38. SOUTH ATLANTIC CENOZOIC PALEOCEANOGRAPHY[1]

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ABSTRACT

Several suites of undisturbed cores obtained by continuous hydraulic piston coring provided material for the investigation of Cenozoic paleoceanography. The faunal record is complete, except for the early Eocene. The isotope records have gaps in the early Eocene and in the early and middle Miocene. Hundreds of samples were studied and thousands of analyses were carried out. Detailed results and conclusions are presented in several separate chapters of this volume. This report is a synthesis.

Studies of calcite dissolution in pelagic sediments indicate that calcite compensation depth (CCD) and lysocline underwent two types of changes during the Cenozoic. The first-order changes, which have a periodicity of some 10^7 yr., have amplitudes (depth difference from the medium) of more than 1,000 m. The Eocene and Miocene were epochs of high CCD. The deepening of the CCD in the late Eocene and early Oligocene was stepwise, but it was more abrupt in the early Pliocene. We attribute the higher CCD in the Eocene and Miocene to a relatively low production rate of calcareous plankton in the open oceans; during those epochs many of the nutrients had been consumed by phosphate deposition and by siliceous plankton. The second-order changes, which have a periodicity of 10^4 or 10^5 yr., represent CCD variations in response to changes from interglacial to glacial paleoenvironments during the Pliocene and Quaternary; the amplitude was a few hundred meters only. Evidence suggests that increased dissolution due to the more active bottom waters during the glacial stages in the Atlantic far overshadowed the effect of possible increases in fertility.

The oxygen-isotope record shows an overall increase of $\delta^{18}O$ values since the middle Eocene, with a moderate reversal of the general trend during the late Oligocene and early Miocene. Whereas the early Paleogene trend of the Atlantic middle-latitude sites closely simulated that of the southern oceans, vertical and latitudinal gradients began to develop in the Oligocene and continued to steepen with time, so that the present thermal structure of the ocean waters above our sites is similar to that of the equatorial Pacific. The cause of the increased $\delta^{18}O$ values should have been related to increased ice volume on Earth and to temperature declines. However, the relative importance of the two different factors at any given time cannot be ascertained. We favor the hypothesis that the Antarctic Ice Cap started to form in the late Eocene and suggest that the early Oligocene ice volume was larger than it is today. The middle Miocene oxygen shift has also been registered at one of our sites, but its magnitude is smaller, and the corresponding environmental changes should have been less impressive, than those in the late Eocene and early Oligocene.

The carbon-isotope record shows a parallel trend between the planktonic and benthic foraminifers since the beginning of the Oligocene and a divergent trend as well as a steeper gradient during the middle Eocene. Peak values of $\delta^{13}C$ are found in the benthic foraminifers of Eocene and Miocene age. The Eocene peak may have resulted from decreases in terrestrial biomass due to the destruction of forests by advancing glaciers. However, the Miocene peak was probably a consequence of preferential sedimentation and the burial of organic carbon relative to carbonate carbon. Abrupt changes in carbon-isotope values have also been noted: the Oligocene carbon shift is parallel to the oxygen shift, whereas the Pliocene–Quaternary short-term carbon trends are opposite to the oxygen trends, with lighter carbon becoming enriched during glacial stages. We interpret those changes as manifestations of different responses of the terrestrial biomass to changes in climate and sea level. A minute late Miocene carbon shift has been found in our record; the Atlantic was apparently less affected by the reorganization of bottom-water circulation 6 Ma than the Pacific and Indian oceans, which display more significant changes in the isotope compositions of dissolved carbonates.

Studies of fauna from sediments dated by precision stratigraphy indicate that the psychrosphere and the thermohaline circulation of the Atlantic started in the late Eocene, with precursor events in the middle Eocene. Faunal evidence suggests a northern source for the cold bottom water in regions east of the Mid-Atlantic Ridge.

INTRODUCTION

The discovery of Miocene red clays in the South Atlantic was one of the surprises of DSDP Leg 3. The inexperienced shipboard staff, which included the first author of this chapter, was unfamiliar with marine geochemistry. Calcite dissolution was assumed to have a simple relation to paleodepth, and the red clays were interpreted as deposits on subsided ridge crest at times when seafloor spreading slowed down or stopped altogether (Hsü and Andrews, 1970). Shortly after the publication of the cruise report, it was pointed out that the calcite compensation depth (CCD) could not have remained constant through geologic time (Hay, 1970). With the accumulation of deep sea data, it became also clear that the rate of seafloor spreading should have been more or less linear since the late Mesozoic (LaBrecque et al., 1977). The facies variations of the Atlantic pelagic sediments thus became the basis for interpreting the ups and downs of the calcite compensation depth, and the abundance of middle Miocene red clays was cited as evidence of an usually high CCD during that epoch.

[1] Hsü, K. J., LaBrecque, J. L., et al., *Init. Repts. DSDP,* 73: Washington (U.S. Govt. Printing Office).

With the observation that the newly formed ridge crest is 2600 m deep almost everywhere and that the seafloor subsides as an oceanic lithospheric plate moves away from a spreading center (Sclater et al., 1971), the paleodepth at any part of the ocean could be computed on the basis of the backtracking method. The wealth of the deep sea data permitted the reconstruction of the fluctuation of the "carbonate line" on the ocean bottom (cf. snowline in the mountains) during the Cenozoic (e.g., Berger and Winterer, 1974; Van Andel et al., 1977). The Atlantic records suggest that the CCD was deep (at a depth of 4.5–5 km) during the Oligocene but rose to an anomalously shallow depth (less than 3.5 km) during the middle Miocene before plunging to a depth of greater than 4.5 km again in the early Pliocene (see Van Andel et al., 1977). The cause of anomalies in the CCD, however, has remained a mystery.

In 1975, at the start of IPOD, the first author proposed drilling a transect of holes to study the middle Miocene CCD crisis in the Atlantic. The site proposed for drilling was the Mid-Atlantic Ridge, and the holes were to be so positioned as to obtain Miocene sediments deposited at relatively shallow paleodepths (< 3000 m), because it was realized that the high CCD during the Miocene should have dissolved away calcareous fossils at deeper sites, eliminating both stratigraphic and paleoceanographic information. The selection of the exact locations required a reliable record of magnetic anomalies, so we decided to return to 30° S latitude, where the magnetic signatures are clear and the weather is suitable for drilling operations throughout the year. Since the west flank of the ridge was drilled during Leg 3, the new transect was designed for the east flank, to test the symmetry (with respect to the ridge axis) of paleoceanographic processes.

While the planning for the cruise progressed, the hydraulic piston coring device was invented; in 1979 it was tested successfully. The new device made it possible to determine a precision stratigraphy. With undisturbed and nearly complete recovery, we could plan to study short-term (10^4–10^5 yr.) variations as well as to obtain a long continuous record. The Mid-Atlantic Ridge transect was eventually drilled during Leg 73, one of five legs undertaken in 1980 to investigate the paleoenvironments of the South Atlantic.

With 5 years of lead time for planning and the assistance of the hydraulic piston corer (with which we obtained several excellent suites of cores), we had a successful cruise. Our methodology was straightforward: magnetostratigraphy and biostratigraphy provided the time framework. Sampling density depended upon sedimentation rate and the nature of the problems involved, and we were able to study variations of the order of 10^3 or 10^4 yr. during times of paleoceanographic crisis. Analyses of bulk composition (particularly the insoluble residue content), grain size, fossil preservation, and the percentage of benthic microfauna provided quantitative expressions of calcite dissolution. Light stable-isotope data served as a basis for inferring the temperatures, chemistry, and circulation patterns of past oceans. Studies of flora and fauna provided information on paleoecology. The results and conclusions are described in detail in several chapters of this volume (e.g., Weissert et al., McKenzie, et al., Poore and Matthews; Oberhänsli et al.; He, et al.; Parker, et al.; and see also the sections of the site chapters written by Wright and by Hsü). This report is an attempt to integrate the details and to provide an overall view of the Cenozoic paleoceanographic changes.

CALCITE DISSOLUTION

Definition

The calcite compensation depth is the level at which the rate of supply of calcareous tests is approximately equal to the rate of dissolution (Bramlette, 1961). If all the calcium carbonate descended from an overlying water column is dissolved, only the insoluble residue is deposited. Therefore the calcite compensation surface should mark the upper limit of the red clay sedimentation on the ocean bottom at that time. However, a sample of pelagic sediment is an accumulative record of oceanographic conditions over a period of thousands, or many thousands, of years. Therefore the depth of the facies boundary on the present ocean bottom may deviate from that of the momentary CCD by 200 m or so (Berger, 1974).

The lysocline is the upper limit of a zone where the dissolution rate increases markedly over a short depth interval. The lysocline in the Atlantic lies hundreds of meters above the calcite compensation surface (Berger, 1974). The foraminiferal lysocline is defined as the upper limit of the boundary zone between well preserved and poorly preserved foraminiferal assemblages on the seafloor. This lysocline has the properties of a compensation depth in that it can exist independent of the particular shape of a dissolution profile as long as the dissolution rate increases with depth. In the central Atlantic, where the lysocline was originally defined, there is evidence that an acceleration in dissolution rate is associated with the foraminiferal lysocline (Berger 1974, 1976).

We carried out a study of the dissolution facies to estimate the levels of past CCDs and lysoclines. Increased dissolution is manifested by increases in (1) the percentage of insoluble residue, (2) the fragmentation of foraminiferal tests, and (3) the ratio of benthic to planktonic foraminifers. The analyses are reported in detail in the site chapters (sections authored by Wright and by Hsü). A high percentage of terrigenous insoluble residue results from the dissolution of calcareous skeletal tests. The fragmentation of foraminiferal tests finds its expression in grain-size analyses. The degree of dissolution of a pelagic sediment can, therefore, be defined by two numerical parameters: carbonate percentage (or conversely insoluble residue percentage) and the percentage of the sediment made up by the size fractions greater than 62 μm (Hsü and Andrews, 1970). We used these parameters to classify the South Atlantic sediments devoid of siliceous plankton and found that the eolytic and oligolytic pelagic sediments contain more than 1% sand fraction and less than 30% terrigenous detritus.

The mesolytic contains 30 to 70% terrigenous detritus and the pleistolytic more than 70%; the hololytic, on the other hand, is an abyssal clay devoid of calcium carbonate. All these more dissolved pelagic sediments contain less than 1% sand fraction (Violanti et al., 1979). Studies by Finger (this vol.) have confirmed the pattern.

Variations of CCD and of Foraminiferal Lysoclines

The planktonic foraminiferal fauna is fairly well preserved in the sediments of the eolytic and oligolytic facies, but it is much more fragmented and poorly preserved in mesolytic and more dissolved sediments. We may thus conclude that the upper limit of the paleodepth of mesolytic facies may define the foraminiferal lysocline; eolytic and oligolytic sediments were deposited at a depth above the lysocline.

The hololytic sediments are deposited below the CCD, and their paleodepths give an estimate of the minimum CCD. A sample of pelagic clay or pleistolytic sediment from a 1-cm interval represents, however, a span of time of about 10^4 yr., during which the CCD may have varied somewhat. A pleistolytic sample, therefore, may include sediments deposited both above and below the CCD, and its paleodepth should represent a good average of the CCD for the time span represented by the sample.

We used the dissolution data summarized in the site chapters and the backtracking method to determine the paleodepth of the samples studied. The resulting temporal variations and the depth distribution of the dissolution facies are shown in Figure 1. If it is assumed that the lysocline and the CCD are defined by the depths of mesolytic and pleistolytic sediments, respectively, the depth variations of these surfaces with time can be estimated. Also shown by the figure are the paleodepths of the samples that yielded the first benthic microfauna containing significant amounts (10–20%) of *Nuttalides umbonifera.*

The temporal variations of the lysocline and of the CCD in the South Atlantic show the same general trends as observed previously by Van Andel et al. (1977) and by Melguen (1978). Two peaks in the CCD have been identified; the CCD was very shallow during the middle Eocene and the middle Miocene. However, a close comparison of the curves (Fig. 2) reveals significant differences. Oddly enough, the Neogene part of our CCD curve closely simulates that of Melguen, with a CCD peak at a depth of about 3 km some 12 Ma, whereas the Paleogene part of our curve simulates that of Van Andel et al., with a broad valley (CCD at about 4.3 km) for the Oligocene.

We believe that Van Andel et al. underestimated the CCD rise during the middle Miocene. Their interpretation of the Neogene CCD variations was based primarily upon Leg 3 data (Sites 13 to 20), which resulted from Miocene sequences that were not continuously cored. We now have a close sampling from a sequence that is both more accurately dated and more complete, and the high quality of our data justifies a revision. For example, the presence of middle Miocene hololytic sedi-

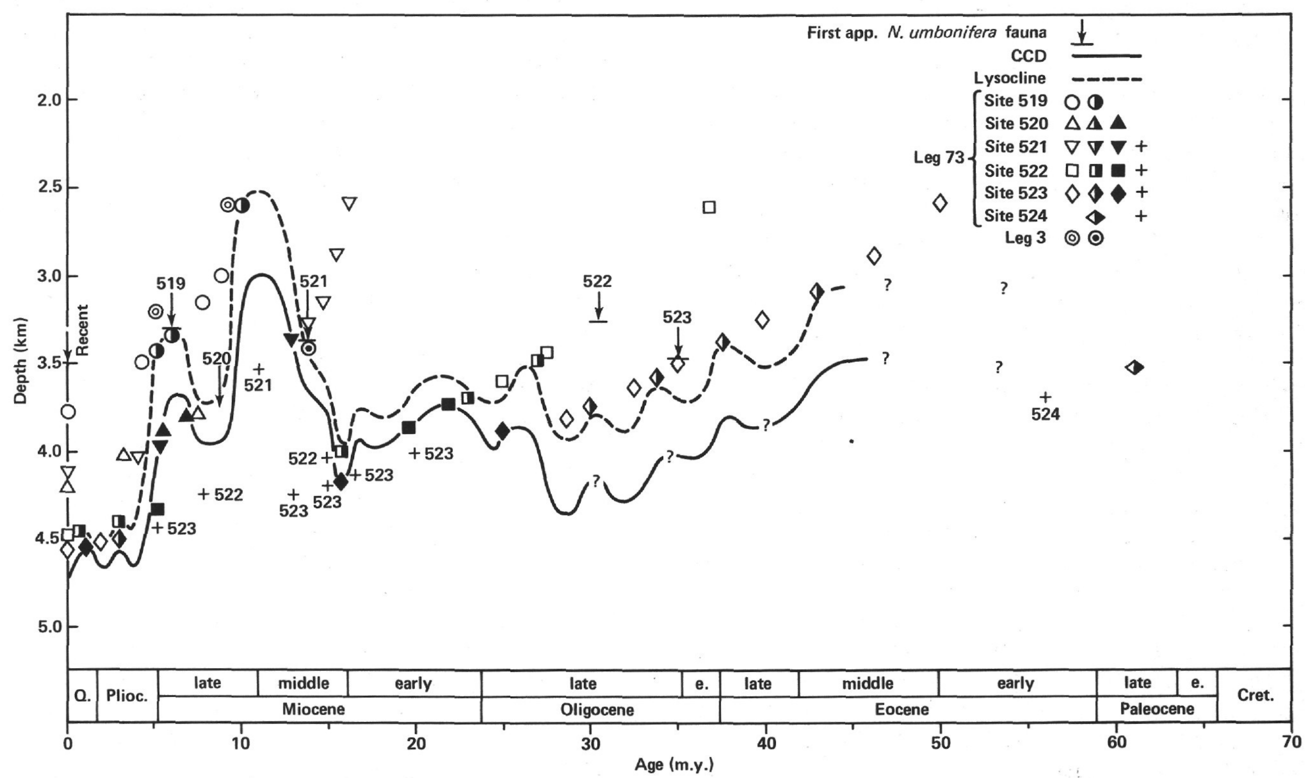

Figure 1. Temporal variation of CCD in South Atlantic. Empty symbols are eolytic and oligolytic sediments, which should lie above lysocline. Half solid symbols are mesolytic sediments, which should lie at or near the lysocline depth. Solid symbols are pleistolytic sediments, which should lie at or near the CCD. Crosses are hololytic sediments, which should lie below CCD.

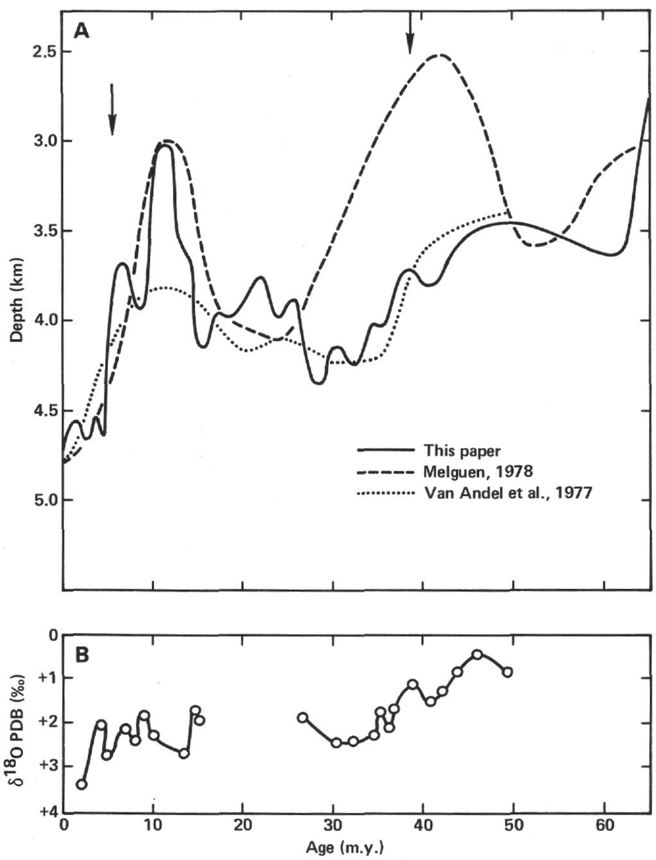

Figure 2. A. Comparison of three different theories of the temporal variation of the CCD in the South Atlantic. B. Temporal variation of $\delta^{18}O$. Arrows indicate end of Eocene and end of Miocene, when CCD became depressed.

ments at Site 521 indicates a CCD not shallower than 3.5 km 12 Ma. The presence of mesolytic sediments on newly formed ridge crest at Site 519 indicates a foraminiferal lysocline as shallow as 2.6 km some 10 Ma. Inasmuch as the CCD is commonly several hundred meters deeper than the lysocline (Berger, 1976), we conclude that the middle Miocene (11–13 m.y.) CCD should have been about 3 km deep at 30°S latitude, and our estimate confirms the conclusion reached by Melguen (1978).

The upper Miocene Messinian sediments at Site 16 on the west flank of the Mid-Atlantic Ridge are chalk oozes, whereas the sediments of the same age on its east flank, which were deposited on bottom 200 or 300 m deeper, are oligolytic or mesolytic marl oozes. This facies difference suggested to us that the lysocline might have been deeper on the west flank and that this lack of symmetry might explain the discrepancy between our curve and that of Van Andel et al. (1977). However, when we plotted the Leg 3 data on our temporal-spatial diagram, the Site 16 oligolytic samples lay above our lysocline curve; thus, the facies difference between the Messinian sediments on the west and east flanks could signify that the sediments at Site 16 were deposited just above, and those at Site 519 just below, the lysocline. A mesolytic sediment of Langhian age from Site 15 was deposited at a paleodepth of 3.4 km on the west flank of the ridge, and the data point lies on our lysocline curve. This fact suggests that there was little difference in the lysocline depth between the two sides of the ridge.

The discrepancy between our estimate and Melguen's of the Paleogene CCD may be a manifestation of the difference in geographical location. Her conclusions were based on data from Sites 360 and 361 in the Cape Basin, and our curve described the change in the CCD in the Mid-Atlantic area at the present 30°S latitude. Paleoceanographical factors could be sufficient to explain the difference in CCD between the two regions. However, we should not exclude the possibility that Melguen overestimated the degree of dissolution during the Paleogene. The Cape Basin sites were drilled on a continental margin, and the high clay content of the Paleogene nannofossil marls and clays may have resulted from a large component of detrital input from land, rather than from dissolution alone.

We are very confident of the reliability of our CCD curve for the last 50 m.y.; we are less certain of the CCD variations during the first 15 m.y. of the Cenozoic. We have no pleistolytic sediments older than Oligocene at Sites 522 and 523, so we constructed the early Paleogene CCD curve on the assumption that it lay at a depth of about 400 m beneath the foraminiferal lysocline. The close similarity between our estimate and that by Van Andel et al. (1977) seems to justify this assumption.

The apparently very high CCD at the very beginning of the Cenozoic has been discussed in another chapter (Hsü, this vol.). We question the hypothesis that the CCD was raised to the photic zone at that time, however, because the boundary clays are absent at a number of DSDP sites (see Supko, Perch-Nielsen, et al., 1977). The data from Site 524 suggest an abrupt and temporary rise of the CCD at that time to a level above 3 km. Meanwhile, the hololytic boundary clay from sections acquired on land (Denmark, Italy, Spain, Tunisia) may have resulted from dissolution by very corrosive CO_2-rich waters in an expanded zone of oxygen minimum (Hsü, this vol.).

Our detailed record permitted us to recognize several subsidiary peaks and valleys in the CCD curve; we also could prove the very rapid rate of changes. For example, the CCD rose more than 1 km during the first few million years after the start of middle Miocene, and it dropped more than 1 km within the first million years of the Pliocene (Fig. 1). The lowering of the CCD after the middle Eocene peak was a stepwise change that simulated somewhat the oxygen-isotope (or paleotemperature) shift. There was no sharp plunge across the Eocene/Oligocene boundary.

Very rapid oscillations of the CCD during the late Pliocene and the Quaternary are also suggested by the alternation of dark marly and light chalky oozes. The dark sediments are glacial in origin and are characterized by signs of more advanced degrees of dissolution (higher insoluble residue content, higher fragmentation of foraminiferal tests, higher percentages of benthic foraminifers). The dark-light couplet has a cyclicity of about 100,000 yr. and has been related by Weissert et al. (this vol.) to oscillations in the CCD during the late Neogene glaciation.

The benthic microfauna can be divided into two major types. After the first appearance of abundant (10–20%) *N. umbonifera,* the benthic fauna is moderately diverse. (In addition to this species, which may constitute a maximum of 60% of the benthic assemblage, the fauna includes *Cassidulina subglobosa, Oridorsalis umbonatus, Epistominella exigua, Planulina wuellerstorfi,* and *Pullenia* spp.) The assemblage dominated by *N. umbonifera* is similar to that found today on the ocean bottom under the Antarctic Bottom Water (AABW). Older faunas at Sites 519, 521, and 522 contain little or no *N. umbonifera* and are dominated by *E. exigua* and *P. wuellerstorfi.* Those Miocene faunas of the South Atlantic Deep Water are similar to a modern fauna typical of the lower North Atlantic Deep Water (NADW), but we are not certain if they can be considered to be equivalent to the Miocene fauna of the NADW. If the *N. umbonifera* fauna lived on ocean bottom under the AABW during the Tertiary as well as today, the first appearance of this fauna at each site should mark the time and paleodepth when the sea bottom at this site subsided to a depth below the top of a deep water mass equivalent to the present AABW.

Figure 1 shows that the *N. umbonifera* fauna first appeared at every site when the site subsided to a depth of 3.25 to 3.75 km. The species first occurred at Site 523 when it had subsided to depth of 3.5 km, deeper (and in that sense later) than at Site 522. The late first occurrence may have resulted from evolutionary development, because the older benthic faunas at this site were dominated by another species of the same genus, *N. truempyi.* If we accept the data from Sites 523 and 522 at face value, however, we may conclude that the top of the AABW rose some 250 m during the first 5 m.y. of the Oligocene because of an intensification of bottom current activity. The Oligocene lysocline lay below the top of the AABW, similar to the situation in the South Atlantic today, where *N. umbonifera* is dominant below 3,500 m (Clark et al., this volume), a level considerably above the lysocline depth there of more than 4 or 4.5 km. On the other hand, the first appearance of *N. umbonifera* during the Miocene occurred when a site subsided to a depth near the top of the lysocline. It seems that the Miocene lysocline may have followed the boundary between the AABW and NADW, as it does in the Central Atlantic today (Berger, 1974).

Cause of CCD Variations

The calcite compensation level is determined by supply and the loss of calcium carbonate through dissolution. The supply is related to the fertility of the calcareous plankton in the open oceans, whereas the loss is influenced to a large extent by the chemistry of the bottom water mass, with which a newly deposited sediment is in contact. In our analyses of CCD changes we see two types of variations: first-order variations involving changes from one geological epoch to another (10^7-yr. cycles) and second-order variations involving fractions of a million years (10^4- or 10^5-yr. cycles).

The second-order variations are exemplified by the light-dark couplets in the Pliocene–Quaternary sediments of the South Atlantic. Wright (see site summaries, this vol.) found that foraminifers in the more marly dark sediments have undergone more dissolution. Weissert et al. (this vol.) cited isotope evidence in concluding that the dark sediments were deposited during glacial stages. They also found that the benthic foraminifer data indicated that the AABW was intensified during glacial times and that this water mass was then enriched in biogenic CO_2 to cause increased dissolution. During the glacial stages this increase must have prevailed over a possible increase of production in the South Atlantic so that there was a rise in the lysocline for the deposition of more marly dark sediments.

The first-order changes appear in both our graphic summary of the CCD during the Tertiary (Fig. 1) and several reconstructions by previous workers (e.g. Hay, 1970; Berger and Roth, 1975; Van Andel et al., 1977; and Melguen, 1978). Melguen (1978) attempted to explain the first-order changes by the same mechanism as the second-order changes, and related the first-order rises of the CCD to intensified activity in the AABW. If the paleodepth of the first appearance of the *N. umbonifera* fauna is any indication, however, the difference in the AABW activities is small. In fact, it seems that the activity in the AABW first started in the late Eocene or Oligocene and became somewhat intensified during the Pliocene and Quaternary. According to the Melguen hypothesis, the CCD should rise during the early Pliocene—a trend opposite to the one commonly observed.

An alternative suggestion emphasizes ocean fertility; van Andel et al. (1977), for example, related the drop of the CCD during the late Eocene and early Oligocene in the South Atlantic to a sharp increase of plankton productivity there. This is a more likely explanation for the first-order changes in the CCD during the Cenozoic.

The productivity of calcareous plankton in the open oceans depends upon the supply of nutrients, particularly phosphorus and nitrogen, and this supply is controlled by (1) fluxes from the continents and (2) consumption by competing sinks for nutrients.

The first-order highs for the CCD occurred during the Eocene and Miocene. Berger et al. (1981) pointed out that the high levels of CCD are correlated to times of transgression, of warmer temperatures, and of enriched $\delta^{13}C$ in the dissolved carbonates of the oceans. Arthur and Jenkyns (1981) further noted a correlation between CCD and the accumulation rate of economic phosphate deposits: the shallow CCDs during the Eocene and Miocene correspond to the epochs of phosphatic giants, whereas the low CCDs during the Oligocene and Pliocene–Quaternary correspond to times when comparatively few phosphates were formed.

It is difficult to judge the temporal changes of the fluxes from continents. More nutrients might be set free by weathering under warmer conditions, but the enlarged land areas during periods of lower sea level might also make more source terrain available for nutrient delivery. The phosphate occurrence indicates, however, that the consumption by competing sinks of nutrients was definitely greater during the Eocene and Miocene. Much of the phosphorus was fixed as phosphate deposits on submerged shelves during those epochs of high sea level. Furthermore, the Eocene and Miocene were epochs of very high productivity of siliceous plankton, which competed with calcareous plankton for nutrients: the Eocene was an epoch when the fertility of radiolarians in the equatorial regions reached its zenith, whereas the Miocene was an epoch when continental margins with upwelling waters were commonly sites of diatomite sedimentation.

We suggest that a low productivity of calcareous plankton in the open oceans explains the high CCD during the Eocene and the Miocene. The correlation with the trends of oxygen and carbon isotopes is indirect. Low $\delta^{18}O$ values correspond on the whole to smaller ice volumes or to transgressive seas and occur at times when abundant phosphates are formed on shelves. Benthic foraminifers with high $\delta^{13}C$ values, as will be discussed later, occur in epochs when the rate of carbonate-carbon production is reduced because of the high CCD.

STABLE-ISOTOPE GEOCHEMISTRY

Stable-isotope analyses were carried out at ETH at Zürich by McKenzie, Oberhänsli, Weissert, and He on foraminiferal samples selected by Wright, Poore, and Oberhänsli. Additional data were made available to us by Poore and Matthews (this vol.).

We made an exhaustive study of the stable isotopes of the sediments near the Cretaceous/Tertiary boundary (He et al., this vol.). We did not have time to study the excellent sequence of Paleocene and early Eocene sediments from Site 524, however; only two samples were analyzed from Nannofossil Zones NP4 and NP5 (64–63 m.y.) and three more from NP10 to NP12 (59–57 m.y.) (Oberhänsli et al., this vol.). The record from the middle Eocene to the late Oligocene (50–35 m.y.) is again of excellent quality (Poore and Matthews; and Oberhänsli et al., this vol.). We had few materials for the stable-isotope analysis of sediments from the early and middle Miocene, but we managed a detailed study of the late Neogene (McKenzie et al.; and Weissert et al., this vol.). The procedures, results, and conclusions of these studies are discussed in detail in separate chapters. The co-authors of this chapter did not always agree on how to interpret the material. This chapter reflects primarily the opinion of the first author; the co-authors state their interpretations of the time intervals they investigated in other chapters.

Oxygen-Isotope Record

The Cenozoic oxygen-isotope record from the Leg 73 sites is summarized graphically in Figure 3. The oxygen-isotope perturbations during the first few hundred thousand years of the Tertiary may be interpreted as paleotemperature changes caused indirectly by the fall of a large extraterrestrial body (Hsü et al., 1982; Hsü, this vol.). The short-term oxygen-isotope perturbations may not be the only consequence of such an impact. A radical change in the composition of the oceans and atmosphere, such as could have occurred after the terminal Cretaceous event, might trigger a long-term temperature decline, one that might continue, with numerous fluctuations, until today.

The Paleocene and early Eocene record showed that the *Subbotina* spp. were deep-living planktonic foraminifers (Oberhänsli et al., this vol.), confirming an observation made by Boersma et al. (1979), who obtained $\delta^{18}O$ values for *Subbotina* spp. very similar to those of the mixed benthic foraminifers in the early Paleocene sediments of Site 384 of the Central Atlantic. The early Paleocene *Morozovella* spp. give values indicative of near-surface conditions. The -1.25 to $-1.53‰$ $\delta^{18}O$ values indicate a Cenozoic climate optimum with surface temperatures up to 18°C.

We have a gap in our record of the Paleogene because the lower Eocene sediments belonging to Chrons C-22 and C-21 (56–50 m.y.) were not cored during the cruise; the record at Site 523 began with earliest middle Eocene sediments (P10, NP15, Chron C-20-R, 50 m.y.). The planktonic and benthic fauna in the oldest middle Eocene samples yields oxygen-isotope values about 1 to 1.5‰ heavier than those of the early lower Eocene sample. This signifies either a gradual decline of temperature or a cooling event. The record from the southern oceans indicates a cooling event, with a +1‰ oxygen shift at the end of the early Eocene (Shackleton and Kennett, 1975). The work by Oberhänsli on DSDP cores from several Indian Ocean sites at high southern paleolatitudes (pers. comm., 1982) also recognized a cooling event of similar magnitude during the time represented by the foraminiferal zone P10, which has a magnetostratigraphic age of about 50 m.y.

The cooling trend at Site 523 continued during the middle Eocene and culminated with another net oxygen shift of about 1‰ at the end of the middle Eocene (41 m.y.). The oxygen shift might represent a change in salinity (Oberhänsli et al., this vol.), or the ocean waters might have cooled by about 3 or 4° during the 8 to 10 m.y. of the middle Eocene.

The isotope records from the Pacific also indicate oxygen-isotope shifts corresponding to a cooling of the bottom and surface waters during the middle Eocene (Savin et al., 1975). However, a difference in the thermal histories of the low and middle-to-high latitudes is revealed by the plot in Figure 4; the Atlantic record at our sites, with a paleolatitude of about 40°S, is very similar to the record at the high-latitude sites from the southern oceans (Shackleton and Kennett, 1975). The isotope records from the Central Pacific sites, on the other hand, reveal the beginning of a gradually widening divergence between surface- and bottom-water temperatures in the ocean waters at low latitudes, a trend that has continued until today. The comparison in Figure 4 shows that the Eocene benthic foraminifers from the Pacific sites yield $\delta^{18}O$ values similar to those from the South Atlantic and the southern oceans but that the

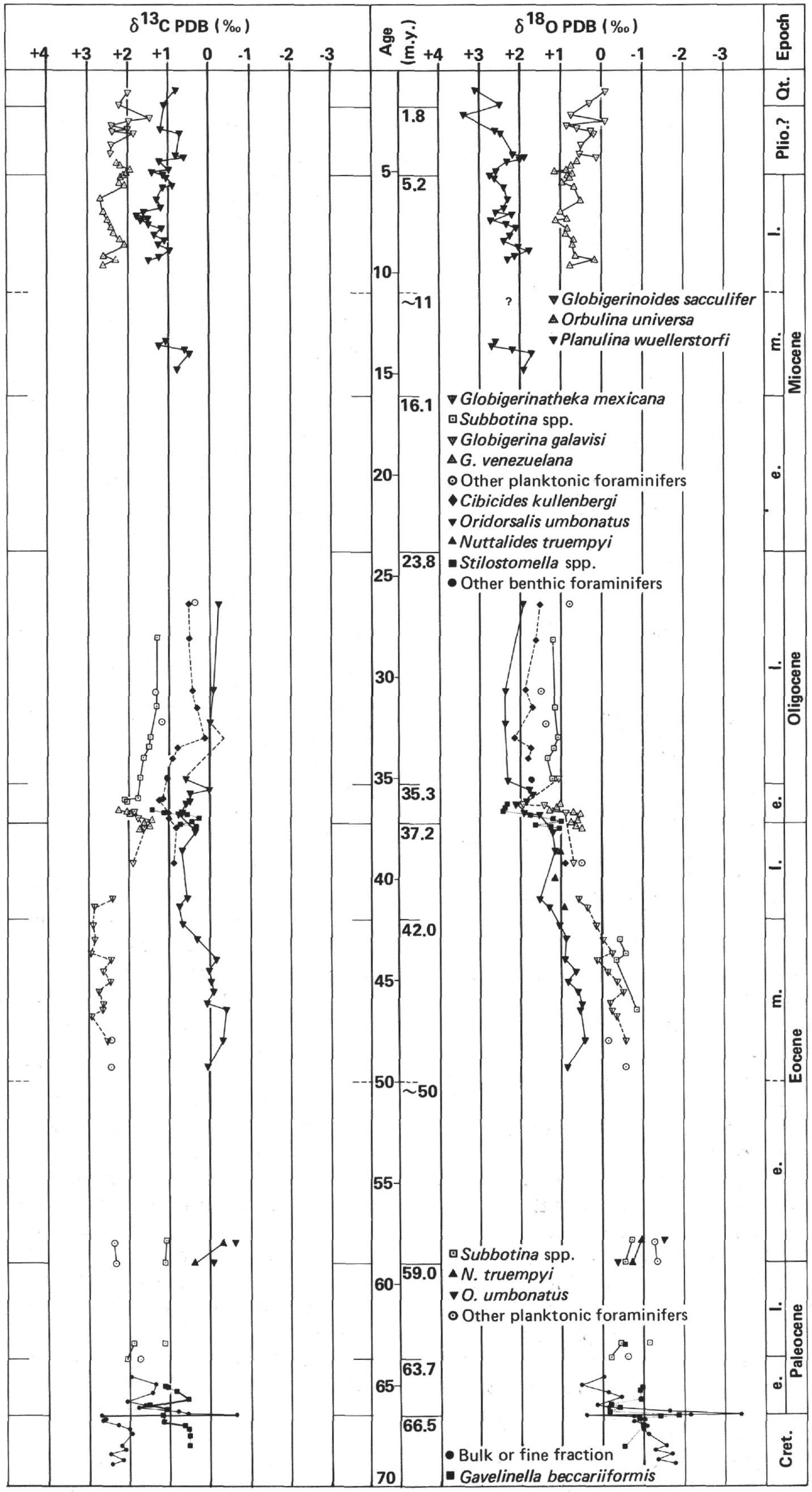

Figure 3. Carbon- and oxygen-isotope trends in the Cenozoic. The information is derived from studies of Leg 73 sites (see McKenzie et al., Poore and Matthews, Oberhänsli et al., and Weissert et al., all this vol.). Solid symbols represent benthic foraminifers, empty symbols planktonic foraminifers.

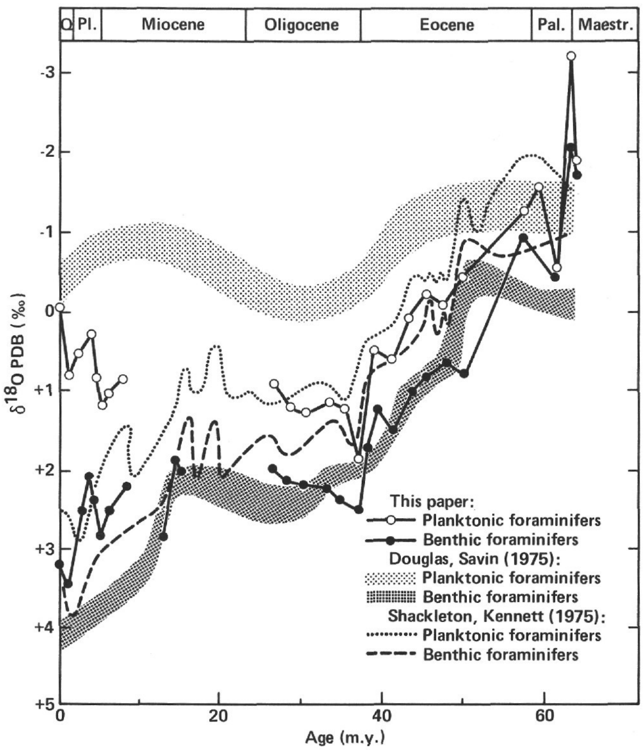

Figure 4. A comparison of temporal variations of oxygen-isotope trends in three oceans. Leg 73 results are compared with those of the southern oceans by Shackleton and Kennett (1975) and those of the Central Pacific by Douglas and Savin (1975).

middle Eocene surface waters in the equatorial regions were a few degrees warmer than those from the higher latitudes.

The upper Eocene and Oligocene samples, which were investigated by Oberhänsli and by Poore and Matthews (this vol.), reveal a dramatic oxygen shift during the earliest Oligocene (NP21). The benthic *Stilostomella* spp. in samples from Site 522 registered a maximum $\delta^{18}O$ shift of about 1‰; and the event, according to magnetostratigraphy, started at 36.7 m.y. and ended 100,000 yr. later (Oberhänsli et al., this vol.). An oxygen shift across the Eocene/Oligocene boundary or in the earliest Oligocene has also been recorded in samples from the Pacific and southern ocean sites: both the planktonic and benthic species exhibit oxygen-isotope shifts at the subantarctic Site 277, but only the benthic foraminifers register the change at the equatorial Pacific Site 292 (Keigwin, 1980). Our record, which is from an intermediate paleolatitude, is intermediate in nature: the oxygen shift in the benthic foraminifers is about the same as at the other two sites, and the planktonic oxygen shift is about half as great as at the subantarctic site.

The environmental changes responsible for the oxygen-isotope shift have been explained by two different models. The ice-free model (so called) assumes that the world was not glaciated before the middle Miocene and assumes the $\delta^{18}O$ value of -1‰ for an ice-free ocean to calculate Paleogene temperatures (Shackleton and Kennett, 1975; Savin et al., 1975). The minimum paleotemperatures for the earliest Oligocene surface and bottom waters at Site 522 (at 33°S paleolatitude) would thus be about 6°C and 2.5°C, respectively. The model used by Matthews and Poore (1980) and Poore and Matthews (this vol.) does assume the presence of ice before the Miocene and incorporates an ice-volume effect in calculating Oligocene paleotemperatures. Noting the diversity of Oligocene nannofossil assemblages, which contain the warm-water genera *Discoaster* and *Sphenolithus,* they assume an Oligocene surface-water temperature of 19°C for Site 522 (a temperature identical to that from 33°S in the South Atlantic today), and they deduce an ice-volume effect of $+1.3$‰ ($\delta^{18}O$). The computed early Oligocene bottom-water temperatures at this site would thus be 12°C, some 10° warmer than the present ocean bottom at 33°S. The much lower temperature gradient during the Oligocene is attributed to the production of warm saline bottom waters (WSBW) for deep circulation (see Peterson et. al., 1981).

The Poore–Matthews model assumes an ice volume during the early Oligocene that is only slightly less than that of the late Pleistocene glacial maximum and an ice volume during the late Eocene that is comparable to that today. This extraordinary interpretation differs radically from most current interpretations, but it should not be dismissed lightly, because recent drilling in the region of the Ross Sea has uncovered evidence of late Eocene glaciation there (L. Frakes, pers. comm., 1982).

As Poore and Matthews pointed out, the conservation of ocean water requires that the continuous production of WSBW be offset by large volumes of less saline waters elsewhere in the world ocean. Whereas an oxygen shift in equatorial surface waters might be attributed to a salinity effect, changes in the heavier $\delta^{18}O$ of the subantarctic surface waters must be related to an ice-volume effect or to cooling. The oxygen-isotope values of the planktonic and benthic foraminifers in southern ocean samples are almost identical, and those in our samples differ only slightly. One would have to assume an improbable coincidence to explain the nearly null oxygen-isotope gradient in the high latitudes; namely, less saline but cooler surface waters and highly saline but warmer bottom waters would have to have identical oxygen-isotope values. It would be more plausible to assume that the isotope gradient is minimum in the southern high latitudes because the cool, saline waters there descended to form bottom waters, the Oligocene Antarctic Bottom Waters. This ancestral AABW flowed northward to the middle latitudes of the South Atlantic, carrying with it the same (or slightly modified) isotope signatures. Thus, it seems more reasonable to assume that the oxygen shift during the early Oligocene included both a cooling and an ice-volume component; we may use the modern SMOW value to obtain more probable estimates of Oligocene paleotemperatures.

We have also carried out detailed investigations on the ecological and vital effects of foraminifers, and on the high-frequency fluctuations in the $\delta^{18}O$ records. Our results indicate that interspecific isotopic trends are not parallel to those established on the basis of investigating the Quaternary fauna (Poore and Matthews, this vol.; Oberhänsli et al., this vol.). High-frequency fluctuations are probably present in the Oligocene, with a periodicity of about 100,000 yr., but a definitive conclusion must await further study (Poore and Matthews, this vol.).

The oxygen-isotope values remained more or less constant during much of the late Oligocene (35–30 m.y.), decreasing toward the end of this epoch. A significant vertical isotope gradient became established during the late Oligocene; the benthic and planktonic foraminifers of this age yield $\delta^{18}O$ values that differ by about 1 or 1.5‰. The presence of this gradient probably signifies the activity of an Oligocene deep water circulation that brought cold, polar bottom waters to the middle and low latitudes. The trend of thermal differentiation of ocean waters at the Leg 73 sites continued throughout the Neogene and reached a maximum during the late Quaternary.

The slight negative oxygen-isotope shift during the latest Oligocene (27–25 m.y.) probably signifies a moderate warming trend. The Pacific record also indicates a terminal Oligocene warming (Savin et al., 1975). The warming took place during the early Miocene (25–20 m.y.) in the southern oceans, however (Shackleton and Kennett, 1975). Unfortunately, we have no suitable lower Miocene samples for isotope analysis. The benthic foraminifers of the earliest middle Miocene sample yield an oxygen-isotope value about the same as the latest Oligocene, but we are not certain if the early Miocene was an epoch of constancy, if an initial rise was followed by a subsequent decrease, or if there were frequent fluctuations.

An oxygen-isotope shift of +1‰ is registered by the fragmentary record of the middle Miocene at Site 521. The cooling event is revealed by the isotope analysis of benthic foraminifers in sediments belonging to nannofossil zone NN5 (magnetostratigraphic Epoch 15); the event started at about 14 m.y. and ended some half million years later. An oxygen-isotope shift of +1.5‰ has also been established by analyzing the middle Miocene benthic foraminifers (N9–N14) in the Central Pacific (Savin et al., 1975; 1981) and by the analysis of southern ocean samples of this age (Shackleton and Kennett, 1975). The older oxygen shifts are assumed to have resulted from cooling, and the middle Miocene shift is usually interpreted as an ice-volume effect, a result of the rapid expansion of an Antarctic continental ice sheet (Shackleton and Kennett, 1975; Kennett, 1982, p. 734). A model that assumes the presence of a large Oligocene ice cap would require a cooling of the bottom waters to be partly or wholly responsible for the middle Miocene oxygen-isotope shift. At any rate, it is generally believed that the East Antarctic ice sheet, the largest in Antarctica, has existed in essentially its present form since the middle Miocene, although isotope data alone cannot determine when the ice sheet first formed.

The highly dissolved sediments of the late middle Miocene have yielded no isotope data, but we do have a relatively continuous and detailed record for the late Miocene and the Pliocene–Quaternary. McKenzie et al. (this vol.) recognized a significant cooling, or continental ice-volume increase, during the Messinian (5.7–5.2 m.y.), which was followed by a period of warming, or ice-volume decrease, in the early Pliocene (5.2–4.3 m.y.). If it is assumed that an ice-volume effect exists, the record at Site 519 indicates a drop in sea level of 40 m in the early Messinian, which may have triggered the Mediterranean salinity crisis. The total decrease of $\delta^{18}O$ from 5.2 to 4.3 m.y. is 0.7‰, corresponding to a rise in sea level of 70 m, which may have resulted from climatic changes after the end of the salinity crisis.

During the early Pliocene (5.2–3.3 m.y.), low-amplitude climatic changes dominated a world that was less glaciated than during the Pleistocene. A cooling or an increase in ice volume at 3.2 m.y. is manifested by a +0.5‰ oxygen-isotope shift in the benthic foraminifers. The middle Pliocene (3.3–2.5 m.y.) is characterized by an increase in the amplitude of the isotope shift and by a significant increase in the vertical isotope gradient (Weissert et al., this vol.). Whereas the oxygen-isotope variation of the benthic foraminifers at the Leg 73 sites follows the trend at southern ocean sites, there is a general decrease of the $\delta^{18}O$ values of the planktonic foraminifers at the Leg 73 sites during the last 10 m.y., so that the late Quaternary planktonic foraminifers here yield oxygen-isotope values similar to those of the Central Pacific (see Fig. 4).

To recapitulate, we might note that the general trend of our data is similar to that found by previous authors (Shackleton and Kennett, 1975, and Douglas and Savin, 1975; see also Berger et al., 1981, and Haq, 1981). However, a detailed comparison of our oxygen shifts with those of the previous authors indicates significant differences (Fig. 4). The southern ocean sites are characterized by a relatively small vertical gradient in oxygen-isotope values throughout the Cenozoic, with an average difference of about 1‰ between the top and the bottom; both the bottom and the surface waters became cooler, from about 15 to 20° in the early Eocene to 2 or 3° in the Quaternary. The Central Pacific record is, however, characterized by a temporal increase in the vertical gradient: while the bottom-water temperatures dropped, the surface-water temperatures fluctuated, so that the difference between oxygen-isotope values of the planktonic and benthic foraminifers increased from about 1‰ in the Eocene to 4 or 5‰ in the Quaternary (see Fig. 4 and also Haq, 1981). This fact has led Berger et al. (1981) to conclude that the Cenozoic cooling trend is chiefly a high-latitude phenomenon. Our isotope record, however, shows a trend that is intermediate between the two extremes, as is to be expected, because our sites are in an intermediate position between the high-southern- and low-latitude sites investigated previously. Our data confirm that the latitudinal gradients of the ocean surface temperatures became much steeper during the Cenozoic. The vertical gradient of the middle-latitude oceans also increased, and the difference between the $\delta^{18}O$ values of the planktonic and benthic foraminifers increased from about 1‰ in the Eocene to about 3‰ in the Quaternary.

Carbon-Isotope Record

A sharp perturbation in the carbon-isotope trend has been detected by analyses of bulk samples across the Cretaceous/Tertiary boundary, and this evidence of mass mortality during a terminal Cretaceous crisis is discussed in another chapter (Hsü, this vol.). We have not yet analyzed our middle Paleocene samples, and we are not certain if the large $\delta^{13}C$ anomaly recorded by the planktonic foraminifers of 60–55 Ma (P4, NP9; see Douglas and Savin, 1971) is present at our site. We begin our systematic record with samples from the beginning of the middle Eocene. The records show both temporal variations in carbon-isotope values and significant changes in the vertical gradient.

Previous studies of samples from the southern oceans and from the Central Pacific have established two broad peaks in the carbon-isotope values since the beginning of the Eocene (Shackleton and Kennett, 1975; Savin et al., 1975; see also Fig. 5). The Miocene peak is manifested by the isotope composition of both benthic and planktonic foraminifers from the southern oceans (dotted lines in Fig. 5); the Miocene maximum there lies some 1.5‰ above the Oligocene and Pliocene–Quaternary low values. The Eocene peak is very broad in the planktonic foraminifers, extending over the whole of the epoch, but the peak value of the benthic foraminifers is sharply defined and restricted to the early Eocene in the southern oceans. The Central Pacific samples show a similarity in both the trends and the absolute values of $\delta^{13}C$; the Miocene and Eocene peaks there are also clearly defined (see Savin et al., 1975).

The carbon-isotope record from the South Atlantic indicates both similarities to, and departures from, the trends established by previous work. The middle Eocene planktonic foraminifers at our sites have very high $\delta^{13}C$ values (+2.5 to +3‰), clearly defining a peak some 1.5 to 2‰ higher than the late Paleocene and late Eocene values. The peak value of the benthic foraminifers was reached only in the late Eocene and earliest Oligocene, however. The late Oligocene low (35–25 m.y.) is well defined despite the gaps in our early and middle Miocene record, and the available data are in agreement with the hypothesis that there was a Miocene carbon peak. A generally high range of $\delta^{13}C$ values for the Miocene has also been revealed by the analysis of bulk samples and benthic foraminifers from Site 400 in the North Atlantic (Vergnaud Grazzini, et al., 1978).

There is a striking similarity between the benthic and planktonic carbon-isotope trends during the Cenozoic, with the notable exception of a section in the middle and late Eocene (Shackleton and Kennett, 1975, p. 754). Our results revealed the same striking parallelism for the Oligocene and the Neogene trends and the same divergence for the Eocene. Furthermore, there seems to be a convergence of carbon-isotope values (of the bottom and surface faunas) during the terminal Cretaceous and the terminal Eocene crises.

The Pliocene–Quaternary parallelism between the carbon-isotope trends of the planktonic and benthic foraminifers took place during the time when the oxygen-isotope trends of the two diverged, whereas the Eocene divergence of the carbon-isotope trends are synchronous to parallel oxygen-isotope trends (see Fig. 3). Yet both the carbon and oxygen values of the two faunas tend to converge at times of crisis.

The carbon-isotope values of foraminifers are influenced by many factors. Changes may result from changes in the isotopic composition of the carbonates dissolved in the oceans. An excessive burial of organic carbon rich in ^{12}C could lead to an enrichment of $\delta^{13}C$ in oceanic carbonates. A change in the steady-state proportion of the terrestrial and marine biomass may also result in a $\delta^{13}C$ variation. If either of these mechanisms is primarily responsible, there should be parallelism in the carbon-isotope shifts of the planktonic and benthic foraminifers. With the exceptions noted above, parallelism prevailed during much of the Cenozoic, giving credence to the hypothesis of worldwide temporal changes in the Cenozoic carbon-isotope composition of dissolved carbonates in the oceans.

The Miocene and Eocene ^{13}C peaks coincide in timing to high levels of CCD. We have no exact information on carbon flux during the Miocene. The common presence of marly sediments in Miocene oceans and of organic-carbon-rich diatomites on Miocene continental margins indicate, however, that unusually slow carbonate-carbon deposition was coupled with rapid organic-carbon burial during this epoch. The Eocene CCD high suggests a less than normal rate of carbonate deposition in the oceans, but the origin of the Eocene $\delta^{13}C$ peak may also have involved other factors. The results from the southern ocean study indicate that there are $\delta^{13}C$ peaks for both the planktonic and benthic faunas in the early Eocene; a middle Eocene high is also manifested, although only by the $\delta^{13}C$ values of the plankton. We suspect that a significant destruction of forests took place on land (terrestrial biomass) during the late Eocene cooling, when glaciation was initiated in the Antarctic. This development may have resulted in an enrichment of oceanic ^{12}C, as indicated by the relatively low $\delta^{13}C$ values for the late Eocene and for much of the Oligocene at the southern ocean and Atlantic sites. Oceanic ^{12}C has remained enriched, compared to the early Paleogene, until today, except for the reversal in the Miocene, which was probably caused by an unusually high rate of organic-carbon deposition.

Another reservoir of light carbon is the CO_2 in the atmosphere. The $\delta^{13}C$ peaks in the Eocene and Miocene might be taken as evidence of high CO_2 contents in the atmosphere. Such high CO_2 contents could have been the cause of relatively warm climates during those epochs. Unfortunately, we have no way to test this idea.

The difference in the $\delta^{13}C$ values of the planktonic and benthic foraminifers is commonly attributed to the fractionation of carbon isotopes by planktonic organisms in surface waters; the organic tissues prefer ^{12}C, and high productivity leads to a fractionation so that dissolved carbonates in ocean surface waters become impoverished in ^{12}C or enriched in ^{13}C. This difference between the surface and bottom waters is reflected in the difference in the isotopic composition of the foraminiferal shells: the benthic and planktonic species in a sediment may show a $\delta^{13}C$ difference ranging from about 1 to 2‰. The Oligocene and the Neogene samples from our sites show such a difference, attesting to a steady-state fractionation by planktonic productivity. After the

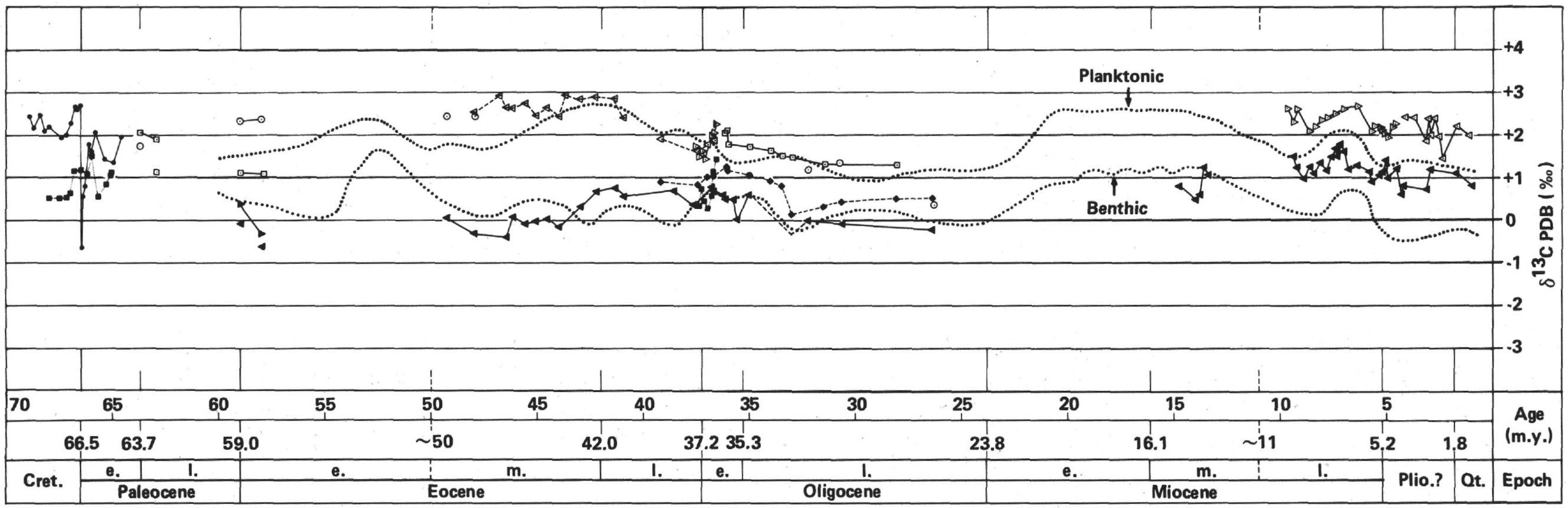

Figure 5. A comparison of the temporal variation of the carbon-isotope trends for the South Atlantic. Data are repeated from Fig. 3; solid symbols denote benthic foraminifers, empty symbols planktonic foraminifers. The curves for the southern oceans are shown by heavy dotted lines, which are based upon three-point sliding averages of the data in Shackleton and Kennett (1975).

terminal Cretaceous event however, there was a mass mortality (Hsü et al., 1982); the plankton fertility of the oceans may have been greatly suppressed for a period comparable in duration to the turnover time of the ocean waters (10^3 years). This results in the anomaly that the earliest Tertiary planktonic foraminifers have $\delta^{13}C$ values equal to or lighter than those of benthic foraminifers (Boersma et al., 1979; Hsü et al., 1982).

An unusually large difference of 2.5 to 3‰ in $\delta^{13}C$ values exists between the planktonic and benthic foraminifers in the middle Eocene sediments. The high CCD argues against an unusually high production rate of calcareous plankton, but the abundant middle Eocene cherts suggest a fertile ocean for siliceous plankton. Fertility influences the rate of carbon-isotope fractionation, but the time available to produce a difference between surface and bottom waters depends also upon the turnover rate of the oceans. A sluggish ocean with a turnover rate twice as long as at the present may be more stratified and allow isotope fractionation to become twice as advanced. This may explain the large difference in the vertical gradient of the carbon isotopes in the Eocene oceans as compared with the vertical gradient in the Oligocene and Neogene oceans, when the vigorous circulation of the AABW and NADW contributed to accelerate the turnover time of ocean waters. This line of reasoning may also explain the very positive carbon-isotope values in the middle Paleocene planktonic foraminifers at Site 47, which are 2 or 3‰ more positive than the extrapolated values for the benthic foraminifers of the same age (see Savin et al., 1975, p. 1509).

A remarkable carbon-isotope shift took place across the Eocene/Oligocene boundary. The maximum shift is +1.5‰ (from about 0 to 1.5‰) in the benthic and slightly more than 1‰ (1.2 to 2.3‰) in the planktonic foraminifers, and the change took place within half a million years (Oberhänsli et al., this vol.). Carbon shifts of similar magnitude across the Eocene/Oligocene boundary have been reported from investigations of southern ocean samples (Shackleton and Kennett, 1975; and Keigwin, 1980). We have no evidence of a preferential dissolution of carbonates or preferential deposition of organic carbon that could explain a positive shift. In fact, the isotope shift was sychronous with a CCD depression, which resulted in an accelerated deposition of carbonate carbon. In addition, the rate was very rapid, and the effect seemed to have been ocean wide. This enrichment of $\delta^{13}C$ could have resulted from an increase in the biomass on land. This is, of course, contrary to the trend of the carbon-isotope changes during the Quaternary, when the destruction of forests by advancing glaciers led to an enrichment of ^{12}C, not ^{13}C, in the dissolved carbonates of ocean waters (Shackleton, 1977). However, if we interpret the oxygen-isotope signal as a manifestation of increased ice volume, it signifies a drop in sea level of 100 m or more. The consequent exposure of the marginal shelves could lead to an enlargement of the land area available for plant growth. Such a scenario might explain the sudden carbon and oxygen shifts across the Eocene/Oligocene boundary. It should be pointed out that the Quaternary glaciers of the Northern Hemisphere covered large tracts of land in the northern temperate climatic zone, where forests grew during the interglacial stages. The Quaternary glaciation also brought about a change from forest to steppe vegetation, thus decreasing the total biomass on land. If the terminal Eocene glaciers grew mainly in the Antarctic, where there had been little forest vegetation, and if the early Oligocene coastal plains were largely sites of forest growth, one can imagine that the effects of the Oligocene and Quaternary decreases in sea level on the $\delta^{13}C$ values of dissolved carbonates in the oceans might be exactly opposite.

Poore and Matthews (this vol.) noted that at Site 522 a negative shift took place about 30 Ma in the $\delta^{13}C$ of the benthic foraminifers. They interpreted this change as a local phenomenon that occurred when the seafloor there subsided from an intermediate water mass into a more corrosive bottom water mass enriched in ^{12}C. The timing of this event in fact coincided with the first appearance of *Nuttalides umbonifera* fauna there (Fig. 1), which, as noted by Wright, may represent the time when the site sank into the deep water mass corresponding to the AABW. The paleoecological evidence thus confirms the geochemical interpretation.

A negative carbon shift of about −1‰ that took place during the late Miocene has been reported from several Pacific and Indian Ocean DSDP sites (Haq et al., 1980). The event occurred 6.1 to 5.9 Ma. At that time Site 519 had a depth of about 3,300 m and should already have subsided into the deep water mass. However, the isotope shift at that site was only −0.14‰, minute compared with the shift reported for other deep sea locations (see McKenzie et al., this vol.).

The late Miocene carbon shift has been related to a major reorganization of bottom-water circulation. At present, the cold, saline waters of the Weddell and Norwegian seas descend into the deep Atlantic and form the AABW and NADW, respectively. Then the deep water masses of the Atlantic flow eastward into the Indian and Pacific oceans. The North Pacific, at the far end of the deep circulation, has a bottom water enriched in biogenic CO_2 and ^{12}C. This pattern was probably first established about 6 Ma, when the carbon shift occurred, because the Pacific $\delta^{13}C$ values have been about 1‰ less than the Atlantic values since that time (Fig. 1). The more positive $\delta^{13}C$ values in the Pacific and Indian oceans prior to the carbon shift suggest that the bottom water there had a larger locally derived component before the present circulation pattern prevailed. The Atlantic bottom waters, on the other hand, have been derived from the Atlantic polar regions before and after the reorganization, so no significant late Miocene carbon shift was recorded by deep benthic foraminifers at Site 519.

Systematic carbon-isotope changes during the Pliocene–Quaternary have been recorded by benthic foraminifers from the dark and light pelagic sediments deposited during glacial and interglacial stages, respectively (Weissert et al., this vol.). A more negative $\delta^{13}C$ signal from the dark sediments confirms the previous conclusion reached by Shackleton (1977), who related Quaternary carbon-isotope values to the size of the terrestrial biomass.

PALEOECOLOGICAL EVOLUTION

The development of the psychrosphere and of thermohaline circulation is a most important event in the history of the Cenozoic oceans (Benson, 1975; Schnitker, 1979; Haq, 1981). The exact timing of this event has been investigated by Parker, Clark, Wright, and Clark, and their results are presented in detail in a separate chapter (Parker et al., this vol.). Their study of the origination, population structure, migration, and extinction of deep benthic fauna provide evidence on the nature and timing of circulation changes in abyssal realms. The results of their study (Parker et al., this vol.) indicate that the initiation of the psychrosphere in the South Atlantic may have occurred in several steps and that the most dramatic step occurred during the late Eocene. The source of the cold deep water for the psychrosphere is commonly believed to have come from as far south as the AABW (e.g., Haq, 1981). The occurrence of some taxa in the eastern Atlantic, but not in the western South Atlantic, suggests, however, a northern source for the deep waters east of the Mid-Atlantic Ridge (Parker et al., this vol.).

The occurrences of several layers of Oligocene *Braarudosphaera* chalk were first discovered in the South Atlantic during the Leg 3 cruise and were subsequently reported from the Indian oceans (Roth, 1974) and from the African Atlantic margin (Melguen, 1978). The chalk is remarkable because one or two species of *Braarudosphaera* dominate the nannoflora, and the chalk layers may attain considerable thickness (e.g., 20 cm at Site 522). Furthermore, *Braarudosphaera* spp. are known for their tolerance of salinity variations, and they are common in coastal waters. The distribution of the chalk in the South Atlantic suggests blooms of *Braarudosphaera* spp. in upwelling waters off the African margin; the nannofloral skeletons were carried westward by a current system that might be designated the ancestral Benguela Current (see Site 522 chapter, this vol.). Differential dissolution, as well as sorting during transport, may have resulted in the accumulation of the monospecific assemblage. The isotope signatures of the chalk are similar to those of benthic foraminifers of the same age, and the presence of euhedral rhombs of calcite in the chalk also suggests recrystallization under deep sea conditions (Wise and Hsü, 1971). Lepispheres of opal C-T have been found in a South Atlantic chalk sample; they are presumably the recrystallization product of siliceous plankton.

A diatom ooze consisting almost exclusively of skeletons of the species *Ethmodiscus rex* is present in the Pliocene sequence of eastern South Atlantic sites. The preservation of the monospecific assemblage has been related to the resistance of *E. rex* to differential solution (Gombos, this vol.), and the distribution of the ooze may mark the path of a westbound branch of the Pliocene Benguella Current. The diatom ooze is absent at western South Atlantic sites.

CONCLUSIONS

We have obtained a relatively complete record of oceanographic change during the last 50 m.y. through the study of hydraulic piston core samples dated by precision stratigraphy. We have confirmed previous observations that the world's climate started to cool in the middle Eocene and continued to deteriorate until today, except for short-term fluctuations and a moderate reversal of the overall trend during the late Oligocene and early Miocene. The cooling may have initiated glacial advances in the Antarctic during the late Eocene. In the early Oligocene the ice volume on Earth may have been greater than it is today. The cooling, which started early in the middle Eocene, resulted in the initiation of thermohaline circulation and the establishment of the psychrosphere in the late Eocene. Significant latitudinal and vertical temperature gradients began to develop during the early Oligocene, and the gradients became steeper with time, leading to the present-day thermal structure of the oceans.

The fertility of calcareous planktons has fluctuated greatly from one geologic epoch to another. The nutrient intake by siliceous plankton and the phosphorus fixation in phosphates may have been responsible for a relatively reduced fertility during the Eocene, which resulted in a high level of CCD. The preferential dissolution of carbonate carbon and burial of organic carbon may have resulted in the high $\delta^{13}C$ values in the benthic foraminifers of the early Eocene. The light carbon in the oceans was then enriched when the terrestrial biomass on the Antarctic continent was largely destroyed by advancing glaciers and cooling in the late Eocene. A sharp perturbation in the Oligocene represented a brief interruption of the trend, when forest growth on shelves exposed at times of low sea level preferentially extracted the light isotope from the world's carbon reservoirs. Another epoch of reduced fertility of calcareous plankton during the Miocene might also be related to coastal processes that competed more successfully for nutrients; diatom blooms and phosphate deposition on continental margins might have greatly limited the supply of nutrients available to the open oceans. The decrease in supply started another epoch of high CCD levels and high $\delta^{13}C$ values.

The middle Miocene oxygen-isotope shift was a sudden event that occurred from 14 to 13.5 Ma. The CCD soon (13–11 m.y.) reached its highest level (< 3,000 m). Another small oxygen shift, signifying perhaps an ice-volume increase, should have lowered the worldwide sea level sufficiently to induce the Messinian salinity crisis. The subsequent rise of sea level at the beginning of the Pliocene followed the end of the crisis. The Pliocene and Quaternary were epochs of high fertility for the calcareous plankton in the open oceans; the CCD suddenly became depressed by some 2 km in the early Pliocene. Oxygen shifts in the middle Pliocene (3.3 and 2.4 m.y) coincided in timing to the postulated start of the Arctic glaciation. Since then a pattern of oscillating glacial and interglacial stages has prevailed in the South Atlantic.

We can only speculate on the ultimate causes of climatic changes by correlating these changes with synchronous events. The world's climate is influenced by the distribution of land and seas, and many hypotheses suggest that tectonic events explain the cooling of the Earth's climate during the Tertiary.

The isotope records showed that the turning point in Cenozoic climatic history was the beginning of the cooling trend that started late in the early Eocene, or a little more than 50 m.y., during Chron C-21. We have no evidence that an unusual extraterrestrial event took place at that time. The terminal Cretaceous event, if it was the impact of a large body, took place some 15 m.y. earlier. The event did cause a large perturbation in climatic conditions, but the effect did not seem to last beyond 10^5 or 10^6 years (Hsü, this vol.); there is no reason to suppose that the Eocene cooling started as a delayed reaction to the terminal Cretaceous crisis. There was, however, a major tectonic event that may have triggered the critical change. Approximately 55 m.y. (Chron C-22), Australia became detached from Antarctica and began to drift northward toward its present position (Weissel and Hayes, 1972). This separation should have started a circum-Antarctic flow of surface waters, and such a paleogeographical change could have influenced precipitation patterns. We theorize that moist air masses might have invaded the Antarctic interior, with the increased precipitation of snow at high latitudes leading to the initiation of glaciation on this continent.

The next milestone in Cenozoic climatic history was the abrupt oxygen shift in the early Oligocene, which signified a major cooling episode and/or a rapid expansion of the Antarctic Ice Sheet. The discovery of microtektites and of an iridium anomaly in the uppermost Eocene sediments indicates the impact of a large body at that time (Alvarez et al., 1982; Glass and Crosbie, 1982). The dating of impact craters of that age strengthens this hypothesis (Bottomley et al., 1979). O'Keefe (1980) proposed that the tektites once formed a ring around the Earth like Saturn's rings, and that the temperature on Earth declined in winter months when half of the Earth was immersed in the shadow of the ring. This is an interesting idea that cannot be tested by our data.

The moderate reversal of the climatic trend in the late Oligocene or early Miocene requires an explanation. A major tectonic event caused the detachment of Tasmania from Antarctica in the late Oligocene (Chron C-8), and this separation permitted the deep circulations around the Antarctic (Kennett et al., 1975). This development diminished the meridional circulation of the world: the circum-Antarctic current trapped much of the cold water masses, the northward flow of the Antarctic Bottom Water as western boundary currents may have diminished in force, and the weakened influences of the cold waters from the south could have been the cause of an amelioration of the climate in the middle and low latitudes.

The middle Miocene climatic change can be explained in various ways. The sharp oxygen shift at about 14 m.y. came some 0.5 m.y. after an impact that created the Ries Crater in Germany and the Moldavite tektite (Gentler and Wagner, 1969). The size of the bolide, as judged by the crater size, seems to have been too small to cause an important global climatic change, however. Furthermore, the middle Miocene cooling was more or less synchronous with a tectonic event that severed the connections between the Atlantic and the Indian oceans (Hsü et al., 1977); one can speculate on the severe consequences of cutting off the surface circulation around the Earth along the equator, which came to an abrupt end some 14 m.y.

The change of oceanographic conditions toward the end of Miocene was apparently related to the Mediterranean salinity crisis. Different scenarios have been suggested, including one that involves a major reorganization of bottom-water circulation (e.g., Bender and Keigwin, 1979; Blanc, 1981). The changes that took place during the Pliocene are also puzzling; one is tempted to explore them in relation to the changes in ocean circulation that came about after the rise of the Panama Isthmus separated the Atlantic from the Pacific.

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