Climate Feedback and Pleistocene Variations in the Atlantic South Equatorial Current

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Abstract: Ice-age cooling of the central equatorial Atlantic Ocean reflects both equatorial upwelling and advection of cool water off the southern-hemisphere eastern boundary, but the largest contribution appears to be advection. This conclusion is based on planktonic foraminiferal assemblages that vary over the last ~300 ky in the tropical Atlantic and Pacific Oceans. Core-top maps of these assemblages reveal relationships to 1) the tropical-subtropical warm pool, 2) equatorial and coastal upwelling, and 3) eastern boundary current advection. The faunal indices are not sensitive to selective dissolution, as they do not correlate with measures of preservation such as calcite fragmentation or water depth. The sequence of changes going into a glacial interval is 1) initial cooling associated with equatorial upwelling driven by trade winds, followed by 2) amplified cooling by intensified meridional winds along the SE Atlantic margin which advected cool Benguela Current water to the central equatorial Atlantic. The Gulf of Guinea maintained its present cyclonic gyre circulation in glacial time, bounded by a front similar to the present Angola-Benguela Front. We prefer a model in which thermal gradients in the southern hemisphere drive changes in the South Equatorial Currents.

Introduction

What caused large ice-age cooling in the South Equatorial Currents? These currents, relatively cool in their eastern reaches in both the Atlantic and Pacific (Fig. 1), are highly sensitive to change on all time frames. Examples of this sensitivity include large oscillations of temperature over annual cycles and inter-annual El Niño - Southern Oscillation phenomena (Philander 1979, 1986). Over longer time intervals, significant cooling occurred in both oceans associated with the peak of the last ice age (CLIMAP 1976, 1981). The South Equatorial Currents, which include upwelling zones near the equator, are major sources of CO₂ to the atmosphere (Tans et al. 1990) and thus their long-term variations are an important element in global climate feedback associated with greenhouse forcing (Jasper et al. 1994).

Although many features of the South Equatorial Currents may change, including current strength, upwelling, thermocline depth, and biological productivity, many studies describe changes in terms of sea-surface temperature. The exact range of glacial-to interglacial temperature contrast in the equatorial oceans, and the cause of the changes, remains controversial. The first detailed map of ice-age sea-surface temperatures, based on microfossil percentages (CLIMAP 1976, 1981) suggested that the South Equatorial Current was cooler than today in the cool upwelling season (August) by 2-4°C in the central Atlantic (or about a 1-3°C annual average change). Molino et al. (1982) argued that in most cases temperature estimates are robust because different faunal and floral groups with unique ecological preferences and different preservation biases give similar results. An exception to this was off SW Africa, and in the equatorial zone, where temperature estimates from different fossil groups were somewhat discordant.

A controversy about tropical temperatures arose when Rind and Peteet (1985) suggested that tropical and subtropical sea-surface temperatures must have been much cooler than the CLIMAP estimates.
Fig. 1. Modern annual mean sea-surface temperatures (from Levitus 1982) illustrate relatively cool equatorial temperatures associated with equatorial upwelling and advection of cool water off the eastern boundaries. Core locations are shown, with small dots representing sites of down-core foraminiferal species data (Table 1) used to define faunal factors, and the star representing core V30-40, which is analyzed in detail here.

at the last glacial maximum, based on an atmospheric climate model and evidence for tropical mountain glaciation. Guilderson et al. (1994) supported this assertion at subtropical locations, using the Sr/Ca thermometer in corals. Other paleothermometers, however, $^{18}$O, SiKs and Keigwin 1994; and oxygen isotopes, Broecker 1986; Mix 1992) support the lower contrast of glacial and interglacial temperature patterns in the low latitudes similar to the CLIMAP reconstructions.

Using methods similar to CLIMAP but a more detailed foraminiferal time series data set, Mix et al. (1986a,b) inferred an annual-average difference of about 4°C between glacial and interglacial extremes in the central equatorial Atlantic (up to 6°C in the upwelling season; Mix 1986; McIntyre et al. 1989). This large change was confined to the central equatorial zone; the temperature changes in the western equatorial Atlantic and the subtropics were as small as the original CLIMAP estimate, about 1-2°C. Mix (1989) suggested that some of the faunal and floral changes in the equatorial systems may reflect changing productivity, as well as temperature. The importance of thermocline depth rather than absolute temperature was noted by Ravelo et al. (1990). Oberhansli et al. (1992) emphasize advection, food, and oxygenation of water as key elements controlling standing stocks of modern foraminifera.

There is as yet no consensus on the true amplitude of tropical sea surface temperature change. In this paper, we focus instead on the oceanographic mechanisms for changing equatorial faunas, using the foraminifera as watermass tracers. If we can identify from the fauna the primary mechanism that causes equatorial cooling, we gain insight into an important climatic feedback, possibly with global consequences related to inter-hemispheric heat transport and the global greenhouse effect.
Modern Oceanographic Summary

Variations in the South Equatorial Currents are driven by, and in turn modify, the low-latitude winds. Sea-surface temperatures are warmest north of the equator, especially in the Western Atlantic (Fig 1). This warm pool is associated with the position of the inter-tropical convergence zone (ITCZ) in the northern hemisphere. The ITCZ approaches the equator in the eastern Atlantic during northern hemisphere winter, when sea-surface temperatures are warmest in the Gulf of Guinea. It occasionally enters the southern hemisphere, in years when the Atlantic SEC is anomalously warm (Philander 1986). Upwelling along the equator responds to the zonal component of the southern trade winds via Ekman divergence. If this zonal component of the winds changes, one would expect to see corresponding changes in equatorial upwelling.

In the eastern equatorial Atlantic the winds are on average from the south (Chelton et al. 1990; Picaut et al. 1985). These southerly winds, which extend as far south as 5-10°S and cover nearly half the width of the Atlantic Basin, are in part a monsoonal effect driven by strong summer heating of the large African land mass north of the equator, but they also reflect the position of the subtropical high. Thus, cool temperatures south of the equator in the Gulf of Guinea reflect a combination of cool eastern-boundary waters advected westward by monsoonal wind forcing (Cane 1979; Philander 1979), and wave-like movements in the thermocline related to seasonal winds in the western Atlantic (Moore et al. 1978; Philander and Pacanowski 1986).

The cool eastern boundary waters of the Benguela Current are presently kept out of the eastern equatorial Atlantic by the Angola-Benguela front (Shannon et al. 1987), the southern boundary of a weak cyclonic gyre in the eastern tropical Atlantic (Gordon and Bosley 1991). Surface currents are dominated by northward and westward flowing Ekman drift (Arnault 1987), but the geostrophic component of currents flows eastward as a subsurface South Equatorial Counter Current (Molinari 1982), and southward in the Angola Current near the African coast (Arnault 1987; Peterson and Stramma 1991). The northward and westward Ekman transport at the surface is small compared to the subsurface eastward advection of relatively warm, salty, subtropical water near the equator. Following the zone of zero wind-stress curl, the cool Benguela Current core leaves the eastern boundary near 15-20°S, and although it advects westward and northward, it is not strong enough at present to cause extreme cooling in the central equatorial Atlantic, where annual average temperatures are > 26°C at present (Fig. 1).

Hypotheses to Explain Ice-age Variations

A vast range of different opinions exists about the processes driving Pleistocene cooling of the SEC in the Atlantic Ocean, and at present there is no consensus. The options for the dominant process are 1) stronger equatorial upwelling related to zonal (Trade) winds, driven by thermal gradients in the southern hemisphere, 2) stronger upwelling along the tropical eastern boundary, related to southerly winds of the North African Monsoon, or 3) influence of cool Benguela Current water, advected to the equator, linked to meridional winds along Southwest Africa. We assess fossil data from the equatorial Atlantic in light of these hypotheses. Cane (1979), and Philander and Pacanowski (1980, 1981) illustrate with ocean models the response of a hypothetical ocean to the first two hypotheses. In both models, zonal (Trade) winds produce circulation symmetrical around the equator, with a shallow thermocline, implying cool temperatures, in the east and along the equator. Southerly (monsoonal) winds drive an asymmetrical circulation with cool water advected off the eastern boundary in the southern hemisphere and along the equator. Although these responses are quite different in their geographic details, with just a few study sites and only sea-surface temperature as a measure, it might be difficult to tell them apart. A better test is to trace the water masses separately, as we do here.

The Trade Wind Effect

Trade wind velocities respond to hemispheric thermal (and thus pressure) gradients. Krauss (1977) and Nicholson and Flohn (1981) argue that expanded glaciation of the high southern latitudes
during the last ice age drove changes in the velocity of the southern trades. This idea was supported by the CLIMAP (1976, 1981) reconstructions, and by time-series studies of sea surface temperatures that associated cool events in the tropics with glacial maxima (Gardner and Hays 1976; Mix et al. 1986a). A problem with this inference was the reconstruction of enhanced seasonal contrast in the glacial tropical Atlantic (Mix et al. 1986b), opposite the reduced seasonal cycle in the southern oceans (Hays 1978). Atmospheric model results do not support a link to continental glaciation. Glaciers do not produce strong changes in the position or intensity of southern-hemisphere wind systems (Manabe and Broccoli 1985). Continental ice sheets alone apparently can not do the job, because the changes in Antarctica were relatively small.

Mix et al. (1986a) argued that a likely causal mechanism for stronger southern-hemisphere winds was greenhouse cooling associated with globally-reduced atmospheric pCO$_2$ of the last glaciation (Jouzel et al. 1993). Using an atmospheric general circulation model coupled to a mixed-layer ocean model, Broccoli and Manabe (1987) tested this idea. They did indeed find a significant cooling effect around Antarctica, and an increase in southern-hemisphere thermal gradients in the atmosphere due to lower pCO$_2$. The feedback mechanisms driving this gradient in the model were 1) expanded sea ice in the southern oceans (which amplified high-latitude cooling by reducing thermal inertia of the sea surface), and 2) convection of the tropical atmosphere (which reduced low-latitude cooling by spreading it through the troposphere). The response of sea-surface temperature in the tropics was small in the model, but because the model ocean did not simulate advection or upwelling, the tropical oceans' responses remain untested.

The Monsoon Effect

Another option for modifying the tropical Atlantic is the African monsoon. Solar heating of the large African land mass north of the equator generates relatively low atmospheric pressures. The contrast of warm land with cool waters (and thus relatively high atmospheric pressures) to the south draws winds across the equator. Changes in the temperatuure of NW Africa would modulate southerly winds in the Gulf of Guinea, and thus advection of cool water off the eastern boundary in the low latitudes of the southern hemisphere.

The African monsoon did vary on glacial-interglacial time frames, based on the occurrence of organic-rich sapropel layers in the Mediterranean Sea (Rossignol-Strick 1983) and wind-blown continental lake deposits in the deep Atlantic (Pokras and Mix 1986; DeMenocal et al. 1995). This process may be partially decoupled from the effect of southern hemisphere thermal gradients on easterly trade winds, if it is driven from the north rather than the south. High early-Holocene lake levels in tropical Africa suggest that northern-hemisphere summer insolation modulates the monsoon (Kutzbach and Street-Perrott 1985; Street-Perrott and Perrott 1993). A low-resolution climate model (Prell and Kutzbach 1987) also supported a significant role for northern-hemisphere insolation, although a higher-resolution model suggested that the insolation effect was small (DeMenocal and Rind 1993). Street-Perrott and Perrott (1990) noted that warmth of the North Atlantic may also have had a strong effect on the monsoon, and this effect was confirmed in a model sensitivity test by DeMenocal and Rind (1993).

Mix et al. (1986b) argued that the African monsoon was weakened during glacial time, and suggested that glacial cooling and an enhanced seasonal cycle in the eastern equatorial Atlantic was due to stronger trade winds and equatorial upwelling linked to reduced monsoonal advection. McIntyre et al. (1989) supported this idea of opposite responses of monsoon winds and trade winds, but argued that three mechanisms (monsoonal winds, equatorial upwelling, and advective heat transport) all contribute to the equatorial Atlantic response. They noted (as did Molfine et al. 1982) increased ice age abundance of a transitional and subpolar fauna in the equatorial Atlantic. More recently, Molfine and McIntyre (1990) analyzed pycnocline depth history using abundance of Florisphaera profunda, a species of coccolithophore, and emphasized the role of equatorial upwelling.

Unresolved is the relationship between southerly monsoon winds and trade winds. Maximum upwelling (minimum $F. profunda$ of Molfine and
McIntyre 1990) does not correlate precisely with minimum monsoon index in the same samples (maximum eolian flux of freshwater diatom Melosira sp., from Pokras and Mix 1985). With the linkage hypothesized by Mix et al. (1986a,b) and McIntyre et al. (1989), one might expect the fauna or flora associated with equatorial upwelling to vary antithetically with that associated with the eastern boundary currents.

**The Benguela Effect**

A third mechanism proposed to cool the SEC is to extend the cool Benguela Current into the equatorial zone (Gardner and Hays 1976). The northward advection of eastern boundary water is related to the position and strength of the subtropical high pressure zone. At present, the Benguela Current leaves the African coast near 20°S, near the northern limit of strongest southerly winds along Namibia, and advects westward well south of the SEC. Gardner and Hays (1976) did not specify how Benguela water would have reached the equatorial Atlantic during glacial time, but implied in a schematic drawing a northward extension of the Benguela current to the eastern-basin source of the SEC, between 5° and 15°S. This would imply northward or eastward displacement of the high-pressure zone, and perhaps elimination of the Angola-Benguela Front. It is also possible that Benguela Current water could leave the coast near its present position, and advect to the central equatorial Atlantic, west of a geostrophic cyclonic gyre that presently occupies the eastern tropical South Atlantic.

**Methods**

To reconstruct past oceanographic conditions, common practice has been to calibrate fossil plankton assemblages in core tops using the modern environment, and to apply these calibrations to faunas from the ancient geologic record. The usual approach is to simplify the core-top fauna into orthogonal Q-mode factors rotated with a varimax criterion (Klovan and Imbrie 1971). Q-mode factor analysis resolves each fossil assemblage into a proportional combination of a smaller number of independent end-members, or factors. Each factor is in turn a linear combination of input species. These assemblages, which generally form oceanographically meaningful map patterns in core tops, are assumed to be ecologically meaningful groupings. The result is a table of factor scores (that define the assemblages as weighted combinations of species), factor loadings (the weighting of each factor for each sample), and communalities (the sum of squares of loadings, representing the fraction of variance in each sample described by the factor model). The next step is often to generate a environmental transfer function which predicts a useful property, such as sea-surface temperature, by regressing modern faunal factor loadings against modern oceanographic properties (Imbrie and Kipp 1971). Finally, the modern factor definitions can be applied to ancient samples to find the relative importance of each factor for samples not included in the calibration. As long as significant evolution or some other process has not modified the ecological groupings, the transfer function can be used to make paleoenvironmental estimates of useful oceanographic indices.

A problem with this approach is that ice-age samples from the eastern equatorial Pacific and Atlantic are not well represented in modern core top assemblages (Moore et al. 1981). This is because the ice-age variation in climates from these areas is larger than the modern range within the tropics. For example, Ravelo et al. (1990) show that the faunal factors defined using only tropical Atlantic core tops explain just 35% of the population variance in similar samples from the last glacial maximum. This is called a "no-analog" condition, which can lead to erroneous estimates of past oceanographic conditions (Hutson 1977).

Here, we improve on this situation by calculating the faunal assemblages using Q-mode factor analysis of down-core samples from the equatorial Atlantic and Pacific Oceans over the past ~300 kyr, rather than from core tops. This guarantees that the assemblages are defined based on species that covary in the past, over a suitably large range of variations. An advantage is that the local down-core faunal assemblages record the regional context of tropical oceanographic processes. Their definitions are not biased by high-latitude core-top faunas that
never existed within the tropics. By including Pacific samples with Atlantic samples in these factor definitions, we attempt to decrease the sensitivity of the faunal assemblages to selective dissolution. This works, because glacial cooling in the Atlantic and Pacific Oceans is similar but dissolution is in the opposite sense. We do not attempt to calibrate these assemblages to yield estimates of sea-surface temperature or any other index, but instead use the assemblages directly as tracers of water masses.

The samples used here for down-core faunal analyses (Table 1) come from four cores in the tropical Atlantic (1057 samples) and six cores in the eastern tropical Pacific (715 samples). Locations are illustrated in Fig. 1. Core-top samples come from the database assembled by Prell (1985) with the addition of Atlantic core tops from Mix (1986) and Pacific core-top samples from Coulbourn (1980), Sverdlove (1983), and Mix (unpublished data). The Pacific results will be discussed in more detail elsewhere (A. Mix, in prep.).

In addition to our definition of factors from down-core samples, two features of our factor analysis differ from common practice. We excluded the dissolution resistant species *G. tumida, G. menardii*, and *G. menardii neoflexuosa*. These species were absent from the Atlantic for long periods of the late Pleistocene (Ericson 1968). They repopulated the Atlantic about 11,000 (14C) years ago (Berger et al. 1985) perhaps when they were advected around South Africa during an episode of anomalous warmth in the southern ocean. This migration effect would introduce bias into local climate reconstructions that included these species as part of the index. Second, our analyses are based on ln(species percentage +1) rather than just percentages. This logarithmic transform amplifies the importance of less abundant species, and minimizes the dominance of a few species. One is added to

### Table 1:
**SOURCE OF 1772 DOWN-CORE PLANKTonic FORAMINIFERAL SPECIES COUNTS TO DEFINE FACTORS**

#### ATLANTIC:

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth</th>
<th>Samples</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>RC24-16</td>
<td>5°02'S</td>
<td>10°12'W</td>
<td>3543 m</td>
<td>255</td>
<td>McIntyre et al. (1989)</td>
</tr>
<tr>
<td>V22-174</td>
<td>10°04'S</td>
<td>12°49'W</td>
<td>2639 m</td>
<td>333</td>
<td>Imbrie et al. (1989)</td>
</tr>
<tr>
<td>V30-36</td>
<td>5°21'N</td>
<td>27°19'W</td>
<td>4245 m</td>
<td>217</td>
<td>Mix (1989)</td>
</tr>
<tr>
<td>V30-40</td>
<td>0°12'N</td>
<td>23°09'W</td>
<td>3706 m</td>
<td>252</td>
<td>McIntyre et al. (1989)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1057 TOTAL ATLANTIC</td>
</tr>
</tbody>
</table>

#### PACIFIC:

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth</th>
<th>Samples</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>ODP-846</td>
<td>3°06'S</td>
<td>90°49'W</td>
<td>3307 m</td>
<td>239</td>
<td>Le et al. (in press)</td>
</tr>
<tr>
<td>RC10-62</td>
<td>3°20'N</td>
<td>101°43'W</td>
<td>3120 m</td>
<td>60</td>
<td>Mix (in prep)</td>
</tr>
<tr>
<td>RC13-110</td>
<td>0°06'N</td>
<td>95°39'W</td>
<td>3231 m</td>
<td>180</td>
<td>Mix (in prep)</td>
</tr>
<tr>
<td>V19-29</td>
<td>5°35'S</td>
<td>83°56'W</td>
<td>3157 m</td>
<td>89</td>
<td>Sverdlove (1983)</td>
</tr>
<tr>
<td>Y71-9-101P</td>
<td>6°23'S</td>
<td>106°56'W</td>
<td>3175 m</td>
<td>104</td>
<td>Mix (in prep)</td>
</tr>
<tr>
<td>Y71-6-12P</td>
<td>16°26'S</td>
<td>77°34'W</td>
<td>2734 m</td>
<td>43</td>
<td>Mix (in prep)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>715 TOTAL PACIFIC</td>
</tr>
</tbody>
</table>
each percentage value to avoid the log of zero. The log transform is applied to all species percentage data (down-core and core top) prior to calculation of factor loadings or scores. As noted by May (1975) regarding other biological systems, this makes relative abundances of species more Gaussian, which is an underlying assumption in the statistical calculations.

We assume closure (100%) around 26 species and morphotypes, listed in Table 2. Only species without significant taxonomic controversies are included. We included specimens referred to by Kipp (1976) as "P-D" intergrade with N. dutertrei, as this is how much of the Pacific core-top data set was counted by several investigators. Also, G. theyeri is grouped with G. scitula, because the former species was not widely recognized in earlier core-top studies. Morphotypes of G. sacculifer with and without a terminal chamber are grouped together, as are pink and white varieties of G. ruber.

To gain insight into the meaning of down-core faunal factors, we apply them to core-top samples. No-analog conditions are apparent as low communalities in core top samples. Because they occur in the modern samples rather than in the down-core samples, these poor analogs can be understood more clearly and excluded from the environmental calibrations. In all cases, our down-core samples are well described by our down-core faunal factors. Communalities in the down-core samples analyzed here are always >0.70, and average 0.90. Rather than explaining 35% of the population variance in down-core samples from the equatorial Atlantic and Pacific Oceans, as was the case for the application of core-top factors to down-core samples, they explain 90%. Thus, variations of our down-core factors are much more representative of oceanographic changes in the tropical water masses than were previous estimates based on core-top factors.

One core from the equatorial Atlantic (V30-40) is analyzed in detail. This site is on the equator and more than 1000 km from the eastern boundary. The Atlantic data were reported earlier by Mix et al. (1986a,b) and McIntyre et al. (1989). To express down-core variations as a function of time, age models developed elsewhere based on oxygen isotope stratigraphy (Mix 1992; Imbrie et al. 1984) were applied to the faunal data.

Cross-spectral analysis techniques assess the significance relationships of various time series generated in this study (Jenkins and Watts 1968). The time series were first interpolated at 3-ka intervals with a 9-ka Gaussian smoothing window. The spectra and cross spectra were calculated on a time series from 0-270 ka, with a bandwidth of 0.011 ka-1. The phase ranges and coherence significance were calculated at the 80% confidence level. For the purposes of calculating phase, we follow the SPECMAP conventions of Imbrie et al. (1989), in which the sign of all faunal indices is changed such that higher positive values would correspond to warmer temperatures or full interglacial conditions. For example, because lower values of δ18O are associated with interglacial episodes, the oxygen isotope index is multiplied by negative one.

Results

Down-Core Faunal Factors and Core-Top Patterns

Q-mode factor analysis of the down-core faunas in the 10 sites used as input here resolves three significant faunal factors, which combine to explain 90% of the pooled population variance (Table 2). Core-top studies of foraminifera over a broad range of latitudes typically resolve five or more factors (Imbrie and Kipp 1971; Kipp 1976). Our results indicate, as did Ravelo et al. (1990), that just three factors are sufficient to describe the long-term variations of planktonic foraminifera in the equatorial current systems. A four-factor solution, which we reject, adds less than 2% additional information, and splits the first factor into two groups that are not ecologically meaningful.

We apply the factor scores from Table 2 to the core-top foraminiferal species percentages to assess the modern distributions of the assemblages. This yields three maps of loadings (strength) of these factors (Fig. 2). When mapping these factor loadings, we exclude any core-top samples with factor communalities less than 0.5, as these are not well described by the fauna present in the equatorial oceans. In nearly all cases, the low-commu-
Table 2: Q-mode Varimax Factor Scores based on 1772 down-core faunal counts in 10 cores from the tropical Atlantic and Pacific Oceans, calculated from species data expressed as ln(%) + 1. Species are grouped by their dominant factor, and scores ≥0.25 are noted in bold type. Larger numbers indicate greater weighting in the factor index.

<table>
<thead>
<tr>
<th>SPECIES</th>
<th>FACTOR 1 WARM TROPICAL</th>
<th>FACTOR 2 UPWELLING</th>
<th>FACTOR 3 E. BOUNDARY</th>
</tr>
</thead>
<tbody>
<tr>
<td>G. ruber</td>
<td>0.564</td>
<td>-0.025</td>
<td>0.086</td>
</tr>
<tr>
<td>G. sacculifer</td>
<td>0.471</td>
<td>-0.094</td>
<td>0.129</td>
</tr>
<tr>
<td>G. glutinata</td>
<td>0.380</td>
<td>0.087</td>
<td>-0.170</td>
</tr>
<tr>
<td>G. aequilateralis</td>
<td>0.224</td>
<td>-0.055</td>
<td>0.013</td>
</tr>
<tr>
<td>P. obliquiloculata</td>
<td>0.226</td>
<td>-0.024</td>
<td>-0.080</td>
</tr>
<tr>
<td>G. rubescens</td>
<td>0.203</td>
<td>-0.070</td>
<td>-0.153</td>
</tr>
<tr>
<td>G. tenellus</td>
<td>0.182</td>
<td>-0.060</td>
<td>-0.135</td>
</tr>
<tr>
<td>G. falconensis</td>
<td>0.143</td>
<td>-0.050</td>
<td>-0.029</td>
</tr>
<tr>
<td>G. congobatus</td>
<td>0.101</td>
<td>-0.047</td>
<td>-0.117</td>
</tr>
<tr>
<td>G. calida</td>
<td>0.105</td>
<td>0.051</td>
<td>0.063</td>
</tr>
<tr>
<td>G. scitula + G. theyeri</td>
<td>0.091</td>
<td>-0.023</td>
<td>-0.016</td>
</tr>
<tr>
<td>G. digitata</td>
<td>0.030</td>
<td>-0.004</td>
<td>0.016</td>
</tr>
<tr>
<td>S. dehiscens</td>
<td>0.022</td>
<td>-0.002</td>
<td>-0.030</td>
</tr>
<tr>
<td>N. datertrei</td>
<td>0.170</td>
<td>0.774</td>
<td>-0.163</td>
</tr>
<tr>
<td>G. bulloides</td>
<td>-0.011</td>
<td>0.469</td>
<td>0.098</td>
</tr>
<tr>
<td>N. pachyderma (right-coiling)</td>
<td>-0.107</td>
<td>0.340</td>
<td>-0.461</td>
</tr>
<tr>
<td>Q. unversa</td>
<td>0.052</td>
<td>0.102</td>
<td>0.097</td>
</tr>
<tr>
<td>G. hirsuta</td>
<td>0.000</td>
<td>0.004</td>
<td>0.000</td>
</tr>
<tr>
<td>G. inflata</td>
<td>0.060</td>
<td>-0.055</td>
<td>0.675</td>
</tr>
<tr>
<td>G. truncatulinoides (right-coiling)</td>
<td>0.144</td>
<td>-0.102</td>
<td>0.297</td>
</tr>
<tr>
<td>G. crassaformis</td>
<td>0.122</td>
<td>-0.024</td>
<td>0.244</td>
</tr>
<tr>
<td>N. pachyderma (left coiling)</td>
<td>-0.009</td>
<td>0.020</td>
<td>0.087</td>
</tr>
<tr>
<td>G. quinquloba</td>
<td>-0.004</td>
<td>0.013</td>
<td>0.035</td>
</tr>
<tr>
<td>G. truncatulinoides (left-coiling)</td>
<td>0.025</td>
<td>-0.012</td>
<td>0.012</td>
</tr>
<tr>
<td>G. conglomera</td>
<td>0.021</td>
<td>0.040</td>
<td>-0.117</td>
</tr>
<tr>
<td>G. hexagona</td>
<td>0.072</td>
<td>0.025</td>
<td>-0.106</td>
</tr>
</tbody>
</table>

*G. menardii, G. menardii flexuosa, and G. tumida excluded from the percentage calculation and the factor analysis (see text).

such as the NW Pacific and Antarctic, which have little relevance to interpretation of the tropical data. Factor 1 contains seven abundant species, in order of importance G. ruber, G. sacculifer, G. glutinata, G. aequilateralis, P. obliquiloculata, G. rubescens, and right-coiling G. truncatulinoides. All of these species are common in warm tropical and subtropical environments (Kipp 1976; Parker and Berger 1971). In core-top samples this factor (Fig. 2a) is clearly associated with warm, salty subtropical and tropical water masses in the western Atlantic, and warm subtropical water masses in the Pacific where the thermocline is deep. Factor 1 also dominates the Panama Basin, where sea surface temperatures are warm, salinities are low, and the thermocline is shallow (Levitus, 1982). Warm temperature is the variable common to all these locations, so we refer to Factor 1 as the "Warm Tropi-
Fig. 2. Faunal factors applied to core-top samples. Values plotted are factor loadings of all samples with communality >0.50. a) Factor 1, Tropical Assemblage, b) Factor 2, Upwelling Assemblage, c) Factor 3, Eastern-Boundary Assemblage.
cal Assemblage". This is the relatively warm end member with which the other two factors, with cooler aspect, are contrasted.

The second factor is dominated by *N. dutertrei*, with secondary contributions from *G. bulloides* and right-coiling *N. pachyderma*. These species are common in tropical upwelling environments and at high latitudes (Kipp 1976; Parker and Berger 1971; Prell and Curry 1981). The core-top projection of Factor 2 (Fig. 2b) is associated with the equatorial upwelling zone in the Pacific, and is also present in the eastern tropical Atlantic and in the Benguela upwelling system. We infer that this assemblage records the strength of cool upwelling systems with relatively high biological productivity, and refer to it as the "Upwelling Factor". The species making up this factor are strongly associated with equatorial upwelling and high productivity. For example, their abundance in plankton tows that sampled cool upwelling conditions of the 1991 "La Niña" event in the central equatorial Pacific (Watkins et al. 1994) supports our inference based on the geologic record.

In the third factor, *G. inflata* is the most important contributor, followed by right-coiling *N. pachyderma* (which is shared with Factor 2), right-coiling *G. truncatulinoides* (which is shared with Factor 1), and *G. crassaformis*. The high-latitude species left-coiling *N. pachyderma* also is associated with Factor 3. The equatorial species *N. dutertrei* (which is common in Factor 2) has a significantly negative score in Factor 3. This means it is negatively correlated with this factor. Tropical samples which have high loadings for Factor 3 contain both abundant high-latitude species and low percentages of *N. dutertrei*. In the core-top samples, Factor 3 penetrates into the low latitudes in the eastern boundary currents (Fig. 2c). This factor is not strongly associated with productive upwelling centers. When present near the equator (i.e., with positive factor loadings) it appears to reflect advection off the eastern boundary. Thus we refer to it as the "Eastern Boundary Factor".

Species without significant loadings in any of these three factors are *S. dehiscens, G. digitata, G. hirsuta*, left-coiling *G. truncatulinoides*, and *G. quinqueloba*. These species are not abundant in the tropical water masses, and could be ignored in the analysis without changing the results.

**Evaluating Preservation Bias**

A possible problem in interpreting down-core variations in faunal factors is the role of preservation in modifying the fauna. Berger (1970) and Berger et al. (1982) note that fragile, lightly calcified species such as *G. ruber, G. sacculifer*, and *G. glutinata* dissolve more rapidly than robust species such as *G. tumida, P. obliquiloculata*, and *G. truncatulinoides*. If the loadings of the faunal factors change appreciably in response to partial dissolution, it would be impossible to interpret them in terms of surface-ocean climate change.

We attempted to minimize the effects of dissolution in our factors by combining Atlantic and Pacific down-core samples in the analysis. We test for the success of this strategy in two ways. First, we plot in Fig. 3 the loadings of each factor in core tops, as a function of water depth in the Pacific and Atlantic. If dissolution were a problem, we would expect factor loadings to change as a function of increasing water depth (resistant faunal loadings up and fragile faunal loadings down). This is because increasing pressure at greater water depths makes calcite more soluble. In all cases, there is no systematic change in factor loadings as a function of water depth. The loadings of Factor 1 are desensitized to this effect because they contain both resistant (*P. obliquiloculata* and *G. truncatulinoides*) and fragile (*G. ruber, G. sacculifer*) components. At most, only 5% of the variance in core top factor loadings (Factor 1 in the Pacific) is related to water depth, and in the Atlantic the sense of change with depth is reversed. This is insignificant relative to the large down-core variations in the cores studied here. There is no correlation of either the upwelling assemblage (Factor 2) or the eastern boundary assemblage (Factor 3) with depth in either ocean (Fig. 3).

A second test of dissolution artifacts is to compare down-core factor loadings with a dissolution index in the same samples (Fig. 4). We use the percentages of foraminiferal fragments as dissolution index, as it is well known that partial dissol-
ution induces fragmentation (Thunell 1976; Le and Shackleton 1992; Le et al. in press). Of the down-core data examined here, we have fragment data from Pacific cores RC10-62, RC13-110, and Y71-9-101P. All these data were generated in a consistent way in our laboratory at Oregon State University. In Fig. 4, we find no significant correlation of any of the faunal factors or commnuniy with fragmentation in the same samples. The strongest relationship is between the warm tropical Factor 1 and the percentage of fragments, with a correlation coefficient $r = -0.19$ (not significantly different from zero). At most, 4% of the variance in Factor 1 loadings (and a smaller percentage of variance in the loadings of the other faunal factors) is linearly related to fragmentation in down-core samples. Again we find no evidence for significant control of the faunal loadings by dissolution. We conclude that our faunal factors are well buffered against dissolution bias, and that we can use them to infer history of upper ocean circulation.

The Last Glacial Maximum

An important test of the causes of large-scale climate change in the South Equatorial Current is to map the loadings of our faunal factors at the last glacial maximum. On this time horizon there is sufficient spatial coverage of samples that we can clearly see water masses. Fig. 5 illustrates the glacial assemblage patterns, which are best interpreted by comparing them to the core-top patterns.

Loadings of Factor 1, the tropical assemblage (Fig. 5a), are slightly reduced at the glacial maximum relative to at present along the equator in the eastern Atlantic. They change little in the western Atlantic, and actually increase at the glacial maximum in the eastern Gulf of Guinea. A strong gradient to less tropical fauna, along the eastern boundary between 15° and 20°S, demonstrates that the front separating cool Benguela waters from the eastern Gulf of Guinea existed in glacial time. This is consistent with the findings of Jansen et al. (this volume) and Schneider et al. (this volume) in cores near the African margin.

The upwelling assemblage of Factor 2 (Fig. 5b) is slightly stronger at the glacial maximum, and extends farther west along the equator, than at present. Strong concentrations of this factor off NW Africa and SW Africa show that these eastern-boundary upwelling centers were maintained in approximately their present configuration in both glacial and interglacial time.

The changes in the first two faunal factors are significant, but subtle. The largest changes we observe are in Factor 3, the eastern boundary assemblage. At present (Fig. 2c) this fauna is most common at higher latitudes. High loadings are drawn to lower latitudes in the eastern boundary currents, but are presently very low ($< 0.1$) along the equator. At the glacial maximum (Fig. 5c), this fauna is clearly drawn into the South Equatorial Current, with loadings greater than 0.4 in the Central Equatorial Atlantic. Note that loadings in the Gulf of Guinea remain relatively low at the glacial maximum. We infer that cool waters from the Benguela system leave the eastern boundary between 10° and 20°S, and transit the westward edge of a cyclonic gyre in the Gulf of Guinea (south and west of the convergent Angola-Benguela Front). The strong presence of the eastern boundary fauna along the equator in the central Atlantic implicates this water mass as a source of cooling in the tropics. It is possible that the eastern boundary fauna is advected to the equator in the shallow subsurface, and blooms along the equator where a cool productive upwelling system provides it with a suitable food supply. Even if this late-bloom effect is at work, the only way we can imagine this fauna reaching the equator is through much stronger advection of Benguela water in glacial time.

Down-Core Variations

In Fig. 6 we illustrate variations of the down-core species percentages as a function of age in core V30-40, and in Fig. 7 the down-core faunal factors interpolated and smoothed for the time-series analysis, next to the oxygen isotope variations used for stratigraphic control. Lower values of $\delta^{18}O$ (plotted to the right) indicate "interglacial" episodes of low ice volume at high latitudes, and high values indicate glacial events.
Fig. 3. Factor loadings of the tropical, upwelling, and eastern-boundary assemblages in core top samples are plotted separately in Atlantic and Pacific samples, as a function of water depth of the core sites. The lack of significant correlations with water depth indicate lack of sensitivity of these assemblages to dissolution, which increases with depth.
In core V30-40, the upwelling and eastern boundary factors tend to covary. Both are higher during glacial episodes. This modifies the inference of Mix et al. (1986a) that zonal and meridional components of the equatorial winds vary antithetically due to monsoonal "steering". Loadings of the upwelling factor vary between extremes of about 0.3 and 0.6, while the eastern boundary factor varies from 0 to +0.5. The warm tropical factor changes in the opposite sense, with low loadings (down to 0.6) during glacial events and high values (up to 0.9) during interglacial events. This is consistent with past interpretations of cool glacial conditions in the equatorial Atlantic (CLIMAP 1976; Mix et al. 1986a; McIntyre et al. 1989). The new information here is that the greatest variability is in the eastern boundary assemblage (with total variance about four times that of the upwelling assemblage).

Discussion

Environmental change of South Equatorial Current is more complicated than just a change in equatorial upwelling. The relationships between the different oceanographic processes contributing to cooling in the Atlantic are expressed via spectral and cross-spectral analyses (Fig. 8), which compare the faunal changes in V30-40 associated with the upwelling and eastern boundary faunas to ice volume as reflected in the oxygen isotope record.

Fig. 8 shows that both the upwelling and eastern boundary faunas in the equatorial Atlantic are dominated by variance near the 100-ky period of orbital eccentricity, and the 23-ky period of orbital precession, and at both these periods the faunal changes are coherent with ice volume as recorded by δ¹⁸O. We infer from this that both upwelling and
Fig. 5. Faunal factors applied to samples from the last glacial maximum. Values plotted are factor loadings of all samples with communality >0.50. a) Factor 1, Tropical Assemblage, b) Factor 2, Upwelling Assemblage, c) Factor 3, Eastern-Boundary Assemblage. Compare with modern distributions in Fig. 2. Note the large changes in Factor 3, extending to the central equatorial Atlantic, west of an apparent cyclonic gyre in the Gulf of Guinea.
**Fig. 6.** Down-core species percentages vs. age in equatorial Atlantic core V30-40.
advection conspire to cause equatorial Atlantic climate changes. The lack of coherent variations of either the upwelling factor or the eastern boundary factor at the period of orbital tilt, 41 ky, which is clearly present in the oxygen isotope record, supports our inference that carbonate dissolution is not strongly affecting the faunal factors, because dissolution imparts a 41-ky rhythmic signal in calcite preservation indices similar to that of δ¹⁸O in this region (Verardo and McIntyre 1994).

Variations in the eastern boundary fauna are larger than those of the upwelling fauna. This is apparent in Figs. 7 and 8, and is quantified by the coherent amplitudes in the orbital frequency bands, calculated following Imbrie et al. (1989) in Table 3. For all assemblages, the largest amplitude is in the 100-ky period, which dominates the ice age cycles, and the smallest is in the 41-ky period related to orbital tilt. For all frequency bands, the largest amplitude is in the eastern boundary assemblage; about a factor of two larger than those of the tropical or upwelling assemblages. We infer that although both upwelling and advection contribute, the eastern boundary fauna traces the primary cause of equatorial cooling to advection of water from eastern boundary watermasses.

The phase spectra, summarized in Table 4 for the orbital bands, reveal key information about the sequence of processes contributing to climate change. Here, negative phases indicate changes in fauna lead ice-volume changes as recorded by δ¹⁸O. Note that the phases are calculated such that zero
Fig. 8. Cross spectra, upwelling and eastern boundary faunal factors relative to $\delta^{18}O$. In all cases, time series are 270 kyr long, interpolated at 3 kyr intervals. Bandwidth is 0.011 kyr$^{-1}$. a) Log of power spectral density, upwelling factor and $\delta^{18}O$, note that values for the upwelling factor are multiplied by 100 for scaling. b) Coherency of upwelling factor vs. $\delta^{18}O$. c) Phase angle, with negative values indicating a lead of minimum upwelling fauna ahead of full interglacial conditions. d) log of power spectral density, eastern boundary factor and $\delta^{18}O$. Values for the eastern boundary factor are multiplied by 10 for scaling. e) Coherency of eastern boundary factor vs. $\delta^{18}O$. f) Phase angle, with negative values indicating a lead of minimum eastern boundary fauna ahead of full interglacial conditions. Both the upwelling and eastern boundary assemblages have power concentrated near the 100-kyr (orbital eccentricity) and 23-kyr (orbital precession) bands.

Phase indicates alignment of interglacial events with contribution of each index to warming (i.e., high tropical assemblage, low upwelling or eastern boundary assemblages). In the 23-kyr band, all factors lead ice volume by 30-40$^\circ$ (i.e., 2-3 kyr), and the phase error bars do not allow us to separate the different processes. Faunal changes are, however, in phase with atmospheric pCO$_2$ as recorded in the Vostok ice core (using the age model of Jouzel et al. 1993). This may suggest a link between greenhouse cooling and wind forcing in this band. In the 100-kyr band, minimum upwelling leads interglacial conditions by about 20$^\circ$ (i.e., ~6 kyr). The upwelling phase is similar to that of pCO$_2$ (low pCO$_2$, high upwelling), but leads changes in the tropical and eastern boundary assemblages,
TABLE 3: Amplitudes of variation of each faunal assemblage. Values here, in factor loading units, are the portion of amplitude that coherent with δ¹⁸O in each of the main orbital bands. Bandwidth for the integration is 0.09 ky⁻¹.

<table>
<thead>
<tr>
<th>ITEM:</th>
<th>100 kyr</th>
<th>41 kyr</th>
<th>23 kyr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Factor 1: Tropical</td>
<td>.062</td>
<td>.021</td>
<td>.055</td>
</tr>
<tr>
<td>Factor 2: Upwelling</td>
<td>.066</td>
<td>.019</td>
<td>.050</td>
</tr>
<tr>
<td>Factor 3: Eastern Boundary</td>
<td>.066</td>
<td>.031</td>
<td>.068</td>
</tr>
</tbody>
</table>

which both reach peak values (maximum tropical assemblage, minimum eastern boundary assemblage) at the interglacial extremes.

This sequence of events, with upwelling changing first, followed by eastern boundary advection, leads us to speculate on a long-term feedback mechanism to account for the ice age cooling of the tropical Atlantic (Fig. 9). Prior to the transition into an ice age, reduction of atmospheric CO₂ and initial greenhouse cooling in the southern oceans strengthened hemispheric thermal gradients. The resulting compression of atmospheric pressure gradients drove stronger zonal trade winds and equatorial upwelling, which began to cool the equator. Oceanic cooling relative to a warm African land mass shifted the thermal equator northward, and intensified the land-sea atmospheric pressure gradients along the southern-hemisphere eastern boundary. Stronger southerly winds here enhanced the Benguela Current, drawing these cool waters off the eastern boundary between 15° and 20°S, to reach the South Equatorial Current in the central Atlantic, causing further (and we think much stronger) cooling. The cyclonic gyre presently detected in the Gulf of Guinea as a geostrophic feature continued to be active during the glacial maximum, and this suggests that the point of zero wind-stress curl stayed well south of the equator in the eastern Atlantic. This inference of an active cyclonic gyre in the eastern tropical Atlantic bounded by and ice-age version of the Benguela-Angola Front was also made by Schneider et al. (1995). Our down-core record in the central equatorial Atlantic is west of the cyclonic gyre, which may have varied independ-

TABLE 4: Phase Angle of orbital parameters (Berger and Loutre, 1991), ice-core pCO₂ (Jouzel et al., 1993), and equatorial Atlantic faunal assemblages relative to δ¹⁸O in core V30-40. Zero phase means index is coincident with interglacial maximum, negative values indicate index leads interglacial maximum. "nc" indicates no significant coherency. Phase of orbital parameters is equivalent to that of northern-hemisphere insolation strength on June 21 for the 41 kyr and 23 kyr bands (Berger and Loutre, 1991).

<table>
<thead>
<tr>
<th>ITEM:</th>
<th>100 kyr</th>
<th>41 kyr</th>
<th>23 kyr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orbital parameters</td>
<td>-13 ± 12</td>
<td>-80 ± 12</td>
<td>-87 ± 9</td>
</tr>
<tr>
<td>Vostok pCO₂</td>
<td>-26 ± 12</td>
<td>2 ± 13</td>
<td>-33 ± 23</td>
</tr>
<tr>
<td>Factor 1: Maximum Tropical</td>
<td>-4 ± 13</td>
<td>-19 ± 29</td>
<td>-36 ± 7</td>
</tr>
<tr>
<td>Factor 2: Minimum Upwelling</td>
<td>-23 ± 12</td>
<td>nc</td>
<td>-37 ± 12</td>
</tr>
<tr>
<td>Factor 3: Minimum East. Boundary</td>
<td>10 ± 22</td>
<td>nc</td>
<td>-21 ± 11</td>
</tr>
</tbody>
</table>
Fig. 9. Schematic diagram illustrating the inferred sequence of events in a "generic" ice age cycle in the South Atlantic near-surface circulation. a) Interglacial event: moderately strong equatorial upwelling. Benguela Current water weak or absent in the central equatorial Atlantic. b) As polar glaciation begins, southern ocean cooling strengthens hemispheric thermal gradients, driving stronger zonal trade winds and equatorial upwelling, which cool the equator. c) Near glacial maximum, thermal equator northward, and land-sea atmospheric pressure gradients enhanced the Benguela Current, drawing cool waters off the eastern boundary between 15 and 20°S, to reach the South Equatorial Current in the central Atlantic, amplifying cooling. The cyclonic gyre presently detected in the Gulf of Guinea as a geostrophic feature continued to be active during the glacial maximum, and this suggests that the point of zero wind-stress curl stayed well south of the equator in the eastern Atlantic. d) Ice melting phase, warmth in the southern ocean reduces thermal gradients, trade winds and equatorial upwelling diminish. Advection of Benguela water lessens, and the central equatorial Atlantic reaches peak warmth prior to full interglacial time.
ently in response to local monsoonal winds over the Gulf of Guinea.

Summarizing, we infer that the central equatorial system is driven by thermal gradients in the southern hemisphere, first by meridional gradients associated with cooling in the high southern latitudes which modulates the zonal component of the southern trades, and second by a feedback process associated with land-sea contrast, which modulate meridional winds in the SE Atlantic that draws water offshore from the Benguela current toward the equator. This system may be separate from that affected by the monsoon of northern Africa, which may be driven from the north by summer insolation, and changing temperatures of the north Atlantic. This supports and expands on the views of Gardner and Hays (1976) and McIntyre et al. (1989), who thought that cold eastern-boundary water could reach the equatorial zone (but didn’t distinguish the local cyclonic gyre), and with that of Molfino and McIntyre (1990) which emphasized equatorial upwelling driven by zonal trade winds.

Conclusions

Changes in Pleistocene planktonic foraminiferal faunas in the tropical Atlantic can be described in terms of three assemblages, a warm tropical fauna, an upwelling fauna, and an eastern boundary fauna. These assemblages, found by analyzing down-core faunas in the tropical Atlantic and Pacific Oceans, appear to be relatively insensitive to preservation effects associate with moderate dissolution.

Pleistocene variations of the “upwelling” and “eastern boundary” assemblages in the central equatorial Atlantic are concentrated in 100,000-yr and 23,000-yr cycles similar to orbital eccentricity and precession. Most of the variance, both in time series and in the difference between core top and glacial maximum maps, is in the eastern boundary fauna, suggesting that advection of cool Benguela Current water is a key process in large-scale climate change at the equator. The sequence of faunal events at the equator, with early changes in the upwelling assemblage preceding glaciation, followed a few thousand years later by changes in the eastern boundary assemblage, suggests a succession of events in a coupled atmosphere and ocean.

We infer that early changes in the zonal component of the southern trades, perhaps driven by greenhouse cooling at higher southern latitudes, began the ice-age cycle. A later effect of a cool ocean next to a still-warm African continent intensified meridional winds that drew cool Benguela water from its source off SW Africa, toward the equator in the central Atlantic. The eastern Gulf of Guinea appears to be excluded from this large-scale variability, because of a local cyclonic gyre that maintains a strong front near 15°S.

Acknowledgements

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References


Gardner JV, Hays JD (1976) Responses of sea-surface temperature and circulation to global climate changes during the past 200,000 years in the eastern equatorial Atlantic Ocean. Mem Geol Soc Amer 145:221-246


Guilderson TP, Fairbanks RG, Rubenstone JL (1994) Tropical temperature variations since 20,000 years ago: Modulating inter-hemispheric climate change. Science 263:663-665


May RM (1975) Patterns of species abundance and diversity. In: Cody ML, Diamond JM (eds) Ecol-
McIntyre A, Ruddiman WF, Karlin K, Mix AC (1989)
Surface water response of the equatorial Atlantic Ocean to orbital forcing. Paleoceanogr 4:19-55
Mix AC (1986) Late Quaternary paleoceanography of the Atlantic Ocean: Foraminiferal faunal and sta-
ble isotopic evidence. PhD Dissertation, Columbia University, New York, 738 pp
Mix AC (1989) Pleistocene paleoproductivity: evidence from organic carbon and foraminiferal spe-
New York, pp 313-340
Mix AC (1992) The marine oxygen isotope record: Constraints on the timing and extent of ice-growth
events (120-65 ka). In: Clark PU, Lea PD (eds) The last interglacial-glacial transition in North
Mix AC, Ruddiman WF, McIntyre A (1986a) Late
Quaternary paleoceanography of the tropical Atlantic, 1: Spatial variability of annual mean sea-
surface temperatures, 0-20,000 years B.P. Paleoceanogr 1:43-66
Mix AC, Ruddiman WF, McIntyre A (1986b) Late
Quaternary paleoceanography of the tropical Atlantic, 2, The seasonal cycle of sea-surface temperatures, 0-20,000 years B.P. Paleoceanogr 1:339-353
Molfino B, Kipp NG, Morley JJ (1982) Comparison of
foraminiferal, coccolithophorid, and radiolarian paleotemperature equations: Assemblage coherency and estimate concordancy. Quat Res 17:279-313
Molfino B, McIntyre A (1990) Precessional forcing of
nutricline dynamics in the equatorial Atlantic. Science 249:766-769
Molinari RL (1982) Observations of eastward currents
Moore DW, Hisard P, McCreary J, Merle J, O'Brien J,
Moore TC Jr, Hutson WH, Kipp N, Hays JD, Prell WL,
general atmospheric circulation. IAHS Publ 131, pp 295-301
Oberhänsli H, Bénier C, Meinecke G, Schmidt H,
Philander SGH (1979) Variability of the tropical
oceans. Dynamics of Atmospheres and Oceans 3:191-208
Philander SGH (1986) Unusual conditions in the tropical
Picaut J, Servain J, LeCompte P, Seva M, Lukas S,
Pokras EM, Mix AC (1986) Earth's precession cycle and Quaternary climate change in tropical Africa. Nature 326:486-487
Ravelo AC, Fairbanks RG, Philander SGH (1990) Re-
constructing tropical Atlantic hydrography using planktonic foraminifera and an ocean model. Paleoceanogr 5:409-431
Rind D, Peteet D (1985) Terrestrial conditions at the
last glacial maximum and CLIMAP sea-surface temperature estimates: Are they consistent? Quat Res 24:1-22
Sverdrup MS (1983) Planktonic foraminiferal ecology of the eastern equatorial Pacific Ocean: Including a paleoceanographic reconstruction of the Panama Basin for the last 320,000 years. PhD dissertation, Univ. Cincinnati, Cincinnati, OH, USA, 317 pp
Verardo DJ, McIntyre A (1994) Production and destruction: Control of biogenous sedimentation in the tropical Atlantic 0-300,000 years B.P. Paleoclimatog 9:63-86