32. PALEOCEANOGRAPHIC IMPLICATIONS OF STABLE-ISOTOPE DATA FROM UPPER MIOCENE–LOWER PLIOCENE SEDIMENTS FROM THE SOUTHEAST ATLANTIC (DEEP SEA DRILLING PROJECT SITE 519)[1]

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ABSTRACT

A stable-isotope stratigraphy was established for planktonic and benthic foraminifers from upper Miocene–lower Pliocene pelagic sediments from the Mid-Atlantic Ridge. A correlation of stable-isotope and biostratigraphic data with magnetostratigraphic age revealed the following: (1) the late Miocene carbon-isotope shift in the South Atlantic bottom waters was minute compared with the shift reported for other deep-sea locations (Haq et al., 1980), (2) a significant cooling or continental ice-volume increase occurred between 5.7 and 5.2 Ma, and (3) a period of warming or ice-volume decrease followed, with the rate of warming increasing beginning at 4.5 Ma and reaching a climax at 4.3 Ma. The timing of these paleoceanographic events is correlated with the onset and termination of the Messinian salinity crisis in the Mediterranean Sea.

INTRODUCTION

The primary objective of DSDP Leg 73 was to study the Cenozoic paleoceanography of the South Atlantic, with emphasis on the time intervals when major oceanographic events occurred. This paper presents a study of the stable-isotope stratigraphy of the late Miocene and early Pliocene and attempts to correlate magnetostratigraphic, biostratigraphic, and isotope stratigraphic evidence to evaluate two major oceanographic events that are recorded in marine sediments of this age: (1) a pronounced global decrease in the $\delta^{13}C$ value of calcitic tests of benthic foraminifers between 6.10 and 5.90 Ma (Haq et al., 1980) and (2) a major increase in Antarctic glaciation during the latest Miocene (~5 Ma) (Shackleton and Kennett, 1975; and Kennett, 1977). The timing of these events corresponds to the period of the isolation and desiccation of the Mediterranean Sea.

It has often been suggested that an expansion of the Antarctic ice cap and lowering of the global sea level may have caused, or at least enhanced, the isolation and eventual desiccation of the Mediterranean Sea (Hsü et al., 1973; Van Couvering et al., 1976; Adams et al., 1977; and McKenzie et al., 1979/1980). Further, the inflow of Mediterranean deep water (MDW) is a major component of the modern thermohaline circulation in the Atlantic Ocean, and it greatly affects the carbon-isotope signal of the North Atlantic Deep Water (NADW; Kroopnick, 1980). Total cessation of this inflow during the Messinian salinity crisis could have resulted in major changes in the circulation pattern of the Atlantic Ocean and, hence, the carbon-isotope distribution (Bender and Keigwin, 1979). Blanc (1981) has proposed that the closure of the Mediterranean outlet resulted in the formation of deep waters in the South Atlantic, which then flowed northward to replace the diminished NADW.

The purpose of this study was to investigate the above-mentioned paleoceanographic events as recorded in pelagic sediments deposited on the eastern flank of the Mid-Atlantic Ridge. The calcite compensation depth (CCD) was elevated during the late Miocene, so a drill site was selected where the sediment had been deposited near the ridge crest during this time period. It was thought that this material would show the least amount of dissolution and would therefore optimize our ability to correlate the biostratigraphy, the isotope stratigraphy, and the magnetostratigraphy at the site. Precision magnetostratigraphy and biostratigraphy, if they could be acquired, would define the age of the sediments, and the biostratigraphy and stable-isotope stratigraphy could then be used to delineate the paleoceanographic changes that took place in the southeast Atlantic during the late Miocene and early Pliocene. It was further hoped that the exact timing of these events would allow for a correlation with the Messinian salinity crisis in the Mediterranean Sea.

SITE DESCRIPTIONS

DSDP Site 519 was located on the edge of a ponded facies on the eastern side of the Mid-Atlantic Ridge (26°8.20′S, 11°39.97′W; Fig. 1). Water depth was 3769 m. A 151-m-thick sequence of pelagic sediments was continuously cored with the hydraulic piston coring apparatus. Recovery was excellent (91.3%). The upper Miocene–lower Pliocene sediments, which were predominantly nannofossil oozes, showed a greater degree of dissolution than the overlying Pliocene–Quaternary sediments, in spite of the fact that the latter were deposited in deeper water. The upper Miocene sedimentation rate is less than 1 cm/10^3 yr. See the site chapters for more detailed information.

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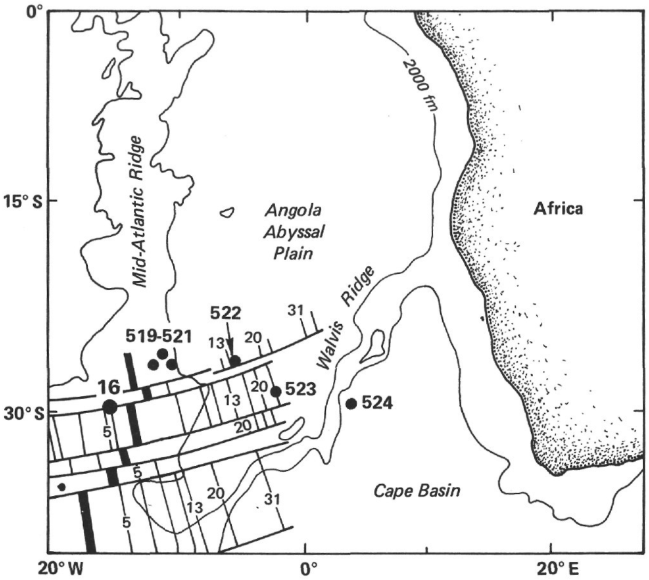

Figure 1. Map of southeast Atlantic Ocean showing location of drill sites from DSDP Legs 3 (Site 16) and 73 (Sites 519–524). The smaller numbers indicate magnetic anomalies.

The apparent dissolution and the suspicion that there was a hiatus[6] in the upper Miocene sediments from Site 519 led us to sample contemporaneous sediments from a sequence of cores from DSDP Leg 3, Site 16, which was located on the western side of the Mid-Atlantic Ridge (30°20.15′S, 15°42.79′W). The water depth at that site was 3526 m. The upper Miocene sediments there also comprised nannofossil ooze but showed somewhat less dissolution. The upper Miocene sedimentation rate was approximately 1.0 cm/10^3 yr. (Maxwell et al., 1970). Although Site 16 was rotary cored, recovery was excellent, allowing us to obtain comparative isotopic record from a second site in the southeast Atlantic.

STABLE-ISOTOPE RESULTS

Methodology

The foraminifers *Planulina wuellerstorfi* and *Orbulina universa* were selected for this isotope study to represent conditions in the bottom and surface waters, respectively. Samples of other benthic foraminifers, from Site 519 and the nearby Site 521, *Oridorsalis umbonatus* and *Nuttalides umbonifera*, were analyzed in conjunction with another isotope study (Weissert et al., this vol.), and the results are included in this study. Samples were picked for monospecific assemblages, crushed and treated with methanol to remove contaminant particles, and roasted under vacuum at 450°C for 30 min. The microsamples were reacted at 50°C by using the traditional phosphoric acid method (McCrea, 1950), and the released CO_2 gas was analyzed immediately after the completion of the reaction. Stable-isotope analyses were made by using a V. G. Micromass 903 triple collecting mass spectrometer. The results, which are tabulated in Table 1, are reported relative to the PDB international standard using the δ-notation:

$$\delta\ (‰) = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 10^3$$

where $R = {}^{13}C/{}^{12}C$ or ${}^{18}O/{}^{16}O$.

Carbon Isotopes

The $\delta^{13}C$ values for the *Planulina wuellerstorfi* and *Orbulina universa* samples from Site 519 are plotted with respect to depth sub-bottom in Figure 2. The magnetostratigraphic ages (in Ma) as determined by Tauxe et al. (this vol.) are also included but on a nonlinear scale. The time period of primary interest to this study, the Messinian stage of the late Miocene (6.6 to 5.2 Ma), is represented by about 5.5 m of sediment. In the interval between 6.38 and 5.69 Ma (121.7 to 120.3 m sub-bottom), the carbon-13 isotope content of *P. wuellerstorfi* decreases by 0.37 ‰. The average $\delta^{13}C$ change recorded for samples above and below 121 m sub-bottom, which corresponds to a magnetostratigraphic age of 6.1 Ma, is somewhat greater. The average $\delta^{13}C$ value for 22 samples from above 121 m sub-bottom is +0.97 (±0.24) ‰, whereas the average $\delta^{13}C$ value for 15 samples below is +1.39 (±0.23) ‰. The difference between these two average values could represent a late Miocene carbon-isotope shift of −0.42 ‰. A limited isotopic sampling of *Oridorsalis umbonatus* from above and below 121 m sub-bottom indicates a carbon-13 shift of −0.32 ‰. For comparison, six samples of *O. umbonatus* from the nearby Site 521 have an average $\delta^{13}C$ value in the Pliocene of −0.64 ‰, while in the middle Miocene, the average value for five samples is more enriched at +0.80 ‰. Data for a third benthic foraminifer, *Nuttalides umbonifera*, are available only for samples above 121 m sub-bottom, that is, after the carbon-isotope shift. In the case of the planktonic species, *Orbulina universa*, the $\delta^{13}C$ values increase steadily by 0.6 ‰ from about 9 to 6 Ma. This trend toward increasing values tends to reverse itself before the Miocene/Pliocene boundary, which is located at 114 m sub-bottom. Such carbon-isotope fluctuations in the surface waters may reflect changes in productivity and be totally unrelated to the bottom-water events recorded in benthic foraminifers.

In Figure 3, the $\delta^{13}C$ values for *P. wuellerstorfi* and *O. universa* from Site 16 are plotted against sub-bottom depth. No magnetostratigraphy is available for this site, but there is some biostratigraphic control on the age of the sediments in the interval of interest to this study. The first appearance of *Amaurolithus amplificus* occurs at 112.7 m or at 5.65 Ma (Haq et al., 1980). The first common occurrence of *A. primus* is at 137.4 m or at 6.25 Ma (Haq et al,. 1980). Approximately halfway between these two datum levels lies 5.95 Ma, which would essentially separate carbon-isotope values before and after the late Miocene shift. A slight carbon-isotope shift of −0.2 to −0.3 ‰ does apparently occur in *P. wuellerstorfi*. As at Site 519, the $\delta^{13}C$ value of *O. universa* also tends to fluctuate systematically. The absolute isotope values for carbon and oxygen in both *P. wuellerstorfi*

6 One of us (R.Z.P.) is convinced that a hiatus does exist. However, its presence would not affect the results presented herein.

Table 1. Stable-isotope data.

Foraminifer	Sample (interval in cm)	Sub-bottom depth (m)	Age (Ma)	$\delta^{18}O$ PDB (‰)	$\delta^{13}C$ PDB (‰)
P. wuellerstorfi	519-9-1, 35–37	31.6	0.98	+3.11	+0.79
	519-10-2, 50–52	37.3	1.70	+2.53	+1.12
	519-12-3, 38–40	47.5	2.22	+3.42	+0.40
	519-16-2, 49–51	63.7	2.98	+2.61	+1.21
	519-16-2, 105–107	64.2	3.01	+2.73	+1.05
	519-16-3, 49–51	65.2	3.07	+2.48	+0.68
	519-22-2, 63–65	90.2	4.18	+2.21	+0.84
	519-23-2, 115–117	95.2	4.34	+1.92	+0.60
	519-24-1, 34–36	97.2	4.39	+2.01	+0.61
	519-24-1, 55–57	97.4	4.40	+2.17	+0.82
	519-24-2, 130–132	99.7	4.45	+2.34	+1.16
	519-25-1, 14–16	101.4	4.49	+2.30	+1.22
	519-27-1, 9–11	110.2	4.93	+2.48	+0.90
	519-27-1, 24–26	110.4	4.94	+2.59	+1.15
	519-27-1, 65–67	110.8	4.96	+2.59	+1.39
	519-27-2, 88–90	112.5	5.05	+2.52	+1.04
	519-28-1, 107–109	115.6	5.21	+2.72	+1.09
	519-28-1, 109–111	115.6	5.21	+2.64	+1.11
	519-28-2, 79–81	116.8	5.29	+2.61	+1.05
	519-28-2, 119–121	117.2	5.34	+2.63	+1.07
	519-29-2, 30–32	120.3	5.69	+2.35	+0.91
	519-29-2, 90–92	120.9	5.76	+2.42	+1.14
	519-29-3, 23–25	121.7	6.38	+2.32	+1.28
	519-30-1, 115–117	124.1	6.83	+2.44	+1.24
	519-30-2, 90–92	125.3	6.99	+2.59	+1.58
	519-30-3, 56–58	126.5	7.15	+2.21	+1.74
	519-31-1, 100–102	127.9	7.33	+2.64	+1.68
	519-31-1, 130–132	128.2	7.37	+2.27	+1.52
	519-31-2, 26–28	128.6	7.42	+2.70	+1.71
	519-31-3, 34–36	130.2	7.63	+2.34	+1.50
	519-32-1, 20–22	131.5	7.80	+2.10	+1.17
	519-32-3, 20–22	134.5	8.11	+2.26	+1.36
	519-33-2, 61–63	137.8	8.42	+2.41	+1.11
	519-34-1, 36–38	140.5	8.74	+2.06	+1.27
	519-34-2, 16–18	141.8	8.93	+1.77	+0.97
	519-34-3, 36–38	143.5	9.18	+2.14	+1.24
	519-35-1, 44–46	144.6	9.34	+2.29	+1.50
	16-4-1, 56–58	55.5	—	+2.08	+0.93
	16-4-3, 66–68	58.5	—	+2.14	+0.53
	16-5-1, 56–58	86.4	—	+2.59	+0.97
	16-5-2, 118–120	88.4	—	+2.45	+0.94
	16-5-5, 65–67	92.3	—	+2.59	+1.07
	16-6-1, 60–62	104.6	—	+2.59	+1.13
	16-6-2, 107–109	106.5	—	+2.51	+1.16
	16-6-3, 66–68	107.6	—	+2.48	+1.29
	16-6-4, 58–60	109.0	—	+2.73	+0.81
	16-6-5, 57–59	110.6	—	+2.37	+1.15
	16-6-6, 58–60	112.1	—	+2.69	+0.94
	16-6-7, 59–61	113.8	—	+2.54	+0.94
	16-7-1, 118–120	114.3	—	+2.59	+0.83
	16-7-2, 58–60	115.9	—	+2.45	+0.98
	16-7-3, 34–36	116.5	—	+2.28	+0.93
	16-7-3, 110–112	117.3	—	+2.59	+1.17
	16-7-4, 60–62	118.3	—	+2.54	+1.14
	16-7-5, 38–30	119.6	—	+2.61	+1.35
	16-7-5, 97–99	120.2	—	+2.58	+1.13
	16-7-6, 29–31	121.0	—	+2.40	+0.99
	16-8-1, 35–37	125.8	—	+2.38	+1.29
	16-8-3, 44–46	128.2	—	+2.18	+0.95
	16-9-1, 124–126	135.8	—	+2.54	+1.33
	16-9-3, 105–107	138.5	—	+2.43	+1.31
	16-9-5, 136–138	141.8	—	+2.44	+1.31
N. umbonifera	519-9-1, 35–37	31.6	0.98	+3.49	+0.86
	519-13-2, 100–102	51.0	2.44	+2.62	+0.85
	519-14-2, 80–82	56.2	2.66	+2.62	+0.85
	519-16-2, 49–51	63.7	2.98	+2.63	+0.92
	519-16-2, 105–107	64.2	3.01	+2.78	+0.78
	519-16-3, 49–51	65.2	3.07	+3.08	+0.59
	519-18-1, 100–102	71.5	3.35	+2.64	+0.87
	519-20-2, 70–72	81.5	3.73	+2.68	+0.80
	519-23-2, 115–117	95.2	4.34	+1.80	+0.42
	519-24-1, 55–57	97.4	4.40	+2.27	+0.69
	519-24-3, 113–115	101.4	4.48	+2.54	+0.61
	519-28-1, 107–109	115.6	5.21	+2.72	+0.61
	519-29-1, 64–66	119.6	5.67	+1.69	+0.15
	519-29-2, 64–66	121.0	6.09	+2.09	+0.71
O. umbonatus	519-28-1, 107–109	115.6	5.21	+3.13	−0.19
	519-28-2, 107–109	117.1	5.33	+2.90	−0.43
	519-29-2, 64–66	121.0	6.09	+3.21	+0.19
	519-29-2, 144–146	121.8	6.50	+2.65	−0.16
	519-29-3, 27–29	122.2	6.58	+2.87	+0.00
	521-8-1, 133–135	31.3	3.11	+3.55	−0.46
	521-8-3, 100–102	34.0	3.32	+2.99	−0.47
	521-9-1, 140–142	35.9	3.48	+2.87	−0.55
	521-9-2, 140–142	37.4	3.67	+2.33	−0.86

Table 1. (Continued).

Foraminifer	Sample (interval in cm)	Sub-bottom depth (m)	Age (Ma)	$\delta^{18}O$ PDB (‰)	$\delta^{13}C$ PDB (‰)
	521-10-1, 100–112	40.1	4.00	+2.55	−0.64
	521-10-2, 110–112	41.6	4.18	+2.00	−0.83
	521-17-1, 54–56	66.5	14.25	+2.64	+1.09
	521-17-1, 107–109	67.1	14.27	+2.75	+1.29
	521-17-1, 39–41	67.9	14.31	+2.24	+0.60
	521-17-3, 34–36	69.3	14.37	+1.78	+0.48
	521-18-3, 33–35	73.8	14.79	+1.93	+0.81
O. universa	519-24-1, 34–36	97.2	4.39	+0.69	+2.34
	519-25-1, 14–16	101.4	4.49	+0.60	+2.26
	519-26-1, 38–40	106.2	4.72	+0.76	+2.19
	519-27-1, 24–26	110.4	4.94	+0.81	+1.94
	519-27-1, 65–67	110.8	4.96	+1.17	+2.15
	519-27-2, 88–90	112.5	5.05	+0.74	+2.02
	519-28-1, 109–111	115.6	5.21	+0.84	+2.14
	519-28-2, 79–81	116.8	5.29	+0.77	+2.19
	519-29-1, 40–42	118.9	5.57	+0.96	+2.41
	519-29-2, 30–32	120.9	5.76	+0.68	+2.22
	519-29-3, 23–25	121.7	6.38	+0.52	+2.66
	519-30-2, 90–92	125.3	6.99	+1.02	+2.61
	519-31-1, 130–132	128.2	7.37	+0.84	+2.49
	519-31-2, 12–12	128.6	7.42	+1.13	+2.56
	519-32-1, 20–22	131.5	7.80	+0.84	+2.39
	519-32-2, 100–102	133.8	8.04	+0.89	+2.36
	519-33-1, 130–132	137.0	8.35	+0.69	+2.21
	519-33-3, 82–84	139.5	8.60	+0.71	+2.09
	519-34-3, 36–38	143.5	9.18	+0.63	+2.59
	519-35-1, 44–46	144.6	9.34	+0.18	+2.33
	519-35-2, 115–117	146.8	9.66	+0.78	+2.61
	16-5-1, 56–58	86.4	—	+0.48	+2.03
	16-5-2, 118–120	88.4	—	+1.11	+2.36
	16-5-3, 109–111	89.8	—	+0.98	+2.38
	16-5, 65–67	92.3	—	+0.80	+2.19
	16-6-1, 60–62	104.2	—	+0.83	+2.62
	16-6-1, 118–120	105.2	—	+1.02	+2.36
	16-6-4, 58–60	109.2	—	+0.95	+2.23
	16-6-5, 57–59	110.6	—	+0.83	+2.49
	16-6-6, 58–60	112.1	—	+0.89	+2.23
	16-7-1, 59–61	113.8	—	+0.70	+2.54
	16-7-1, 118–120	114.8	—	+0.95	+2.19
	16-7-2, 58–60	115.2	—	+0.94	+2.18
	16-7-2, 120–122	115.9	—	+0.95	+2.34
	16-7-3, 34–36	116.5	—	+0.59	+2.26
	16-7-3, 110–112	117.3	—	+0.81	+2.39
	16-7-4, 60–62	118.3	—	+1.03	+2.45
	16-7-5, 38–40	119.6	—	+0.95	+2.48
	16-7-5, 97–99	120.2	—	+0.96	+2.50
	16-7-6, 29–31	121.0	—	+1.07	+2.39
	16-8-1, 35–37	125.8	—	+0.82	+2.74
	16-8-3, 44–46	128.8	—	+0.60	+2.56
	16-8-5, 35–37	131.8	—	+0.59	+2.85
	16-9-1, 124–126	135.8	—	+0.87	+2.86
	16-9-3, 105–107	138.5	—	+0.72	+2.86
	16-9-5, 136–138	141.8	—	+0.75	+2.60

and *O. universa* are the same at Sites 16 and 519. This is certain to reflect similar surface and bottom-water conditions at the two sites during the late Miocene.

In Figure 4, the carbon-isotope data for benthic foraminifers (*P. wuellerstorfi, N. umbonifera*, and *Oridorsalis umbonatus*) from Sites 519 and 521 are replotted against the magnetostratigraphic ages determined by Tauxe et al. The graph shows a tendency toward more negative carbon-isotope values after 6 Ma for all three species. Thus, the late Miocene carbon-isotope shift recorded in the sediments of the southeast Atlantic sites is apparent, although very small.

Oxygen Isotopes

The oxygen-isotope data for planktonic and benthic foraminifers from Site 519 are plotted in Figure 2. The upper Miocene $\delta^{18}O$ values for *Orbulina universa* and *Planulina wuellerstorfi* fluctuate in a similar manner, that is, from a low value at about 9 Ma (~143 m sub-bottom) to a maximum value at 7.4 Ma (128.6 m sub-

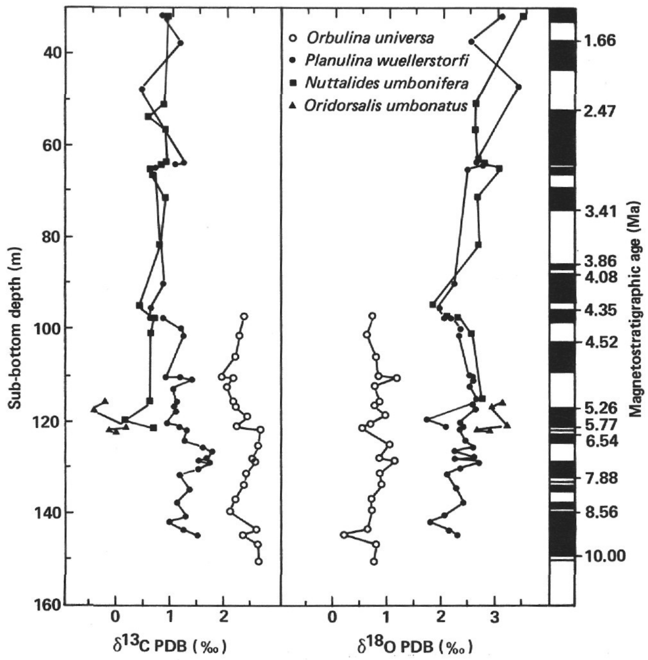

Figure 2. Stable-isotope stratigraphy for Site 519: oxygen- and carbon-isotope content of planktonic foraminifers (open symbols) and benthic foraminifers (filled symbols) versus sub-bottom depth. Magnetostratigraphic age is as determined by Tauxe et al. (this vol.).

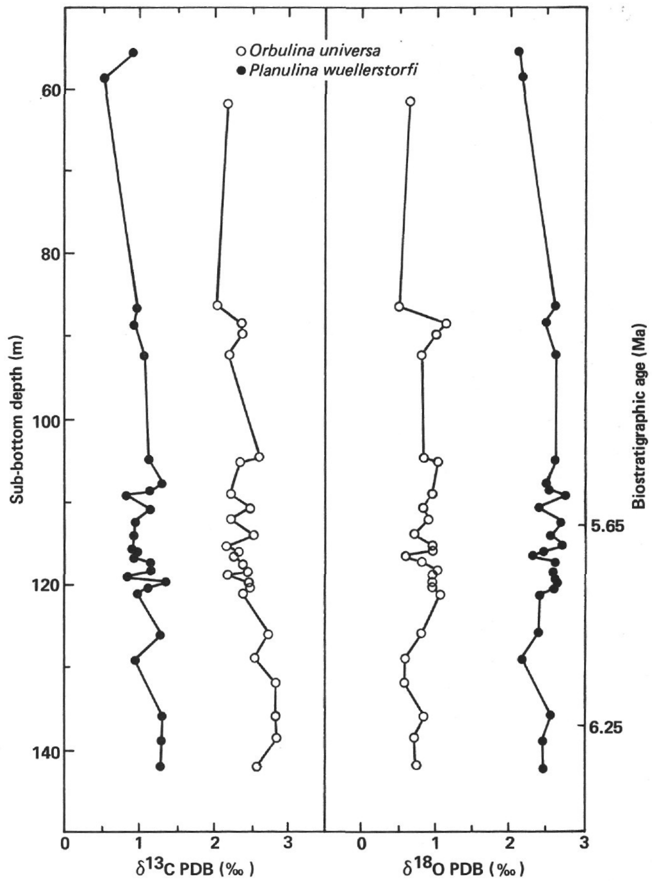

Figure 3. Stable-isotope stratigraphy for Site 16: oxygen- and carbon-isotope content of planktonic and benthic foraminifers versus sub-bottom depth. Symbols as in Fig. 2. Biostratigraphic age is based on FAD of *Amaurolithus primus* at 6.25 Ma and FAD of *A. amplificus* at 5.65 Ma (Haq et al., 1980).

bottom) to a second, but lesser, minimum at about 6.4 Ma (121.7 m sub-bottom). There is then a second maximum between 5.2 and 5.0 Ma (115.8 and 110.8 m sub-bottom). In the early Pliocene, the $\delta^{18}O$ value of *P. wuellerstorfi* declines significantly, reaching a minimum at 4.3 Ma (95.15 m sub-bottom). Afterwards, the $\delta^{18}O$ value again increases, reflecting Northern Hemisphere glaciation, as discussed by Weissert et al. (this vol.). The equivalent lower Pliocene data for *O. universa* are not available for comparison.

In the case of Site 16, the age control is limited to two biostratigraphic levels designated by the first common occurrence of *Amaurolithus primus* and the first appearance of *A. amplificus*. The oxygen-isotope stratigraphy of *P. wuellerstorfi* for the late Miocene of Site 16 (Fig. 3) shows that from a minimum value at 130 m sub-bottom (~6 Ma) the value reaches a maximum at 120 m sub-bottom (~5.8 Ma). From this point, the upper Miocene $\delta^{18}O$ values tend to show smaller-scale fluctuations while remaining near the maximum value. This tendency probably continues at least until 86.4 m sub-bottom, which is certainly near the end of the Miocene, inasmuch as sediments from this depth contain the nannofossil zones NN11/NN12. The isotope stratigraphy for *O. universa* shows a similar pattern.

In Figure 4, the oxygen-isotope data for benthic foraminifers from Sites 519 and 521 are plotted against magnetostratigraphic age. The graph illustrates the isotopic changes recorded in the southeast Atlantic bottom waters from the middle Miocene to the Pliocene. It is assumed that these global changes reflect continental ice-volume fluctuations, with increases corresponding to coolings and decreases to warmings. A very limited number of analyses of *Oridorsalis umbonatus* from the middle Miocene (15 to 14 Ma) indicate a sharp cooling ($\Delta\delta^{18}O = +1.0$ ‰), which has been previously observed for this time (e.g., Kennett, 1977). Between 14 and 9.5 Ma, we have no data. Our data continue at 9.3 Ma with values for *P. wuellerstorfi*. Around 9 Ma, there appears to have been a warming followed by a steady cooling; the cooling reached a plateau $\delta^{18}O$ value at 7.5 Ma, declined somewhat, and then peaked again at 5.2 Ma. This latter cooling is also seen in the limited data for *Nuttalides umbonifera* and *O. umbonatus*. Afterwards, the $\delta^{18}O$ value declines, with a very sharp warming between 4.5 and 4.3 Ma. This is clearly followed by a cooling trend in the middle Pliocene, which is discussed by Weissert et al. (this vol.).

DISCUSSION

Late Miocene Carbon-Isotope Shift

This study was undertaken not only to determine the stable-isotope stratigraphy for the southeast Atlantic sites but also to attempt to correlate this stratigraphy with the paleoceanographic events that occurred in the late Miocene and early Pliocene. It would be impossible to attempt such an exact correlation without the corre-

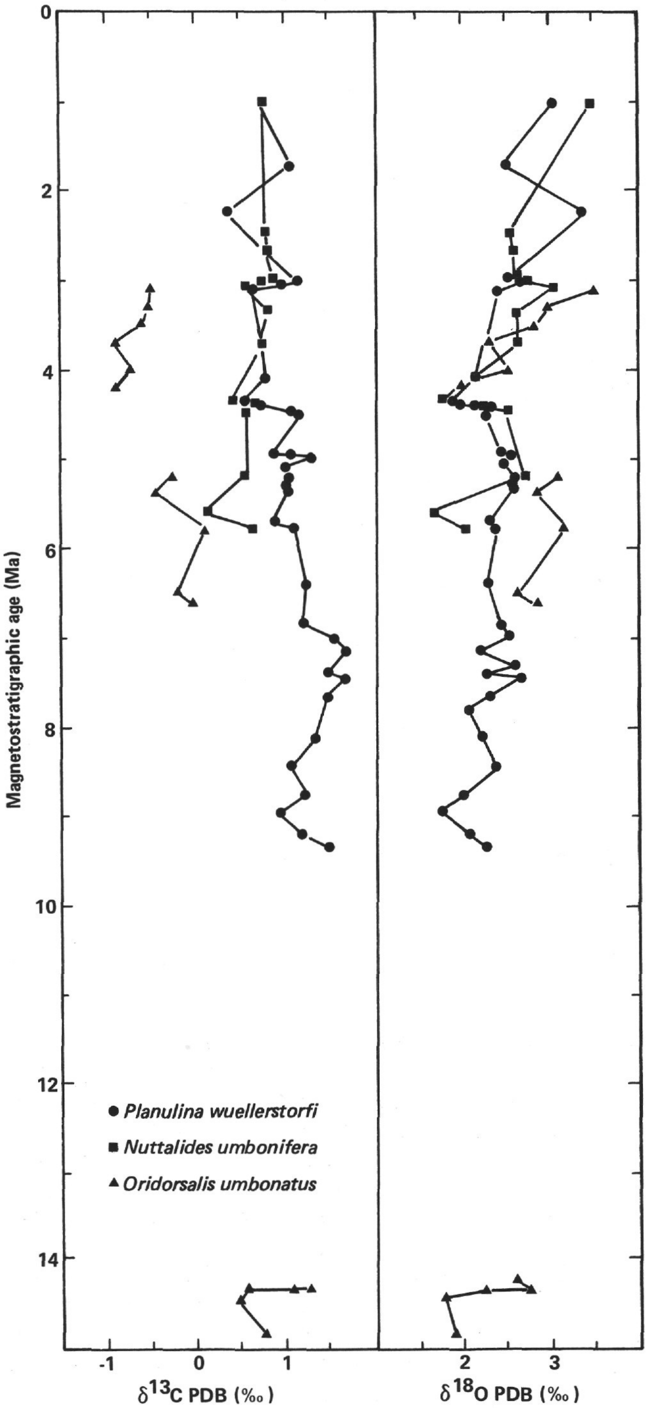

Figure 4. Stable-isotope stratigraphy for Sites 519 and 521: oxygen- and carbon-isotope content of benthic foraminifers versus magnetostratigraphic age as determined by Tauxe et al. (this vol.).

sponding magnetostratigraphy, which enables us to place the isotope stratigraphy within a time frame. Figure 5 is a compilation of the biological and geochemical data that have accrued from our study of sediments from Site 519 plotted against the magnetostratigraphic ages determined by Tauxe et al. (this vol.). The first appearance datums (FAD) of three species further aid in the delineation of our time scale. The Miocene/Pliocene boundary can be placed between the FAD of *Globorotalia crassaformis* at 5.05 Ma (112.4 m sub-bottom) and the FAD of *G. cibaoensis* at 5.21 Ma (115.55 m sub-bottom). The FAD of *Amaurolithus amplificus* probably occurs at 5.74 Ma (120.76 m sub-bottom) in the lowermost part of Epoch 5 (Hsü et al., this vol.), although Haq et al. (1980) place the FAD of *A. amplificus* at 5.65 Ma.

In the study of changes in the carbon-isotope content of benthic fauna, it may be critical to know whether the water mass has changed during the period of interest. The work of Kroopnick (1980) has demonstrated that different water masses have different $\delta^{13}C$ values. In the South Atlantic, the NADW can be as much as 0.5 ‰ more enriched in carbon-13 than the AABW. Site 519 is situated on the flank of a subsiding mid-ocean ridge and, with subsidence, it undoubtedly passed through different water masses. The transition from deep to bottom water is marked by a significant increase in the percentage of *Nuttalides umbonifera* in the sediments at 125 m sub-bottom (6.95 Ma) (site chapter, this vol.). From subsidence curves, it was estimated that this transition occurred at a water depth of approximately 3000 m. In addition, above 125 m sub-bottom, there is an abrupt change in the mineralogy of the noncarbonate components. In particular, the amount of quartz increases and the X-ray reflection peak for smectite shifts from 14 Å to a normal 15 Å; both of these facts indicate a change in water mass (Karpoff, this vol.). Apparently, by 7 Ma Site 519 had entered into the AABW, where it has remained until today. Between 7.15 and 6.85 Ma, a decrease of 0.5 ‰ in the $\delta^{13}C$ value of *Planulina wuellerstorfi* was measured. This decrease could also represent the transition from deep to bottom waters. Perhaps the deep waters were more enriched in $\delta^{13}C$ than the bottom waters, as in the modern South Atlantic (Kroopnick, 1980).

The occurrence of the late Miocene carbon-isotope shift has been placed between 6.10 and 5.90 Ma (Haq et al., 1980). If this time interval is examined in Figure 5, it can be seen that the $\delta^{13}C$ value of *P. wuellerstorfi* does decrease, but the decrease is very slight, and the $\delta^{13}C$ value remains essentially at the more negative value until the end of the Miocene. In comparison to the constant late Miocene value, the $\delta^{13}C$ value at Site 519 shows wide fluctuations after the Miocene/Pliocene boundary (Fig. 5).

The magnitude of the carbon-isotope shift reported for other deep-sea sites, mainly in the Pacific but also from the Indian and South Atlantic oceans, is generally about -1 ‰ (Haq et al., 1980). The relatively small carbon-isotope shift measured at Site 519 indicates that the $\delta^{13}C$ value of AABW was not significantly affected by the change(s) that caused the global readjustment of the carbon signal at other locations. On the other hand, the carbon-isotope shift (~ -0.8‰) reported for DSDP Site 357 in the South Atlantic (Haq et al., 1980) was found in sediments from the Rio Grande Rise. Site 357, presently at a water depth of 2086 m, would have been located within the deep-water mass between 6.10 and 5.90 Ma, and, therefore, benthic foraminifers should

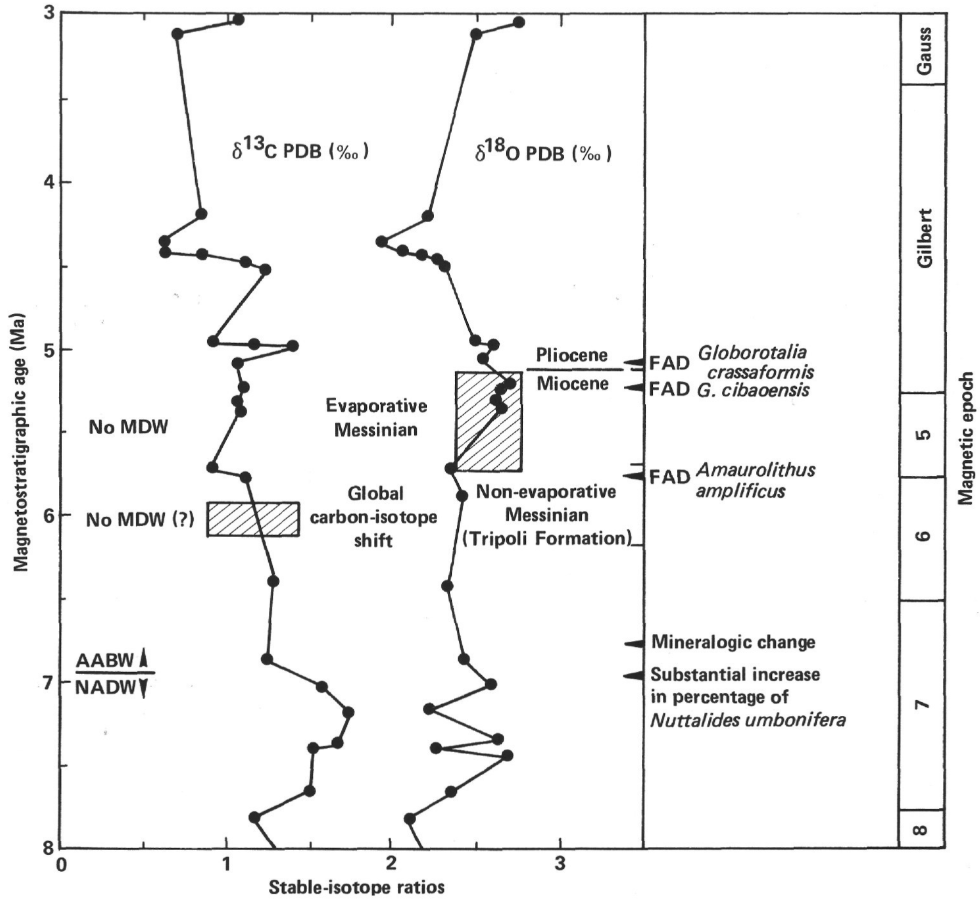

Figure 5. Compilation of late Miocene–early Pliocene paleoceanographic events with respect to stable-isotope changes, as recorded in *Planulina wuellerstorfi*, versus magnetostratigraphic age (Tauxe et al., this vol.). See text for discussion.

give an isotopic signal representing changes in deep-water circulation. The magnitude difference of the late Miocene carbon-isotope shift between Site 357 and Site 519 (and Site 16) is thus apparently related to the different water masses, NADW versus AABW, in which the benthic foraminifers lived. This, then, lends credibility to the idea that the cessation of the inflow of Mediterranean deep water (MDW) into the Atlantic resulted in changes in the deep-water circulation pattern and ultimately produced the carbon-isotope shift (Bender and Keigwin, 1979). A change in the deep-water circulation pattern would not necessarily affect South Atlantic bottom-water circulation. Therefore, the carbon-isotope signal at Site 519 could remain unchanged, while that at Site 357 could reflect the proposed change in deep-water circulation.

Messinian Isolation and Desiccation of the Mediterranean

If the isolation of the Mediterranean Sea from the Atlantic Ocean resulted in the changes in deep-water circulation that produced the carbon-isotope shift, it becomes significant to attempt to stratigraphically correlate events in land sections from the evaporating Mediterranean with the deep-sea record. An ideal candidate for such a correlation is the extensively studied Messinian sequence from Sicily (Ogniben, 1957; and Decima and Wezel, 1973). Evaporite sedimentation was preceded by the deposition of the Tripoli Formation, which is comprised of cyclic marlstones and diatomites. A comparison of these cyclic sediments with glacial–interglacial cycles suggests that Antarctic glaciation largely controlled the passage of water into the Mediterranean through sea-level fluctuations during the deposition of the Tripoli Formation (McKenzie et al., 1979/1980). In Figure 5, this period of cyclic sedimentation is designated the non-evaporative Messinian. At some point, the influx of Atlantic water halted completely, and the evaporative phase of the Mediterranean Messinian began, as demonstrated by the abundance of evaporite sediments. In the Mediterranean region, the Miocene/Pliocene boundary marks the end of the desiccation phase, with the recommencement of pelagic sedimentation.

In the present circulation pattern, surface waters flow into the Mediterranean from the Atlantic and deep, more saline waters return to the Atlantic across the Gibraltar sill. This inflow of dense Mediterranean water is a major contributor to the formation of the NADW (Sverdrup et al., 1942). It may be that during the time when the Tripoli Formation was deposited, surface waters continued to enter the Mediterranean from the Atlantic but the return flow to the Atlantic was inhibited. This change would permit the cyclic deposition of marine diatomites in the Mediterranean but hinder the formation of the NADW in the Atlantic. The coincidence of the timing of the late Miocene carbon-isotope shift with the deposition of the Tripoli Formation is critical. After making their stable-isotope study of the late Miocene Carmona–Dos Hermanos section in southwestern Spain, Loutit and Keigwin (1982) proposed that the global decrease in $\delta^{13}C$ was correlated with the lowermost Tripoli Formation. Gersonde (1980) estimated, by using biostratigraphic correlations based on diatom assemblages, that the Tripoli Formation was deposited between the upper part of magnetic Epoch 6 and the upper part of the lowest normally magnetized interval in Epoch 5, that is, between 6.2 and 5.7 Ma, the period of time that includes the carbon-isotope shift (Fig. 5). Therefore, the period from 5.7 Ma until the end of the Miocene between 5.2 and 5.1 Ma, which is the period in which the Mediterranean was totally desiccated and evaporites were deposited, is the evaporative Messinian. This period correlates with the late Miocene glaciation, which is seen in the oxygen-isotope record of *Planulina wuellerstorfi* at Site 519 (Fig. 5). Oxygen-isotope variations of benthic fauna living in cold bottom waters reflect changes in the amount of water stored in continental ice. The increased storage of water on the continents, as seen in the increase in $\delta^{18}O$ between 5.7 and 5.1 Ma, could have been sufficient to lower the global sea level to a critical point that essentially cut off the surface inflow of Atlantic water into the already semi-isolated Mediterranean.

Shackleton (1977) estimates that a 0.1 ‰ increase in $\delta^{18}O$ represents a 10-m drop in sea level. The Site 519 record shows an increase of approximately 0.4 ‰, or a sea-level drop of 40 m. From their paleontological studies of the Andalusian Stage stratotype located in the Atlantic-connected Guadalquivir Basin of southwestern Spain, Van Couvering et al. (1976) proposed that the water level of the basin was reduced from 70–100 m to approximately 30 m in the uppermost levels of the foraminiferal zone N17, that is, in the latest Miocene. They link the Spanish regression to the Messinian salinity crisis but recognize that the magnitude of the sea-level drop was not sufficient to isolate the Mediterranean Basin unless it had already been nearly isolated by tectonism. They follow Ruggieri (1967), in suggesting that the Messinian in the Mediterranean had two phases: (1) an initial deep-water phase between 6.5 and 5.5 Ma when the sill between the Atlantic and Mediterranean was tectonically elevated and (2) a phase of total isolation between 5.5 and 5.0 Ma caused by the worldwide lowering of sea level. Phase 1 is the time when the Tripoli Formation was deposited; Phase 2 represents the period of evaporite deposition.

Early Pliocene Sea-Level Rise

In the Mediterranean, the end of the Messinian is marked by the restoration of pelagic sedimentation that resulted from the reinvasion of marine waters from the Atlantic. The oxygen-isotope record from Site 519 suggests that a warming trend began at the Miocene/Pliocene boundary and that the rate of warming increased at 4.5 Ma, reaching a climax at 4.3 Ma. The depletion of oxygen-18 in benthic foraminifers indicates the melting of continental ice and, hence, the rising of sea level. The total decrease in $\delta^{18}O$ from 5.2 to 4.3 Ma is 0.7 ‰, suggesting a sea-level rise of 70 m. Global sea-level curves based on seismic stratigraphy (Vail et al., 1977) indicate a relatively low stand in the late Miocene and a high stand in the early Pliocene. After 4.3 Ma, the climate deteriorated again, as seen in the oxygen-isotope record (Figs. 4 and 5).

Initial data from DSDP Leg 71 in the southwest Atlantic Ocean (Ludwig et al., 1980) indicate trends for the late Miocene and early Pliocene that are similar to those noted at Site 519. Unconformities and very low species diversity in planktonic foraminifers and calcareous nannofossils attest to severe climatic conditions during the late Miocene; siliceous microfossils indicate a warming period in the latest Miocene and earliest Pliocene. Previous paleontological studies in the Antarctic seas (Ciesielski and Weaver, 1974) suggest that surface water temperatures between 4.3 and 3.95 Ma were too warm for the existence of the West Antarctic ice sheet. The Messinian salinity crisis in the Mediterranean may have ended when the basin was flooded by water rising as a result of the melting of the West Antarctic ice sheet, or the end of the salinity crisis may have triggered climatic changes that caused a retreat of Antarctic glaciers.

CONCLUSIONS

By combining magnetostratigraphic, biostratigraphic, and isotope stratigraphic evidence we were able to correlate our data from the southeast Atlantic Ocean with paleoceanographic events that occurred in the late Miocene and early Pliocene. In the case of the global late Miocene carbon-isotope shift, only a rather minor carbon-isotope change was identifiable in our record. The minuteness of this carbon-isotope shift relative to those measured elsewhere (Haq et al., 1980) indicates that the carbon-isotope content of the Antarctic bottom waters (AABW) was not significantly altered between 6.10 and 5.90 Ma. Further, the existence of a carbon-isotope shift in sediments from the Rio Grande Rise (Haq et al., 1980) suggests that the carbon-isotope change may have been confined to South Atlantic deep waters. On the other hand, the late Miocene cooling and early Pliocene warming, or sea-level fall and rise, respectively, were clearly recorded in the oxygen-isotope values of benthic foraminifers. The magnetostratigraphic ages for these events correspond exceedingly well with the interval proposed for the Messinian salinity crisis in the Mediterranean. Their contemporaneity strongly suggests that (1) the late Miocene carbon-isotope shift was caused by the cessation of the influx of dense Mediterranean deep water into the Atlantic Ocean, which produced changes in deep-water circulation, (2) the final isolation and desiccation of the Mediterranean was caused by the lowering of sea level due to the expansion of the Antarctic ice sheet and (3) the abrupt end of the Messinian salinity crisis was brought about by the melting of the West Antarctic ice sheet. In general, the results of our correlation demonstrate the feasibility of using the high-resolution stratigraphy that can be developed for deep-sea sediments obtained with the hydraulic piston corer to study paleoceanographic events in greater detail.

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