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2. MID- TO LATE-QUATERNARY VARIATIONS IN THE OXYGEN ISOTOPE SIGNATURE OF *GLOBIGERINOIDES RUBER* AT SITE 1006 IN THE WESTERN SUBTROPICAL ATLANTIC[1]

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ABSTRACT

The 1.4-m.y.-long stable oxygen isotope record of Site 1006 in the low-latitude North Atlantic Ocean shows large glacial/interglacial amplitude changes caused by a combination of temperature and salinity fluctuations. A trend of increased sea-surface temperatures during the interglacial periods is present in the record beginning at isotopic Stage 11 and ultimately leading to the lightest $\delta^{18}O$ values in isotopic Stages 9, 5, and 1. Maximum $\delta^{18}O$ values are recorded during glacial isotopic Stages 6 and 8. Stable isotopic variability increased during the Brunhes Chron at the 100-ka time scale. The large amplitude changes can best be explained by global and regional ocean circulation changes. Increased strengthened return flow of warm salty water from the Pacific may have occurred during interglacial periods since isotopic Stage 11, which was largely reduced during glacial periods. The large climate fluctuations had a profound effect on the shallow-water carbonate production of the Great Bahama Bank. The aragonite content of the sediments shows fluctuations that follow the $\delta^{18}O$ record. The leeward side of the Great Bahama Bank received increased input of platform material during sea-level highstands when the sea-surface waters were warm.

INTRODUCTION

Past low-latitude climate changes are still under considerable debate. Different techniques to extract past sea-surface temperatures are being used, yet all are subject to potential errors and uncertainties. The best approach is to use a multitude of techniques; however, it is prohibitively time consuming. Therefore, at this stage, we present the stable isotope record of Site 1006 to obtain a first impression of climate variability in the low-latitude northern Atlantic, with the intention of producing additional proxy records in the future.

Site 1006 was drilled in 658 m of water in the northern portion of the Santaren Channel, ~30 km from the western platform edge of the Great Bahama Bank (Fig. 1). It is the most distal member of the five transect sites located westward of the Great Bahama Bank, positioned on a thick sequence of Neogene-age drift sediments. The drift sediments interfinger with neritic material derived from the carbonate platform. The result is an expanded section suitable for paleoceanographic studies because of its high temporal resolution. The advanced hydraulic piston corer (APC) system was used to retrieve the Neogene sequence, of which the upper seven cores of Hole 1006A were studied. The cores include 166-1006A-1H through 7H, which cover the upper 65 m of sediment.

The age model for Site 1006 was constructed by directly correlating its stable oxygen isotope record to the astronomically tuned record of Site 806 from the western equatorial Pacific (Berger et al., 1993). This provided the opportunity to make direct comparisons between different oceans in the low latitudes to highlight similar and dissimilar features of both records as a response to glacial/interglacial variability. As we shall show, the low-latitude glacial/interglacial variability is about twice as large, or even larger, in the Atlantic than in the western Pacific throughout the last 1.4 m.y.

Site 1006 is situated in an interesting position in 'the source area of the Gulf Stream' and thus the stable oxygen isotope record reflects its sea-surface conditions. The Gulf Stream originates from the western equatorial Atlantic, and its low-latitude variability may have possibly influenced northern-latitude climate variations. We shall show that considerable interglacial warming has taken place since stable oxygen isotopic Stage 11. Transport of additional heat into the North Atlantic during the last five interglacial periods may have contributed to the rapid melting of the large ice sheets that developed during the late Quaternary glacial periods.

Finally, the platform production of the Great Bahama Bank is explored at this site as a response to sea-level changes and sea-surface conditions. This is more fully explained in the chapter by Rendle et al. (Chap. 6, this volume). Here we use the aragonite records to aid in the reconstruction of the age model because the aragonite cycles perfectly match those of the stable oxygen isotope record and are sometimes better developed than the oxygen isotope cycles.

METHODS

Hole 1006A was sampled on board the *JOIDES Resolution* during Leg 166 with the idea of sampling additional selected intervals of Hole 1006B at the Core Repository in Bremen to bridge missing intervals in Hole 1006A. Sampling densities throughout Hole 1006A were typically one sample every 20 cm, although the sampling density was greater in several intervals. This study was designed to develop the stable isotope record to ~1.4 Ma, while our shipboard colleagues would extend the stable isotope record farther down into the Pliocene and Miocene. Biostratigraphic datum levels used in this study are given by T. Sato (unpubl. data).

Oxygen Isotope Analysis

Splits of each sample were oven dried at 60°C, weighed, and washed over a 63-µm stainless-steel mesh sieve. The coarse sediment fraction (>63 µm) was dry sieved through 250- and 355-µm mesh sieves. Tests of the planktonic foraminiferal species *Globigerinoides ruber* (d'Orbigny, 1839) were hand picked under a binocular microscope from the 250- to 355-µm sediment fraction. We have chosen *G. ruber* because it dwells in the mixed layer for all size fractions (Kroon and Darling, 1995). Tests were soaked in methyl alcohol (analytical reagent) for several minutes and later cleaned in an ultrasonic

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bath to remove adherent contaminants. Following ultrasonic cleaning, excess methyl alcohol was drawn off with tissue paper and any residual alcohol was allowed to evaporate. Foraminiferal sample weights were typically <0.1 mg and composed of 5–8 specimens. Following cleaning, foraminiferal tests were reacted in orthophosphoric acid (specific gravity = 1.9) at 90°C, and the resulting CO_2 gas was analyzed using a VG Isogas precision isotope ratio mass spectrometer located at the University of Edinburgh. Precision for oxygen isotope analysis was 0.08‰ (standard deviation for 100 analyses of an ‘in-house’ standard carbonate [SM1] conducted over several months). The oxygen isotope data for Holes 1006A and 1006B are given in Tables 1 and 2.

X-ray Diffractometry

Marine sediments accumulating on basin and ocean floors adjacent to shallow-water carbonate banks and platforms (<200 m water depth), termed “periplatform ooze” by Schlager and James (1978), consist predominantly of a mixture of carbonate platform-derived and planktonic-derived sediments (Boardman et al., 1986). The shallow-water carbonates have a different mineralogical composition than the deep-water carbonates. Modern tropical shallow-water carbonate sediments are primarily composed of aragonite and high-Mg calcite (HMC), with subordinate amounts of low-Mg calcite (LMC) (Friedman, 1965; Land, 1967; Milliman, 1974; Ginsburg and James,

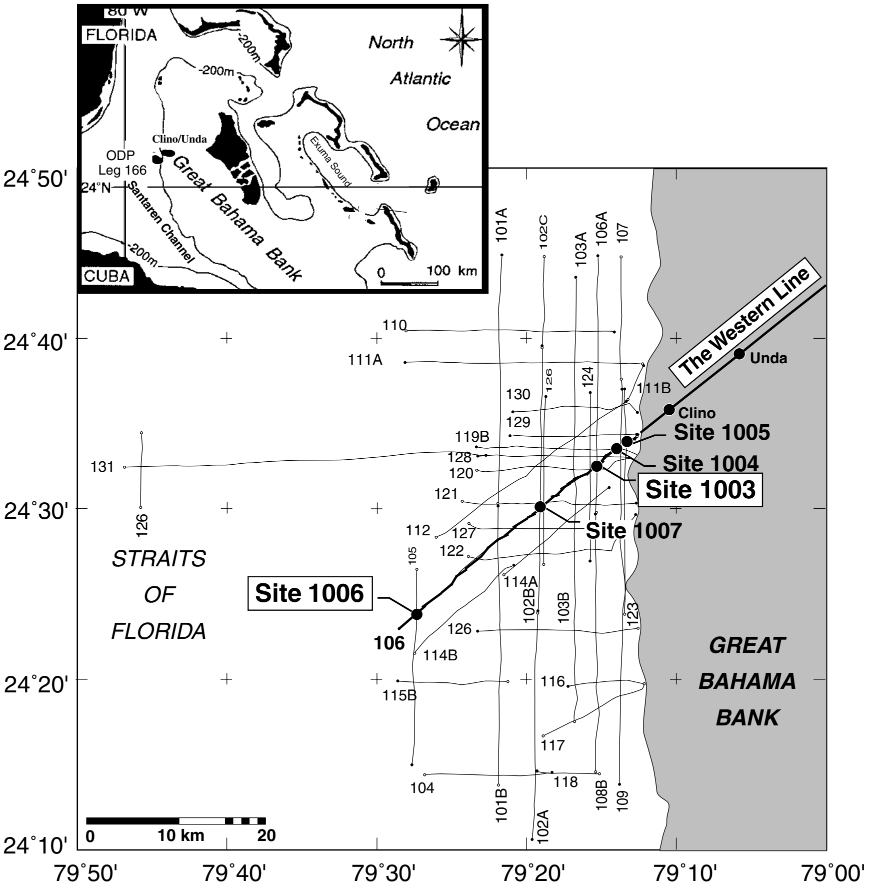

Figure 1. Positions of individual sites drilled during Leg 166. Site 1006 is the deepest site located in 658 m of water in the northern portion of the Santaren Channel on the leeward side of the Great Bahama Bank.

Table 1. Hole 1006A isotope and aragonite data.

Core, section, interval (cm)	Depth (mbsf)	$\delta^{18}O$ (PDB)	Aragonite (wt%)
166-1006A-			
1H-1, 15-17	0.15	–1.54	76.72
1H-1, 35-37	0.35	–1.94	79.00
1H-1, 55-57	0.55	–1.80	83.25
1H-1, 75-77	0.75	–1.76	86.95
1H-1, 95-97	0.95	–1.76	88.75
1H-1, 115-117	1.15	–1.54	90.47
1H-1, 135-137	1.35	–1.67	88.26
1H-2, 15-17	1.65	–1.44	79.28
1H-2, 35-37	1.85	–1.45	61.37
1H-2, 55-57	2.05	–0.87	55.19

Note: PDB = Peedee belemnite standard.

This is a sample of the table that appears on the volume CD-ROM.

Table 2. Hole 1006B isotope and aragonite content data.

Core, section, interval (cm)	Depth (mbsf)	$\delta^{18}O$ (PDB)	Aragonite (wt%)
166-1006B-			
2H-3, 5-7	8.55	0.91	
2H-3, 25-27	8.75	1.18	51.90
2H-4, 45-47	8.95	0.82	25.06
2H-3, 65-67	9.15	0.98	
2H-3, 85-87	9.35	0.29	65.24
2H-3, 105-107	9.55	–0.84	74.46
2H-3, 125-127	9.75	–0.50	81.64
2H-3, 145-147	9.95	–0.86	77.37
2H-4, 15-17	10.15	–0.64	66.84
2H-4, 35-37	10.35	–0.15	69.43

Note: PDB = Peedee belemnite standard.

This is a sample of the table that appears on the volume CD-ROM.

1974). This mineralogy reflects the composition of the skeletal and nonskeletal components that dominate the shallow-water setting. In contrast, the mineralogy of deep-water carbonate sediments is largely controlled by the production of pelagic-driven carbonate and by the selective dissolution of the more soluble carbonate minerals (Friedman, 1965). Deep-water carbonate-rich sediments contain abundant LMC with minor amounts of aragonite and HMC.

The relative abundance of a variety of carbonate minerals present in Site 1006 sediments was quantified by X-ray diffraction (XRD). These minerals include aragonite, LMC, HMC, and dolomite. XRD was typically performed on ~0.5 g of the <63 µm fraction of the sediment. Each sample was oven dried at 60°C before being hand ground in an agate mortar (with analytical grade acetone) for exactly 4 min (Milliman, 1974). Individual samples were drawn up in a pipette, dispersed on a glass slide (2-cm diameter), smeared out to produce a sediment slurry of standard thickness and diameter, and dried at room temperature before being subject to X-rays. Each individually prepared sample was then analyzed using a Philips PW 1011/1050 automatic powder diffractometer (PW 1808 sample changer) at 40 kV and 50 mA through a scan from 5° to 40° 2q (Cu-ka) at a low scanning speed of 0.040° per second for optimal resolution. A calibration curve was produced for aragonite using a modified version of the sample-spiking method proposed by Gunatilaka and Till (1971) and Alexander (1996). The aragonite percentage data of Hole 1006A and 1006B are given in Tables 1 and 2, respectively.

AGE MODEL FOR SITE 1006

We attempted to correlate the Site 1006 oxygen isotope record to the Site 806 profile as provided by Berger et al. (1993) to facilitate low-latitude Atlantic-Pacific comparisons. The stable isotope record of Site 806 was tuned by using the obliquity cycles apparent in the record (Berger et al., 1993). Site 806 is located in an area where equatorial upwelling influences are minimal; therefore, sea-surface temperatures were not affected by upward transport of cool, deeper waters.

The oxygen isotope and aragonite profiles of Hole 1006A were first used for identifying the oxygen isotopic stages (Fig. 2). Hole 1006B provided a means to fill in apparent gaps in the oxygen isotope record or condensed sections for those portions of the record that the Leg 166 Shipboard Scientific Party (1997) indicated were missing in Hole 1006A. Gaps were identified based on correlations between the magnetic susceptibility profiles of Holes 1006A and 1006B. We did not use the shipboard version of the meters composite depth (mcd) scale directly because it was obvious at the time of construction that potential errors were made and that in most places Hole 1006A was more complete than originally thought.

High-amplitude δ^{18}O variations are apparent throughout the entire record of Hole 1006A and the amplitude seems to increase within the top 17 m of sediment. U/Th-series dating and nannofossil biostratigraphy were used to constrain the assignment of stable oxygen isotopic stages, keeping in mind the position of the core breaks. The influence of the core breaks on the completeness of the record at this time is unknown. Absolute dates based on the U/Th-series were provided by Henderson et al. (Chap. 3, this volume). The three given dates indicate the position of oxygen isotopic Stages 1, 5, and 9. This gives us an immediate problem because isotopic Stage 7 seems to be missing because of the first core break (Fig. 2). Sediments from Hole 1006B were used in an attempt to fill the gap and an interval was found, evidently representing isotopic Stage 7, that was not present in Hole 1006A. Unfortunately it is not entirely clear whether the isotopic Stage 7 to 8 transition in Hole 1006B is complete because the interval used from Hole 1006B was also terminated by a core break; therefore, part of the record might still be missing. There is no distinct overlap with Hole 1006A.

The nannofossil biostratigraphy was used to further constrain the assignment of isotopic stages. Nannofossil datum levels were provided by T. Sato (unpubl. data). The last occurrence (LO) of *Pseudoemiliania lacunosa* indicates the position of isotopic Stage 12 in the top of Core 166-1006A-3H. Isotopic Stages 10 and 11 in the lower half of Core 166-1006A-2H and isotopic Stages 13, 14, 15, 16, and 17 in Core 166-1006A-3H were apparent. The second and third core breaks straddle the transitions between isotopic Stages 12 and 11 and Stages 18 and 17, respectively. However, inspection of the magnetic susceptibility record of the Hole 1006B shows that little or no sediment is missing at these boundaries. Core 166-1006A-4H contains isotopic Stages 18, 19, 21, and 22, the latter indicated by the LO of *Reticulofenestra asanoi.* The mid-Pleistocene revolution, as defined by Berger et al. (1993), can now readily be seen in the Hole 1006A δ^{18}O record at ~33 meters below seafloor (mbsf) (Fig. 2).

The lower part of the Hole 1006A record is well constrained by three excellent nannofossil events. The base of *Gephyrocapsa parallela* occurs in the middle of Core 166-1006A-5H and indicates the position of isotopic Stage 30. Recognition of this datum shows that the core break between Cores 166-1006A-4H and 5H has not significantly affected the completeness of the record, as all isotopic stages from 22 to 32 could be clearly recognized. The next deeper nannofossil event, the base of *R. asanoi,* occurs at the lower end of Core 166-1006A-5H, and the LO of large *Gephyrocapsa*, which indicates isotopic Stage 37, lies within Core 166-1006A-6H. Inspection of Hole 1006A shows that little sediment is missing across the core break between Cores 166-1006A-5H and 6H.

After identification of most of the isotopic stages, the Hole 1006A depth scale was ready for conversion to a time scale. Before doing so, the interval from Hole 1006B representing isotopic Stage 7 was patched into Hole 1006A to create a new depth scale (mcd) for Site 1006 (Fig. 3; Table 3). This was the only portion of the Hole 1006B record we have used, because it appeared that not much of the record in Hole 1006A had disappeared across the other core breaks. The stratigraphic tie points used to correlate to Site 806 are indicated in Table 4. A plot of linear sedimentation rates based on this age model is presented in Figure 4.

OXYGEN ISOTOPE RECORD

The lightest oxygen isotope values in the Site 1006 record were found in isotopic Stages 1, 5, 9, and 11. The Holocene values are very similar to the core-top *G. rube*r values as reported by Slowey and Curry (1987) from near Site 1006. The δ^{18}O values of the planktonic foraminiferal calcite from the Holocene section are in equilibrium with modern temperature and salinity values of the surface waters using the δ^{18}O-salinity relationship for the North Atlantic (Fig. 5) as given by Ganssen and Kroon (in press). Our observation that the most depleted values are seen in the Holocene section is unusual for deep-sea records; normally isotopic Stage 5e typically shows the lightest values, which makes the Holocene unique in the Bahamian area. It seems that this trend toward lighter values during the periods of maximum deglaciations began around isotopic Stage 11.

In general, the amplitude changes in the oxygen isotope record are large, especially in the top 17 m (Figs. 2, 3) because of the light values of δ^{18}O recorded during sea-level highstands since isotopic Stage 11 and maximum values of isotopic Stages 6 and 8 during sea-level lowstands. The Holocene/last glacial maximum transition shows a maximum isotopic shift of 2.15‰. The maximum interglacial-glacial amplitude difference is used here rather than an average (Slowey and Curry, 1987). We prefer the maximum amplitude δ^{18}O difference because bioturbation may have smoothed the record. About 1.1‰–1.3‰ can be ascribed to the ice volume/sea-level effect (Berger and Gardner, 1975; Chapell and Shackleton, 1986; Fairbanks, 1989), which leaves a residual δ^{18}O component of ~1‰ that can be attributed to salinity and temperature variations. Unfortunately, we cannot discriminate between these two parameters because independent proxies have not as yet been produced on this core. However, temperature reconstructions from other records in the western equatorial Atlantic have been a useful aid. The temperature proxies (Sr/Ca and $\delta^{18}O$) from the Barbados coral record suggest a temperature shift of 5°C from the last glacial maximum to the Holocene (Guilderson et al., 1994), although it is not clear how unambiguous the interpretation of such a record is. Nonetheless it would fit with a residual component of ~1‰ in the Hole 1006A record (0.2‰1°C, Epstein et al., 1953). This result means that salinity changes were small at the isotopic Stage 2/1 transition and had no significant effect on the $\delta^{18}O$ record. On the other hand, other proxies such as estimates from planktonic foraminiferal assemblages and alkenone analysis show relatively little cooling (1°–2°C) in the west equatorial Atlantic (CLIMAP, 1981; McIntyre et al., 1989; Sikes and Keigwin, 1994, Billups and Spero, 1996). For each 0.45‰ increase in $\delta^{18}O$ (Standard Mean Ocean Water [SMOW]), there is an increase of 1 practical salinity units (psu) in salinity (Fig. 5; Ganssen and Kroon, in press). This would imply a salinity effect of ~1 psu between the last glacial maximum and Holocene low-latitude Atlantic Ocean assuming a 2°C change, a result also found by Broecker (1989). Thus, the glacial Atlantic Ocean was saltier than today in this scenario. Unfortunately, it is not known which temperature proxy is more correct; this leaves us with the two different scenarios or a combination of the two, although in the discussion we argue that the temperature change based on coral data is probably more likely.

The isotopic shifts across the other transitions since isotopic Stage 12 are larger than the isotopic Stage 2/1 shift. The largest shifts occur from isotopic Stage 9 to 8 and from Stage 7 to 6. The maximum change of ~3‰ was measured from isotopic Stage 9 to 8 (Fig. 3), which could potentially imply a temperature drop of 9°C (assuming no salinity change) since the temperature optimum of isotopic Stage 9. However, some diagenetic effects may have altered the original

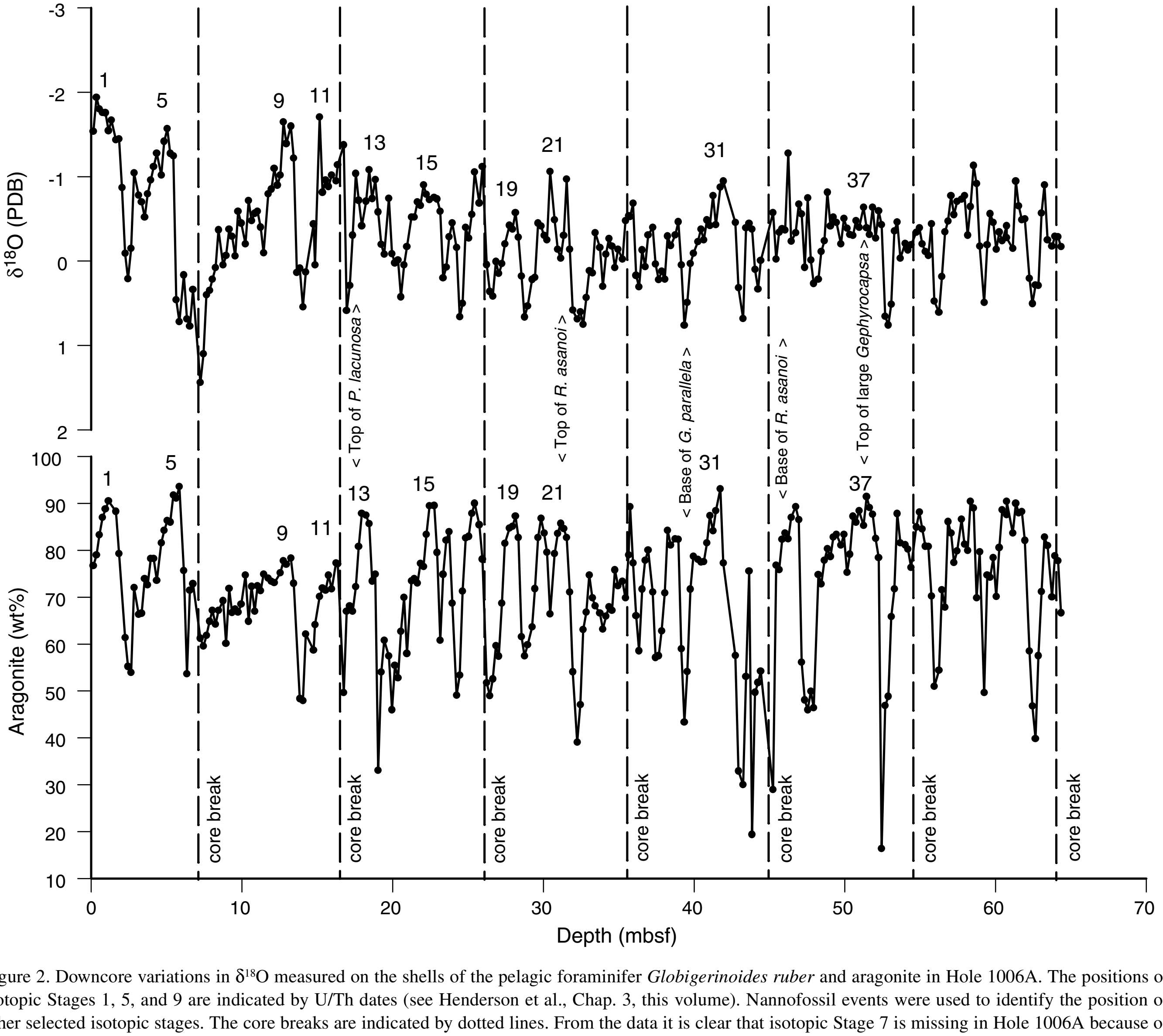

Figure 2. Downcore variations in $\delta^{18}O$ measured on the shells of the pelagic foraminifer *Globigerinoides ruber* and aragonite in Hole 1006A. The positions of isotopic Stages 1, 5, and 9 are indicated by U/Th dates (see Henderson et al., Chap. 3, this volume). Nannofossil events were used to identify the position of other selected isotopic stages. The core breaks are indicated by dotted lines. From the data it is clear that isotopic Stage 7 is missing in Hole 1006A because of the first coring gap and that the remainder of Hole 1006A is essentially complete. It is possible that small portions of the record are missing below isotopic Stage 11 at the core breaks, but inspection of the magnetic susceptibility and aragonite content profiles of Holes 1006A and 1006B shows that the potential coring gaps are insignificant in Hole 1006A. The aragonite cyclicity is clear; the cycles follow the oxygen isotope cycles and are thus being controlled by sea-level variations. During sea-level highstands, aragonite is produced on the platform and exported toward the deeper basins around the platform.

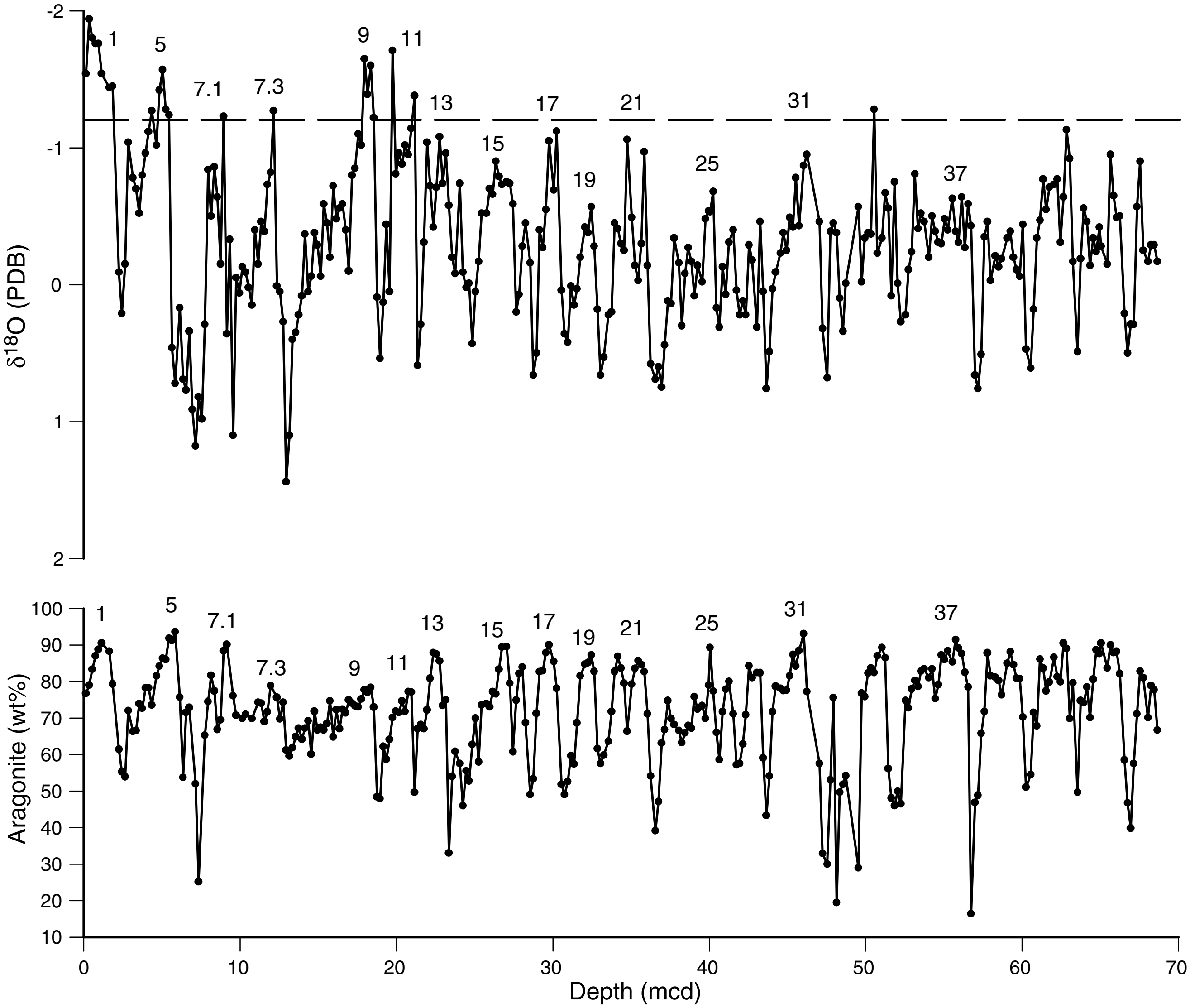

Figure 3. The $\delta^{18}O$ and aragonite records versus depth. Selected isotopic stages are indicated. The glacial/interglacial variations throughout the record are large relative to typical low-latitude records showing that sea-surface conditions varied significantly throughout the mid-late Quaternary. The largest change was found at the isotopic Stage 9/8 transition (~3‰). However, the heavy glacial Stage 8 $\delta^{18}O$ values may have been altered slightly by diagenesis. Amplitude change increased after isotopic Stage 12. The interglacial $\delta^{18}O$ values decreased after isotopic Stage 12, which is indicated by the horizontal dashed line.

Table 3. Site 1006 isotope and aragonite data.

Depth (mcd)	Age (ka)	$\delta^{18}O$ (PDB)	Aragonite (wt%)
0.15	1.5	–1.54	76.72
0.35	3.5	–1.94	79.00
0.55	5.4	–1.8	83.25
0.75	7.4	–1.76	86.95
0.95	9.3	–1.76	88.75
1.15	11.3	–1.54	90.47
1.65	16.2	–1.44	88.26
1.85	18.1	–1.45	79.28
2.25	22.1	–0.09	61.37
2.45	24.0	0.21	55.19

Note: PDB = Peedee belemnite standard.

This is a sample of the table that appears on the volume CD-ROM.

isotope values of the foraminiferal shells in the samples from the glacial Stage 8 and 6 sediments. The glacial sediments are much coarser in this part of the record. Coarse-grained sediments facilitate diagenesis (Rendle et al., Chap. 6, this volume).

The trend toward lighter values over the course of the Quaternary can best be explained by a temperature shift (see “Discussion” section, this chapter). The trend toward lighter values was also seen in the oxygen isotope record of Deep Sea Drilling Project Hole 502B farther to the south in the Caribbean (Prell, 1982), although amplitude changes were slightly smaller because another species, *Globigerinoides sacculifer,* was used to generate the isotope record. The records from Holes 502B and 1006A suggest that the warming is potentially characteristic for the Caribbean and Bahamian area.

The $\delta^{18}O$ values of the Hole 1006A interglacial periods prior to isotopic Stage 11 are on average 0.75‰ lower than the values of the interglacial periods after Stage 12, or ~3.5°C cooler, assuming no salinity change. The interglacial/glacial variability prior to isotope Stage 11 is large compared to the changes at Site 806 (Fig. 6). Maximum amplitude changes at Site 1006 are ~1.5‰, implying large

Table 4. Stratigraphic tiepoints of Site 1006 to Site 806.

Depth (mcd)	Age (ka)
2.65	24
3.55	68
5.05	128
5.85	144
8.75	212
10.75	248
11.75	284
17.55	342
17.75	344
18.35	356
19.45	404
21.05	428
21.25	440
22.75	500
24.85	548
27.05	580
28.25	624
28.55	630
29.75	692
30.25	708
30.75	748
32.45	788
33.05	804
35.45	859
36.55	900
37.35	908
38.25	916
40.05	952
40.65	976
41.55	988
41.75	1016
42.55	1032
43.65	1056
46.05	1080
47.55	1128
49.55	1172
51.85	1208
54.45	1256
56.75	1296
60.25	1340
63.55	1380
66.95	1428

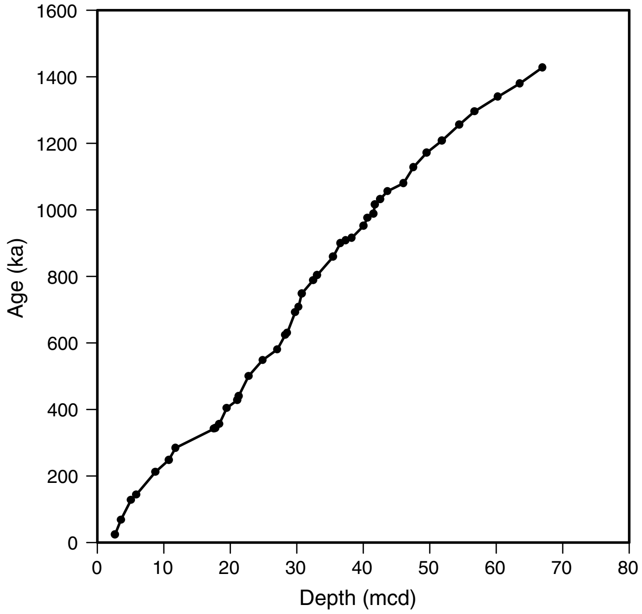

Figure 4. Age versus depth plot showing sedimentation rate changes at Site 1006. The age model is based on stratigraphic tielines to the astronomically tuned record of Site 806 (Berger et al., 1993). The selected tielines are given in Table 4.

temperature shifts on a glacial/interglacial basis unless salinity changes play a major role.

ARAGONITE CONTENT RECORD

The aragonite content record (in wt%) shows regular cycles that follow the oxygen isotope pattern (Figs. 2, 3). This trend in the Bahamian area has been previously described in Droxler et al. (1983, 1988) and Reijmer et al. (1988). During sea-level highstands, the platform sheds fine-grained material into a basin consisting mainly of the metastable mineral aragonite. Site 1006 was drilled on the leeward side of the Great Bahama Bank in 658 m of water, well above the depth of dissolution of aragonite (Droxler et al., 1991). Thus the aragonite content curve can be interpreted as the sole result of aragonite input (Boardman and Neumann, 1984). Maximum input of aragonite occurs consistently during the sea-level highstands. The carbonate fraction is clearly dominated by aragonite during interglacial periods (up to 85 wt%), while LMC and HMC prevail during glacial periods at 41 and 5 wt%, respectively (Rendle et al., Chap. 6, this volume). However, platform delivery was not entirely switched off during the lowstands, resulting in increased total sediment input during the highstands compared to the lowstands. Therefore, our sample density in the glacial intervals is low (normal pelagic sedimentation vs. elevated rates by platform input during interglacial periods); this explains the spikiness in the oxygen isotope and aragonite content records, and indicates that the true glacial-interglacial $\delta^{18}O$ amplitude changes are possibly underestimated here.

The fact that aragonite fluctuations at Site 1006 are entirely controlled by input from the platform and not by dissolution on the seafloor, coupled with the record's completeness, makes this site ideal for use as a standard site in studying dissolution patterns at other deeper sites. However, that is beyond the scope of this paper. Rendle et al. (Chap. 6, this volume) use Site 1006 as the most distal site of the Leg 166 transect to study the sedimentation patterns along the flank of the Great Bahama Bank.

DISCUSSION

The $\delta^{18}O$ record of low-latitude Atlantic Ocean Site 1006 appears to be much more variable than the low-latitude Pacific Ocean Site 806 record from the Ontong Java Plateau throughout its entire 1.4 m.y. (Fig. 6). The small interglacial/glacial amplitude changes in the Site 806 record may not be representative, however, of the western Pacific. Sedimentation rates are relatively low at Site 806, implying that bioturbation may have smoothed the isotope record. Other isotope records from the western Pacific show large amplitude changes as summarized in Martinez et al. (1997). Some western Atlantic sites exhibit the large amplitude changes such as shown at Site 1006. For instance, Curry and Oppo (1997) record a 2.1‰ change across the Holocene/glacial transition at Ceara Rise, very similar to the amplitude change recorded at Site 1006. High-resolution cores taken from drift sediments are most likely less affected by bioturbation, revealing the maximum Holocene/glacial amplitude.

Broecker (1989) noticed that the amplitudinal change is much larger for the last glacial maximum/Holocene transition in the Atlantic, and he concluded that a larger salinity contrast existed between the Atlantic and Pacific during the glacial periods than today, assuming that temperature change was on the same order of magnitude in both tropical oceans. The Atlantic-Pacific contrast appears to have been significantly larger during several glacial periods such as glacial Stages 6, 8, 14, 20, 22, and also during several glacial periods before 900 ka (Fig. 6). The glacial $\delta^{18}O$ difference between the Pacific and Atlantic is on average ~1.5‰. The salinity contrast (3.3 psu) during the cold periods must have been larger than the salinity contrast today, taking the average 1.5‰ difference in $\delta^{18}O$ and assuming a similar temperature regime in both oceans. This would imply increased water vapor loss from the Atlantic, making the waters around the Great Bahama Bank extremely salty. The salinity is difficult to estab-

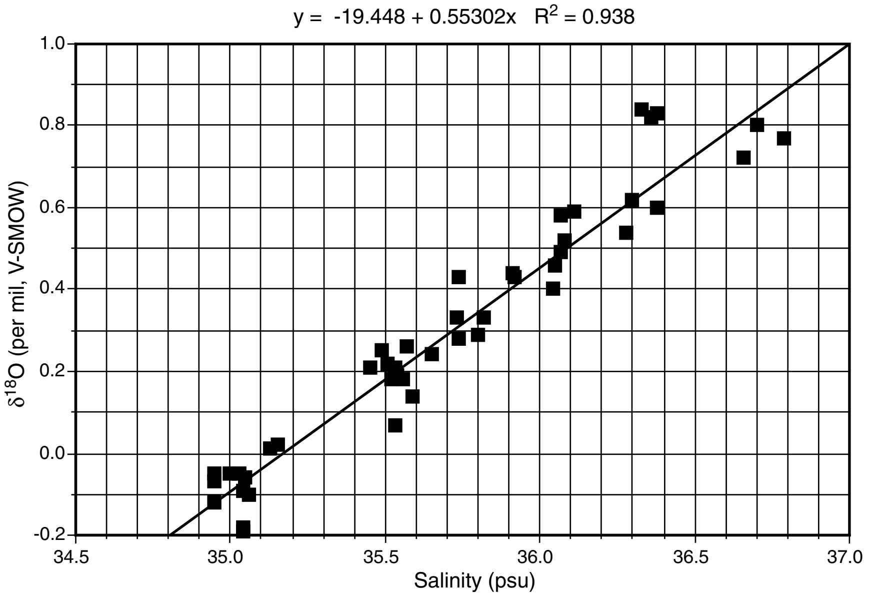

Figure 5. Plot of $\delta^{18}O$ vs. salinity in modern Atlantic surface waters (after Ganssen and Kroon, in press).

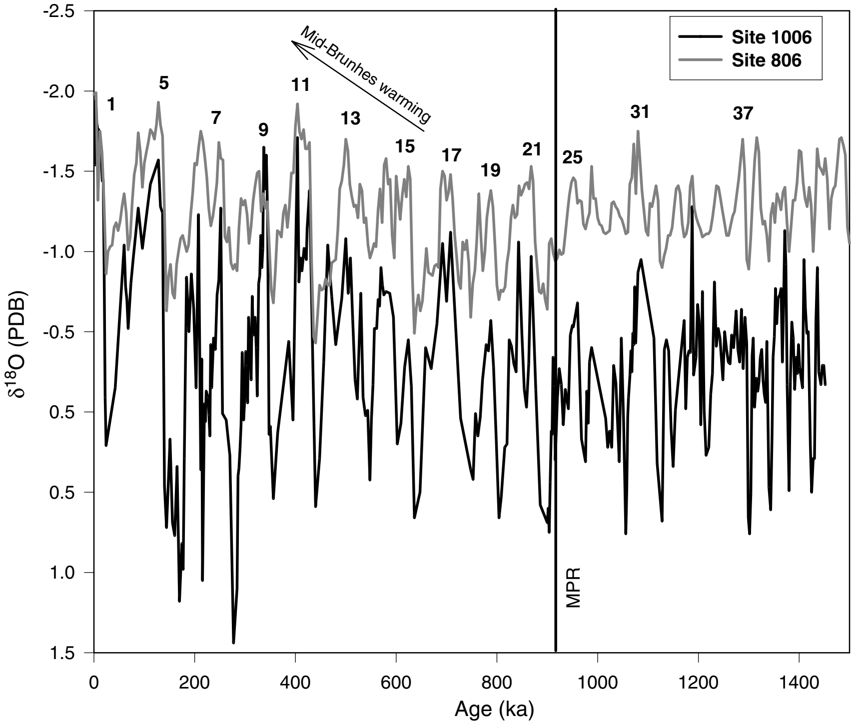

Figure 6. The $\delta^{18}O$ records of Site 1006 and western Pacific Site 806 are plotted against time for a direct comparison of the low-latitude Atlantic and Pacific sites. The oxygen isotope variations are much larger in Site 1006. The mid-Pleistocene revolution and mid-Brunhes warming are indicated. Mid-Brunhes warming occurred at a time when solar insolation was low (Imbrie and Imbrie, 1980) and another mechanism needs to be called upon to explain it. Maximum return flow of surface waters from the Pacific around the tip of Africa is a possibility.

lish because of potential differential temperature changes in the Atlantic and Pacific, and because Site 806 probably does not record the maximum interglacial/glacial amplitude in $\delta^{18}O$ in the western Pacific.

Differential temperature changes between the Pacific and Atlantic during glacial periods are not unlikely when taking into account potential past changes in global oceanic circulation patterns. Today, warm Pacific surface waters return to the southeastern Atlantic via the low-latitude Indian Ocean and the Agulhas Current. The Benguela system transports the warm waters farther north via the South Equatorial and the North Brazil Currents through the Caribbean into the Gulf Stream. One of the bottlenecks in this general oceanic circulation system is the transport of warm waters around the Cape of South Africa before the warmer waters can be transported northward. Past northward movements of the subtropical convergence may have prevented or reduced the influence of Agulhas waters coming around, thus reducing heat delivery to the north. Reduction of transport of warm water around the Cape would have happened most certainly during glacial periods. A reduction in warm-water planktonic foraminiferal abundances during the last four major glacial periods indicates cooler temperatures around the Cape (Imbrie et al., 1973; Berger et al., 1985; Little, 1997) during maximum northerly expansion of the Antarctic Polar Front. Another cooling mechanism of the surface waters in the southern Atlantic relates to the Benguelan upwelling system. The southeastern trade wind system was stronger during the cooler periods forcing increased upwelling along the southwestern coast of Africa which made the Benguelan surface waters cooler (Little et al., 1997). The glacial surface waters in the Benguelan system would have moved northward, eventually affecting the warm pool waters of the Caribbean through cross-equatorial flow.

A consequence of cross-equatorial flow of glacial cool water is that glacial northward heat transport was reduced when a mixture of colder waters of subantarctic origin flowed from the west through Drake Passage, replacing the warmer waters from the east. Cross-equatorial flow of this relatively cool water would lead to heavier $\delta^{18}O$ values in planktonic foraminiferal shells in the western equatorial Atlantic Ocean. However, the relatively low salinity values of southern origin would make the $\delta^{18}O$ values lighter. Assuming that generally cooler waters crossed the equatorial Atlantic, the temperature difference between the Atlantic and Pacific Oceans would increase and the salinity difference would decrease because the present low-latitude Atlantic is saltier than the Pacific; therefore, at least a large part of the glacial $\delta^{18}O$ difference between the Atlantic and Pacific could have been controlled by temperature. Temperature is the more likely factor in the glacial $\delta^{18}O$ difference between the Atlantic and Pacific because sufficiently increased salinity in the Gulf Stream area would lead to a strong Atlantic conveyor circulation. In contrast, glacial deep-sea ventilation in the Atlantic was reduced (Boyle and Keigwin, 1982; Curry and Lohmann, 1983; Curry et al. 1988; Sarnthein et al., 1994).

One of the implications of changes in global circulation at the glacial/interglacial time scale is that the interglacial/glacial temperature range in the low-latitude Atlantic was larger than in the Pacific. The coupled oxygen isotope records of Sites 1006 and 806 would suggest a larger temperature range: Site 1006 shows a much larger isotope variability than does Site 806. Martinez et al. (1997), however, documented that the fluctuations in $\delta^{18}O$ on the interglacial/glacial time scale are much larger at other sites in the western Pacific, which would indicate that the oxygen isotope records are much more dependent on regional evaporation-precipitation cycles and temperature changes. It is, therefore, impossible to know how valid an Atlantic-Pacific $\delta^{18}O$ comparison based on Sites 1006 and 806 is. Another problem is that bioturbation may have affected the records depending on sedimentation rates. We conclude that if global circulation changes have modified the oxygen isotope record of Site 1006, temperature changes largely explain the isotopic variability.

The interglacial $\delta^{18}O$ values of Site 1006 are mainly more positive than at Site 806. The interglacial $\delta^{18}O$ values should be heavier in the low-latitude Atlantic than in the Pacific, assuming that a salinity contrast similar to that of the present day existed during all interglacial periods. There is no temperature difference between the two tropical oceans during the summer, although a temperature contrast may exist during winter as the Bahamian area cools ~5°C and the Ontong Java Plateau area does not. *G. ruber* mainly calcifies its shell during the summer (Cifelli and Smith, 1974), although Deuser (1986) shows that *G. ruber* is present year round in subtropical areas. A combination of slightly cooler temperatures and higher salinity explains why the Site 1006 oxygen isotope curve is heavier in general than the Site 806 profile. In addition, the $\delta^{18}O$ record of Site 1006 was generated using *G. ruber,* while the $\delta^{18}O$ profile of Site 806 used *G. sacculifer. G. ruber* tends to generate slightly lighter $\delta^{18}O$ values than *G. sacculifer,* which may further explain why the Site 806 curve shows lighter $\delta^{18}O$ values than the Site 1006 curve. This general feature can be seen in all interglacial periods except for the Holocene and Stage 9. We don't know why the $\delta^{18}O$ difference between *G. ruber* (Site 1006) and *G. sacculifer* (Site 806) disappears during the Holocene and Stage 9 interglacial periods.

The $\delta^{18}O$ difference prior to Stage 11 was in general larger (on average 0.7‰) during the sea-level highstands. A larger salinity contrast by increased vapor export from the Atlantic to the Pacific may possibly explain this, or an increased temperature difference. A temperature difference seems to be the more likely scenario in light of the above discussion. Unfortunately, there are no paleotemperature data off South Africa prior to Stage 11 available to determine whether somewhat cooler waters came around the Cape during the sea-level highstands before Stage 11. The oxygen isotope record of Site 1082A, drilled in the Benguelan system, recently produced by Jahn et al. (1999) shows trends very similar to the Site 1006 isotope record with lighter $\delta^{18}O$ values during the interglacial periods since isotopic Stage 11. The similarity between the two records shows that global circulation patterns are involved in modifying the temperature and salinity regimes that influence the oxygen isotope records. The $\delta^{18}O$ record of Site 806 also shows a slight trend toward lower values in isotopic Stage 11 (Fig. 6). Another western Pacific example comes from the Great Barrier Reef area where Sites 820 and 823 were drilled. Both records show a $\delta^{18}O$ decrease during isotopic Stage 11 (Peerdeman et al., 1993; Alexander, 1996). The similarity between the Pacific and Atlantic stable isotope records of a $\delta^{18}O$ decrease since isotopic Stage 11 suggests that temperature increased in both areas, possibly as a result of strengthening of the warm pool of surface ocean water in both areas.

The $\delta^{18}O$ records of benthic foraminifers and also the SPECMAP curve show slightly heavier values for the pre-Stage 11 highstands, implying slightly lower sea levels and perhaps a slightly larger Antarctic ice sheet during these periods than today (Kuijpers, 1989). Although it is difficult to imagine why the Antarctic ice sheet may have been slightly larger, there is ample evidence that subantarctic waters were cooler before Stage 11 (Keany and Kennett, 1972). Cold subantarctic waters extended farther northward and displaced the Antarctic Polar Front to the north, possibly reducing return flow of warmer waters from the Indian Ocean. Jansen et al. (1986) postulated that the climatic belts were displaced northward by ~5° latitude, which would have hindered the intrusion of warm waters into the Atlantic. Therefore, in this context, we interpret the Stage 11 $\delta^{18}O$ decrease as a warming and not as a result of reduced salinity. The pre-Stage 11 interglacial $\delta^{18}O$ values in Site 1006 are on average ~0.6‰ heavier than post-Stage 11, which would imply ~3°C warming.

Late Pleistocene warming during the mid-Brunhes Chron seems to coincide with a period when the earth's climatic system was mainly dominated by 100-k.y. cycles, although warming of surface waters as recorded by a decrease in stable oxygen isotope values in sediments from isotopic Stage 11 occurred much later than the onset of 100-k.y. cycles as defined by the mid-Pleistocene revolution at ~1 Ma (Berger et al., 1993). It is not known which phenomenon triggered the increase in amplitudinal change in climate around isotopic Stage 11 characterized by very cold glacial and very warm interglacial stages. Direct solar insolation was minimal at this time because of weakened precessional cycles when eccentricity was at a low (Imbrie and Imbrie, 1980). Jansen et al. (1986) speculated that the mid-Brunhes changes in temperature were triggered by the 400-k.y. cycle, but other possibilities such as ocean circulation changes cannot be excluded. It is simply not known which feedback mechanisms controlled the increase in global climate change since isotopic Stage 11. We speculate that increased return flow of surface waters from the Pacific to the Atlantic may have been involved, or perhaps partly involved, as one of the mechanisms because the Sites 1006 and 806 oxygen isotope curves converge during the sea-level highstands, pointing to a more homogeneous world ocean with similar surface conditions in the low latitudes.

This trend seen in the $\delta^{18}O$ record of amplification of climate change during the mid-Brunhes (younger than isotopic Stage 12) is also developed in the carbonate mineralogy and grain-size distribution profiles at various other Leg 166 sites. There is a marked increase in the average concentrations of aragonite (from 66 to 71 wt%) and in HMC (from 2 to 4 wt%), while LMC decreases at Site 1003 (from 29 to 15 wt%; Rendle et al., Chap. 6, this volume). These values are calculated from the extreme values within each stage for both sedimentary units. Quartz shows continuous representation until the Stage 11 transition; above that its occurrence is sporadic. This change is probably related to a change in the ocean circulation pattern, with a reduced input of southern material (Lidz and McNeill, 1998). This change accompanies a general coarsening of the sediments at Site

1003 (Rendle et al., Chap. 6, this volume). Regional changes in circulation patterns may be responsible for this, and thus may have contributed to the mid-Brunhes oxygen isotope changes.

We conclude that the Site 1006 isotopic variability can at least in part be explained by global circulation changes affecting the temperature regime in the western equatorial Atlantic, although regional changes in circulation may have affected the area as well. The glacial/interglacial amplitude is large throughout the record and is proposed to be due to temperature variability. The interglacial periods since isotopic Stage 11 show the lightest $\delta^{18}O$ values, which we interpret as mid-Brunhes surface-water warming. This trend is contemporaneous with changes in the mineralogy and grain-size distribution as described by Rendle et al (Chap. 6, this volume), suggesting that regional climate changes and ocean circulation changes could have altered the stable isotope record of Site 1006 as well.

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