

25. STABLE ISOTOPE RECORD OF THE PLIOCENE AND PLEISTOCENE PLANKTONIC FORAMINIFERS FROM LEG 78A, SITES 541 AND 543, ON THE BARBADOS RIDGE COMPLEX¹

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ABSTRACT

The stable isotope study of monospecific planktonic foraminifer samples recovered at Sites 541 and 543 during Deep Sea Drilling Project Leg 78A indicates a warming during the early Pliocene about 4.7 to 4.3 Ma. The changes in the late Pliocene oxygen isotope record around 2.9 to 2.7 Ma coincide with changes in the circulation pattern resulting from the closure of the Panama seaway and the beginning of the Northern Hemisphere glaciation. The Pleistocene record is characterized by 0.5 to 1.0‰ fluctuations in the $\delta^{18}\text{O}$ record. These fluctuations reflect salinity changes, rather than temperature changes, as indicated by *Globigerinoides ruber* and *G. sacculifer* abundances. The salinity changes may be explained by a drifting of (1) the highly saline Central Water Mass of the southern Sargasso Sea, and (2) lower-salinity ocean water displaced by the northward shift of the Intertropical Convergence Zone into the Caribbean region during cooler intervals.

INTRODUCTION

Stable isotope studies are a valuable means of tracing the glacial history in late Cenozoic oceans (Emiliani, 1955; Shackleton and Opdyke, 1973; and many others). Changes in microfossil assemblages suggest a substantial cooling during the latest Miocene (Ingle, 1967; Kennett and Watkins, 1976). Oxygen isotope data support the postulate of an expansion of the Antarctic ice sheet during the late Miocene (Shackleton and Kennett, 1975). Shackleton and Opdyke (1977) postulated that the arctic glaciation started in the later Pliocene (3.2–3 Ma). Since that time, alternations between cooler and warmer stages characterize the Earth's climatic history. The paleoceanographic system of the Caribbean area (Sites 541 and 543) and of the northern Atlantic was further affected by the closure of the Panama seaway, which occurred in the later Pliocene, around 3 Ma (Keigwin, 1979) at about the time of the onset of Northern Hemisphere glaciation.

In this chapter, we focus our main interest on the glacial-interglacial fluctuations during the Pliocene and Pleistocene, as revealed by isotopic measurements made on materials recovered during Leg 78A from the Barbados Ridge. Further, we attempt to understand the changes in the salinity pattern of the ocean surface water, comparing $\delta^{18}\text{O}$ information from *Globigerinoides sacculifer* and *G. ruber* with their respective abundance patterns.

STRATIGRAPHY

Sites 541 and 543 are situated at the seaward edge of the Barbados Ridge complex and on the adjacent abyssal

sal plain (Figs. 1 and 2). At Site 541 the cored section is continuous, reaching the lower Miocene at 460 m (*Discoaster berggrenii* Subzone; Bergen, this volume) (Fig. 3). Several stratigraphic repetitions occur, the result of tectonic juxtaposition of sediments above a décollement at the toe of the Barbados Ridge (see Site 541 report, this volume). The drilled sediments consist mainly of foraminifer-bearing nannofossil mud. In the Pleistocene a series of darker gray nannofossil mud layers up to 20 cm thick alternate with layers of light gray nannofossil mud up to 50 cm or more in thickness. The boundaries are often sharp. In several intervals the color changes are less pronounced than in others. According to Weissert et al. (in press), the sharpness of these color changes is a function of the carbonate compensation depth (CCD). The pelagic sequence is locally interrupted by redeposited carbonate mud (Site 541 report, and Hemleben and Auras, this volume) and by ash layers (Natland, this volume). The redeposited calcareous sediments were formed by subaqueous gravity flows and probably also by intensified bottom currents. Some color changes within the section may also be the result of gravity flows (Hemleben and Auras, this volume). Sediments of Cores 541-1 through 4 are highly affected by drilling disturbance (Site 541 report, this volume).

Hemleben and Auras (this volume) concluded from their study of the foraminiferal dissolution index (Berger, 1979) and the abundance of *Globorotalia menardii* that the Pleistocene of Site 541 covers the fossil Zones P to V? of Ericson and Wollin (1968) and Briskin and Berggren (1975) (Fig. 4).

The Pliocene/Pleistocene boundary, defined by the first occurrence of *Globorotalia truncatulinoides*, is recorded between the top of Section 541-11-4 and Sample 541-11-3, 131–132 cm. The base of fossil Zone P, which marks the first *G. menardii* peak (Sample 541-11-4, 8–10 cm; see Hemleben and Auras, this volume), occurs

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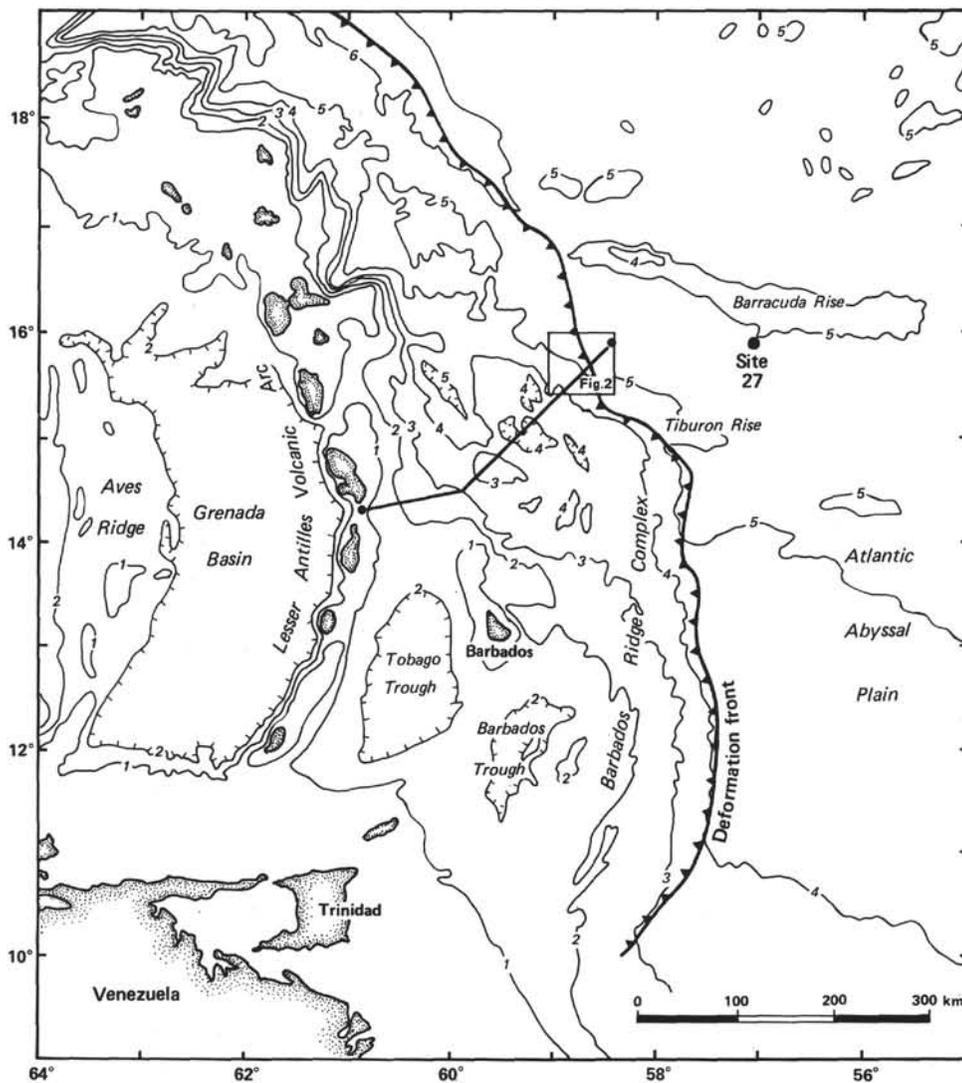


Figure 1. Regional location map; subduction front east of Barbados Ridge complex; water depth contours are in kilometers (adapted from Moore et al., 1982).

slightly below the Pliocene/Pleistocene boundary. The base of fossil Zone P is documented at both Sites 541 and 543. The boundaries of the fossil zones in the upper Pleistocene may be somewhat different than shown in Figure 4 (from Hemleben and Auras, this volume). The extinction of *Pseudoemiliana lacunosa* occurs in Site 541 between Section 541-1, CC and Sample 541-2-1, 70–72 cm (Bergen, this volume). This extinction, coincident with the fossil Zone U/V boundary occurs within isotope stage 13 (Emiliani, 1966; Imbrie et al., 1973; Shackleton and Opdyke, 1973) at about 3.8×10^5 y. (Ericson and Wollin, 1968; Broecker and van Donk, 1970; Gartner, 1972; van Donk, 1976). However, Sample 541-2-1, 27–28 cm shows still a *G. menardii* peak and therefore represents a climatically warmer stage, which could be either fossil Zone V or T. The senior author prefers the latter interpretation, which is consistent with the nannofossil data, but poor preservation of the fauna, drilling disturbance, and a poor paleomagnetic record (Wilson, this volume) prevent a definitive statement.

At Site 543, which lies in deeper water (Figs. 1 and 2), only 70 m of light dark gray foraminifer-bearing nannofossil muds near the top of the hole were cored above pelagic clays and other poorly fossiliferous sediments. Their age ranges from the early Pleistocene (lower *Pseudoemiliana lacunosa* Zone) to the late early Pliocene (*Reticulofenestra pseudoumbilica* Zone; Bergen, this volume).

The deposition of calcareous sediments started at Site 541 in the later Miocene, whereas at the deeper Site 543 it started only in the later early Pliocene. The underlying sediments accumulated at both sites below the CCD. The preservation of the planktonic foraminifers of the Miocene to Pleistocene interval of the two sites indicates that the sediments mostly accumulated below the lysocline (Hemleben and Auras, this volume).

METHODS

For the isotopic record at Sites 541 and 543 monospecific planktonic foraminifer samples of the fraction $>250 \mu\text{m}$ were picked. We studied samples of *Globigerinoides sacculifer*, *G. ruber*, *G. obliquus*, *Pulle-*

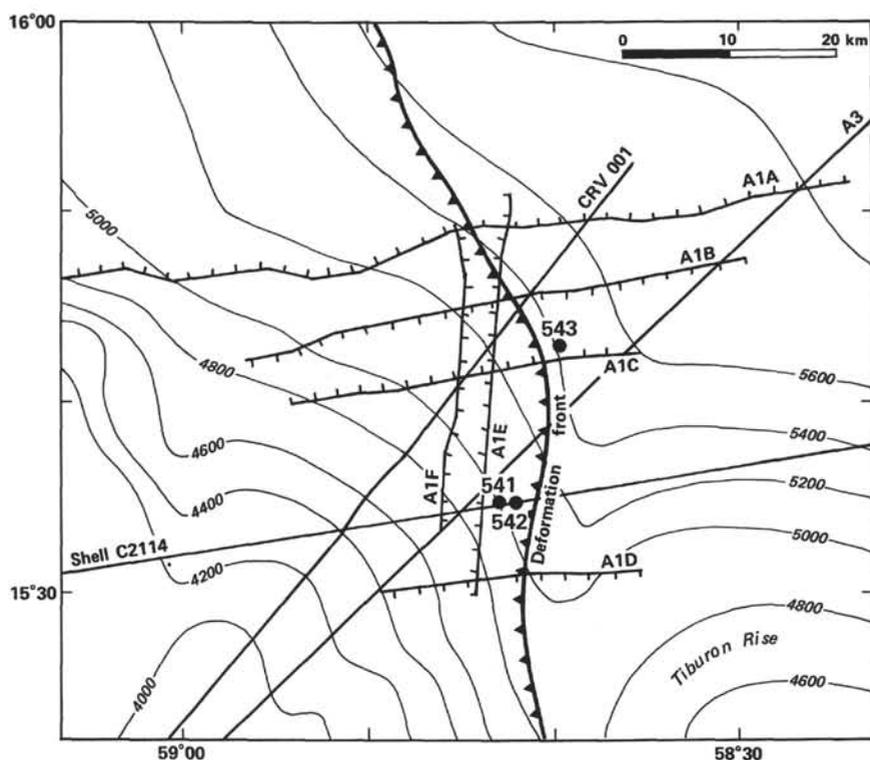


Figure 2. Site location map (identified in Fig. 1) showing the position of the deformation front and seismic reflection profiles used for the site survey (adapted from Moore et al., 1982).

niatina obliquiloculata, and *Sphaeroidinella dehisces*. These samples were roasted *in vacuo* at 400°C for 30 min. in order to remove organic contaminants. The carbon dioxide was released by reaction with 100% orthophosphoric acid at 50°C. The isotopic composition of the gas was measured by a Triple collector mass-spectrometer, Micromass 903 C, at the Geological Institute of the ETH, Zürich. The isotope ratios are reported in the δ notation as per mil deviation from the PDB-Standard. For oxygen and carbon isotope ratios the analytical precision is better than +0.1 per mil. Data are reported in Table 1. Further, we report the percentages of *G. sacculifer* and *G. ruber* in the total planktonic assemblage in Tables 1 and 2.

RESULTS

The oxygen and carbon isotope results are plotted in Figures 4 and 5. Figure 4 shows the Pliocene–Pleistocene isotope record at Site 541. Because the samples were obtained from two principal age-equivalent tectonic units juxtaposed by faulting (see Site 541 report), we have plotted the data as a function of biostratigraphic sequence, combining the samples from the tectonic units.

In our study *G. ruber*, as a typical surface species, generally shows the lightest $\delta^{18}\text{O}$ values. They are close to equilibrium values. In recent oceans, they reflect changes in the surface water temperature (Deuser et al., 1981). *Globigerinoides sacculifer* also lives in the very top surface-water layer, but in the lower part of the section (Site 541) the $\delta^{18}\text{O}$ record of this species reveals slightly heavier $\delta^{18}\text{O}$ values than does *G. ruber*. *Pulleniatina obliquiloculata* and *Sphaeroidinella dehisces* show the heaviest $\delta^{18}\text{O}$ record in most investigated intervals. They reflect mean annual temperatures (Deuser et al., 1981; Hemleben, unpublished data). The $\delta^{13}\text{C}$ records of *G. ruber* and *G. obliquus* are generally lighter than that of

G. sacculifer. The $\delta^{13}\text{C}$ values of *S. dehisces* are intermediate between high and low values in the earlier Pliocene. During the later Pliocene and Pleistocene, values are rather similar to those of *G. sacculifer*.

The Pliocene oxygen isotope record of all species is rather smooth. This is partly a consequence of wide sample spacing, required by the often advanced dissolution. But we note significant changes in the $\delta^{18}\text{O}$ plot of *S. dehisces* and *G. ruber*. In the early Pliocene the $\delta^{18}\text{O}$ values of *S. dehisces* decrease by 0.6‰. In the late Pliocene from about 2.9 to 2.6 Ma the $\delta^{18}\text{O}$ values of *G. ruber* show a positive excursion by 0.5‰. The oxygen record of *S. dehisces*, a species with a deeper living habitat than the *Globigerinoides* species, shows a somewhat divergent evolution. The $\delta^{18}\text{O}$ gradient between the surface inhabitants and *S. dehisces* increases slightly since 2.6 Ma. The $\delta^{13}\text{C}$ signals of the *Globigerinoides* species and *S. dehisces* are fluctuating in the same horizons as we observed in the $\delta^{18}\text{O}$ plot. The $\delta^{13}\text{C}$ excursions range between 0.5 and 0.75‰. In general the changes in $\delta^{18}\text{O}$ and the $\delta^{13}\text{C}$ signals show a positive correlation.

In the Pleistocene samples, oxygen isotope ratios of *G. ruber* fluctuate significantly at both Sites 541 and 543. At Site 541 the $\delta^{18}\text{O}$ values of *G. ruber* range between -2.0‰ and -1.0‰ , whereas those of Site 543 oscillate between -2.0‰ and -1.0‰ , whereas those of Site 543 oscillate between -1.5‰ and -0.5‰ . The lightest $\delta^{18}\text{O}$ values of *G. ruber* occur in the lowermost part of the Pleistocene, where the most pronounced change of 1‰ toward heavier $\delta^{18}\text{O}$ values also occurs (Cores 541-9-2 and 543-3-4 through 3-3). Similar trends

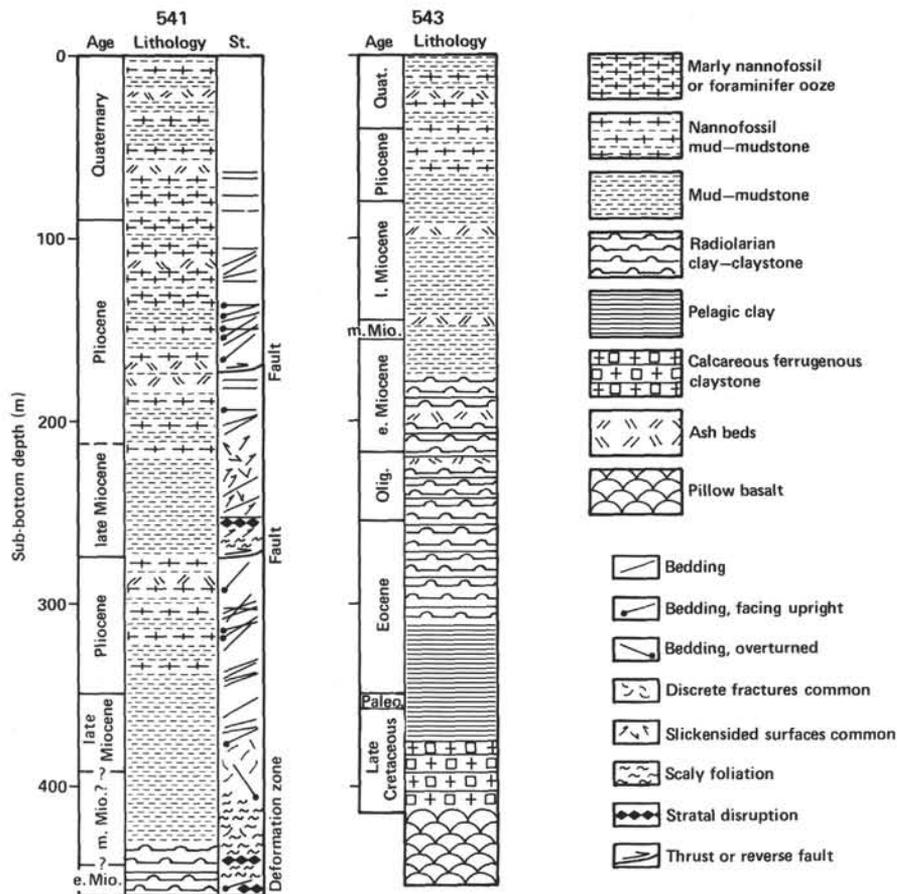


Figure 3. Lithostratigraphy and structural geology at Sites 541 and 543 (adapted from Moore et al., 1982).

are recorded for *G. sacculifer*. The excursion toward lighter $\delta^{18}\text{O}$ values of *G. ruber* coincide in some cases with drastic color changes. From dark gray layers the oxygen isotope signal of *G. ruber* is by 0.4 to 0.6‰ more positive than the signal from lighter waves. *G. ruber* samples, with a more positive $\delta^{18}\text{O}$ imprint, often show extremely heavy $\delta^{13}\text{C}$ values (Samples 541-3-2, 143-145 cm; 541-3-4, 43-45 cm, and many others).

DISCUSSION

Under equilibrium condition the oxygen isotope ratio of shells reflects the oxygen isotope composition and the temperature of the ambient water mass (Craig, 1965). During glacial and interglacial time intervals, the oxygen isotope signal of the ocean water is strongly affected by changes in the global ice volume (Shackleton and Opdyke, 1973). During ice ages, O^{16} is preferentially stored in glaciers. Consequently, the average composition of seawater shifts to more positive values. The surface planktonic record in tropical sections mainly documents changes in the $\delta^{18}\text{O}$ values of the seawater (resulting from ice volume changes) rather than significant temperature decreases (van Donk, 1976). According to Bé et al. (1976), temperature changes between the last glacial and the postglacial stages in the tropical western Atlantic did not exceed 2 to 3°C (maximum deviation of winter tem-

peratures). The $\delta^{13}\text{C}$ composition of the shells reflects the isotopic composition of the dissolved carbon dioxide in the seawater. Furthermore, the $\delta^{13}\text{C}$ ratio of the planktonic foraminifers is strongly controlled by the organic productivity in the surface-water layer and by metabolic processes.

Pliocene

The Pliocene oxygen isotope record is mainly characterized by two events.

1. A slow but steady decrease in the $\delta^{18}\text{O}$ values of *S. dehiscentis* in the early Pliocene (Fig. 4; Samples 541-22-2, 3-5 cm to 541-33-6, 45-47 cm) may indicate a climatic warming. A similar observation was made for the benthic and planktonic foraminifer record of the southern Atlantic Site 519. Weissert and others (in press) found a 0.5‰ shift toward lighter $\delta^{18}\text{O}$ values that occurred 4.4 Ma, in the middle of the magnetic C 1 Event (Mankinen and Dalrymple, 1979). During the latest Miocene the ice volume in the Antarctic was waxing significantly. But already, in the early Pliocene, climatic fluctuations caused the volume of glaciers to pulsate (Shackleton and Kennett, 1975). Compared to the Pliocene-Pleistocene, pulsation rates during the early Pliocene were considerably slower and probably even less pronounced. The climatic event 4.45 Ma is also documented in a study on silicoflagellates by Ciesleski and Weaver (1974). Those

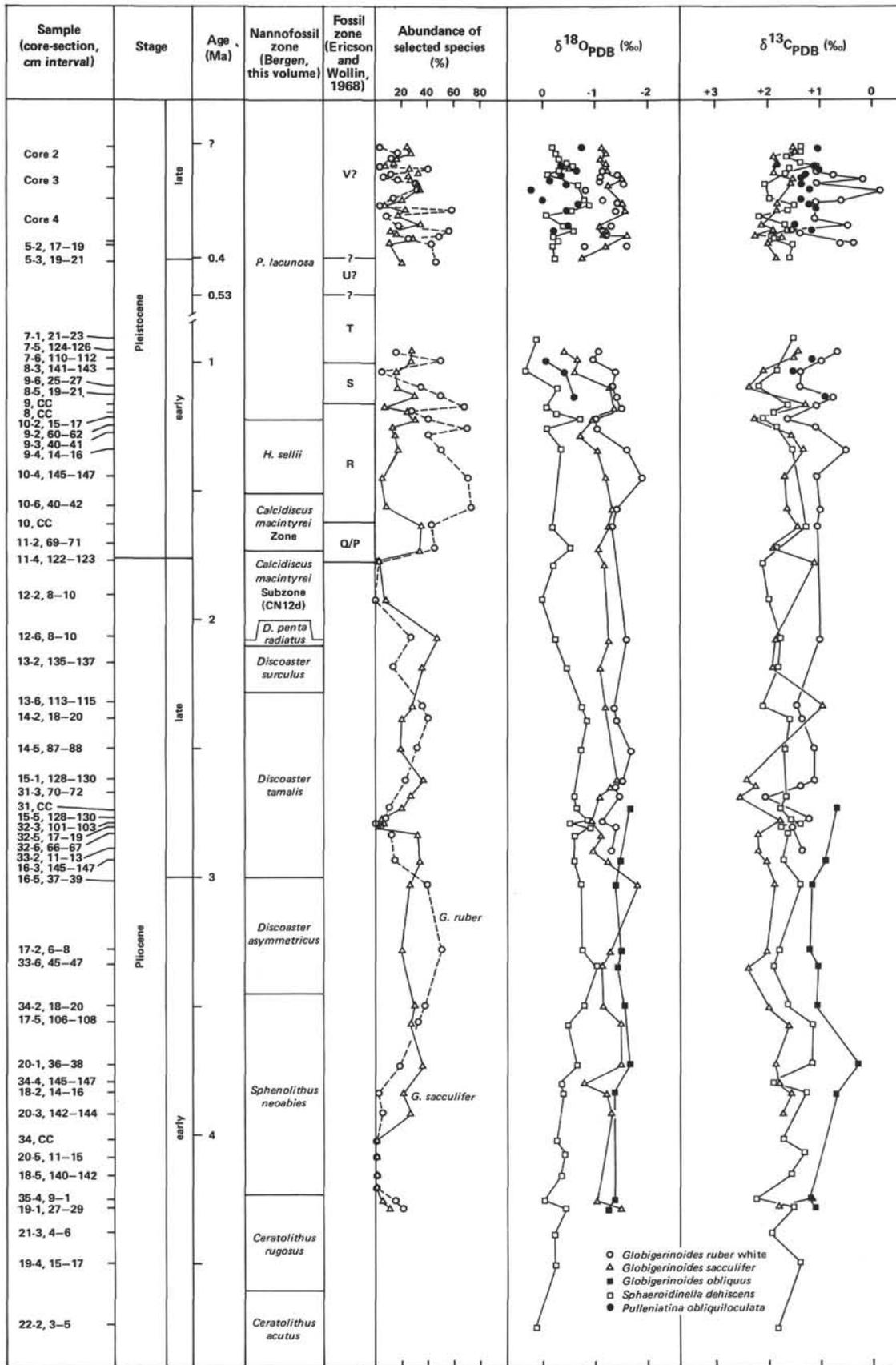


Figure 4. Compilation of the Pliocene to Pleistocene stable isotope stratigraphy and selected species abundance from Site 541.

Table 1. Abundance (%) at Site 541 of *globigerinoides sacculifer* and *Globigerinoides ruber* and oxygen and carbon stable isotope data of different planktonic foraminifers.

Sample (core-section, interval in cm)	Abundance of <i>G. ruber</i> / <i>G. sacculifer</i> (%)	<i>Globigerinoides</i> <i>sacculifer</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Globigerinoides</i> <i>ruber</i> white $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Globigerinoides</i> <i>obliquus</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Pulleniatina</i> <i>obliquiloculata</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Sphaeroidinella</i> <i>dehiscens</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$
2-1, 27-29	2/24	-1.13/+1.54			-0.72/+1.03	-0.17/+1.40
2-2, 83-85	17/26	-1.21/+1.98				-0.24/+1.39
2-2, 150-151		-1.11/+1.93				
2-3, 127-129	11/15	-1.19/+1.86				-0.30/+1.64
2-4, 140-142	10/13	-0.51/+1.85			-0.29/+1.12	-0.44/+1.39
2-5, 70-71	3/6					-0.56/+1.08
3-1, 30-32	41/25	-1.19/+1.90	-1.14/+1.11			-0.30/+1.59
3-1, 140-142	13/32		-1.44/+0.79		-0.61/+1.02	
3-2, 46-48	16/27		-1.12/+1.38		-0.32/+1.30	
3-2, 129-131	6/25	-1.50/+1.54	-1.08/+0.22		-0.09/+1.37	-0.11/+1.67
3-3, 84-86	30/32	-1.27/+1.55	-1.53/+1.12		-0.45/+1.37	-0.67/+2.04
3-4, 43-44	32/33	-1.16/+1.56	-0.78/-0.15		-0.25/+1.19	
3,CC	14/20		-1.15/+0.65		-0.01/+1.35	-0.78/+1.99
4-1, 58-60	3/6	-1.50/+1.84	-1.57/+1.13		-0.64/+1.19	-0.88/+1.50
4-2, 30-32	59/23	-1.58/+1.84	-1.39/+1.11		-0.64/+1.07	-0.56/+1.63
4-4, 88-90	8/16					-0.07/+2.16
4-5, 112-114	17/34	-1.11/+2.13	-1.30/+0.49		-0.45/+1.48	-0.42/+1.68
4-6, 106-108	55/13	-1.18/+1.54	-1.14/+1.54		-0.19/+1.15	-0.60/+1.63
4-6, 131-132	48/17	-1.18/+2.26	-1.19/+1.39			-0.23/+2.01
4-7, 8-10	22/29	-1.60/+1.74				-0.27/+1.92
5-2, 17-19	42/10	-1.22/+2.00	-0.78/+0.41			-0.18/+1.53
5-3, 19-21	44/20	-0.75/+1.84	-1.11/+0.87			-0.26/+1.65
7-1, 21-23						+0.10/+1.50
7-5, 124-126	20/29	-0.42/+1.45	-1.03/+0.68			
7-6, 110-112	51/28	-0.67/+1.00	-0.94/+1.00		-0.06/+1.22	
8-3, 141-143	5/15	-0.61/+2.10	-1.38/+1.43		-0.37/+1.55	+0.35/+1.84
8-5, 19-21	53/30		-1.43/+0.79		-0.56/+0.95	
8,CC	26/28					-0.25/+1.91
9-2, 60-62	70/13		-1.05/+1.11			-0.05/+1.84
9-3, 40-41		-0.66/+1.55				-0.35/+1.54
9-4, 14-16	51/18	-1.02/+1.36	-1.60/+0.51			-0.24/+2.19
9-6, 25-27	35/17	-1.24/+2.39	-1.27/+1.43			-0.06/+1.64
9,CC	69/7	-1.35/+1.31	-1.51/+1.08			-0.69/+2.11
10-2, 15-17	38/31	-0.96/+2.25	-0.98/+1.65			-0.57/+1.78
10-4, 145-147	71/6	-1.21/+1.70	-1.88/+1.12			-0.27/+1.40
10-6, 40-42	73/9	-1.33/+1.66				-0.19/+1.27
10,CC	45/37	-1.26/+1.46	-1.27/+1.10			-0.52/+1.84
11-2, 69-71	47/34	-1.04/+1.90				-0.21/+2.11
11-4, 122-123	4/4	-1.14/+1.13				0.0/+1.98
12-2, 8-10	1/9					-0.24/+1.74
12-6, 8-10	27/47	-1.24/+1.81	-1.62/+1.03			-0.43/+1.81
13-2, 135-137	14/36	-1.07/+1.93				-0.73/+2.09
13-6, 113-115	36/29	-1.18/+0.99	-1.36/+1.45			-0.83/+1.57
14-2, 18-20	41/20		-1.40/+1.34			-0.75/+1.68
14-5, 88-89	30/16		-1.67/+1.12			
15-1, 128-130	23/36	-1.42/+2.43	-1.53/+1.13			-0.91/+1.56
15-5, 128-130	6/5	-0.93/+1.73	-1.16/+1.24	-1.48/+0.92		-0.59/+1.68
16-3, 145-147	15/34	-1.26/+2.03		-1.38/+1.16		-0.74/+1.41
16-5, 37-39	41/27	-1.79/+1.88		-1.48/+1.20		-0.77/+1.81
17-2, 6-8	50/20	-1.29/+2.02				-0.52/+1.14
17-5, 106-108	32/27	-1.49/+1.61		-1.39/+0.70		-0.42/+1.27
18-2, 14-16	2/23	-1.25/+1.64				-0.42/+1.27
18-5, 140-142				-1.28/+1.11		-0.55/+2.11
19-1, 27-29	21/11	-1.49/+1.80				-0.47/+1.48
19-2, 27-29						-0.27/+1.40
19-4, 15-17				-1.60/+0.36		-0.62/+1.18
20-1, 36-38	17/34	-1.48/+1.90		-1.29/+0.83		
20-3, 142-144	4/26	-1.32/+1.75				-0.42/+1.34
20-5, 11-13						-0.24/+1.93
21-3, 4-6						+0.17/+1.81
22-2, 3-5						+0.02/+1.10
23-2, 57-59						
28-5, 24-26						+0.08/+1.18
30-7, 18-20	31/10	-1.26/+2.26				
31-3, 70-74	23/27	-1.09/+2.54				-0.59/+1.65
31,CC	10/20			-1.67/+0.71		-0.65/+1.75
32-3, 101-103	0/9					-0.50/+1.41
32-5, 17-19	1/4					-0.96/+1.71
32-6, 66-67	12/32	-0.51/+2.22				-0.61/+1.64
33-2, 11-13		-0.99/+2.22				
33-6, 45-47		-1.16/+2.41		-1.55/+1.04		-1.04/+1.91
34-2, 18-20	37/29	-1.10/+2.03		-1.53/+1.14		-0.76/+1.64
34-4, 145-147		-0.79/+1.81				-0.37/+1.90
34,CC	1/2					-0.27/+1.74
35-4, 9-11	16/6	-0.45/+1.16		-1.40/+1.16		-0.05/+2.20

Note: Blank spaces indicate no data available.

authors also showed that warmer temperatures existed in the southern ocean during the early Pliocene.

The interpretation of waxing and waning glaciers correlates partly with the foraminiferal dissolution index (FDX) (Berger, 1979) in Site 541 (Hemleben and Auras, this volume). But the usefulness of the FDX in part may

be compromised within certain horizons because of faulting.

2. At Sites 541 and 543 the isotopic changes around 2.9 to 2.7 Ma. (Figs. 4 and 5) indicate a further climatic deterioration. By this time the Northern Hemisphere ice sheets were already formed. Shackleton and Opdyke

Table 2. Abundance (%) at Site 543, of *Globigerinoides sacculifer* and *Globigerinoides ruber* and oxygen and carbon stable isotope data of different planktonic foraminifers.

Sample (core-section, interval in cm)	Abundance <i>G. ruber</i> / <i>G. sacculifer</i> (%)	<i>Globigerinoides sacculifer</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Globigerinoides ruber white</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$	<i>Sphaeroidinella dehiscens</i> $\delta^{18}\text{O}/\delta^{13}\text{C}$
1-3, 47-49	10/31	-0.64/+2.02	-0.70/+0.16	-0.90/+1.61
1-4, 38-39	10/24	-0.51/+1.74		-0.06/+1.73
1-5, 100-102	26/25	-0.65/+1.44	-1.20/+0.23	
1-7, 47-49	22/27	-0.67/+1.94	-0.96/+0.57	-0.32/+1.90
2-2, 6-8	21/33	-0.97/+1.87	-1.18/+0.42	-0.23/+2.14
2-3, 37-39	6/28	-1.13/+1.92	-1.16/+1.12	-0.03/+2.02
2-4, 69-71	12/29	-1.30/+1.73	-1.37/+1.00	-0.20/+1.98
2-5, 37-39	3/11	-1.05/+1.05		-0.01/+1.75
2, CC	18/38	-0.89/+2.35	-1.34/+0.83	-0.13/+2.12
3-2, 29-31	12/14			+0.15/+2.38
3-3, 80-82	21/25	-0.44/+1.75	-0.56/+0.50	+0.21/+2.06
3-4, 30-32	69/7	-1.08/+1.57	-1.60/+0.93	-0.04/+2.08
3-4, 110-112	8/7			+0.41/+1.59
3-5, 27-29	34/37	-1.04/+1.75	-1.49/+1.63	-0.27/+2.32
3-5, 128-129	0/2			-0.05/+2.10
3-7, 36-37	19/25	-0.73/+1.13	-1.60/+0.28	+0.09/+1.75
3, CC	39/35	-0.70/+2.03	-1.52/+1.04	-0.07/+1.96
4-6, 147-149	35/38	-0.47/+2.45	-1.47/+0.27	-0.69/+1.89
5-1, 120-122	18/49	-0.80/+1.72	-1.68/+0.25	-0.44/+1.82
5-2, 27-29	13/54	-0.50/+1.73	-1.60/-0.27	-0.53/+2.06
6-3, 60-62	0/0			-0.79/+1.76
6-7, 34-36	0/9			-0.36/+1.70
7-2, 10-12	0/2			-0.21/+1.88
7-3, 110-112	0/34	-0.51/+1.56		

Note: Blank spaces indicate no data available.

(1977) show the first significant positive oxygen isotope deviation ($\Delta^{18}\text{O} = 0.5\text{‰}$) in a paleomagnetically well dated section at 3.2 Ma (lower part of the Gauss anomaly). This date is also documented by ice-rafted sediments in the North Atlantic DSDP sites (Berggren, 1972). At Site 541 the interval from Cores 18 to 20 (*Sphenolithus neobabies* Subzone, CN11a) probably documents the initiation of the Northern Hemisphere glaciation. Most of these samples reveal a FDX between 6 to 8, which corresponds to a poor to very poor preservation (Hemleben and Auras, this volume). The poor preservation within this interval may further be a result of (1) a changed circulation pattern or (2) vertical tectonic movements associated with the closure of the Panama seaway. Both would significantly affect the position of the CCD. The interruption of Pacific bottom-water influx would allow Atlantic deep water to invade the Caribbean basins. Possibly even transient upwelling currents were active. To date we do not have isotopic records from benthic foraminifers or other paleontological data adequate to distinguish between either of these hypotheses.

The excursion in $\delta^{18}\text{O}$ 2.9 to 2.7 Ma marks the beginning of the glacial-interglacial fluctuations, which lasted till the latest Pleistocene. According to Shackleton and Opdyke (1977), the 0.5 to 1‰ fluctuations in the oxygen record began in the Pacific during the later Gauss

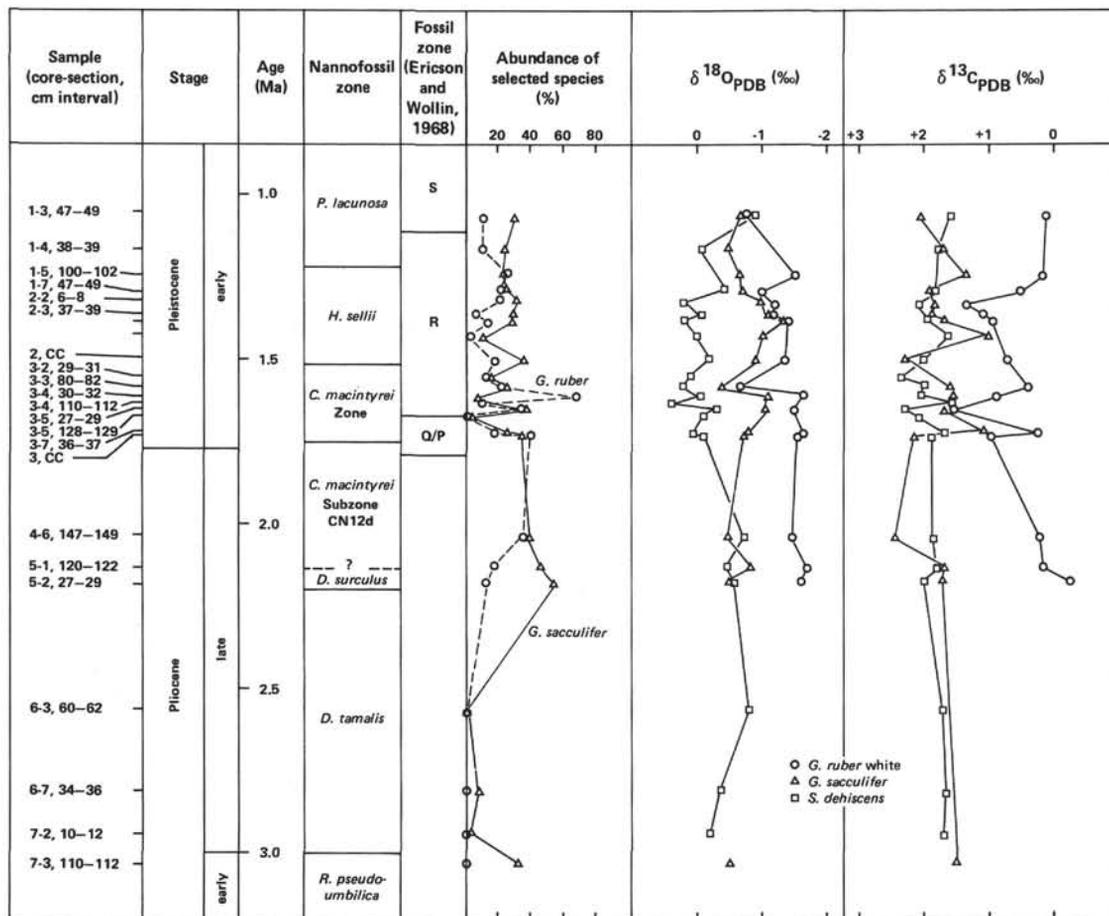


Figure 5. Late Pliocene to early Pleistocene stable isotope stratigraphy and selected species abundance from Site 543.

anomaly (2.7 Ma). The steeper oxygen isotope gradient between surface (*Globigerinoides ruber* and *G. sacculifer*) and deeper dwellers (*Sphaeroidinella dehiscens*) could also signify some glacial expansion in the Northern Hemisphere as vertical temperature gradients increased. Published $\delta^{18}\text{O}$ records (Shackleton and Opdyke, 1976) of surface indicators living during the major glacial stages up until the Holocene range between 1 and 1.5‰. The $\delta^{18}\text{O}$ fluctuations recorded at Sites 541 and 543 during the late Pliocene are relatively small ($\Delta^{18}\text{O} = 0.5\text{‰}$), although both hemispheres were glaciated. Oxygen isotope patterns similar to those at our sites were reported from the equatorial Pacific and the Atlantic, as well as at South Atlantic sites (Keigwin, 1979; Vergnaud Grazzini and Rabussier-Lointier, 1980; Weissert et al., in press). This may indicate that the glacial ice effect on the $\delta^{18}\text{O}$ value is to a certain extent compensated by a warming in low-latitude areas (Ruddiman and McIntyre et al., 1976). Weissert et al. (in press) suggest that a warming of 2°C in low latitudes occurred during a time of growing glaciers. This phenomenon may have been restricted to low latitudes. Bé et al. (1976) studied the last glacial interglacial intervals in equatorial Atlantic cores. By applying the transfer functions developed by Imbrie and Kipp (1971), Bé et al. calculated winter and summer sea-surface temperatures from the abundance data of planktonic foraminifers; they found no warming in the tropics at the beginning of the last glacial stage. Thus the small $\delta^{18}\text{O}$ fluctuations in samples from tropical Sites 541 and 543 cannot be explained by temperature increases, which would partly compensate the glacial decrease in $\delta^{18}\text{O}$ value. Other factors such as dissolution and salinity changes may have played a role, as discussed later.

Pleistocene

The isotopic record of the Pleistocene sediments of Site 541 cover the interval from the *Calcidiscus macintyrei* Zone to the *Pseudoemiliana lacunosa* Zone. At Sites 541 and 543, fluctuations in the $\delta^{18}\text{O}$ record of *G. ruber* range from 0.5 to a maximum of 1‰. These figures correlate with results from Core V 16-205 from the tropical Central Atlantic Ocean (van Donk, 1976). In this core, van Donk distinguishes thirteen interglacial stages from the basal Pleistocene to the top of the *P. lacunosa* Zone. A correlation of single stages between this core and those from our sites is not feasible for several reasons. Our sampling is partly too widely spaced, because of drilling disturbance, reworking, or the advanced levels of dissolution. Nor do we have a complete magnetostratigraphic records (Wilson, this volume); thus the age control is not accurate enough for a correlation.

The pronounced $\delta^{18}\text{O}$ excursions within the later Pliocene and Pleistocene intervals occur commonly in connection with color changes in the nannofossil mud and changes in the FDX. They may therefore reflect real climatic fluctuations. At South Atlantic Leg 73 sites, Weissert et al. (in press) show from the benthic and planktonic oxygen isotope records that darker layers document glacial intervals, whereas the brighter layers correspond to interglacial stages. Their study of the foraminiferal assemblages shows higher dissolution rates

in the darker than in the lighter layers. They attributed this phenomenon to the transient influx of cooler, more corrosive bottom water.

Which factors caused the fluctuation of the late Pliocene and Pleistocene $\delta^{18}\text{O}$ record of the surface water in this tropical region? Was it (1) temperature changes, (2) changes in the $\delta^{18}\text{O}$ composition of the water mass, or (3) dissolution?

Faunal analysis from western equatorial Atlantic cores reveals that the differences in the winter surface temperatures did not exceed 3.6°C between the last glacial stage and the Holocene (Bé et al., 1976). The smallest difference (0.1°C) they noted in a core (V 16-205, about 17°N, 50°E) near the southern Sargasso Sea. We may therefore also conclude that during former glacial-interglacial cycles, temperature fluctuations of surface waters did not significantly exceed 1°C in the area of Sites 541 and 543. Thus temperature changes may not be the main reason for the fluctuations in the $\delta^{18}\text{O}$ values of the surface dwellers.

G. sacculifer and *G. ruber* are dominant warm-water surface inhabitants. *G. ruber* (var. white and pink) comprises 20 to 50% of the planktonic foraminiferal assemblage in well-preserved samples of Sites 541 and 543 (Figs. 4 and 5), and in exception instances may exceed 60%. *G. sacculifer* never exceeds 50%. On the average, this species contributes 20 to 30% of the planktonic assemblage. Similar values are reported from the western equatorial Atlantic cores (Bé et al., 1976). In most of these cores significant changes in the abundance of *G. sacculifer* are not obvious between glacial and interglacial stages. In a few cores Bé et al. (1976) noted anticyclic fluctuations in the abundance of these two species. The abundance of *G. sacculifer* and *G. ruber* is, among other factors, controlled by salinity changes (Bé and Tolderlund, 1971). According to these authors, *G. sacculifer* is abundant at normal salinities (34.5 and 36‰). *G. ruber* is most abundant at salinities from 36 to 39‰ and tolerates low-nutrient environments better than does *G. sacculifer*. Today, *G. ruber* is a typical inhabitant of the more saline (36‰) watermass of the southern Sargasso Sea and of the low-salinity water (<34‰) off Barbados. The low-salinity component in the southeastern Caribbean is the result of seasonal fresh-water discharge from the Amazon River (Kidd and Sander, 1979).

However, *G. sacculifer* is the dominant species in the high-salinity water (>41‰) of the Gulf of Elat (Reiss et al., 1980). Culture experiments with *G. ruber* and *G. sacculifer* indicate a different salinity range for each species. *G. ruber* survives salinity conditions between 25 and 42‰, whereas *G. sacculifer* tolerates 29 to 45‰ (Hemleben and Spindler, unpublished data).

The abundance of *G. sacculifer* and *G. ruber* at Sites 541 and 543 is also strongly controlled by dissolution. Berger (1979) assigned *G. ruber* and *G. sacculifer* a low index of resistance to dissolution, whereby *G. ruber* is less resistant than *G. sacculifer*. Below the lysocline both are among the first species to disappear. Dissolution not only affects the faunal assemblage but also influences the oxygen and carbon isotope composition of the different species (Berger and Gardner, 1975; Barthelémy-Bonneau,

1978). Barthelémy-Bonneau shows that the $\delta^{18}\text{O}$ value of *G. ruber* is considerably more affected by partial dissolution than the $\delta^{18}\text{O}$ value of *G. sacculifer*. *G. ruber* and *G. sacculifer* show an enrichment in $\delta^{18}\text{O}$ of 0.8 and 0.4‰, respectively.

The abundance of *G. sacculifer* and *G. ruber* is high throughout the upper Pliocene and Pleistocene. A low abundance of these two species is recorded in intervals with a high FDX. We explain the anticyclic abundances of these two species, which is drastic in some intervals, by salinity changes. During intervals with high abundance of *G. ruber* and relatively heavier $\delta^{18}\text{O}$ values (e.g., 1.22 Ma), we assume a higher salinity concentration (>36‰) in the surface seawater. During intervals with high abundance of *G. ruber* and light $\delta^{18}\text{O}$ values we assume that a lens of lower-salinity seawater was present in the Caribbean. This correspondence is strongest from the late Pliocene to the top of faunal Zone S. Above Section 541-7, CC this correlation is less obvious because of drilling disturbance and dissolution.

During the last glacial stage (18,000 y. ago), the Caribbean Sea had surface waters similar to those of the Central Sargasso Sea today (Prell and Hays, 1976; McIntyre et al., 1976). This high-salinity water shifted during glacial intervals by about 15 to 20° WSW. The Central Water Mass of the Sargasso Sea is not only characterized by a relative high salinity but is also nutrient-poor. Water masses of low nutrient content have a deficiency in ^{12}C . Thus we may also explain the positive $\delta^{13}\text{C}$ imprint during periods with relatively heavy $\delta^{18}\text{O}$ values by a water-mass change. Hence, during cooler intervals in the earliest Pleistocene or even earlier, drifting of the Sargasso Sea water mass into the Caribbean region had already started.

The rather light $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ signals of *G. ruber* between 1.5 and 1.3 Ma we ascribe to a fresh-water and/or a low-salinity seawater admixture. Today, the surface-water salinity off Barbados shows seasonal changes of 3 to 5‰. This is the result of tropical rainfall from South America and/or the northern drift of the Intertropical Convergence Zone (ITCZ). The drift of the ITCZ allows the Benguela Current to bring lower-salinity water into the western South Atlantic. Both water sources could enrich the ambient seawater with ^{16}O and nutrients and therefore produce lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. We imagine that similar circulation patterns already existed during earlier glacial-interglacial cycles.

CONCLUSIONS

The early Pliocene is characterized by a warming from about 4.7 to 4.3 Ma. The late Pliocene indicates a climatic deterioration around 2.9 to 2.7 Ma. The changes in the oxygen isotope signal coincide with changes in the circulation pattern resulting from the beginning of Northern Hemisphere glaciation. A steepening of the $\delta^{18}\text{O}$ gradient between surface dwellers and intermediate dwellers may be a further hint of climatic deterioration. The Pleistocene isotopic record is characterized by regular fluctuations caused by salinity changes, rather than temperature changes, and is indicated also by *G. ruber* and *G. sacculifer* abundances. The salinity changes may be

explained by a drifting of the highly saline Central Water Mass of the Sargasso Sea into the Caribbean region during cooler intervals. This current pattern between glacial-interglacial stages has existed in the western Atlantic at least since the earliest Pleistocene. Occasionally, lower-salinity water also drifted into the Caribbean area; this water may have been derived from a fresh-water discharge from the Amazon River and/or the drifting Intertropical Convergence Zone. By means of northward drift, the South Atlantic Benguela Current, which wells up off Namibia and South Africa, may influence water compositions as far north as Caribbean Sites 541 and 543.

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