-scale atmospheric water transport across the Central American Isthmus (Weyl, 1968) are important underlying factors promoting the development of the modern circulation system with a strong NADW source. The Pliocene ocean clearly differed from the modern ocean in one very significant way: upwelling was greatly enhanced in the eastern South Atlantic (Dean and others, 1984), about the Antarctic (Brewster, 1980), and probably elsewhere. The cause of this intensification of upwelling has yet to be determined. Changes in the thermohaline circulation during the Pliocene have been discussed by Weissert and Oberhänsli (1985) and Murray and others (1986).

During the late Miocene (Messinian), a special event occurred—the "salinity crisis" which occurred as the Mediterranean became desiccated and extracted about $1 \times 10^6$ $km^3$ of salts from the world ocean (Ryan, Hsü, and others, 1973; Cita, 1982; Cita and McKenzie, 1986; Mueller and Hsü, 1987). The fall of sea level associated with build-up of the Antarctic ice cap was a contributing factor in causing the isolation of the Mediterranean. The Red Sea had become a salt-extraction basin in the earlier Miocene, and these two salt extractions may have lowered the salinity of the world ocean by as much as 7% (Southam and Hay, 1981).

Paleoceanography of the Miocene ocean on a global scale was the subject of GSA Memoir 163 (Kennett, 1985). The interpretations are based largely on analysis of oceanic microfossils and stable isotopes. It was during the Miocene that the deep-water passage from the Pacific into the Indian Ocean became restricted as Australia–New Guinea collided with the Indonesian Arc. Kennett and others (1985) have determined that the Pacific Equatorial Undercurrent developed during the middle Miocene, and they concluded that this was related to the closing of the Indonesian Seaway, which also increased the intensity of gyral circulation in the Pacific. Intensification of the Countercurrent is associated with increased productivity and output of biogenic silica. The middle Miocene was also the time of re-establishment of a strong North Atlantic Deep Water source after it has weakened in the early Miocene (Schnitker, 1980, 1986). The Miocene was a time of increasing latitudinal temperature gradients and decrease in bottom-water temperature. These both reflect the increasing isolation of the Antarctic continent with the development of the West Wind Drift and the expansion of the ice cap. The sharp middle Miocene cooling of oceanic deep waters is assumed to be in response to an increasing buoyancy flux from a source in the Southern Ocean or along the Antarctic margin. Woodruff (1985) has shown how benthic foraminifera can be used to recognize Miocene water masses in the Pacific and to estimate their relative oxygen concentrations.

The Oligocene was the age during which the ocean took on its modern aspect, when the deep waters became significantly cooler than the surface waters (Corliss and Keigwin, 1986). During some short episodes, the bottom waters may have cooled to temperatures comparable to those of today (Miller and Fairbanks, 1985). Ice may have been present on the Antarctic continent by the early Oligocene (Wise and others, 1987), and deep-water production in the surrounding seas may have been restricted by fresh-water runoff. It is not clear what other deep-water sources may have existed during the Oligocene.

In the Eocene, Paleocene, and Cretaceous, the ocean was fundamentally different from that of today (Kraus and others, 1978; Barron and others, 1981; Hay, 1983; Oberhänsli and Hsü, 1986; Roth, 1986; Barron, 1987); the circulation appears to have been halothermal rather than thermohaline. Brass and others (1982a, 1982b) have suggested that warm saline waters that formed in marginal seas in the arid zones filled the ocean basins. Barrera and others (1987) have presented evidence for Antarctic shallow-water temperatures and concluded that the polar regions were sites of deep-water formation for at least the Pacific Basin during the Late Cretaceous and Paleocene, although they acknowledged that the Atlantic basins may have had local sources. Miller and others (1987b) found that the Antarctic region was the source of bottom water for the Cape Basin during the late Paleocene, but that the supply was reduced or eliminated by the early Eocene. They also noted that the nutrient-depleted Antarctic waters failed to enter the western basins of the South Atlantic, probably being blocked by the Islas Orcadas Rise.

As noted above, warm saline waters contain only about one-half the dissolved oxygen of young modern deep waters; hence, intermediate and deep waters that formed in arid marginal or shelf seas would have been much more likely to become anoxic. The great variability in the CCD between different ocean basins during the Cretaceous (Fig. 2) is readily understood if the ocean depths were being filled from numerous, relatively small and weak, locally warm, saline-water sources. Based on the oxygen-isotopic evidence from inoceramids reported by Barron and others (1982), Hay (1983) suggested that in the Late Cretaceous South Atlantic, surface-water temperatures were similar to those of today, but bottom waters were about 10–12 °C, and intermediate waters were about 8 °C (see Fig. 4B). This implies that the intermediate waters were produced at higher latitudes than were the deep waters (Figure 5B shows a general scheme for halothermal circulation). Bottom waters were formed in the mid-latitude arid regions; the source areas would be in shallow seas or nearly isolated basins on the eastern margins of the oceans. In each hemisphere, these waters would flow poleward, following the eastern basins or as contour currents along the eastern margin of each ocean basin. These warm saline waters would return to the surface in the polar regions. On reaching the surface, the waters would be chilled by conductive cooling and, more importantly, by evaporation in the relatively warm polar regions. They could then sink and flow equatorward as intermediate waters; during their equatorward flow, they would be concentrated as contour currents on the western margins of the oceans. It is interesting to note that in the Late Cretaceous the over-all salinity of the ocean was about the same as it is today; but in the middle Cretaceous, after the very large salt extraction in the opening South Atlantic, over-all salinities may have been 20% higher (Southam and Hay, 1981), so that the average salinity would have been about 41.6%. This means that the major water masses would fall into a temperature/salinity field near and beyond the right side of Figures 4A and 4B.

## THE FUTURE

The most exciting prospects for the future of paleoceanographic studies lie in attempting to understand how plate tectonics, atmospheric and oceanic circulation, and biologic evolution of marine animals and plants are interrelated. Do plate-tectonic cycles drive the climate and ocean systems on long-term time scales as proposed by Worsley and others (1986)? It is particularly important to understand why the Earth has different climatic states and how the ocean has responded in the past. What was the nature of oceanic circulation in warm geological periods when the deep ocean was filled with relatively warm water? Because most of the older ocean floor has been subducted, the data base for the oceans proper is very incomplete. To understand the oceans of the Eocene, Paleocene, Mesozoic, and Paleozoic, it will be necessary to understand the relation of the record preserved in marginal seas and on the continents. The separation between those working in the deep sea and those working on epeiric seas must end. To understand how the Earth has operated in the past, we make use of all of the available data in a global context. Modeling oceanographic circulation for the earlier Paleogene and Mesozoic will undoubtedly generate many new ideas. Before modeling can be realistic enough to provide output that can be tested against geological observations, however, the shape of the surface of the Earth in the past must be reconstructed as accurately as possible. Plate-tectonic reconstructions show the positions of the major continental blocks and bathymetry of the ocean floor based on simple subsidence curves. The present reconstructions are, unfortunately, gross oversimplifications because they do not take into account the many large terranes which have moved independently of the major continental blocks and because they do not include the residual depth anomalies (areas of the ocean floor which are more than 500 m shallower or deeper than predicted by the thermal decay curve) which form a significant part of the sea floor. Preliminary reconstructions by Wilson (1987) and Wilson and others (in press) reveal that drifting terranes obstruct many passages which appear to be clear on maps that show only the major continental blocks. Coupled atmospheric and oceanic models run on such a realistic paleogeographic base are necessary to achieve the better resolution needed before the output can be used for site-specific geologic predictions.

Much of paleoceanographic interpretation by geologists in the past has reflected a lack of familiarity of the principles of physical oceanography as they are understood today, although many recent works have begun to bridge the gap and integrate the two fields. The languages of the two fields are sufficiently different that most physical oceanographic literature is not readily accessible to geologists. Fortunately several general works have appeared in recent years which summarize modern concepts in physical oceanography and provide the background necessary to understand developments in that field: Harvey (1976), which provides a general introduction; Warren and Wunsch (1981), which contains a series of summary papers; Pickard and Emery (1982); Pond and Pickard (1983), which includes a glossary of commonly used symbols and a review of the mathematical notation commonly used; and Tolmazin (1985), which has a good discussion of flow resulting from pressure gradients especially applicable to studies of marginal seas.

The new field of paleoceanography was born three decades ago, but it is still very much in its infancy. Only recently have we come to realize that the oceans of the past may have operated in modes very different from that of today. The new proxy indicators being used to investigate distribution of properties and materials in the ancient oceans will make it possible to understand major processes at a level which could not have been anticipated when the field was born.

## SUMMARY AND CONCLUSIONS

Paleoceanography, a field born little more than 30 yr ago, has become a major branch of geology. It is an interdisciplinary field making use of stratigraphy, geochemistry (particularly stable isotopes), geophysics (particularly reflection seismics and paleomagnetics), paleontology, sedimentology, and other aspects of geology and relating these to progress in physical, chemical, and biological oceanography. Coring of deep-sea sediments began after World War II and developed into the ocean drilling and coring programs, notably the Deep Sea Drilling Project and the Ocean Drilling Program. These have provided a rich sample library and data base, permitting the development of a global science. Among the most significant advances in the field are the following.

1. Development of a refined stratigraphic base incorporating micropaleontology, magnetostratigraphy, and isotope stratigraphy so that resolution of 100,000 yr or less is now possible for parts of the Cenozoic.

2. Documentation and interpretation of the thermal history of the ocean, from the equable climates of the Mesozoic through the development of the pole-to-equator temperature gradient during the course of the Cenozoic to the present condition of alternating glacials and interglacials.

3. Documentation of fluctuations of the calcite compensation depth and development of an understanding of the complex interrelation of factors which control the oceans' buffering capacity and how these are related to changes in the concentration of $CO_2$ in the atmosphere.

4. Investigation of the accumulation of organic carbon-rich sediments in the deep sea, and their implications for changes in surface productivity and the nature of deep-ocean circulation.

5. The discovery that the deep sea is not the site of continuous sedimentation as had been assumed by earlier geologists, but that the deep-sea record is riddled by hiatuses. The hiatuses testify to the development of strong bottom currents and corrosive deep waters, and they offer valuable clues to ancient deep circulation.

6. Documentation of changes in paleobiogeography of ocean plankton which indicate changes in surface currents as the Earth changed from a planet with latitudinal oceanic circulation in the subtropics (Tethys) to one with meridional oceans connected in the subpolar region.

7. Insight into alternative modes of circulation of the ocean in which the deep circulation was not thermohaline, that is, driven by cold waters sinking in the subpolar or polar regions, but may have been halothermal, driven by dense plumes of warm saline waters formed in marginal seas and shelves of the subtropics.

8. The field of paleoceanography is burgeoning with new ideas and offering a view of the history of the Earth's surface in a global context. It offers a bright future for documentation and modeling of global oceanic processes and the opportunity to relate marine geologic observations to atmospheric and oceanic climate and circulation models, and relation of atmospheric and ocean climate studies of the deep circulation.

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