

Paleoenvironment of the North-East Atlantic during the Cenozoic : oxygen and carbon isotope analyses at Dsdp sites 398, 400 A and 401

North Eastern Atlantic
Cenozoic
Oxygen isotopes
Carbon isotopes
Foraminifera
Carbonates
Nord-Est Atlantique
Cénozoïque
Isotopes de l'oxygène
Isotopes du carbone
Foraminifères
Carbonates

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ABSTRACT

Oxygen and carbon isotope analyses of bulk carbonates and foraminiferal material from North Eastern Atlantic Dsdp sites 398 D, 400 A and 401, have provided information on the evolution of water masses during the Cenozoic era; though this evolution is greatly affected by diagenesis and selective dissolution at site 398, it is clear that Cenozoic time is characterized, at all sites, by a net cooling of the sea water. Surface temperatures decrease from maximal values around 19-20°C in the late Paleocene, to 10-16°C during the Oligocene, and 8°C during the glacial episodes of the Pleistocene. Bottom temperatures, which range between 12 and 16°C in the late Paleocene, decrease to 6-8°C in the Oligocene, 5°C in the middle late Miocene, and to an average value of 2.5°C in the Pleistocene.

This evolution of temperature patterns is related to the establishment, beginning at about -50 myrs, of a deep circulation current in the North Atlantic. The establishment of the psychrosphere at ca. -38 myrs, and the major ice accumulation in the Antarctic at ca. -11 myrs, are not well depicted by the isotopic curves. This may be due to the fact that, in this part of the North Atlantic, Antarctic bottom waters may have competed with warmer waters of Northern or Tethysian origin.

Fluctuations in the nutrient supply, as deduced from the $^{13}\text{C}/^{12}\text{C}$ ratio variations, are also correlated with the influx, during the middle Eocene, late-middle Miocene and Pleistocene epochs, of cold deep waters.

The overall paleoclimatic evolution of the North Eastern Atlantic throughout the Cenozoic era is similar to that of the Pacific Ocean, but the isotopic record is complicated by the fact that in the Atlantic, deep and surface waters alike resulted from a mixture of waters of different origins.

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RÉSUMÉ

Paléoenvironnement de l'Atlantique Nord-Est au Cénozoïque : analyses isotopiques de l'oxygène et du carbone dans les sites Dsdp 398, 400 A et 401

L'évolution des compositions isotopiques de l'oxygène et du carbone de la totalité de la fraction carbonatée, ainsi que des tests isolés de foraminifères, dans les sédiments prélevés aux sites 398 D, 400 A et 401 du Dsdp (bien que très affectée par les phénomènes de diagenèse-dissolution sélective au site 398 D), s'explique par une détérioration climatique progressive du milieu marin au cours du Cénozoïque. Les températures de surface, maximales au Paléocène supérieur (19 à 20°C) décroissent jusqu'à 10-16°C à l'Oligocène et 8°C pour les périodes froides du Pléistocène. Les températures du fond,

comprises entre 12 et 16°C au Paléocène supérieur, décroissent jusqu'à 6-8°C à l'Oligocène, 5°C à la fin du Miocène moyen, et 2,5°C au Pléistocène. Cette évolution est à mettre en relation avec l'installation d'une circulation profonde débutant il y a - 50 ma en Atlantique. Elle est en bon accord avec les données isotopiques concernant le Pacifique. Toutefois, dans cette partie de l'Atlantique, le signal isotopique est compliqué par le fait que les eaux de surface, aussi bien que de fond, résultent d'un mélange en proportions variées d'eaux d'origine nordique et téthysienne entre autres.

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INTRODUCTION

The variation through time of the oxygen and carbon isotopic compositions of biogenic carbonates (primarily Foraminifera) has frequently been used as a basis for the reconstruction of Quaternary paleoclimates. The amount of isotope paleoclimate for the pre-Quaternary period is much smaller. An initial attempt to define the climatic evolution of the global ocean was made by Douglas and Savin (1973) and Savin, Douglas and Stehli (1975), using sediments cored from low and middle latitudes in the Pacific (Dsdp Legs 6 and 17), and from South Atlantic middle latitudes (Dsdp Leg 3). Saito and van Donk (1974) published isotopic data on materials of late Cretaceous to early Tertiary origin from the South Atlantic. Subsequently, Shackleton and Kennett (1975) published a deep sea paleotemperature record for the past 54 myrs, based on Dsdp sites 277, 279 A and 281 in the South Antarctic, and proposed a timing of the principal advances of Antarctic glaciers and ice caps, together with an estimation of the change in the $^{18}\text{O}/^{16}\text{O}$ ratio of sea water produced by the accumulation of Antarctic ice during the Tertiary period. Shackleton and Kennett (1975) also proposed that Antarctic glaciers first formed at sea level at the beginning of the Oligocene. This would have led to the sinking of cold saline waters, as the first stage in the formation of deep Antarctic waters. This circulation of deep cold water would be responsible for a drastic temperature drop in the late Eocene-early Oligocene. A second important event was the creation of a permanent ice cap on Antarctica during the late middle Miocene. The storage at the pole of a great quantity of water depleted in ^{18}O ($\delta^{18}\text{O}$ around -50‰) would have induced a mean oceanic enrichment in ^{18}O of about $+1,2\text{‰}$ (Shackleton, Kennett, 1975). Before that time, the overall isotopic composition of ocean water was more negative and the isotopic variations registered by biogenic carbonates in deep waters resulted from thermal variations only.

Differences may, however appear in the precise timing of these events and in the velocity and intensity of the isotopic changes; these differences might be the result of diagenetic dissolution processes; but they might also be due to differences in depth of deposition, sampling intervals and distances from the zone of production of Antarctic bottom waters, or to the existence of topographic obstacles to the northward penetration of the AABW.

Such differences appear clearly when one compares the Subantarctic paleotemperature record from site 277-279 A-281 (Shackleton, Kennett, 1975) and the paleotemperature curves obtained for the Tertiary in the

South Atlantic site 357 (Boersma, Shackleton, 1977 *a*). As far as the paleo-oceanographic evolution of these North Atlantic is concerned one may expect these differences to be of a complex nature, due to:

- the evolution through time in the southern part of this area of topographical barriers which may have prevented the penetration of Antarctic deep waters;
- the uncertain timing of the onset of a deep circulation current originating in the Northern basins (Labrador Sea), as soon as the early Tertiary according to Berggren and Hollister (1974, 1977), but not before the subsidence of the Iceland-Faeroe Ridge during the middle Miocene according to others (see Thiede, in press, for references);
- variations in the regional westward influence of the Tethys;
- the regional influence of the Pyrenean orogeny during the Eocene;
- the extension and vicinity of the northern ice cap during the Pleistocene epoch.

THE STUDY AND ITS OBJECTIVES

Isotopic analyses of oxygen and carbon were carried out on carbonate tests of planktonic and benthic foraminifera selected from levels corresponding to time intervals of 2 to 4 myrs. Bulk carbonate was also analysed. Bulk carbonate is not commonly used, in paleotemperature studies, as its isotopic record may be altered by diagenesis or/and the presence of detrital carbonates. Nevertheless, when it is possible to ensure that these two factors do not play a rôle (by Scanning Electron Micrographs and smear slide observations), it may be assumed that the sample consists mainly of biogenic carbonates (coccoliths and foraminifera). From a weight-point of view, coccoliths are predominant.

Although a biological (non equilibrium) fractionation due to environmental factors might in certain cases influence the isotopic composition of coccoliths (Anderson, Cole, 1975), precipitation occurs in most cases, near thermodynamic equilibrium with surface ocean water (Margolis *et al.*, 1975). It may thus be assumed that bulk carbonate isotopic compositions reflect the temperature and salinity conditions of superficial waters, at least in a first approximation. Their variations should parallel those displayed by planktonic foraminifera.

RESULTS

The isotopic record. Effects of diagenesis and dissolution

The effects of dissolution and diagenesis on the isotopic record are not yet well understood: faunal assemblages

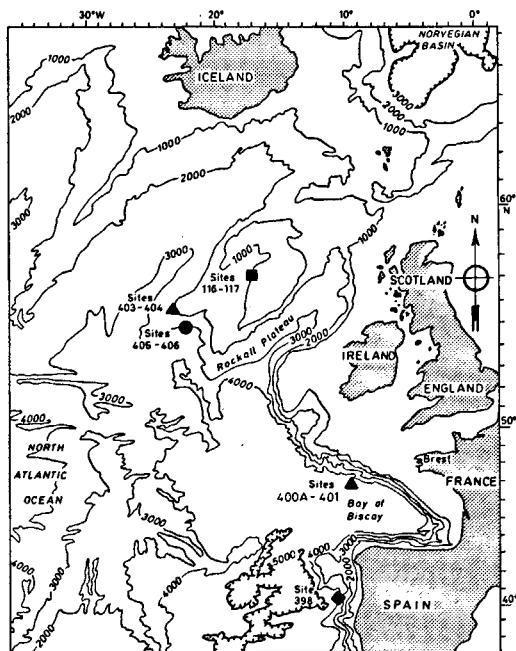


Figure 1
Location
of sites
studied.

are modified by the selective dissolution of thin-walled forms which are generally shallow-dwelling, hence warmer, species. In such cases, the mean isotopic composition no longer reflects the original isotopic record, and the apparent isotopic temperature may be altered by several ($^{\circ}\text{C}$) (Douglas, Savin, 1973; Berger, Killingley, 1977). Diagenetic recrystallization also modifies the isotopic record. The new isotopic ratios $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ depend on the temperature and on the corresponding isotope ratios of the water in which the recrystallization occurred. In deeply buried sediments, pore water isotopic composition, relative amounts of pore water, carbonate permeability and the geothermal gradient determine the new isotopic composition of the carbonates. In most instances, $^{18}\text{O}/^{16}\text{O}$ ratios are lowered during recrystallization. Temperatures calculated from isotopic compositions will hence be higher than growth temperatures. But in the case of calcite biogenically precipitated in near surface waters (planktonic foraminifera and nannofossils), secondary calcite formed at colder temperatures may occur in the form of overgrowths. In such cases, the $^{18}\text{O}/^{16}\text{O}$ ratio will be greater than that of unaltered planktonic material. This seems more frequent for nannofossils (Douglas, Savin, 1975). The comparison of sites located in the same region may yield some information concerning a possible alteration of the original isotopic signal.

Site 398

SEM examination of some samples as well as the study of variations in strontium content (Letolle, in press), have shown that at site 398, biogenic carbonates have been strongly recrystallized, at least in cores of upper Miocene age (core 2) and older.

Sites 400 A and 401

The data corresponding to relatively short episodes in the early to middle Eocene are rather puzzling: at site 400 A, located at deeper waters at the present time, the oxygen isotopic values are lower on the average (1‰) for bulk carbonates and for planktonic foraminifera

than those measured from the same levels at site 401. Furthermore, for the period corresponding to the late Early Eocene-early Middle Eocene, the oxygen isotopic compositions measured on benthonic foraminifera and nannofossils at both sites are not substantially different.

Diagenetic processes, involving dissolution as well as recrystallization, might be responsible for a change in the original isotopic signal. At site 401, during the Eocene, the $\delta^{18}\text{O}$ of bulk carbonates remain, on average, higher than the $\delta^{18}\text{O}$ of planktonic foraminifera. At both sites, through zones P8-P10, the $\delta^{18}\text{O}$ of bulk carbonates is higher than that of benthonic foraminifera. These abnormal results remain to be explained, although, at site 400 A, decreasing content in strontium before the Miocene epoch indicates the existence of a diagenetic recrystallization (Renard, Letolle, in press).

Variations of isotopic compositions with time

Results are reported in Tables 1 to 9 and displayed in Figures 2 and 3. From an isotopic point of view, the Cenozoic sequences may be divided into three major units which may be recognized in all three sites:

- *the Paleocene-Eocene sequence*: isotopic compositions display large fluctuations around a low $\delta^{18}\text{O}$ average. The lowest $\delta^{18}\text{O}$ values from -1 to -2‰ are found in this unit (at site 400 A);
- *the Oligocene-Miocene sequence*: isotopic variations are smaller and fluctuate around an average $\delta^{18}\text{O}$ value of $+1\text{‰}$;
- *Pliocene to Quaternary sediments*: isotopic variations are large. The highest $\delta^{18}\text{O}$ values, up to $+4\text{‰}$ for benthic foraminifera, are found here.

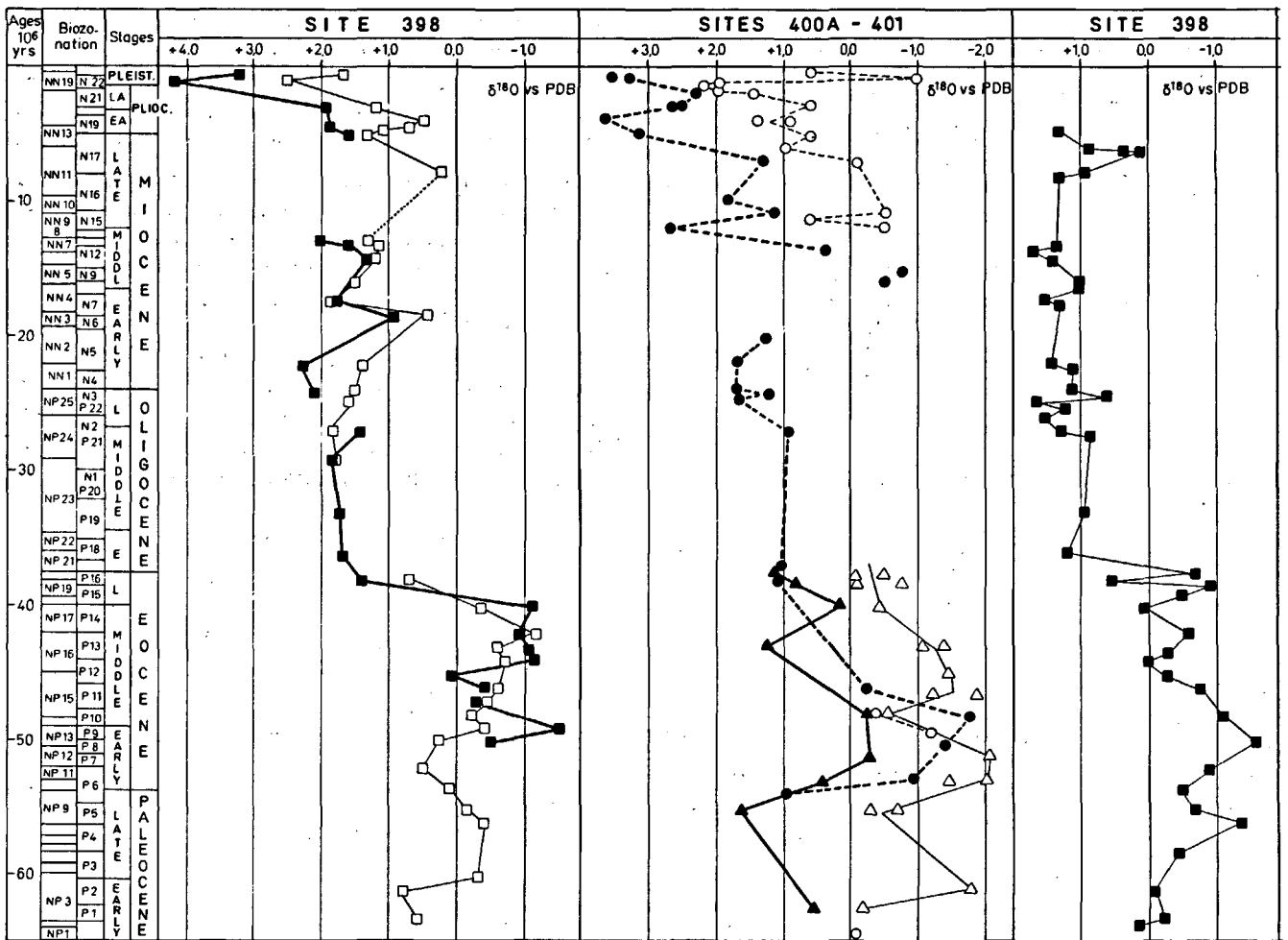
INTERPRETATION

The Paleocene-Eocene sequence

Bottom temperatures

During the late Paleocene, there is a rise in bottom temperatures, reaching a peak through zones P8-P10. Isotopic paleotemperatures may be calculated, using a value of -1.2‰ for the isotopic composition of sea water (following Shackleton and Kennett, 1975). At site 401, bottom temperatures reach values around 12.5°C ; this value is similar to that reported by Boersma and Shackleton (1977 a) at site 357. Thus, the warm pool of Atlantic water may be detected at greatly different depths and latitudes, and seems to have had a considerable extent during this period. Other evidence of warm water conditions at that time may be found in the literature [see Bryozoan assemblages of the late Paleocene from Rockall Bank (Cheetham, Hakansson, 1972)]. Bottom temperatures start to decrease near zone P10, at least in the Bay of Biscay, where isotopic temperatures reach very low values (around 7.5°C) during the late Eocene.

At site 398, oxygen isotope oscillations are difficult to interpret as a result of the diagenetic alteration (Vergnaud Grazzini, in press) Nevertheless, during the late Eocene, bottom temperatures suddenly fall to close to 8°C .



The drop in bottom temperatures which has been reported by Shackleton and Kennett (1975), Kennett and Shackleton (1976), Savin *et al.* (1975), Douglas and Savin (1973) for Pacific sediments of middle and high latitudes (Dsdp sites 167, 277) is believed to be Oligocene and most probably corresponds to the same event as the one observed here in the North-East Atlantic and in the South Atlantic by Boersma and Shackleton (1977 *a*). But in the Atlantic, cooling occurs during the middle to late Eocene, before zone P17, and appears to be more progressive. Moreover, isotopic temperatures evaluated for this part of the North Atlantic during the late Eocene appear to be higher than those evaluated for the South Atlantic or Pacific Oceans. The drop in temperature, which is to be related mainly to the formation of very cold waters (near freezing point) around Antarctica, also points to some "local" event (on the Atlantic scale) that may have also influenced this evolution. We are not in a position to assess whether this local event was the production of cold deep water of Northern origin as early as the middle Eocene. Although the Norwegian Sea (formed by the separation of Greenland from Scandinavia) already existed during the Eocene (Harland, 1969; Vine, Hess, 1971), the Iceland-Faeroe Ridge prohibited exchanges between Northern basins and the Northern part of the Atlantic. According to Berggren and Hollister (1974, 1977), the Northern flank of the Iceland-Faeroe Ridge had no direct connexion with the North Atlantic and the Norwegian Sea area may have been virtually isolated from the North Atlantic prior to the middle Eocene.

The Labrador basin attained its present size during the late Eocene-early Oligocene. It thus seems difficult to postulate, in the present state of our knowledge, the existence of a deep circulation of cold waters of Northern origin. These observations notwithstanding, the early middle Eocene was a period of significant changes in the characteristics of water masses in the North Atlantic. These changes are also documented by the evolution of benthic microfaunas. An important change in the ostracods fauna during the middle Eocene, where an increase in the number and diversity of species is observed, is reported by Ducasse and Peypouquet (in press). But the most important change occurs during the late Eocene when species representing the present modern fauna appeared. This fauna lives today at cold temperatures, less than 8-10°C. These temperature estimations agree well with isotopic temperatures.

Surface temperatures

The difficulty in estimating surface temperatures has been discussed in an earlier work (Vergnaud-Grazzini *et al.*, in press).

In the Bay of Biscay, near zone P6, there is a rise in "surface" isotopic temperatures, which reach their maximal values between zones P6 and P8. Thus, this maximum occurs earlier than the thermal maximum of bottom waters. If we use a δ¹⁸O value of -1.2‰ to calculate the temperatures of surface waters (it should be noted however that the evaporation-precipitation balance may have affected this parameter in surface), we obtain surface isotopic temperatures between 16.5

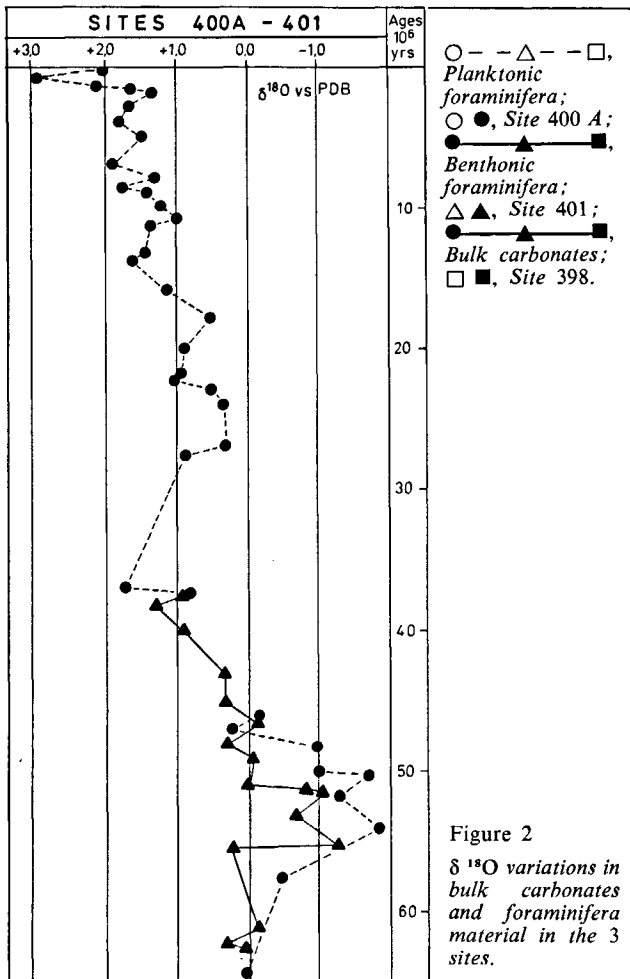


Figure 2
 $\delta^{18}\text{O}$ variations in
 bulk carbonates
 and foraminifera
 material in the 3
 sites.

and 20°C. These values are not substantially different from those prevailing today at these latitudes, in summer.

At site 398, the warmest surface temperatures that we may calculate for the early to middle Eocene are around 17°C. Although diagenesis and dissolution might have altered the original isotopic compositions, this temperature evaluation agrees well with the surface temperatures found in the Bay of Biscay.

The $^{18}\text{O}/^{16}\text{O}$ ratios start to increase near zone P8 at site 401, near zone P9 at site 400 A and near zone P14 at site 398. Since diagenesis has much affected sites 400 A and 398, these time-lags may only be artificial. This increase continues until the late Eocene, by which time temperatures have already dropped to values ranging between 12.5°C (as indicated by planktonic foraminifera) and 8°C (as indicated by bulk carbonates at site 401). At site 398, they drop to 10°C. Although the general trend during the Eocene is the same for different kind of planktonic material and indicates a cooling of the waters, the isotopic curves obtained for planktonic foraminifera and bulk carbonates of the same site are not exactly parallel. This may simply indicate that bulk carbonate is not an adequate material for analyses of this kind. Diagenetic alteration of coccoliths with calcitic overgrowths might explain the relatively high $\delta^{18}\text{O}$ values of bulk carbonates (Vergnaud Grazzini *et al.*, in press).

Thus, by the end of the Eocene epoch, temperatures had dropped through the whole water column by 4.5-5°C for bottom waters and 7-8°C for surface waters.

The same difference between the thermal evolution of deep and surface waters was found by Boersma and Shackleton (1977 *a*) at South Atlantic site 357. However, bottom temperatures were cooler in the South Atlantic than in the North Eastern Atlantic, where temperatures may have been influenced by the warm waters of the Tethys.

Carbon isotopic data

Carbon isotopic compositions are less sensitive than oxygen to diagenetic alterations. At each site, $\delta^{13}\text{C}$ decrease between -50 and -47 myrs for surface and deep biogenic carbonates alike. This decrease in $\delta^{13}\text{C}$ may be related to an increase in surface productivity resulting itself from the massive influx of cold waters which supplied large amounts of nutrients to the euphotic zones. During the middle Eocene, ostracod assemblages also indicate an increase in nutrients, and thus in productivity, in bottom waters (Ducasse, Peypouquet, in press). This decrease in $\delta^{13}\text{C}$ does not appear in carbon isotopic data reported by Shackleton and Kennett (1975) for the South Pacific Ocean. Sedimentological studies (Auffret, Pastouret, in press) also indicate increasing surface water productivity from the late Paleocene to middle Eocene in the Bay of Biscay. This coincides approximately with the Pyrenean orogeny.

The Oligocene - Miocene sequence

Bottom temperatures

During the early and middle Oligocene, isotopic compositions remain relatively stable at site 398. Since dissolution processes cannot have an isotopic selective effect, one may suppose when dealing with organisms which secrete calcite at the same temperature, that the isotopic temperatures calculated at site 398 for bottom waters correspond to the true temperatures. Further more, these isotopic compositions are of the same magnitude as those reported for equatorial latitudes (Boersma, Shackleton, 1977 *b*) and for southern middle latitudes (Boersma, Shackleton, 1977 *a*); they are slightly lower than 2‰ and correspond to temperatures around 4°C. The very sparse sampling of Oligocene sediments in the Bay of Biscay does not permit precise evaluation of the climatic fluctuations of this period. However, the few $\delta^{18}\text{O}$ values obtained are around 1‰, indicating that temperatures were higher in the Bay of Biscay than elsewhere in the Atlantic. This may be related to the influence of warm waters of Tethysian origin; the reason why this influence is not expressed at site 398 remains to be explained.

Later in the Oligocene, bottom temperatures drop, near zone P22-N4. This drop is much more pronounced in the Bay of Biscay, where temperatures reach values around 4.5-5°C. At site 398 there is a slight decrease and temperatures reach values around 3°C in the lower Miocene; this value agrees with the evaluations made in South Atlantic middle latitudes (Boersma, Shackleton, 1977 *a*).

During the early Miocene, near zone N6, bottom temperatures increase; the very strong decrease in $\delta^{18}\text{O}$ values observed through zones N8 to N12 at site 400 A may very well be an artefact resulting from diagenetic alteration.

During the late middle Miocene, temperatures decrease again in the Bay of Biscay and the very high $\delta^{18}\text{O}$ values observed near zone N15 may result from the building up of the Antarctic ice cap. We can arrive at a plausible evaluation of bottom temperatures only by assuming that the isotopic composition of the sea water was higher than -1.2‰ in the PDB scale.

Surface temperatures

The scanty information available for the Oligocene epoch indicates that all sites yield temperatures warmer than those of today. At site 398, the $\delta^{18}\text{O}$ values obtained for planktonic foraminifera are nearly superimposed with those obtained from benthonic foraminifera, indicating a strong dissolution effect. At this site, during the late Oligocene, $^{18}\text{O}/^{16}\text{O}$ ratios in bulk carbonates indicate a slight cooling, while at site 400 A, bulk carbonates indicate a warming of surface waters (near zones N3-N4).

During the Miocene, in the Bay of Biscay, the oxygen isotopic composition for planktonic foraminifera as well as for bulk carbonates continue to increase by stages. $^{18}\text{O}/^{16}\text{O}$ ratios measured on bulk carbonates remain higher than those measured on planktonic foraminifera at site 400 A. Since, after the early middle Miocene, the $^{18}\text{O}/^{16}\text{O}$ ratio variations are determined both by the building up of the ice cap at the poles and by variations in sea water temperature, the paleotemperature curves are difficult to interpret.

At site 398, this trend towards higher $\delta^{18}\text{O}$ values throughout the Miocene period is not evident, as a result of strong diagenetic processes. However, at all sites, we may note two large oscillations towards warmer temperatures, corresponding to the zones N7-N9 ($-16/-18$ myrs) and N16 ($-9/-10$ myrs).

Berggren and Hollister (1974), Shackleton and Kennett (1975), Barker and Burrell (1977) and Savin (1977) have stated that a number of major events affecting oceanic circulation occurred during the Miocene: setting of the circum-antarctic circulation towards -23 myrs; closure of the Tethys towards -16 myrs; greater extension of the Southern ice caps during the middle Miocene (-11 to -12 myrs). The warming postulated $-16/-18$ myrs ago could be related to the closure of the Tethys; according to Berggren (1972) this would have strengthened the Gulf Stream gyre, and a North Atlantic current might have developed and warmed the West coasts of Europe. The apparent cooling occurring $-11/-12$ myrs ago which is recorded both by planktonic and benthic fossils, could also correspond to the growth of the Antarctic ice cap. Total disappearance at these levels of warm nannoplanktonic forms such as Discoasters (Muller, in press) accords well with this interpretation.

Carbon isotopic data

A number of general features emerge which also appear in the $\delta^{13}\text{C}$ curves published for Pacific sites (Margolis *et al.*, 1977) and South Atlantic site 357 (Boersma, Shackleton, 1977 *a*). These features, which show up well in the isotopic curves obtained from bulk carbonates at sites 398 and 400 A and from foraminifera at site 400 A, include:

- heavy values in the middle Miocene ($-14/-16$ myrs);
- lighter ^{13}C contents in surface waters during the period starting at about $-6/-8$ myrs (late Miocene to present).

The fact that the lightest ^{13}C contents are to be found in a period starting at ca. -8 myrs may be related to the influx of cold waters of Northern and Southern origin, due to the growth of polar ice caps. These cold waters may have been rich in nutrients. This is also the period of the Alpine orogeny.

The Pliocene and Pleistocene sequence

The oscillations observed in the isotopic curves are the result both of temperature changes and of the growing ice sheets, first at the South Pole and then at the North Pole.

The strong $\delta^{18}\text{O}$ increase observed in deep waters only and which culminates at ca. -4 myrs in the Bay of Biscay, may reflect the development of an Antarctic ice sheet much thicker than at present. It also correlates with a major regression (Kennett, 1967; Adams *et al.*, 1977) and a global marine cooling (Kennett, Vella, 1975).

Lower temperatures also prevail in surface waters during the upper Pliocene. At site 400 A, a greater decrease in surface water temperature near the NN16 zone is indicated by the near extinction of Discoasters in the North Atlantic (Muller, in press).

The highest $^{18}\text{O}/^{16}\text{O}$ ratios are reached in the Pleistocene epoch ($+4.24$ for *Uvigerina* at site 398, $+2.21$ for *Orbulina universa* at site 400 A). Detailed $\delta^{18}\text{O}$ oscillations which occurred in bulk carbonates during the Pleistocene epoch and which may be deduced from an examination of Table 6, have not been reported on the paleotemperature curves. This indicates that surface waters may have undergone thermal oscillations as high as 7.5°C (Letolle *et al.*, in press). This value closely agrees with an estimation by McIntyre *et al.* (1976) of summer temperatures 8°C lower at 18 000 years than those of the present day at similar latitudes. This high thermal effect may be related to the immediate proximity of the Northern ice shelf (McIntyre *et al.*, 1976).

Large variations in the ^{13}C signal may also be related to the Northern ice sheet pulsations, with increasing upwelling during glacial advances.

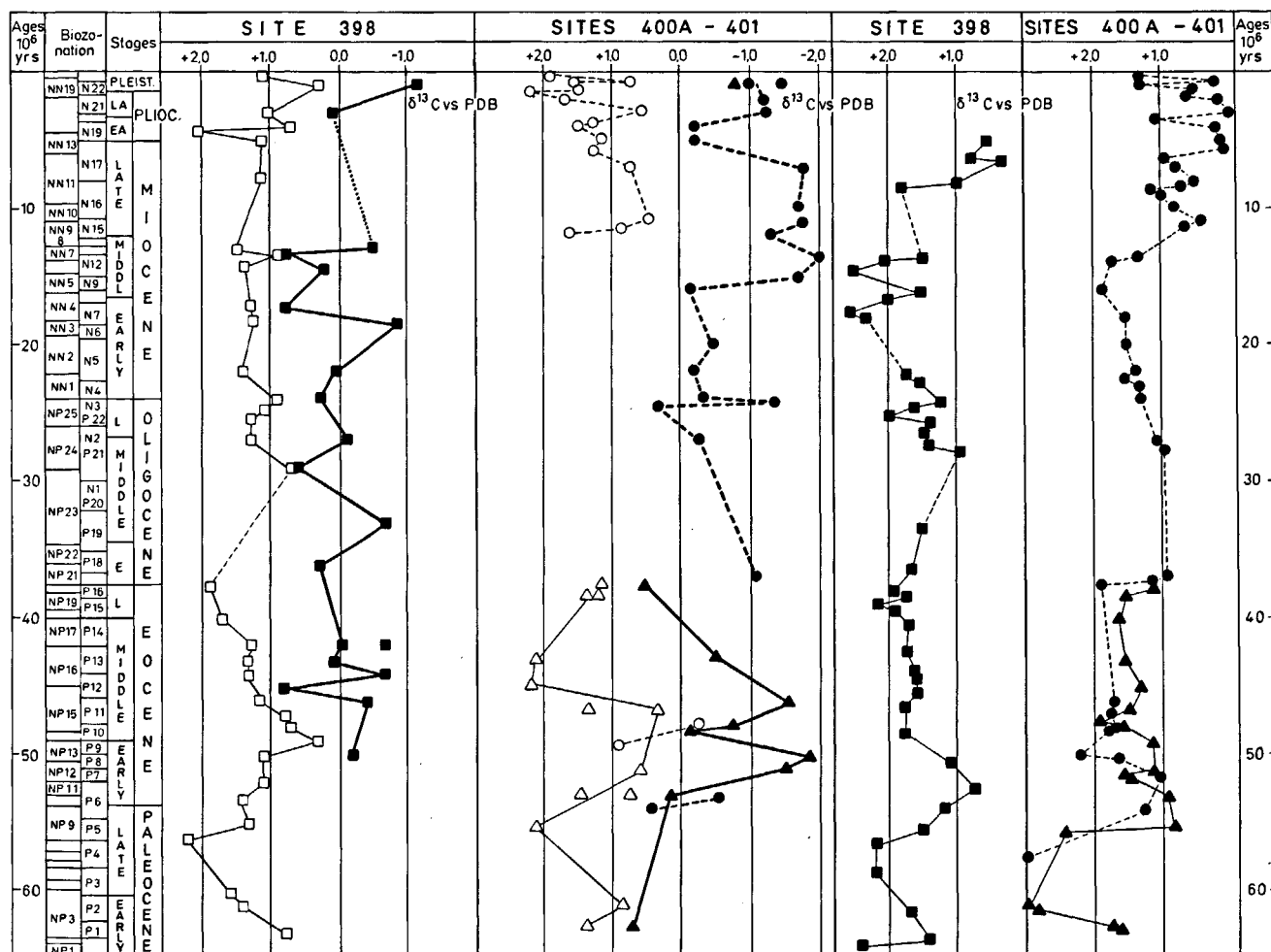
CONCLUSIONS

Oxygen and carbon isotope analyses performed on biogenic carbonates from sites 398, 400 A and 401 were used to reconstruct the paleoclimatic history of the North Eastern Atlantic during the Cenozoic era.

After a warming up during the Paleocene-early to middle Eocene, occurring in surface as well as bottom waters, the paleoclimatic evolution of this part of the Atlantic corresponds to a net cooling, with an average drop in temperatures of 10°C , interrupted by warmer episodes.

The major changes in isotopic temperatures are related to major paleo-oceanographic events:

- during the middle Eocene ca., surface and bottom temperatures start to decrease. This may correspond



○ — △ — □, Planktonic foraminifera; ● — ▲ — ■, Benthonic foraminifera; ● — ▲ — ■, Bulk carbonates; ○ ●, Site 400 A; △ ▲, Site 401; □ ■, Site 398.

Figure 3
δ¹³C variations in bulk carbonates and foraminifera material in the 3 sites.

to the establishment of a cold deep circulation. The fact that bottom temperatures remained higher than elsewhere might indicate that in the North Atlantic, cold Antarctic bottom waters were in competition with warmer waters of different origin such as the Tethys Ocean or the Labrador Sea. The early and middle Eocene also correspond to an increase in productivity, depicted by low ¹³C contents in biogenic carbonates, high surface-to-bottom δ¹³C gradient, and corroborated by the occurrence of precise ostracods species. This increase in productivity indicates that nutrients may have been supplied at some dates either by the influx of cold deep waters or through terrigenous supply (related to the vicinity of the Pyrenean orogeny);

– isotopic paleotemperature does not appear to drop drastically at the Eocene-Oligocene boundary in the Bay of Biscay. Oligocene temperatures are cooler than those of the Eocene, and bottom temperatures are lower than 8°C in the Bay of Biscay, lower than 4°C at site 398. At this time, a cold fauna of ostracods appears, corresponding to the “psychrospheric assemblages”. Thus, at that time, a cold deep circulation was well established;

– during the Miocene and Pliocene epochs, isotopic paleotemperatures continue to decrease. Some warmer episodes may be observed, however, at the Oligo-Miocene boundary in surface waters, between 14-18 and

8-10 myrs ca. Although, in this part of the Atlantic, the isotopic effect of ice accumulating on the Antarctic continent may have been superimposed with the isotopic effect corresponding to the definitive opening of Northern basins (by subsidence of the Iceland-Faeroe Ridge) which allowed cold Arctic bottom waters to flow into the North Atlantic, this is not clearly depicted on our isotopic curves. The cooling of the middle-late Miocene, however (ca. -11/-12 myrs) is probably due to the substantial growth of the Antarctic ice sheet at this time. During the Pleistocene, northern hemisphere glaciations were responsible for both large thermal oscillations and isotopic changes of the water.

We should add that diagenetic processes obliterate the isotopic records to an extent that cannot be predicted here, and prohibit any further discussion on detailed isotopic fluctuations of minor amplitude.

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Table 1

Oxygen and carbon isotope ratios * for planktonic foraminifera at Dsdp hole 398.

Ages (zones)	Samples (levels in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	× Dissolution + Diagenesis
398					
NN 20	2-3, 84	<i>Orbulina</i>	+1.64	+1.16	
	3-5, 20	<i>Globoquadrina</i> + <i>Globorotalia</i>	+2.52	+0.27	
NN 19	4-3, 125	<i>Orbulina</i>	+1.12	+0.85	
398 B					
NN 12	1-3, 51	<i>Orbulina</i>	+0.54	+0.75	
398 D					
N 19-20	1-6, 96	<i>Orbulina</i>	+0.65	+1.86	
	2-2, 65	<i>Orbulina</i>	+1.11	+1.46	
NN 10	2-4, 20	<i>Orbulina</i>	+1.29	+1.10	
	4-6, 141	<i>Orbulina</i>	+0.13	+1.12	
N 6-N 8		<i>Globorotalia</i>	+0.33	+1.10	
		<i>Globoquadrina</i>	+0.16	+1.14	
	5-2, 24	Mixed species	+1.34	+1.48	
	5-2, 124	<i>Globoquadrina</i>	+1.18	+0.86	
	6-5, 69	Mixed species	+1.17	+1.39	
	7-2, 74	Mixed species	+1.46		×
	8-1, 123	Mixed species	+1.85	+1.13	
	8-3, 76	Mixed species	+0.43	+1.11	×
	9-1, 25	Mixed species	+1.42	+1.39	
	12-1, 106	Mixed species	+1.53	+0.96	
NP 24/25	12-3, 94	Mixed species	+1.62	+1.07	
	13-6, 30	Mixed species	{ +0.12 +0.20 } { +0.26 +0.29 }		× ×
N 2	15-2, 11	Mixed species	+1.84	+1.23	
	15-3, 13	Mixed species	+1.80	+0.64	
NP 22/21	20-3, 135	Mixed species	+0.71	+1.95	× +
	22-2, 120	Mixed species	-0.39	+1.68	
	23-3, 113	Mixed species	-1.18	+1.24	
NP 16	24-1, 111	Mixed species	-0.59	+1.31	
	26-1, 64	Mixed species	-0.77	+1.33	
	28-5, 102	Mixed species	-0.64	+1.15	
	29-5, 140	Mixed species	-0.40	+0.83	+
NP 13-14	30-4, 54	Mixed species	-0.14	+0.69	× +
	31-1, 33	Mixed species	-0.42	+0.26	×
	32-3, 121	Mixed species	+0.27	+1.11	
	33-5, 20	Mixed species	+0.49	+1.06	
NP 9	34-6, 95	Mixed species	+0.09	+1.38	
	35-2, 16	Mixed species	-0.12	+1.29	
NP 7	37-3, 113	Mixed species	-0.39	+2.24	+
NP 3	39-3, 131	Mixed species	-0.36	+1.61	+
NP 1-2	40-3, 84	Mixed species	+0.83	+1.43	×
	41-1, 7	Mixed species	+0.57	+0.78	+

* Expressed as deviations per mil from PDB standard.

Table 2

Oxygen and carbon isotope ratios * for benthonic foraminifera at Dsdp hole 398.

Ages (zones)	Samples (levels in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	Dissolution
398					
NN 20	2-3, 84	<i>Melonis</i>	+3.17	-	
	3-5, 20	<i>Uvigerina</i>	+4.24	-1.16	
NN 19	4-3, 125	{ <i>Pyrgo</i> }	+2.89	+0.14	
		{ + <i>Melonis</i> }			
398 D					
N 19-20	1-6, 96	<i>Gyroidina</i>	+1.93	-	
	2-4, 20	{ <i>Melonis</i> } { + <i>Uvigerina</i> }	+1.64	-	
N 12-13	5-2, 24	Mixed species	+2.04	+0.62	
	5-2, 124	Mixed species	+1.57	-0.70	
	6-5, 69	Mixed species	+1.30	+0.02	
N 6-N 8	8-1, 123	Mixed species	+1.80	+0.58	
	8-3, 76	Mixed species	+0.86	-0.09	×
	9-1, 25	Mixed species	+2.31	+0.03	
	12-3, 94	Mixed species	+2.14	+0.31	
N 2	15-2, 11	Mixed species	+1.43	-0.27	
	15-3, 13	Mixed species	+1.80	+0.64	
	17-3, 23	Mixed species	+1.65	-0.71	
	19-4, 79	Mixed species	+1.72	+0.34	
	20-3, 135	Mixed species	+1.38		
	22-2, 120	Mixed species	-1.14	-	
	23-3, 113	Mixed species	-0.93	-0.07	×
	24-1, 111	Mixed species	-1.12	+0.05	
	26-1, 64	Mixed species	-1.20	-0.73	
	27-4, 117	Mixed species	+0.03	+0.84	
	28-5, 102	Mixed species	-0.43	-0.39	
	29-5, 140	Mixed species	-0.33	-	
31-1, 33	Mixed species	-1.58	-		
32-3, 121	Mixed species	-0.52	-0.23		

* Expressed as deviation per mil from PDB standard.

Table 3

Carbon and oxygen isotopic data for bulk carbonates at Dsdp site 398 D.

Sample (interval in cm)	$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)	Sample (interval in cm)	$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)
2.2.41.45	+0.56	+1.32	19.3.50.51	+1.66	+1.19
2.2.65.67	-0.10	+0.73	20.1.80.82	+1.90	-0.75
2.3.10.14	+1.11	+1.20	20.3.67.70	+1.74	+0.55
3.1.96.98	+1.08	+1.31	20.3.118.120	+2.15	-0.94
3.2.70.71	+0.80	+0.84	21.4.49.51	+1.90	-0.48
3.3.44.45	+1.14	+0.70	22.3.23.25	+1.66	-0.03
3.3.65.67	+1.29	+1.26	23.2.41.45	+1.72	-0.62
3.4.68.70	+1.02	+0.84	25.1.102.104	+1.61	-0.33
3.5.62.64	+0.25	+0.81	26.2.39.41	+1.57	-0.02
3.6.72.74	+0.37	+0.07	27.4.89.91	+1.55	-0.27
4.2.60.62	+0.96	+0.94	28.2.54.56	+1.75	-0.80
4.3.107.111	+1.80	+1.30	30.4.132.134	+1.74	-1.12
4.5.107.111	+1.87	+1.04	32.1.95.97	+1.04	-1.60
5.2.40.42	+1.53	+1.35	33.2.141.142	+0.71	-0.93
5.2.60.62	+1.63	+1.08	34.1.53.55	+1.16	-0.44
5.4.44.46	+1.60	+1.44	35.2.77.79	+1.44	-0.75
5.5.130.134	+2.04	+1.67	36.2.100.102	+2.55	-0.85
6.3.40.42	+2.40	+1.32	37.3.69.71	+2.12	-1.40
6.3.60.62	+2.60	+1.52	38.2.56.58	+2.15	-0.45
6.4.31.33	+2.65	+1.68	40.1.9.11	+1.63	-0.11
7.2.58.60	+1.54	+1.03	41.2.8.10	+1.38	-0.25
7.5.80.82	+2.01	+1.01	41.2.31.33	+0.90	-0.26
8.2.50.51	+2.56	+1.53	41.2.38.40	+1.61	-1.08
8.5.77.79	+2.29	+1.34	41.2.45.47	+2.19	-0.27
9.3.99.101	+1.73	+1.42	41.2.65.68	+2.24	-0.37
9.5.79.81	+1.50	+1.07	41.2.131.133	+2.38	+0.12
12.1.110.111	+1.20	+1.14	43.2.16.18	+2.53	-0.43
12.2.100.101	+1.72	+1.63	45.1.18.20	+2.02	-0.59
12.3.44.47	+1.58	+0.60	46.2.28.30	+1.69	-0.43
12.5.45.46	+1.69	+1.55	48.1.13.15	+1.83	-0.74
12.5.100.101	+1.97	+1.63	49.1.36.38	+1.40	-0.67
13.2.60.61	+1.35	+1.17	56.5.88.90	+1.61	-2.62
13.3.62.64	+1.54	+1.11	103.3.29.31	+1.66	-1.38
13.5.66.68	+1.12	+1.56	106.1.36.38	-0.38	-1.11
14.1.35.37	+1.53	+1.54	133.2.51.53	+1.31	-1.77
15.2.45.46	+1.37	+1.30	134.2.78.80	+1.50	-1.47
15.3.67.69	+0.99	+0.79	135.2.27.29	+1.09	-1.74
15.5.62.64	+0.90	+0.85	137.2.24.26	+1.10	-1.46
17.1.32.34	+1.51	+0.97	138.1.38.40	+1.08	-1.57

Table 4

Oxygen and carbon isotopic data for planktonic foraminifera at Dsdp site 400 A*.

Ages	Sample (interval in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
NN 19	1.1,74	<i>Orbulina universa</i>	+0.58	+1.90
	2.2,78	<i>Orbulina universa</i>	-1.00	+0.69
	2.4,56	<i>Orbulina universa</i>	+1.95	+1.52
	2.6,60	<i>Orbulina universa</i>	+2.21	+1.43
	3.4,84	<i>Orbulina universa</i>	+1.95	+2.22
	5.1,26	<i>Orbulina universa</i>	+1.44	+1.56
	10.1,57	<i>Orbulina universa</i>	+0.59	+0.50
	13.4,64	<i>Orbulina universa</i>	+0.94	+1.24
			<i>Orbulina universa</i>	+1.38
NN 12	14.3,22	<i>Orbulina universa</i>	+0.62	+1.07
	16.1,143	<i>Orbulina universa</i>	+1.00	+1.27
	18.3,52	<i>Orbulina universa</i>	-0.13	+0.73
	24.6,67	<i>Orbulina universa</i>	-0.56	+0.42
NN 9	25 cc	<i>Orbulina universa</i>	+0.64	+0.86
	26.1,114	{ <i>Globigerinita glutinata</i> }	-0.52	+1.61
NP 15	50 cc	Mixed <i>Globigerina</i>	-0.27	-0.34
	53 cc	{Mixed planktonic species}	-1.13	+0.93
Cretaceous	62.6,8	{ <i>Rotalipora ticinensis</i> }	-1.57	+1.01

* Expressed as deviation per mil from PDB.

Table 5

Oxygen and carbon isotopic data for benthonic foraminifera at Dsdp site 400 A.

Ages	Sample (interval in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	
NN 19	2.2,114	<i>Melonis</i>	+3.16	-1.43	
	2.2,114	<i>Oridorsalis</i>	+3.51	-0.98	
NN 12	5.1,86	Mixed species	+2.28	-1.15	
	10.1,57	<i>Melonis</i>	+2.63		
	10.1,57	<i>Oridorsalis</i>	+2.49	-1.25	
	13.4,64	Mixed species	+3.59	-0.18	
	14.3,22	Mixed species	+3.14	-0.21	
	18.3,52	<i>Oridorsalis</i>	+1.24	-1.83	
	23 cc	<i>Oridorsalis</i> + <i>Melonis</i>	+1.79	-1.70	
	24.6,67	<i>Oridorsalis</i> + <i>Melonis</i>	+1.11	-1.75	
	26.3,65	<i>Oridorsalis</i> + <i>Melonis</i>	+2.63	-1.28	
	29.1,114	<i>Oridorsalis</i> + <i>Melonis</i>	+1.61	-0.33	
NN 10	30.1,98	<i>Oridorsalis</i>	+0.36	-1.98	
	34 cc	<i>Oridorsalis</i> + <i>Melonis</i>	-0.80	-1.73	
	35.3	Mixed species	-0.55	-0.15	
	NN 2	37.1	<i>Oridorsalis</i>	+1.23	-0.49
		39.1,117	<i>Oridorsalis</i>	+1.72	-0.22
	43.2,65	<i>Oridorsalis</i>	+1.69	-0.36	
	43.5,65	Mixed species	+1.18	-1.41	
	43 cc	Mixed species	+1.64	+0.34	
	44.1,69	Mixed species	+0.94	-0.32	
	45.5,20	Mixed species	+0.94	+0.13	
NP 21	45 cc	<i>Oridorsalis</i>	+1.03	-1.10	
	47.6,97	<i>Oridorsalis</i> + <i>Melonis</i>	-0.24	-1.58	
	51.6,93	<i>Nuttalides</i>	-1.77	-0.13	
	54 cc	<i>Oridorsalis</i> + <i>Nuttalides</i>	-1.39	-1.88	
NP 11	56.2,60	<i>Oridorsalis</i> + <i>Nuttalides</i>	-0.94	-0.58	
	57.2	<i>Melonis</i> + <i>Oridorsalis</i>	+0.93	+0.41	

Table 6

Oxygen and carbon isotopic data for bulk carbonates at Dsdp site 400 A.

Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
1.1.13.14	+0.39	+0.83	4.4.90	+0.09	+1.37
1.1.48.50	+1.29	+2.01	4.5.16	-0.12	+1.39
1.2.16.18	+1.83	+2.07	4.5.148	-0.52	+1.35
1.5.86	-0.25	-0.87	4 cc	-0.31	+1.44
2.1.54	+0.48	-2.31	5.1.30.31	+0.16	+1.36
2.1.91	+0.60	+1.09	5.1.50	-0.94	+0.16
2.2.30	+0.19	+0.12	5.1.148	+0.32	+1.79
2.2.54.56	+0.17	+3.04	5.2.35		+1.91
2.2.100	+0.34	+1.53	6.1.130	+0.24	+2.51
2.3.54.56	+0.91	+3.00	6.2.30	+0.59	+2.08
2.3.8	+0.42	+0.62	6.2.100	-0.24	+1.72
2.4.15	+0.75	+1.63	6.3.30		+1.87
2.4.54.56	+1.29	+4.20	6.3.100	+0.17	+1.71
2.4.108	-0.08	+1.55	6 cc		+1.88
2.5.54.56	+0.89	+2.17	7.1.140	+0.42	+1.72
2.5.90	+0.61	+1.95	7.2.30	+0.28	+1.93
2.6.15	+0.31	+1.86	7.2.110	+0.72	+1.94
2.6.60.62	+0.51	+2.12	7.3.30.32	+0.23	+1.49
3.1.84.86	+0.67	+2.29	8.1.100	-0.23	+1.67
3.1.90	+0.28	+1.75	8.3.34.36	+0.49	+1.37
3.2.40	+0.52	+2.06	8.3.40	+0.07	+1.69
3.2.84.86	+1.08	+2.97	8.5.112.114	+0.30	+1.85
3.2.100	+0.45	+1.54	9.3.30	-0.21	+1.20
3.3.50	+0.51	+2.25	9.3.40.42	+0.53	+1.58
3.3.78.79	-0.52	+0.11	9.5.100	-0.07	+1.29
3.3.84.86	+0.24	+3.86	9.6.100	+0.48	+2.54
3.3.100	-0.14	+1.07	10.2.118.120	-0.07	+1.66
3.4.30.32	-0.17	+1.61	11.2.18.20	+1.08	+1.76
3.4.84.86	+0.39	+1.64	12.3.47.49	+0.81	+2.02
3.4.90	+0.27	+1.80	13.4.57.59	+0.23	+1.77
3 cc	+0.06	+1.09	14.1.63.65	+0.17	+1.45
4.1.75	+0.15	+1.59	15.2.60.62	+0.12	+1.45
4.2.90.92	-0.23	+2.32	17.2.113.115	+0.94	+1.52
4.3.100.102	+0.16	+1.60	18.3.73.75	+0.78	+1.89
4.4.50	+0.69	-1.31	19.2.86.88	+0.52	+1.30

Table 6 (continued)

Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
20.3.95.97	+0.68	+1.76	55.2.89.91	+1.01	-1.34
21.4.43.47	+1.16	+1.74	57.2.17.19	+1.21	-1.91
22.2.65.67	+1.00	+1.42	59.1.23.24	+2.96	-0.52
23.3.90.92	+0.77	+1.21	59.1.68.71	+2.90	-0.54
24.4.59.61	+0.43	+0.97	60.6.141.143	+2.54	0.00
25.6.109.111	+0.63	+1.35	60.7.19.21	+2.43	-0.25
30.1.58.60	+1.35	+1.42	61 cc	+2.58	-0.24
31.1.50.52	+1.74	+1.61	62.1.16	+2.05	-1.65
35.2.62.64	+1.86	+1.16	62.1.49.52	+1.84	-2.27
36.1.45.47	+1.54	+0.53	62.1.70.73	+1.80	-2.55
37.1.31.32	+1.49	+0.85	62.2.24.25	+1.80	-2.31
39.1.135.137	+1.36	+0.94	62.5.24.27	+2.11	-2.10
40 cc.71.76	+1.47	+0.98	62.6.8.11	+2.21	-2.10
41.1.16.18	+1.29	+0.49	63.1.104.105	+1.98	-2.31
43.3.104.106	+1.27	+0.29	63.3.24.27	+2.33	-2.28
44.1.85.87	+1.06	+0.31	63.4.109.110	+1.94	-1.38
44.1.105.106	+0.92	+0.85	64.5.70.71	+2.25	-2.51
45.2.78.80	+0.90	+1.74	66.1.13.15	+2.60	-1.96
45.3.110.112	+1.09	+0.79	66.1.33.35	+3.01	-1.69
46.4.71.73	+1.80	+0.98	68.3.9.10	+2.90	-2.25
46.5.140.142	+2.13	+0.87	69.1.103.104	+3.79	-1.40
47.6.106.108	+1.66	-0.22	69 cc	+4.15	-1.26
48.2.109.111	+1.72	+0.20	71.1.98.99	+2.34	-1.48
50.4.119.121	+1.63	-0.70	71.2.36.37	+2.03	-1.82
51.4.137.141	+1.77	-1.00	72.4.1.2	+2.51	-1.43
53.1.69.73	+2.17	-1.05	74.1.22.23	+2.10	-2.91
54.3.103.105	+1.62	-1.70			

Table 7

Oxygen and carbon isotopic data for planktonic foraminifera at Dsdp site 401*.

Ages	Sample (interval in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
NP 21	2 cc	<i>Catapsydrax unicava</i>	-0.05	+1.15
		<i>Globigerinatheca index</i>	-0.46	-1.08
	3 cc	<i>Globigerinatheca index</i>	-0.69	+1.38
		<i>Catapsydrax unicava</i>	-0.07	+1.19
	4 cc	<i>Catapsydrax unicava</i>	-0.39	-1.08
		<i>Globigerinatheca index</i>	-1.29	+2.05
NP 15	5.3,119	<i>Globigerina linaperta</i>	-1.01	-
		<i>Globigerina linaperta</i>	-1.38	+2.09
	7.5,87	<i>Globigerinatheca index</i>	-1.83	+1.14
		<i>Globigerina linaperta</i>	-1.14	+0.30
9.6,72	<i>Globigerina linaperta</i>	-0.47	-1.18	
	<i>Globigerina linaperta</i>	-2.00	+0.60	
NP 10	11.4,95	<i>Globigerina linaperta</i>	-1.95	+0.73
		<i>Subbotina</i>	-1.41	+1.44
NP 9	14 cc	<i>Globigerina venezuelana</i> } + <i>Globigerina linaperta</i> }	-0.58	+2.12
		Mixed planktonic species	-0.24	+2.11
NP 3	16 cc	<i>Globigerina triloculinooides</i>	-1.71	+0.84
		<i>Subbotina</i>	-0.12	+1.41
	17 cc	Mixed species	-0.17	+1.40

* Expressed as deviation per mil from PDB.

Table 8

Oxygen and carbon isotopic data for benthonic Foraminifera at Dsdp site 401

Ages	Sample (interval in cm)	Species	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
NN 19	1.1,11	<i>Melonis</i>	+2.10	-0.78
NP 21	2 cc	<i>Nuttalides</i>	+1.13	+0.55
NP 19	3 cc	<i>Nuttalides</i> + <i>Oridorsalis</i>	+0.82	-
		<i>Oridorsalis</i>	+0.10	-
NP 15	5.3,119	<i>Oridorsalis</i> + <i>Nuttalides</i>	+1.26	-0.49
		<i>Nuttalides</i>	-0.25	-0.77
NP 10	9.6,72	Mixed species	-0.28	-1.57
		<i>Oridorsalis</i> + <i>Nuttalides</i>	+0.41	+0.10
NP 9	13 cc	<i>Oridorsalis</i> + <i>Nuttalides</i>	+1.63	+2.42
NP 3	14 cc	<i>Oridorsalis</i> + <i>Nuttalides</i>	+1.63	+2.42
NP 3	17 cc	Mixed species	+0.55	+0.73

Table 9

Oxygen and carbon isotopic data for bulk carbonates at Dsdp site 401.

Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Sample (interval in cm)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
2 cc	+1.10	+0.90	14.5.44.45	+2.41	+0.22
3.1.16.17	+1.54	+1.29	16.1.11.12	+3.02	-0.22
4 cc	+1.59	+0.87	16.3.9.10	+2.79	-
5.2.40.42	+1.56	+0.35	17.1.81.82	+1.73	+0.31
6.2.73.74	+1.27	+0.31	17.3.20.21	+1.61	-0.02
7.4.98.100	+1.44	-0.19	18.1.15.17	+2.58	-0.16
8.4.91.92	+1.87	-0.05	18.1.46.47	+2.41	-0.31
9.6.102.103	+1.59	+0.27	19.1.53.54	+2.68	-0.52
10.2.21.23	+1.15	-0.12	19.1.67.68	+2.67	-0.55
11.1.77.78	+1.15	0.00	19.1.122.123	+3.02	+0.39
11.2.139.140	+1.57	-0.86	20.1.10.12	+0.66	-0.77
11.4.36.37	+1.43	-1.07	21.1.12.13	+1.94	+0.12
13.3.79.80	+0.87	-0.68	23.1.3.4	+1.12	-0.33
14.3.71.72	+0.80	-1.31	24.1.10.11	-1.97	-0.42

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