## 17. PALEOTEMPERATURE HISTORY OF THE CENOZOIC AND THE INITIATION OF ANTARCTIC GLACIATION: OXYGEN AND CARBON ISOTOPE ANALYSES IN DSDP SITES 277, 279, AND 281

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## ABSTRACT

An oxygen and carbon isotopic history based on analyses of benthonic and planktonic foraminifera in three overlapping subantarctic sections is presented for the last 55 m.y. with a sampling interval of less than 1 m.y. Surface temperature at Site 277, on the Campbell Plateau, was about 19°C in the early Eocene, about 13°C in the middle Eocene, about 11°C in the late Eocene, and about 7°C in the Oligocene. Declines in temperature appear to have been rather rapid and are separated by episodes of relative temperature stability. Bottom temperature at Site 277 was about 1°C below surface temperature in the Paleocene and about 2°C below surface temperature in the Oligocene.

Site 279, on the Macquarie Ridge, records an early Miocene warming of over 2°C followed by a cooling and a second similar temperature rise in the middle Miocene. Bottom temperature at this somewhat deeper site was about 3°C below surface temperature and was probably as low as 4°C during part of the early Miocene. Comparisons between Sites 277 and 279 suggest that from the early Oligocene temperatures of deep water were low like the present day, implying that the mean annual temperature in high southern latitudes was near freezing by the beginning of the Oligocene (but certainly no earlier). From this time glaciers would have descended to sea level, and there would have been sea-ice production. If an ice sheet were present, it could not have been more than a small fraction of its present-day size.

Site 281, on the South Tasman Rise, extends the record into the middle and late Miocene during which time the major East Antarctic ice sheet accumulated. A significant rise in surface temperature during the late Miocene did not cause the melting of this ice sheet, demonstrating that by this time it had already achieved its present invulnerability to climatic change. Since temperatures during much of the Miocene were significantly above any Pleistocene values, it is extremely unlikely that any climatic change in the geologically near future will significantly affect the stability of the East Antarctic ice sheet.

#### INTRODUCTION

The purpose of stable isotope analyses from Leg 29 sites is to evaluate the high-latitude climatic record of the Cenozoic and to investigate the history of Antarctic glaciation in this interval. It has been established that the Antarctic ice sheet was present at about its present dimensions throughout the Pliocene and that the deep temperature structure of the oceans as we now know it was also established by the late Miocene (Shackleton and Kennett, Chapter 20, this volume). The purpose of studying Sites 277, 279, and 281 (Figure 1, Table 1) is to establish the earlier development of the Antarctic ice sheet and the related deep temperature structure of the oceans. The relatively close spacing of samples analyzed also provides a more detailed paleoclimatic sequence than the two main studies previously carried out (Devereux, 1967; Douglas and Savin, 1973).

TABLE 1 Location of Sites

Site	Latitude	Longitude	Water Depth (m)	Cored (m)	Recovered (m)
277	52°13.43'S	166°11.48'E	1222	434.5	258.5
279A	51°20.14'S	162°28.10'E	3368	110.0	79.8
281	47°59.84'S	147°45.85'E	1591	169.0	105.6

#### ANALYTICAL PROCEDURE

Samples of both planktonic and benthonic foraminifera were extracted from nearly every core; some cores were sampled twice. Monospecific samples of planktonic foraminifera were analyzed where possible. It was not feasible to analyze monospecific samples of benthonic foraminifera except in a few samples. Ages of samples are based on the site summaries (this volume)



and are principally based on abundant planktonic foraminifera and calcareous nannofossils.

Samples consisted of about 0.1 and 0.4 mg carbonate after cleaning. Adhering fine-grained matter was removed by ultrasonic cleaning in AnalR grade methanol. Samples were roasted in vacuo at 450°C for 30 min to eliminate possible volatile contaminants, and carbon dioxide evolved by the action of 100% orthophosphoric acid at 50°C (Shackleton and Opdyke, 1973; Shackleton, in press). Isotope analyses were performed in a VG Micromass 602C mass spectrometer and calibrated by analysis of standard carbonate samples under identical conditions. The overall reproducibility of the analytical system is estimated to be  $\pm 0.05\%$  on the basis of previous analyses of 100 aliquots of a standard carbonate. Calibration to the PDB standard for oxygen isotopes is effected via the Emiliani B-1 standard, assuming its isotopic composition versus PDB to be  $\pm 0.29$  per mil (Friedman, personal communication; Shackleton, in press). Calibration to the PDB standard for carbon isotopes is effected via NBS 20 standard, assuming its value to be -1.06 per mil versus PDB (Craig, 1957). Analytical results are given in Tables 2-7 and displayed in Figures 2-4.

CENOZOIC PALEOTEMPERATURES

 TABLE 2

 Oxygen and Carbon Isotopic Data for Planktonic

 Foraminifera at DSDP Site 277<sup>a</sup>

Sample (Interval in cm)	Species	<sup>δ18</sup> 0	$\delta_{13}$ C
1-5, 140	Mixed Globigerina	+1.12	+0.96
2-6, 140	Mixed Globigerina	+1.23	+0.96
5-6,140	Mixed Globigerina	-	
6-6, 140	Mixed Globigerina	+1.28	+0.93
7-6, 140	Mixed Globigerina	+1.27	+0.97
8-6, 140	Mixed Globigerina	+1.31	+1.08
9-6, 140	Mixed Globigerina	+0.89	+1.11
10-6, 140	Mixed Globigerina	+1.11	+1.18
11-6, 140	Mixed Globigerina	+1.19	+1.50
12-6, 140	Mixed Globigerina	+0.90	+1.32
13-6, 140	Mixed Globigerina	+1.18	+0.99
14-6, 140	Mixed Globigerina	+1.15	+1.12
15-5, 142	Mixed Globigerina	_	1
16-5, 140	Mixed Globigerina	+1.00	+1.23
17-6, 140	Mixed Globigerina	+1.01	+1.23
18-3, 140	Mixed Globigerina	+.136	+1.28
19-2, 140	Mixed Globigerina	+1.12	+1.54
20-6, 140	Mixed Globigerina	+0.37	+1.14
21-3, 140	Mixed Globigerina	+0.35	+1.29
22-3, 140	Globigerapsis index	+0.23	+2.77
23-1, 140	Globigerapsis index	-	-
24-3, 140	Mixed Globigerina and Globigerapsis index	+0.17	+1.28
25-2, 140	Mixed Globigerina and Globigerapsis index	+0.20	+1.89
26-4, 140	Globigerapsis index	-0.17	+2.38
27-1,92	G. index	_	-
28-2, 140	G. index	-0.40	+2.63
29-3, 140	G. index	-0.44	+2.16
30-5, 140	G. index	-0.14	+2.35
31-2, 140	G. index	-0.23	+2.78
32-3, 140	G. index	-0.42	+1.94
33-2, 140	G. index	-0.13	+2.03
34-1, 135	G. index	-0.50	+1.83
35-2, 125	Acarinina primitiva	-0.16	+1.66
36-3, 144	A. primitiva	-0.47	+2.17
37-3, 135	A. primitiva	-0.71	+1.73
38-3, 143	A. primitiva	-0.75	+1.93
39-2, 140	A primitiva	-1 39	+0.87
40-2, 113	A. primitiva	-0.84	+1 70
41-3, 101	A. primitiva	-1 13	+1.64
42-2, 122	Acarinina primitiva and similar forms	-1.80	+1.60
43-1, 118	Mixed Acarinina and Truncorotaloides	-1.85	+1.92
44-2, 136	Mixed "Globigerina"	-1.46	+1.22
45-6, 123	Mixed "Globigerina"	-1.55	+1.35

<sup>a</sup>Expressed as deviation per mil from PDB.

## ISOTOPIC COMPOSITION OF CENOZOIC OCEAN WATER

In order to estimate paleotemperatures from the oxygen isotope data, Equation 1 is used.

 $T = 16.9 - 4.38 (\delta_{\rm c} - \delta_{\rm w}) + 0.10 (\delta_{\rm c} - \delta_{\rm w})^2 \quad (1)$ 

This equation (Shackleton, in press) represents an expression of isotopic equilibrium between water and calcite (O'Neil et al., 1969) It is expressed in the same form as that derived by Epstein et al. (1953) in the first evaluation of the isotope method for paleotemperature estimation. This equation is used because it is valid for the low temperatures found at the ocean floor.

TABLE 3 Oxygen and Carbon Isotopic Data for Planktonic Foraminifera at DSDP Hole 279<sup>a</sup>

Sample (Interval in cm)	Species	$\delta_{18}$ o	$\delta_{13}$ C
2-1,96	Globorotalia cf. miozea	+0.84	+2.36
2-2,90	G. cf. miozea	+0.57	+2.14
3-1, 100	G. cf. miozea	+1.09	+2.63
3-4,90	G. cf. miozea	+0.80	+2.48
4-1,95	G. cf. miozea	+1.06	+2.57
4-2,60	G. cf. miozea	+0.77	+2.85
5-2, 147	G. cf. miozea	+0.46	+2.46
5-4,90	G. cf. miozea	+0.66	+2.45
6-1,84	G. cf. miozea	+0.51	+1.99
6-3,65	Globoquadrina dehiscens	+0.75	+1.65
10-1, 120	G. dehiscens	+1.00	+1.14
9-1,90	G. dehiscens		
9-3,90	G. dehiscens	-	
10-3,90	G. dehiscens	+1.12	+1.20
11-1,60	G. dehiscens	+1.26	+1.29
11-3, 55	G. dehiscens	+1.07	+1.29
8-1,90	Globorotalia sp.	_	

<sup>a</sup>Expressed as deviation per mil from PDB.

TABLE 4 Oxygen and Carbon Isotope Data for Planktonic Foraminifera at DSDP Site 281<sup>a</sup>

Sample (Interval in cm)	Species	<sup>δ18</sup> 0	$^{\delta 13}\mathrm{C}$
1, CC	Globorotalia inflata	+2.88	+0.79
2-1, 10	G. inflata	+2.16	+0.96
2-3,90	G. inflata	+2.77	+0.99
3-1, 10	G. puncticulata	+2.98	+1.19
3-3,90	G. puncticulata	+2.27	+1.35
5-1, 10	G. conomiozea	+1.98	+0.77
5-2,90	G. conomiozea	+1.75	+0.77
6-2, 50	G. cf. miotumida	+1.54	+1.99
6-3,90	G. cf. miotumida	+1.49	+1.41
7-2, 20	G. cf. hirsuta	+2.23	+1.02
7-3,90	G. cf. hirsuta	+2.05	+1.29
2, CC	G. cf. hirsuta	+1.92	+1.26
9-1, 10	G. cf. hirsuta	+1.62	+1.24
10-3, 53	G. cf. crassaformis	+1.64	+2.00
10-4,93	G. cf. crassaformis	+1.61	+2.09
11-1, 10	G. cf. peripheracuta	+0.97	+2.51
11-3,90	G. cf. peripheracuta	+0.94	+2.38
12-1, 10	G. cf. peripheracuta	+1.26	+1.58
12-3,90	G. cf. peripheracuta	+1.31	+1.65

<sup>a</sup>Expressed as deviation per mil from PDB.

In order to estimate temperatures, the past value of  $\delta_w$  (the isotopic composition of the water inhabited by the foraminifera) must be evaluated. Today, the overall average isotopic composition of ocean water, expressed in terms of PDB, is about -0.28 per mil (Craig, 1965). However, prior to the accumulation of the Antarctic ice sheet, the value was isotopically more negative, since the ice now stored in Antarctica is isotopically more negative. The magnitude of this effect was estimated by Emiliani (1955) at 0.3 per mil and by Craig (1965) at 0.50 per mil. Since then a great deal of new data has been obtained from Antarctica, and a better estimate may be made.

 TABLE 5

 Oxygen and Carbon Isotope Data for Benthonic

 Foraminifera at DSDP Site 277<sup>a</sup>

Sample (Interval in cm)	Species	δ18 <sub>0</sub>	$\delta_{13}$ C
1-5, 140	Uvigerina	+1.43	-0.30
2-6,140	Mixed benthics	_	-
3-6, 143	Mixed benthics	+1.73	-0.04
5-6, 140	Mixed benthics	+1.82	-0.17
6-6, 140	Mixed benthics	+1.73	-0.23
7-6, 140	Mixed benthics	+1.89	-0.40
8-6, 140	Mixed benthics	+1.57	+0.07
9-6, 140	Mixed benthics	+1.60	-0.15
10-6, 140	Mixed benthics	+1.77	-0.16
11-6, 140	Mixed benthics	+1.73	-0.34
12-6, 140	Mixed benthics	+1.43	-0.37
13-6, 140	Mixed benthics	+1.53	-0.12
14-6, 140	Mixed benthics	+1.48	+0.04
15-5, 142	Mixed benthics	+1.23	+0.22
16-5, 140	Mixed benthics	+1.41	+0.75
17-6, 140	Mixed benthics	+1.48	+0.60
18-6, 140	Mixed benthics	+1.69	+0.70
19-2, 140	Mixed benthics	+1.72	+0.70
20-6, 140	Mixed benthics	+0.71	+0.45
21-1,92	Mixed benthics	+0.77	+0.17
21-3, 140	Mixed benthics	+0.88	+0.58
22-3, 140	Mixed benthics	+0.92	+0.98
23-1, 140	Mixed benthics	+0.59	+0.70
23-1, 140	Uvigerina	+0.71	+0.56
24-3, 140	Mixed benthics	+0.62	+0.04
25-2, 140	Mixed benthics	+0.58	-0.05
24-3, 140		0.00	0.00
25-2, 140	Uvigerina	+0.93	+0.50
26-4, 140	Mixed benthics	+0.54	-0.28
26-4, 140	Uvigerina	$+2.00^{a}$	+0.94b
28-2, 140	Mixed benthics	+0.57	-0.15
29-3, 140	Mixed benthics	+0.61	-0.10
30-5, 140	Mixed benthics	+0.28	+0.51
31-2, 140	Mixed benthics	+0.31	+0.54
32-3, 140	Mixed benthics	-0.40	+0.66
33-2, 140	Mixed benthics	+0.44	+0.46
34-1, 135	Mixed benthics	-0.03	-0.01
35-1, 125	Mixed benthics	+0.23	+0.55
36-3, 135	Mixed benthics	+0.09	+0.51
37-3, 135	Mixed benthics	-0.45	+0.74
38-3, 143	Mixed benthics	-0.58	+0.96
39-2, 140	Mixed benthics	-1.19	+0.32
40-2, 113	Mixed benthics	-0.89	+1.29
41-3, 100	Mixed benthics	-0.77	+1.02
42-2, 122	Mixed benthics	-0.88	+0.19
43-1, 118	Mixed benthics	-0.64	-0.03
44-2, 136	Mixed benthics	-0.54	+0.10
45-6, 127	Mixed benthics	-1.01	+0.82

<sup>a</sup>Expressed as deviation per mil from PDB.

<sup>b</sup>Anomalous result (3 individuals) contamination suspected.

Olausson (1965) made an estimate of -40 per mil for the isotopic composition of Antarctic ice based on an average over the surface area of the ice sheet. Since the residence time for the most isotopically light ice formed in the center is considerably greater than the residence time for the ice formed on the periphery, this areal average must be less negative than the bulk average of the ice stored. Using the model illustrated in Figures 5 and 6, Shackleton (1967) estimated the isotopic composition of the stored ice to be about -50 per mil. This too may be an underestimate, since there may be an appreciable quantity of ice remaining from the last glacial that is isotopically more negative than predicted by this

TABLE 6 Oxygen and Carbon Isotope Data for Benthonic Foraminifera at DSDP Hole 279<sup>a</sup>

Sample (Interval in cm)	Species	<sup>δ18</sup> 0	$\delta_{13}$ C
2-1,96	Mixed benthics	+1.84	+0.98
22-2,90	Mixed benthics	+1.26	+0.69
3-1, 100	Uvigerina	+2.09	+1.27
3-4,90	Mixed benthics	+2.17	+0.37
4-1,95	Mixed benthics	+1.86	+1.29
4-2,60	Mixed benthics	+1.99	+0.81
5-2, 147	Mixed benthics	+1.69	+0.95
5-4,90	Mixed benthics	+1.55	+0.49
6-1,84	Mixed benthics	+1.50	+0.63
6-3,65	Mixed benthics	+1.46	+0.17
8-1,90	Mixed benthics	+1.98	+0.41
9-1,90	Mixed benthics	+2.02	-
9-3,90	Mixed benthics	+2.13	+0.09
10-1, 120	Mixed benthics	+1.82	
10-3,90	Mixed benthics	+1.78	+0.20
11-1,60	Mixed benthics	+2.05	-
11-3, 55	Mixed benthics	+1.91	

<sup>a</sup>Expressed as deviation per mil from PDB.

 TABLE 7

 Oxygen and Carbon Isotope Data for Benthonic

 Foraminifera at DSDP 281<sup>a</sup>

Sample (Interval in cm)	Species	δ18 <sub>0</sub>	$\delta_{13}$ C
1. CC	Mixed benthics	-	_
2-1, 10	Uvigerina	+3.29	+0.33
2-3,90	Uvigerina	+4.11	-0.46
3-1, 10	Mixed benthics	+3.58	-0.15
3-3,90	Mixed benthics	+2.93	-0.52
5-1, 10	Uvigerina + mixed benthics	+3.09	-0.08
5-2,90	Uvigerina	+2.85	-0.10
6-2, 50	Mixed benthics	+2.85	+0.90
6-3,90	Mixed benthics	+2.89	+0.91
7-2, 20	Mixed benthics	+2.88	-0.18
7-3,90	Mixed benthics	+2.50	+0.42
8. CC	Mixed benthics	+2.64	+0.56
9-3,90	Mixed benthics	+2.56	+0.44
10-3, 53	Uvigerina + mixed benthics	+2.07	+0.71
10-4,93	Uvigerina + mixed benthics	+1.94	+1.36
11-1, 10	Uvigerina + mixed benthics	+1.75	+0.86
11-3,90	Mixed benthics	+1.48	+0.98
12-1, 10	Mixed benthics	+1.71	+0.80
12-3,90	Mixed benthics	+1.65	+0.58

<sup>a</sup>Expressed as deviation per mil from PDB.

model. The result, however, cannot be seriously wrong, since it relies only on the fact that the bulk of the ice stored is composed of snow that fell well within the interior of the ice sheet, characterized by values between -50 per mil at the South Pole (Epstein et al., 1963) and -57 per mil at the Pole of Inaccessibility (Picciotto et al., 1967). Using a value of -50 per mil, the isotopic composition of the ocean prior to the formation of the presently existing ice sheets may be estimated at-1.2 per mil (Table 8).

It must be noted that the value of -1.2 per mil is expressed in relation to the PDB standard and represents a value of -1.0 on the SMOW standard. It is unlikely that





Figure 2. Oxygen isotope data for planktonic foraminifera at Sites 277, 279, and 281, from Tables 1, 2, 3, and 4, deviation per mil from PDB. Stratigraphy from site summaries (this volume). Stratigraphic plot is not based on thickness or age difference between samples, but samples are arbitrarily plotted at constant divisions. Surface temperature values calculated from Equation A, assuming δ<sub>W</sub> to be -1.28 per mil, valid from base of sequence to mid Miocene only. Above Core 10, Site 281, surface temperature estimation is complicated by ocean isotopic changes. (See Figure 8.)



Bottom Water Temperature

Figure 3. Oxygen isotope data for benthonic foraminifera at Sites 277, 279, and 281, from Tables 5, 6, and 7, deviation per mil from PDB. Stratigraphy from site summaries (this volume). Stratigraphic plot is not based on thickness or age difference between samples, but samples are arbitrarily plotted at constant divisions. Bottom temperature as each site calculated using Equation 1, assuming  $\delta_W$  to be -1.28 per mil, valid only from base of sequence in mid Miocene only. Above Core 10, Site 281, the record depicts change in ocean isotopic composition as the Antarctic ice sheet accumulated. From the deposition of Cores 281-7-2 upwards the Antarctic ice sheet was continuously present and changes in Core 281-3 and above represent changes in Northern Hemisphere glacial ice accumulation after t = 2.6 m.y. (Shackleton and Kennett, Chapter 20, this volume).



Figure 4. Carbon isotope data from planktonic and benthonic foraminifera at Sites 277; 279, and 281, deviation per mil from PDB. Stratigraphy from site summaries (this volume). Stratigraphic plot is not based on thickness or age difference between samples, but samples are arbitrarily plotted at constant divisions.



Figure 5. Profile of idealized Antarctic ice sheet, from the model of Nye (1959). Streamlines calculated on the basis of a relationship between annual accumulation and distance from the coast based on data in Cameron (1964).

this value is in error by more than 0.1 per mil, an insignificant amount in terms of paleotemperature estimation in the Tertiary; conversely it differs from the isotopic composition of the present-day ocean by an amount that would affect paleotemperature estimates by a significant 4°C. Thus it is suggested that this value should be used as a standard for paleotemperature estimation prior to the accumulation of the Antarctic ice sheet.

## Local Variations in the Isotopic Composition of Ocean Water

Regional variations in the surface evaporationprecipitation balance give rise to variations in the isotopic composition of seawater. The same factors give rise to changes in salinity. Thus there is a rough correlation between salinity and isotopic composition (Epstein and Mayeda, 1953; Craig and Gordon, 1965). So far as deep and bottom waters are concerned, such isotopic variations may be neglected in most situations.

At the present time in high latitudes, precipitation exceeds evaporation, so that the surface waters are isotopically lighter than mean ocean water; Craig and Gordon (1965) give values of about -1.0 per mil for salinities below 33º/00. Failure to consider such a figure could give rise to paleotemperature estimates as much as 4°C too high. Despite this, it is proposed to neglect this source of uncertainty, for two reasons. First, since latitudinal temperature gradients were lower during much of the Tertiary than today, it is not obvious how extreme the variations in the evaporation/precipitation balance may have been so that any estimate made would be unreliable. Second, another uncertainty of opposite direction arises from the fact that the foraminifera analyzed may have lived below the ocean surface and may register a lower temperature than the true surface value. There is some evidence (Douglas and Savin,



Figure 6. Profile of Antarctic ice sheet as in Figure 5, plotted on a horizontal scale varying as the square of distance from the center. Equal areas on the diagram correspond to equal volumes of ice, so that the volume average isotopic composition of Antarctic ice may be estimated from the average over the area of the figure. Temperatures based on Bently (1964). Isotopic composition between streamlines based on Fonfiantini et al. (1963), Picciotto et al. (1967), Epstein et al. (1965). The bulk average isotopic composition yeilded by this model is -50 per mil.

 TABLE 8

 Isotopic Composition of the Preglacial Ocean

Water Mass	Volume	Isotopic Composition
Antarctica	$24 \times 10^6 \text{ km}^3$	-50 per mil
Other ice	$2.6 \times 10^{6} \text{ km}^{3}$	-30 per mil
Present ocean	$1360 \times 10^{6} \text{ km}^{3}$	-0.28 per mil
Preglacial ocean	$1385.6 \times 10^{6} \text{ km}^{3}$	-1.2 per mil

1973), discussed below, that this source of error is considerably less important in the middle Miocene and earlier than today. The magnitude of the first uncertainty is severely limited by the fact that to take it into account would reduce the measured estimate of vertical temperature gradient, which is already very small. Thus, it is believed that the isotopic paleotemperatures illustrated in Figures 2, 3, 7, and 8 are much closer approximations to true surface temperature estimates than one can derive for the Pleistocene oceans using this method.

## OCEAN SURFACE PALEOTEMPERATURES: THE INFLUENCE OF DEPTH HABITAT AND OF DIFFERENTIAL SOLUTION

It is well known that in the Recent and Pleistocene oceans, there is a depth stratification among planktonic foraminiferal species. In order to estimate surface temperatures by the oxygen isotope method, a species should be selected that inhabits the surface water (Emiliani, 1954; 1955). Savin and Douglas (1973) have pointed out a more subtle aspect of this problem: the general tendency for foraminifera from deeper water to have greater resistance to solution than those that inhabited the surface is operative even within a species. Even if the identical species is analyzed in all samples, there is a likelihood that those samples that have suffered more solution will register lower isotope temperatures.

As a result, there are extensive areas of ocean over which it is not possible to estimate true Pleistocene surface temperatures by the oxygen isotope method, because it is not possible to isolate individuals that actually lived and deposited their tests close to the water surface. Nevertheless, Savin and Douglas (1973) have shown that isotope analyses provide a generally correct picture of surface temperature trends.

Douglas and Savin (1973) have also investigated whether this difficulty also applies to earlier sediments. They analyzed eight separate species from DSDP Site 167-5, CC (middle Miocene) and found a range in oxygen isotopic composition equivalent to 2.5°C. In contrast, they cite the results of several studies (Lidz et al., 1968; Hecht, 1971; Emiliani, 1971; Savin and Douglas, 1973) showing that in Pleistocene sediments a range of some 15°C may be found among different planktonic foraminiferal species. They conclude that uncertain depth habitat does not seriously contribute to the uncertainty in deriving Tertiary surface temperature from oxygen isotope data.

# Departure from Isotopic Equilibrium in Benthonic Foraminifera

Duplessy et al. (1970) have shown that there are systematic differences in isotopic composition between coexisting benthonic foraminifera. This can only be explained in terms of a departure from isotopic equilibrium in some or all species. Shackleton (in press) has shown that bathyal *Uvigerina* spp. deposit their carbonate in, or very close to, isotopic equilibrium. It is thus desirable to analyze monospecific samples where possible and to calibrate the departure from isotopic equilibrium of each species analyzed.

In the present study it was possible to isolate *Uvigerina* in only four samples. Of these, one yielded a clearly anomalous result; comparison of the analyses of the remainder, with the mixed benthic species analyzed at most levels, suggests that samples were selected whose departure from isotopic equilibrium is small, perhaps in

the range 0.2-0.4 per mil. If the observed variations in the planktonic-benthonic difference were attributed entirely to variations in departure from isotopic equilibrium, this would imply that the uncertainty caused by sample-to-sample differences in departure from isotopic equilibrium is about  $\pm 0.2$  per mil or less. Thus estimated bottom temperatures are possibly  $1\pm1^{\circ}$ C higher than actual temperatures.

In an attempt to minimize this source of uncertainty. a certain selection policy was adopted. First, Uvigerina or related genera were selected when present, and second, where possible, a dozen or so individuals of similar size were selected. Large individuals were avoided because these might bias the analysis in favor of the particular fractionation factor of one species. Forms commonly picked include Cibicides, Stilostomella, Gyroidina, Pleurostomella, Bulimina, Melonis, and others. No miliolids were included except for Pyrgo in a few samples. Although it will ultimately be of value to analyze monospecific samples of benthonic foraminifera in Tertiary sequences, it is concluded that the broad picture presented here, as well as by Douglas and Savin (1973) and by Saito and van Donk (personal communication), is substantially correct.

## HISTORY OF HIGH-LATITUDE SURFACE TEMPERATURE DURING THE CENOZOIC

Figure 2 shows the record of surface temperature at Site 277 (Paleocene to late Oligocene) and at Site 279 (early-middle Miocene). Salient features are as follows. In the late Paleocene and early Eocene, surface temperature was 18°C to 20°C. Although Douglas and Savin (1973) did not make the same assumptions about the isotopic composition of seawater, their measurements yield virtually the same temperature values at low latitudes. Apart from a single reversal in Core 39, Section 2, temperature in high southern latitudes appears to have dropped in a series of steps to a value of about 7°C in the Oligocene. The middle Eocene, the late Eocene, and the entire Oligocene, are each characterized by comparative climatic stability, separated by relatively brief intervals of climatic cooling near the boundaries. The most dramatic drop in temperature is closely associated with the Eocene-Oligocene boundary and has previously been recorded in New Zealand marine sections by Devereux (1967).

In the early Miocene there was a rise in surface temperature, followed by a fall and another rise early in the middle Miocene. During warmer episodes, temperatures appear to have been almost as high as during the late Eocene.

The temperature values estimated for Site 281 in the middle Miocene are a degree or so lower than those from Site 279. This is consistent with paleolatitude estimates based on the reconstructions of Molnar et al. (in preparation) and the Australian pole positions of McElhinny et al. (1974). On the basis of these data, D. P. McKenzie (personal communication) estimates a latitude of 49°S at 20 m.y.B.P. for Site 279 and about 52°S for Site 281. By 10 m.y.B.P. both sites had about the same latitude of 51°S.

Temperature estimation for the late Miocene is complicated by the change in ocean isotopic composition as the Antarctic ice accumulated. In Figure 8 the oxygen isotopic difference between planktonic and benthonic species is plotted for this interval of time. It is reasonable to assume that from the time the Antarctic ice started to accumulate, the bottom temperature at the site would have remained more or less constant, so that surface temperature may be estimated by way of the surface-to-bottom gradient. In Figure 8, a temperature scale has been placed on the assumption that there was insignificant Antarctic ice up to the deposition of the sediment in Core 10, Section 4 (middle middle Miocene). Using this assumption, it is apparent that, after the accumulation of the Antarctic ice sheet, the surface temperature at Site 281 rose during the late Miocene to about the same value as the middle Miocene. This occurred without significantly affecting either the ice sheet or the deep-water temperature, and thus is a striking demonstration of the stability of the ice sheet and its control over the oceans.

## HISTORY OF OCEAN DEEP-WATER TEMPERATURES DURING THE CENOZOIC

Figure 7 shows the temperature record at a depth of approximately 1000 meters, since the late Paleocene. Most of the record represents Site 277, but the Miocene portion is from Site 279. The depth at this site (3368 m) is substantially greater than at Site 277 (1222 m). In order to determine the temperature record for a uniform depth, the data from Site 279 have been adjusted by 0.45 per mil (cf. Figure 3). This value represents the difference between the planktonic-benthonic gradient at Sites 277 and 279. Since surface temperature values at the end of the Oligocene at Site 277 are rather similar to those at the beginning of the Miocene at Site 279, it is a reasonable assumption that the change in vertical gradient between the sites is due only to the difference in depth.

Apart from the early Eocene, there is a trend to lower deep temperatures through the Eocene, culminating in a value of about 5°C in the early Oligocene. If the value had not been corrected by assuming the absence of the Antarctic ice sheet, the value yielded at this point would have corresponded to a temperature of about 9°C. Such a value would of course be incompatible with the presence of a major ice sheet extending to the coast. Thus the assumption that the ice sheet was not present is strongly supported by observations.

Although no ice sheet had developed, it is clear that if the temperature at 1000 meters was already as low as 5°C at the beginning of the Oligocene, a value not substantially different from that prevailing today, the surface temperature at the Antarctic coast must have been near freezing. The dramatic drop in temperature near the Eocene-Oligocene boundary must therefore have produced glaciers at sea level and hence represents a critical stage in the development of Antarctic glacial history.

This dramatic temperature event also marks the onset of a deep-water circulation dominated, as today, by the formation of bottom water at a temperature close to freezing around the Antarctic continent. The close correspondence between the planktonic and benthonic isotope record through the Eocene also suggests that bottom water was probably being formed in high southern latitudes before the Oligocene, although not at such low temperatures.

The early Miocene section at Site 279 records an important event; a substantial temperature rise during which bottom temperatures rose about 2°C. This clearly implies that the mean annual sea-level temperature around the Antarctic continent rose significantly above 0°C during this interval, thus raising the temperature of the whole ocean bottom water. The history of this, and a second temperature rise early in the middle Miocene, may prove to have been more complex when more detailed studies are undertaken. After the second event, temperatures fell again, and by the beginning of the late Miocene, isotopic values are substantially more positive than those prevailing throughout the Oligocene.

### FORMATION OF THE ANTARCTIC ICE SHEET

The isotopic transition between Cores 281-10 (early middle Miocene) and 281-7 (early late Miocene) is interpreted to represent the relatively rapid accumulation of an Antarctic continental ice sheet, composed of isotopically light ice, and ultimately extending to sea level. We are unable to register any temperature change associated with this event, since before the ice accumulation bottom-water was already at a temperature limited by formation at about freezing in high latitudes. However, the ice clearly accumulated until, as is the case today, the dimensions of the East Antarctic ice sheet were limited by the coastline of the continent (Hollin, 1962). The West Antarctic ice sheet comprises a rather small proportion of the present-day ice sheet and its buildup would have much less marked effects compared with the East Antarctic ice sheet.

Support for a stable late Cenozoic ice sheet comes from the record, at Site 281, of a significant surface temperature rise later in the late Miocene. At this time despite a rise in surface temperatures to middle Miocene values, the bottom water is totally unaffected. The temperature rise, although substantial, was not sufficient to melt the entrenched ice sheet, so that the near-coast ocean surface from which the bottom water formed did not warm as did the ocean surface on the South Tasman Rise. In a similar manner, the Antarctic ice sheet has not been affected to any significant extent by climatic fluctuations that occurred during the Pleistocene.

It has been demonstrated that the East Antarctic ice sheet has existed in essentially its present form since early in the late Miocene. Its subsequent history has been discussed elsewhere (Shackleton and Kennett, Chapter 20, this volume). Apart from relatively warm episodes in the early and middle Miocene, the basic temperature structure of oceanic deep waters was probably established at the beginning of the Oligocene. The change which marks its inception appears to have been rapid, and indeed on present evidence may have been the most rapid climatic change in pre-Pleistocene times; its close association with an epoch boundary is probably not coincidental as rapid changes in planktonic and benthonic faunas would have resulted from such a climatic change (Bramlette, 1955).



Figure 7. Deep-water temperature record for a depth of about 1000 meters, late Paleocene to middle Miocene, based on Figure 3 after adjusting values from Site 279 by -0.45 per mil to reduce them to the equivalent depth at Site 277.



Figure 8. Surface temperature changes estimated from oxygen isotopic composition of planktonic foraminifera at Site 281, using changes in oxygen isotopic composition of benthonic foraminifera to correct for changes in ocean isotopic composition as ice sheets accumulated.

#### **CARBON ISOTOPE VARIATIONS**

Analytical results from carbon istopes in benthic and planktonic foraminifera are shown in Figure 4. Two general trends can be observed: First, with the exception of a section in the middle and late Eocene, there is a striking parallelism between the benthic and planktonic trends in carbon isotopic composition. Second, comparison with the data obtained by Douglas and Savin (1973, fig. 8) reveals a striking similarity both in trends and in absolute values.

As Douglas and Savin (1973) point out, the factors controlling the isotopic composition of the carbon in foraminiferal tests are complicated; little work has been done to elucidate the complications even in today's oceans, although studies are beginning to throw light on the variations in isotopic composition in dissolved carbonate in seawater (Duplessy, 1971). Nonetheless, it now appears certain that the trends shown in Figure 4 represent changes in the isotopic composition of the bulk of available carbon in the ocean reservoir, or at least in the Pacific Ocean reservoir. It is clearly desirable to collect data from other oceans to determine whether this phenomenon is one of changing isotopic distribution between the oceans, or whether the whole reservoir changed. Today, there is a difference in isotopic composition between dissolved carbon in Pacific and Atlantic deep waters amounting to nearly 1 per mil (Duplessy, 1971).

#### CONCLUSIONS

The surface temperature at Site 277 (originally at 50°S latitude) was about 20°C in the early Eocene, falling to about 13°C in the middle Eocene, and about 11°C in the late Eocene. Deep water at the site, now at a depth of a little over 1000 meters, was formed to the south and had a temperature about a degree lower.

At the beginning of the Oligocene a further temperature drop occurred, to about 7°C at the surface, while the deep-water temperature at this site dropped to between 5°C and 6°C. The ocean bottom waters by the early Oligocene thus developed at about the same temperature as today.

Temperatures rose by about 3°C during the early Miocene, followed by a fall, and a later rise at the beginning of the middle Miocene. After this, they fell once again and a major ice sheet accumulated and stabilized on the Antarctic continent. During a surface temperature rise of about 3°C in the late Miocene, the ice sheet did not melt, and the ocean deep-water temperature was not affected.

By the beginning of the Oligocene, glaciers on the Antarctic continent probably descended to sea level. Throughout much or all of the Eocene there were probably many glaciers at higher altitudes. Thus, the glacial history of Antarctica spans tens of millions of years, although the ice sheet as we know it today has only been present for a little over 10 m.y.

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