Similar glacial and Holocene Southern Ocean hydrography

Katsumi Matsumoto,[1] Jean Lynch-Stieglitz, and Robert F. Anderson

Lamont-Doherty Earth Observatory and Department of Earth and Environmental Sciences, Columbia University
Palisades, New York, USA

Abstract. We present new Holocene and glacial δ^{18}O data measured on planktonic foraminifera *Neogloboquadrina pachyderma* and benthic foraminifera *Cibicidoides* species from the Atlantic and Pacific sectors of the Southern Ocean in order to better understand its glacial hydrography. Combined with previously published data, the latitudinal δ^{18}O distributions of these foraminifera show no appreciable difference during the Holocene and the Last Glacial Maximum. This suggests that the basic Southern Ocean hydrography, including stratification and the mean position of the Antarctic Circumpolar Current frontal system, remained largely unchanged, although the small number of Pacific data render any conclusion in that sector preliminary.

1. Introduction

The Southern Ocean is a critical part of the global climate system. The continental shelves of the Weddell and Ross Seas are sites of Antarctic Bottom Water (AABW) formation. The highly barotropic Antarctic Circumpolar Current (ACC) reaches the ocean floor and is able to mix efficiently the North Atlantic Deep Water (NADW) and deep waters from the Indian and Pacific Oceans. The mixture of these deep waters, the Circumpolar Deep Water (CDW), then spreads back into the other ocean basins in one form or another. The formation of Antarctic Intermediate Water (AAIW) by Ekman pumping occurs in the convergence zone at the Polar Front and to the north in the subantarctic zone. The northward advection of temperatures and salinities of AABW, CDW, and AAIW from the Southern Ocean to other ocean basins has a profound effect on the global ocean circulation and thus on global climate.

In addition, the surface waters of the Southern Ocean are key in controlling the atmospheric CO_2 content because they link the atmosphere and the large carbon reservoir of the world deep ocean. The Southern Ocean surface waters have thus been a focus of attention in connection with the investigations of low atmospheric CO_2 content during the Last Glacial Maximum (LGM) [*Barnola et al.*, 1987] as well as with the forecasts of future atmospheric pCO$_2$ levels [e.g., *Sarmiento et al.*, 1998]. Specifically, the relatively abundant surface macronutrients of the modern Southern Ocean provide an opportunity for enhanced biological pump (export of surface nutrients and CO_2 to the deep ocean) to reduce the atmospheric CO_2 content [*Martin and Fitzwater*, 1988]. Also, reducing the vertical mixing of Southern Ocean surface waters can lower the atmospheric pCO$_2$ by limiting the transfer of CO_2 from the deep ocean to the atmosphere [*Ennever and McElroy*, 1985; *Toggweiler and Sarmiento*, 1985; *Wenk and Siegenthaler*, 1985].

In order to better understand the glacial Southern Ocean hydrography near the surface and at depth we examine the latitudinal distributions of δ^{18}O in the calcite test of planktonic foraminifera *Neogloboquadrina pachyderma* and benthic foraminifera *Cibicidoides* species. Their δ^{18}O reflects the seawater temperature and salinity in which they calcify their shells. The planktonic foraminiferal carbon isotope ratios associated with this data set are

discussed extensively by *Kohfeld* [1998]. Also, benthic foraminiferal carbon isotope ratios in the Southern Ocean have been discussed recently at length [*Matsumoto and Lynch-Stieglitz*, 1999; *Ninnemann and Charles*, 1997; *Oppo and Horowitz*, 2000]. Our new data in combination with those in the literature show that the spatial distribution of δ^{18}O in planktonic and benthic foraminiferal tests remained nearly the same during the LGM as during the Holocene, a conclusion similar to that reached by *Charles and Fairbanks* [1990] on the basis of a subset of the planktonic foraminiferal δ^{18}O data shown here. The additional work in the Southern Ocean since that of *Charles and Fairbanks* [1990] more strongly constrains the LGM Southern Ocean hydrography, and we also infer that the circulation was similar during the Holocene and the LGM.

2. Materials and Methods

For the new data presented here, oxygen and carbon isotope measurements were made on *N. pachyderma* and *Cibicidoides* species (including *Planulina wuellerstorfi*) from cores in the Atlantic and Pacific sectors of the Southern Ocean (Table 1). Ten or more individual shells of *N. pachyderma* larger than 150 μm were used to make a single isotope measurement. For benthic foraminifera, measurements were made on groups ranging from one to four individuals larger than 212 μm. All measurements were made at Lamont-Doherty Earth Observatory on Micromass Optima coupled with a Multiprep carbonate preparation device. Repeat analyses of in-house standards indicate a precision of ±0.07‰ for δ^{18}O and ±0.04‰ for δ^{13}C for samples larger than 30 μg. The precision including smaller samples is ±0.12‰ and ±0.05‰ for δ^{18}O and δ^{13}C, respectively.

Foraminiferal calcite δ^{18}O (δ^{18}O$_{calcite}$) was predicted using the equation for *Cibicidoides* species [*Lynch-Stieglitz et al.*, 1999], which is nearly the same as for inorganically precipitated calcite, presumed to be equilibrium precipitation [*Kim and O'Neil*, 1997], and for cultured planktonic foraminifera [*Bemis et al.*, 1998] for the temperature range over which they overlap. None of these relationships were calibrated below 4°C. It is worth noting that the "paleotemperature" equations in these three studies all produce slightly higher δ^{18}O values than the older experiments with inorganic precipitation [*O'Neil et al.*, 1969] and the *Uvigerina*-based paleotemperature equation of *Shackleton* [1974]. It was necessary to add +0.64‰ [*Shackleton and Opdyke*, 1973] or +0.83‰ [*Herguera et al.*, 1992] to *Cibicidoides* δ^{18}O in order to bring the values into agreement with those predicted from the *O'Neil et al.* [1969] experiments. There no longer appears to be

Table 1. New Stable Isotope Data[a]

| | | | | Neogloboquadrina pachyderma | | | | Cibicidoides spp. | | | |
| | | | | Holocene | | LGM | | Holocene | | LGM | |
Cores	Latitude	Longitude	Depth, m	$\delta^{18}O$	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C$
South Atlantic											
RC11-86	35°S	19°E	2829	0.84	−0.41	1.82	0.09				
V34-157	41°S	26°E	3636							3.98[b]	−0.18[b]
RC11-80	46°S	0°	3656							3.85	−0.04
V29-105	48°S	18°E	4350	2.32	1.01	3.25	0.48			3.90	−0.28
South Pacific											
NBP9802 St.8	59°S	170°W	4324	2.55	1.05			3.06	0.27		
NBP9802 St.7	60°S	170°W	4005	2.62	1.45			2.65[b]	0.43[b]		
NBP9802 St.5	63°S	170°W	2940	3.15	1.00	4.55	0.51			4.22[b]	0.07[b]
NBP9802 St.4	64°S	170°W	2688	3.02	0.76	4.48	0.37			4.00[b]	0.07[b]
NBP9802 St.3	66°S	170°W	3232	3.18	0.62	4.22	0.21			3.98[b]	0.00[b]
GL78-16	76°S	163°W	1937	3.86	0.50			3.38	1.07		

[a] All isotope measurements are reported in ‰ versus VPDB.

[b] Measured on *Planulina (Cibicidoides) wuellerstorfi*.

any discrepancy between the *Cibicidoides* $\delta^{18}O$ and the inorganic calcite experiments over the temperature range of 10–26°C [*Bemis et al.*, 1998; *Lynch-Stieglitz et al.*, 1999]. Further experiments are necessary to determine whether *Cibicidoides* precipitates in isotopic equilibrium at colder temperatures. The $\delta^{18}O$ measurements on *N. pachyderma* collected from plankton tows in the South Atlantic [*Kohfeld*, 1998] also appear consistent with the *Kim and O'Neil* [1997] equation. Here we use the $\delta^{18}O$ relationship of *Lynch-Stieglitz et al.* [1999]:

$$\delta^{18}O_{calcite}(PDB) = \delta^{18}O_{SW}(SMOW) - 0.21T + 3.38 - 0.27,$$

where PDB is Peedee belemnite and SW is sea water, because it was calibrated at a lower temperature (4°–26°C) than the *Kim and O'Neil* [1997] experiments, making it more appropriate for application to the low polar temperatures in this study.

Annual mean salinity [*Levitus et al.*, 1994] and temperature [*Levitus and Boyer*, 1994] data at 3000-m water depth were used to calculate $\delta^{18}O_{calcite}$ to compare with *Cibicidoides* $\delta^{18}O$. For comparison with *N. pachyderma* $\delta^{18}O$, near-surface $\delta^{18}O_{calcite}$ were calculated using the austral summer data [*Levitus and Boyer*, 1994; *Levitus et al.*, 1994]. The seawater $\delta^{18}O$-salinity relations for the deep and near-surface waters were based on the regional Geochemical Ocean Sections Study data [*Bainbridge*, 1981; *Broecker et al.*, 1982; *Weiss et al.*, 1983].

Some of the cores that we analyzed from the South Pacific were multicores, which are designed to retrieve the surface sediments. The $\delta^{18}O$ data from multicores are shown to be consistent with the predicted values and are thus presumed to represent the Holocene. Core top data from gravity and piston cores are likewise presumed to represent the Holocene when they are consistent with predicted $\delta^{18}O_{calcite}$. The LGM interval was identified by the downcore maximum in *N. pachyderma* $\delta^{18}O$. Benthic isotope data, when available, were used to confirm the LGM level. For V29-105, *Cycladorphora davisiana* biostratigraphy was also used (J. Hays, personal communication, 1998). For most cores the glacial $\delta^{18}O$ is heavier than the Holocene $\delta^{18}O$ values by the amount expected from a global ice volume change of 1 ∼ 1.3‰ as reported by *Fairbanks* [1989], *Schrag and DePaolo* [1993], and *Schrag et al.* [1996]. For the previously published LGM data we used the intervals chosen by these authors using very similar criteria.

We compiled previously published *N. pachyderma* and *Cibicidoides* species $\delta^{18}O$ data from the South Atlantic and the South Pacific south of 30°S (Figure 1). The benthic $\delta^{18}O$ data compilation includes four data points from the eastern Weddell Sea on the Antarctic continental margin measured on another epibenthic species *Epistominella exigua*. It is not clear whether this species has the same fractionation as observed for *Cibicidoides*. The core top values are heavier than predicted using seawater $\delta^{18}O$ and temperature, but the *Lynch-Stieglitz et al.* [1999] equation did not include data below 4°C. It is also possible that the core top values were not recording latest Holocene conditions. While *Woodruff et al.* [1980] found both *E. exigua* and *E. nuttiloides* to have the same $\delta^{18}O$ as *C. wuellerstorfi*, *Mackensen et al.* [1989] refer to data which suggest heavier $\delta^{18}O$ values for *E. exigua* as well. Despite the ambiguity, *E. exigua* $\delta^{18}O$ data are included in our compilation because they offer the only benthic data from the periphery of the Antarctic continent. A more detailed comparison of these data with those farther to the north will require further investigation into the isotopic fractionation of *E. exigua* relative to *Cibicidoides*. Our compilation is archived at the World Data Center-A for Paleoclimatology (http://www.ngdc.noaa.gov/paleo/data.html).

3. Latitudinal $\delta^{18}O$ Distributions

The Holocene and LGM $\delta^{18}O$ values from the new downcore $\delta^{18}O$ measurements from the Pacific (Figure 2) and the Atlantic sector (Figure 3) are summarized in Table 1. These new time slice values are plotted together with compiled data versus latitude for the Atlantic and Pacific sector in Figures 4 and 5, respectively.

The Holocene *N. pachyderma* $\delta^{18}O$ in the Atlantic sector agrees with the predicted $\delta^{18}O_{calcite}$ at 200 m and shows a gradient of ∼2.5‰ between 30°S and 70°S (Figure 4a), which reflects the summer temperature gradient of ∼15°C between the cold polar and warm subtropical waters at 200 m. The warmer surface mixed layer hydrography is not reflected in *N. pachyderma* $\delta^{18}O$, which is heavier than the predicted surface $\delta^{18}O_{calcite}$ by ∼0.5‰. Plankton tow results in the Atlantic sector of the Southern Ocean as reported by *Kohfeld* [1998] show that planktonic foraminifera are most abundant in the top 200 m but do not reveal a pronounced depth-dependent encrustation for *N. pachyderma*. One possible reason for this may be that the water column stratification was weak at the time of plankton tows [*Kohfeld*, 1998], before the formation of the shallow, seasonal pycnocline (∼50 m). The implication is that plankton tows during the summer may show preferential *N. pachyderma* encrustation in the upper 50 m. While this may, indeed, be the case, *N. pachyderma* associated with the shallow mixed layer appears to contribute little to the *N. pachyderma* preserved in the sediments because surface sediment *N. pachyderma* $\delta^{18}O$ data agree with the $\delta^{18}O$ of *N. pachyderma* captured by plankton tows in the uppermost 300 m when stratification was weak [*Kohfeld*, 1998].
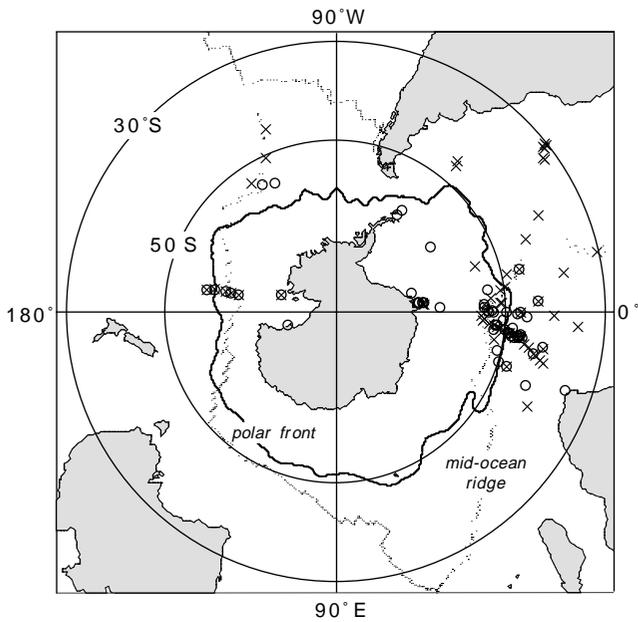
Figure 1. Locations of sediment cores from which new data were obtained or previous data were compiled. Compiled data include $\delta^{18}O$ from planktonic foraminifera *N. pachyderma* [*Charles*, 1991; *Charles and Fairbanks*, 1990; *Charles et al.*, 1996; *Donner and Wefer*, 1994; *Grobe and Mackensen*, 1992; *Grobe et al.*, 1990; *Hodell*, 1993; *Hodell et al.*, 2000; *Keigwin and Boyle*, 1989; *Mackensen et al.*, 1994, 1989; *Niebler*, 1995; *Ninnemann and Charles*, 1997; *Wefer et al.*, 1982] and epibenthic foraminifera *Cibicidoides* species and *E. exigua* from deeper than 2000 m [*Belanger et al.*, 1981; *Boyle*, 1992; *Charles and Fairbanks*, 1992; *Curry et al.*, 1988; *Curry and Lohmann*, 1982; *Duplessy et al.*, 1988, 1984; *Graham et al.*, 1981; *Mackensen et al.*, 1994, 1993; *Matsumoto and Lynch-Stieglitz*, 1999; *Sarnthein et al.*, 1988, 1994]. Also included are unpublished data from P. G. Mortyn et al. (Building a marine sedimentary analog for the Vostok ice core, manuscript in preparation, 2001). Core locations of planktonic foraminiferal data are indicated with circles, and benthic foraminiferal data are indicated with crosses. Positions of the modern Antarctic Polar Front [*Orsi et al.*, 1995] and mid-ocean ridge axes are also plotted.

Waters in the upper 300 m prior to the formation of the seasonal pycnocline correspond to waters below it during the summer, including the subsurface water at 200 m (Figure 4).

The 2.5‰ change in *N. pachyderma* $\delta^{18}O$ with latitude occurs almost entirely to the north of the Polar Front (Figure 4), which in this sector is at ~50°S (Figure 1). Poleward of 50°S, *N. pachyderma* $\delta^{18}O$ is nearly constant and reflects the hydrography of CDW (mostly NADW in the Atlantic sector), which upwells at the Antarctic Divergence.

At depth, CDW has approximately uniform temperature and salinity, which are reflected in the nearly constant Holocene benthic foraminiferal $\delta^{18}O$ and the predicted $\delta^{18}O_{calcite}$ at 3000 m (Figure 4a). The majority of benthic $\delta^{18}O$ data agree with the predicted $\delta^{18}O_{calcite}$, while the *E. exigua* data are heavier by ~0.6‰, as discussed in section 2. *Cibicidoides* $\delta^{18}O$ becomes slightly heavier from 30°S to 55°S, reflecting the presence of increasingly colder waters to the south. For the Holocene this can be explained by the presence of a higher percentage of the colder AABW relative to NADW. It is worth noting that as long as there are sources of deep water flowing into the Southern Ocean with varying densities (e.g., AABW and NADW), the geostrophic flow

of the ACC will be reflected in isopycnals that slope upward to the south, and an increase in density to the south will be reflected in the southward increase of benthic $\delta^{18}O$.

The latitudinal distributions of glacial *N. pachyderma* and *Cibicidoides* $\delta^{18}O$ from the Atlantic sector are structurally very similar to the Holocene (Figure 4b). The glacial benthic $\delta^{18}O$ is nearly uniform. The scatter in the glacial data is larger than for the Holocene, so we do not observe the increase in benthic $\delta^{18}O$ between 30°S and 55°S. However, the scatter in the data is large enough that we cannot rule out that such a gradient and continued formation of AABW somewhere in the perimeter of Antarctica [*Duplessy et al.*, 1996] did exist during glacial time. Also, like the Holocene, glacial *N. pachyderma* $\delta^{18}O$ shows a transition near 50°S in the form of sloping $\delta^{18}O$ equatorward and relatively constant $\delta^{18}O$ poleward, indicating that the LGM position of the Polar Front remains at ~50°S.

Although there is almost no glacial $\delta^{18}O$ data for both the planktonic and benthic foraminifera between 50°S and 70°S, the likely $\delta^{18}O$ values are well constrained to ~4.5‰ by the $\delta^{18}O$ immediately to the north and south of the data gap having comparable values. The apparent lack of change in the $\delta^{18}O$ gradient between the planktonic and benthic data poleward of the Polar Front suggests that CDW continued to upwell because it is the strong vertical exchange that drives the Antarctic surface water temperature and salinity toward those of CDW. If this upwelling were significantly reduced, then the processes, such as precipitation and evaporation, that cause the surface water temperature and salinity to diverge from those of CDW would gain proportionally stronger control of the Antarctic surface water hydrography. Therefore, if the lack of change in the surface-to-deep $\delta^{18}O$ gradient were maintained by processes other than a continued CDW upwelling, then those processes somehow had to change in concert, which appears unlikely.

Despite the structural similarity to the Holocene the glacial *Cibicidoides* $\delta^{18}O$ data are heavier by ~0.5‰ than the $\delta^{18}O_{calcite}$ at 3000 m adjusted for the global ice volume effect (1.0 ~ 1.3‰; see Figure 4b). This suggests that the glacial deep water was colder by 2°C or more [*Labeyrie et al.*, 1987]. Because the deep water poleward of the Polar Front is near 0°C today, the temperature decrease implies a near freezing temperature for the glacial CDW.

In the Pacific sector, much fewer data are available (Figure 5), $\delta^{18}O$ trends are less obvious, and interpretations are more difficult. However, $\delta^{18}O$ patterns observed in the Atlantic sector are recognizable in or appear consistent with the Pacific $\delta^{18}O$ data. The modern Polar Front in this part of the Southern Ocean is located at ~60°S (Figure 1). The Holocene *N. pachyderma* $\delta^{18}O$ to the north of 60°S becomes increasingly light (Figure 5a), reflecting the increasing proportion of warm subtropical waters in the subantarctic zone. *N. pachyderma* $\delta^{18}O$ agrees better with the estimated $\delta^{18}O_{calcite}$ at 200 m than that with the estimated $\delta^{18}O_{calcite}$ at the surface. Finally, latitudinal *Cibicidoides* $\delta^{18}O$ distribution is nearly uniform but becomes heavier toward the pole, probably reflecting the increasing percentage of the colder AABW versus the warmer Pacific Deep Water.

The available glacial *Cibicidoides* $\delta^{18}O$ data in the Pacific sector are nearly uniform with latitude (Figure 5b). However, with so few data it is difficult to assess whether the $\delta^{18}O$ gradient is present. The LGM position of the Polar Front as defined by the latitude at which *N. pachyderma* $\delta^{18}O$ becomes nearly uniform poleward and increasingly light toward the equator is likewise ambiguous with so few data (Figure 5b).

4. Discussion

In light of the limited number of data and the ambiguities in the Pacific sector, we discuss glacial Southern Ocean hydrography and
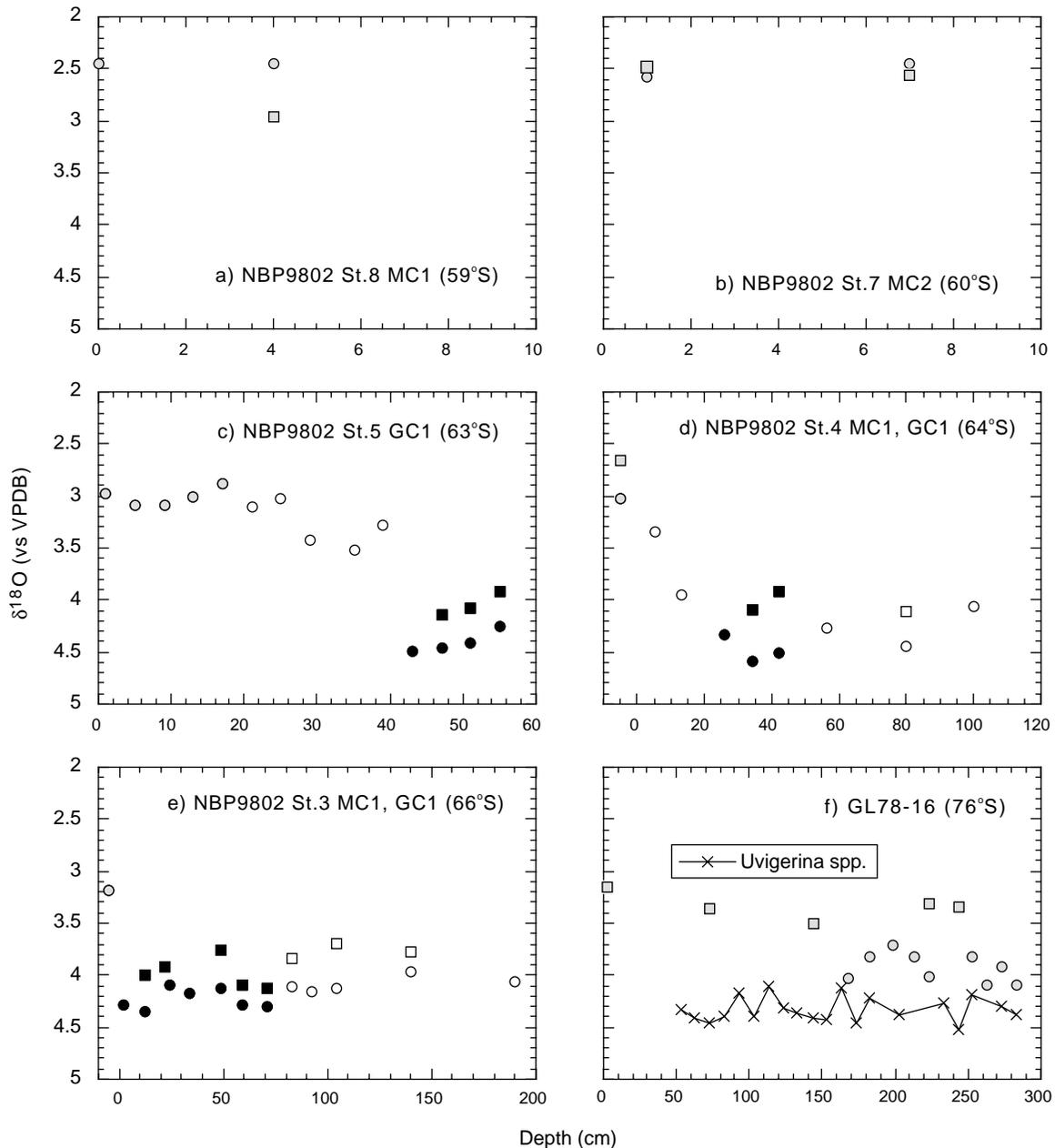
Figure 2. South Pacific $\delta^{18}O$ data. *N. pachyderma* data are indicated by circles, and *Cibicidoides* species data are indicated by squares. Data plotted at negative core depths in Figures 2d and 2e are from multicores. Shaded and solid symbols indicate data used to obtain the Holocene and glacial time slice values, respectively, given in Table 1. Glacial sediments were not reached or were not available in multicores at NBP9802 stations 8 (Figure 2a) and 7 (Figure 2b) and at GL78-16 (Figure 2f).

circulation on the basis of $\delta^{18}O$ data from the Atlantic sector, where there is good evidence for continued upwelling of CDW at the Antarctic Divergence and no change in the mean position of the Polar front.

4.1. Glacial Southern Ocean Circulation

The most simple and fundamental conclusion to be drawn from the latitudinal $\delta^{18}O$ distributions of planktonic and benthic foraminifera is that there was little change in the overall Southern Ocean hydrography and thus in circulation between

the last glacial period and today. The robust Southern Ocean circulation suggests that factors such as bottom topography and land-sea configuration that do not change over the glacial-interglacial timescale exert strong controls over large-scale ocean circulation.

A glaciated continent centered over the South Pole, together with the same sense of the Earth's rotation and geostrophy, would set the conditions for the presence of surface westerly winds over the LGM Southern Ocean. The winds, through Ekman pumping, would in turn cause CDW to upwell where surface water and sea ice diverge. Also, the circumpolar west-
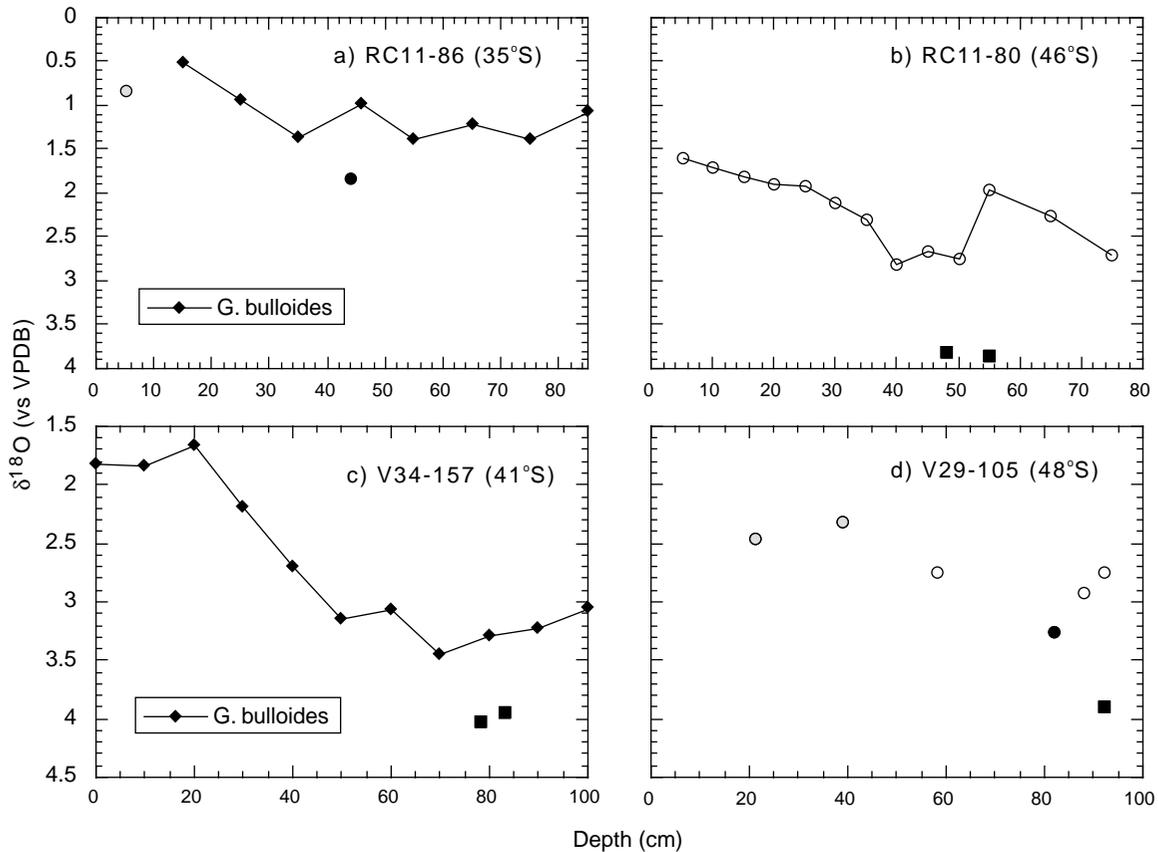
Figure 3. South Atlantic $\delta^{18}O$ data. Symbols are as in Figure 2. *Globigerina bulloides* $\delta^{18}O$ stratigraphies of RC11-86 and V34-157 are from *Verardo* [1995], and RC11-80 *N. pachyderma* $\delta^{18}O$ stratigraphy is from *Charles and Fairbanks* [1990].

erlies would drive the ACC. We expect the glacial ACC and its fronts to have been guided largely by bottom topography like today, as determined from historical hydrographic data [*Belkin and Gordon*, 1996; *Gordon et al.*, 1978], satellite data [*Chelton et al.*, 1990; *Gille*, 1994; *Moore et al.*, 1999], and models [*Gille*, 1997; *Hughes and Killworth*, 1995; *Marshall*, 1995]. Although the mean position of the Polar Front in a simple, one-layer Southern Ocean model is shown to shift 2° to the north in a glacial simulation forced by westerlies increased in strength by 70% and shifted northward by 5° compared to the modern Polar Front [*Klinck and Smith*, 1993], a 2° shift does not appear significant, as it is well within the natural spatial variability of the Polar Front according to a 7-year observation of satellite sea surface temperatures [*Moore et al.*, 1999]. It should be pointed out that it remains to be demonstrated that the westerlies can shift in position independently of the ACC, given the strong atmosphere-ocean coupling in the Southern Ocean. While our $\delta^{18}O$ data do not preclude frontal shifts by ~2° or less, any larger shift as suggested by other studies and discussed in section 4.2 is not supported (Figure 4).

It is important to point out that planktonic foraminiferal $\delta^{18}O$, including the *N. pachyderma* data presented here, does not offer much constraint on sea ice formation and melting. While these processes have important roles in determining the sea surface salinity and thus the surface conditions in polar regions, they do not change the $\delta^{18}O$ of seawater as much as evaporation and precipitation [*Craig and Gordon*, 1965]. Because of this the $\delta^{18}O$ of foraminiferal calcite from the Antarctic is not related to density

in a straightforward way, as it is in the warm waters of the world ocean [*Lynch-Stieglitz et al.*, 1999].

4.2. Consistencies With Other Studies

Our view that large-scale glacial Southern Ocean circulation was similar to modern circulation adds to the constraints offered by previous studies in reconstructing the glacial Southern Ocean. While we are unable to consistently account for all observations, here we discuss our interpretation in relation to other work that relates to the position of the Polar Front, sea ice extent, and surface productivity.

The position of the Polar Front is associated today with high surface productivity [*Lutjeharms et al.*, 1985] and with a sediment boundary that separates the carbonate-rich sediments to the north from the opal-rich sediments to the south. This boundary shifted northward by ~7° during the LGM compared to the Holocene, which has been used as supporting evidence for a northward migration of the Polar Front [*Hays et al.*, 1976]. Additional supporting evidence is provided by the enhanced export productivity (primarily opal flux) during the LGM to the north of the modern Polar Front [*Kumar et al.*, 1995]. However, export production and plankton assemblages reflect environmental conditions, including sea ice cover, surface water stratification, and the supply of nutrients to the surface waters, and they are not tied directly to the Polar Front itself. The $CaCO_3$-opal sedimentary boundary recorded in glacial sediments could reflect a biological response to variables other than the position of the Polar Front. In contrast, foraminiferal $\delta^{18}O$ directly reflect sea-
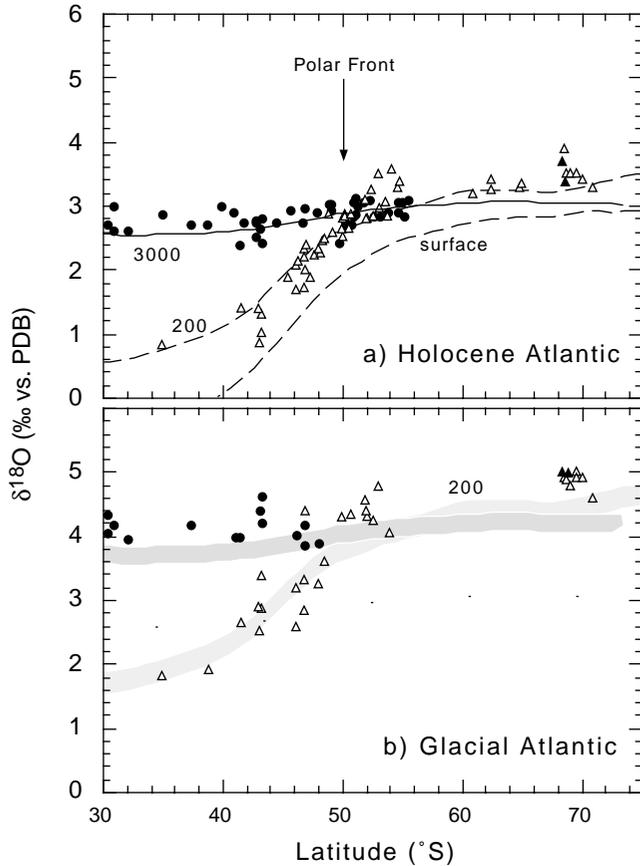
Figure 4. South Atlantic (a) Holocene and (b) LGM latitudinal $\delta^{18}O$ distributions of *N. pachyderma* (open triangles), *Cibicidoides* species (circles), and *E. exigua* (solid triangles). Calcite $\delta^{18}O$ in equilibrium with modern ambient seawater was calculated for the Holocene at the surface and at 200 m (dashed lines) using austral summer hydrographic data and at 3000 m (solid line) using annual mean data. Thick shaded lines in Figure 4b represent the 200- and 3000-m equilibrium calcite $\delta^{18}O$ lines shifted by +1.0‰ [*Schrag and DePaolo*, 1993; *Schrag et al.*, 1996] and by +1.3‰ [*Fairbanks*, 1989]. The approximate position of the Polar Front is indicated by an arrow. The Antarctic Zone is to the south of the Polar Front.

water temperature and salinity, which are the hydrographic properties used to define the Polar Front and other physical oceanographic features.

One way to enhance diatom productivity (and hence opal flux) and to shift the $CaCO_3$-opal sediment boundary without a change in the position of the Polar Front is a stronger northward mixing of relatively fresh, silica-rich Antarctic surface water across the front, as first suggested by *Charles et al.* [1991]. Cross-frontal mixing is evident today in the formation of AAIW with its low salinity signature within the Polar Frontal Zone, lying between the Polar Front and the Subantarctic Front. Also, the nitrogen isotope ratio ($\delta^{15}N$) measurements of surface water nitrate from the subantarctic Indian and Pacific Oceans point to cross-frontal mixing as the main mechanism for supplying nitrate to the surface subantarctic [*Sigman et al.*, 1999]. By analogy with modern nitrate transport an increased cross-frontal mixing of dissolved silica could have supported the increased glacial diatom productivity recorded in sediments north of the modern Polar Front. This mechanism can accomplish the sepa-

ration of the "silica front" as seen in sedimentary opal belt and the "dynamical front" represented by the Polar Front [*Charles et al.*, 1991].

Today, freshly supplied silica and nitrate are supplied from below to the Antarctic surface waters in a molar ratio of roughly 2:1 which drops as surface waters advect away because proportionally more silica compared to nitrate is consumed by diatoms [*Hutchins and Bruland*, 1998; *Takeda*, 1998; *Franck et al.*, 2000]. By the time surface waters approach the Polar Front, silica is significantly depleted, especially in summer [*Sigmon et al.*, 2001]. If this condition had persisted during the last glacial period, even a stronger cross-frontal mixing may not have provided a sufficient amount of silica to the surface waters north of the Polar Front to support an enhanced diatom productivity. Assuming that CDW upwelling was largely unchanged relative to today, as suggested by our $\delta^{18}O$ data, reduced silica utilization south of the Polar Front would have permitted an increased cross-frontal transport of silica during the LGM. A lower silica utilization in the glacial Antarctic surface is consistent with low silicon isotope ratio ($\delta^{30}Si$) measured on sedimentary opal [*De La Rocha et al.*, 1998]. In fact, a box model calculation of surface silica budget to account for observed opal burial rates in the glacial Southern Ocean indicates a less efficient silica utilization in the Antarctic surface waters [*Charles et al.*, 1991], although the opal burial rates used in the model were not ^{230}Th-normalized, which is necessary to account for syndepositional sediment redistribution [*Suman and Bacon*, 1989].

A scenario that can reduce silica utilization by limiting diatom growth to the south of the Polar Front is increased sea ice cover
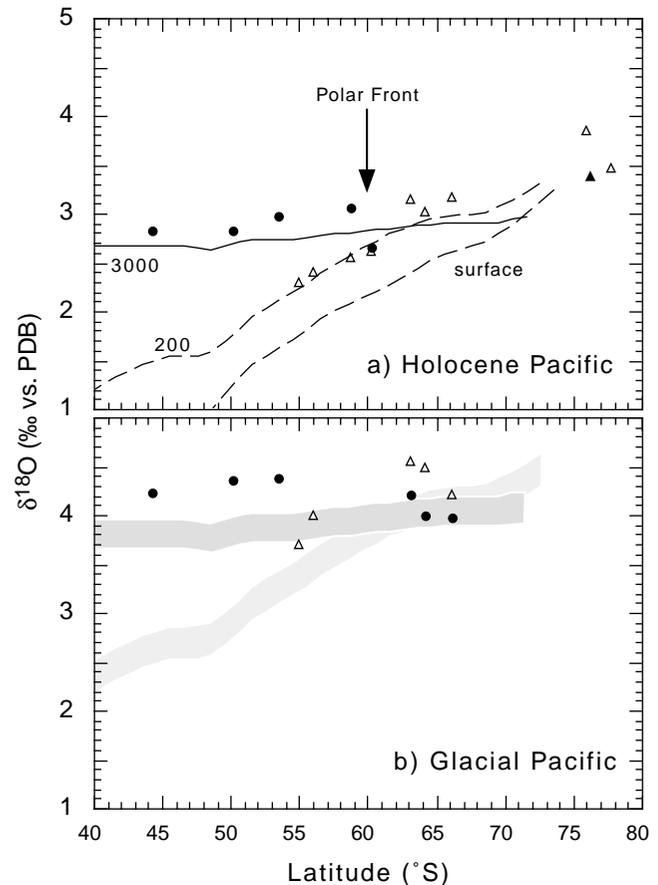

Figure 5. South Pacific (a) Holocene and (b) LGM latitudinal $\delta^{18}O$ distributions of *N. pachyderma*, *Cibicidoides* species, and *E. exigua*. Symbols are as in Figure 4.
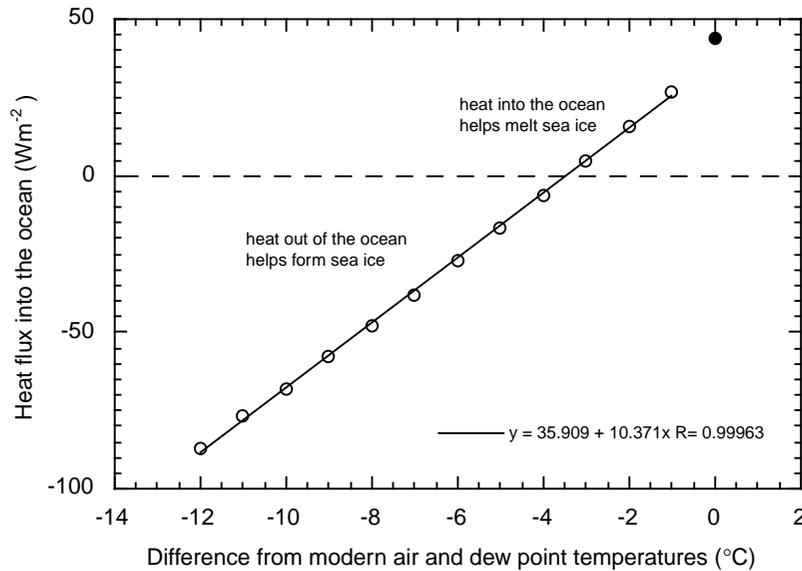
Figure 6. Average total atmosphere-to-ocean heat flux for a 60-day melting period today between mid-November and mid-January in the latitude band from 60°~70°S [after *Gordon*, 1981]. The modern value (solid circle) was calculated using all climatologies describing the modern conditions from *Gordon* [1981], except wind speeds which were taken from 1854–1992 monthly Comprehensive Ocean-Atmosphere Data Set climatology [*Woodruff et al.*, 1987]; the modern value is equivalent to "Qt" in Table 3 of *Gordon* [1981]. Incoming shortwave radiation was adjusted for the Last Glacial Maximum according to *Berger* [1978]. Glacial heat fluxes (open circles) were calculated for equivalent temperature decreases from the modern surface air and dew point temperatures. This simple plot shows, as expected, that as the air temperature and moisture content decrease, the ocean loses more heat to the atmosphere as the colder and drier air extracts larger sensible and latent heat from the ocean.

in space and time [*Charles et al.*, 1991], a notion that appears to be partially consistent with a recent reconstruction of the LGM Southern Ocean sea ice [*Crosta et al.*, 1998]. According to diatom floral assemblages, winter sea ice was more expansive by 5° ~ 8° [*Crosta et al.*, 1998], which was likely a result of a greater rate of sea ice formation facilitated by stronger katabatic winds and colder surface air during the LGM. The more expansive and possibly thicker winter sea ice may have partly persisted through the year and may have occupied a larger area than today, thereby limiting diatom productivity and silica utilization. However, the summer sea ice edge of *Crosta et al.* [1998] shows that the glacial Southern Ocean was as ice-free as today, although this summer reconstruction is not as well constrained by data. To account for the open waters, *Crosta et al.* [1998] call on heat flux from upwelling CDW to melt back the winter sea ice [*Gordon*, 1981]. However, it is unlikely that any significant amount of heat was supplied by upwelling CDW if, indeed, the glacial CDW was near the freezing temperature, as suggested by our δ^{18}O data.

A simple air-sea heat flux calculation using bulk aerodynamic formulae that ignore ocean-atmosphere coupling suggests that it also appears unlikely that any significant heat flux was supplied by the atmosphere (Figure 6). For analysis we assume that the change in surface air temperature over the Southern Ocean between today and the LGM lies somewhere between a minimum reduction of ~5°C, as observed in tropical continental regions [*Stute et al.*, 1995], and a maximum of ~12°C, as observed at Vostok, Antarctica [*Petit et al.*, 1999]. For surface air temperatures from 5°~12°C below the modern temperatures the calculated ocean heat flux for the spring season is all negative (Figure 6), which thermodynamically predicts sea ice formation. We emphasize, however, that this result is speculative because ocean-atmosphere coupling was neglected and it is unclear if

the 5°~12°C reduction in air temperatures over the Southern Ocean is reasonable. However, we note that ice divergence under the influence of the strengthened westerlies, instead of sea ice melting, could have created sufficient leads that allowed "open ocean" diatoms to survive and to have left a mark in the sediments [*Crosta et al.*, 1998]. The presence of a substantial amount of sea ice in the Antarctic surface waters during the LGM summer is consistent with the dominance of *Eucampia antarctica*, a diatom species abundant today only near shore or in a pseudoneritic environment such as floating sea ice, in the glacial sediments [*Burckle*, 1984].

These uncertainties indicate that sea ice reconstruction during the LGM summer is an area which would benefit from more paleoceanographic studies in the future. Also, the role of dynamics (ice divergence) versus thermodynamics (ocean heat fluxes) in determining the modern Southern Ocean sea ice distribution is currently being debated [*Martinson*, 1993], and a better understanding of these modern processes would help extend our understanding of the glacial Southern Ocean.

Two other mechanisms are also capable of producing the silica underutilization south of the Polar Front during the LGM that allows enhanced cross-frontal mixing of silica to explain the northward migration of the sedimentary opal belt [*Hays et al.*, 1976; *Kumar et al.*, 1995]. One is that the increased availability of iron from the dustier glacial Southern Hemisphere atmosphere [*Petit et al.*, 1999] effectively lowered the Si:N uptake ratio by diatoms [*Hutchins and Bruland*, 1998; *Takeda*, 1998]. The second is that other phytoplankton species common in the Southern Ocean, such as *Phaeocystis*, which do not make siliceous shells, outcompeted diatoms in the Antarctic surface waters. It is important to note that the growth of diatoms in the first case and the growth of nonsiliceous phytoplankton in the second would have to be somehow slowed or halted so that nitrate would not

become completely depleted. Otherwise, there would not be enough nitrate to be mixed northward together with silica across the Polar Front.

However silica utilization was diminished to the south of the Polar Front during the LGM, the idea of diminishment gains support from the observation that opaline export productivity was reduced there [*Kumar et al.*, 1995]. This is in addition to the evidence for reduced silica utilization from sedimentary δ^{30}Si [*De La Rocha et al.*, 1998] mentioned above. Furthermore, a reduction in silica utilization is not inconsistent with increased nitrate use, as indicated by bulk sedimentary δ^{15}N [*Francois et al.*, 1997]. If reduced silica use was partly brought about by a lower Si:N uptake ratio by diatoms due to iron availability or ecologically dominant phaeocystis, then increased nitrate utilization would, in fact, be predicted. Therefore reduced silica utilization south of the Polar Front appears to be relatively well supported by observations.

Observations from the north of the Polar Front paint a more mixed picture of the glacial Southern Ocean. According to sedimentary δ^{15}N measurements the increase in opal export productivity [*Hays et al.*, 1976; *Kumar et al.*, 1995] occurred without much difference in nitrate utilization efficiency compared to today [*Francois et al.*, 1997]. To account for these observations, *Francois et al.* [1997] suggest that there was an increased supply of nitrate from below as a result of weaker vertical stratification. However, a weaker stratification seems to be at odds with the diatom's need for a strongly stratified, shallow mixed layer in order to bloom.

Also, a stronger stratification would be expected if there were a larger flux of fresh water to the north of the Polar Front as a result of stronger cross-frontal mixing of Antarctic surface waters that is needed to supply silica for enhanced diatom productivity as recorded in sediments. A large amount of fresh water would be available for cross-frontal mixing in the vicinity of the Polar front from increased sea ice formation due to colder and stronger katabatic winds and decreased melting of sea ice to the south of the polar front due to colder air temperatures and colder CDW.

These considerations indicate that the characterization of the glacial Southern Ocean surface waters to the north of the Polar Front is another area that would benefit from more studies. For example, a better understanding of the modern nutrient supply [*Sigman et al.*, 1999] and use [*Hutchins and Bruland*, 1998; *Takeda*, 1998] as well as more reconstruction of nutrient utilization [*De La Rocha et al.*, 1998; *Francois et al.*, 1997] would help advance our understanding of the glacial Southern Ocean.

5. Conclusions

We presented Holocene and LGM latitudinal foraminiferal δ^{18}O distributions in the Atlantic and Pacific sectors of the Southern Ocean in order to understand its glacial hydrography. *N. pachyderma* and *Cibicidoides* δ^{18}O patterns are nearly the same during the Holocene and the LGM, suggesting that the basic hydrographic structure and thus the circulation remained unchanged. The δ^{18}O data indicate a continued upwelling of CDW at the Antarctic Divergence and no significant change in the mean position of the ACC frontal system. These results suggest that factors that do not change over glacial-interglacial timescale, such as bottom topography and land-sea configuration, exert significant controls over the large-scale Southern Ocean circulation. Our results, when examined together with other LGM observations, suggest that a reduced silica utilization south of the Polar Front and a stronger northward mixing of the underutilized silica across the Polar Front can account for a northward migration of the opal belt without having to shift the Polar Front as called for by previous studies.

References

Bainbridge, A. E., *GEOSECS Atlantic Ocean Expedition*, vol. 1, *Hydrographic Data 1972–1973*, 121 pp., Natl. Sci. Found., Washington, D. C., 1981.

Barnola, J. M., D. Raynaud, Y. S. Korotkevich, and C. Lorius, Vostok ice core provides 160,000-year record of atmospheric CO_2, *Nature*, *329*, 408–414, 1987.

Belanger, P. E., W. B. Curry, and R. K. Matthews, Core-top evaluation of benthic foraminiferal isotopic ratios for paleo-oceanographic interpretations, *Paleogeogr. Paleoclimatol. Paleoecol.*, *33*, 205–220, 1981.

Belkin, I. M., and A. L. Gordon, Southern Ocean fronts from the Greenwich meridian to Tasmania, *J. Geophys. Res.*, *101*, 3675–3696, 1996.

Bemis, B. E., H. J. Spero, J. Bijma, and D. W. Lea, Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations, *Paleoceanography*, *13*, 150–160, 1998.

Berger, A., Long-term variations of daily insolation and Quaternary climatic change, *J. Atmos. Sci.*, *35*, 2362–2367, 1978.

Boyle, E. A., Cd and δ^{13}C paleochemical ocean distributions during the stage 2 glacial maximum, *Annu. Rev. Earth Planet. Sci.*, *20*, 245–287, 1992.

Broecker, W. S., D. Spencer, and H. Craig, *GEOSECS Pacific Ocean Expedition*, vol. 3, *Hydrographic Data 1973–1974*, 137 pp., Natl. Sci. Found., Washington, D. C., 1982.

Burckle, L. H., Ecology and paleoecology of the marine diatom *Eucampia antarctica* (Castr.) Mangin, *Mar. Micropaleontol.*, *9*, 77–86, 1984.

Charles, C. D., Late Quaternary ocean chemistry and climate change from an Antarctic deep sea sediment perspective, Ph.D. thesis, Columbia Univ., New York, 1991.

Charles, C. D., and R. G. Fairbanks, Glacial to interglacial changes in the isotopic gradients of Southern Ocean surface waters, in *Geological History of the Polar Oceans: Arctic Versus Antarctic*, edited by U. Bleil and J. Thiede, pp. 519–538, Kluwer Acad., Norwell, Mass., 1990.

Charles, C. D., and R. G. Fairbanks, Evidence from Southern Ocean sediments for the effect of North Atlantic deep-water flux on climate, *Nature*, *355*, 416–419, 1992.

Charles, C. D., P. N. Froelich, M. A. Zibello, R. A. Mortlock, and J. J. Morley, Biogenic opal in Southern Ocean sediments over the last 450,000 years: Implications for surface water chemistry and circulation, *Paleoceanography*, *6*, 697–728, 1991.

Charles, C. D., J. Lynch-Stieglitz, U. S. Ninnemann, and R. G. Fairbanks, Climate connections between the hemisphere revealed by deep sea sediment core/ice core correlations, *Earth Planet. Sci. Lett.*, *142*, 19–27, 1996.

Chelton, D. G., M. G. Schlax, D. L. Witter, and J. G. Richman, Geosat altimeter observations of the surface circulation of the Southern Ocean, *J. Geophys. Res.*, *95*, 17,877–17,903, 1990.

Craig, H., and L. I. Gordon, Deuterium and oxygen 18 variations in the ocean and the marine atmosphere, in *Stable Isotopes in Oceanographic Studies and Paleotemperatures: Third Spoleto Conference*, edited by E. Tongiorgi, pp. 9–130, V. Lischi and Figli, Spoleto, Italy, 1965.

Crosta, X., J.-J. Pichon, and L. H. Burckle, Reappraisal of Antarctic seasonal sea ice at the Last Glacial Maximum, *Geophys. Res. Lett.*, *25*, 2703–2706, 1998.

Curry, W. B., and G. P. Lohmann, Carbon isotopic changes in benthic foraminifera from the western South Atlantic: Reconstructions of glacial abyssal circulation patterns, *Quat. Res.*, *18*, 218–235, 1982.

Curry, W. B., J.-C. Duplessy, L. D. Labeyrie, and N. J. Shackleton, Changes in the distribution of δ^{13}C of deep water ΣCO_2 between the

last glaciation and the Holocene, *Paleoceanography*, *3*, 317–341, 1988.

De La Rocha, C. L., M. A. Brzezinski, M. J. DeNiro, and A. Shemesh, Silicon-isotope composition of diatoms as an indicator of past oceanic change, *Nature*, *395*, 680–683, 1998.

Donner, B., and G. Wefer, Flux and stable isotope composition of *Neogloboquadrina pachyderma* and other planktonic foraminifers in the Southern Ocean, *Deep Sea Res., Part I*, *41*, 1733–1744, 1994.

Duplessy, J.-C., N. J. Shackleton, R. K. Matthews, W. Prell, W. F. Ruddiman, M. Caralp, and C. H. Hendy, ^{13}C record of benthic foraminifera in the last interglacial ocean: Implications for the carbon cycle and the global deep water circulation, *Quat. Res.*, *21*, 225–243, 1984.

Duplessy, J.-C., N. J. Shackleton, R. G. Fairbanks, L. Labeyrie, D. Oppo, and N. Kallel, Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation, *Paleoceanography*, *3*, 343–360, 1988.

Duplessy, J.-C., L. Labeyrie, M. Paterne, S. Hovine, T. Fichefet, J. Duprat, and M. Labracherie, High latitude deep water sources during the last glacial maximum and the intensity of the global oceanic circulation, in *The South Atlantic: Present and Past Circulation*, edited by W. G. Wefer et al., pp. 445–460, Springer-Verlag, New York, 1996.

Ennever, F. K., and M. B. McElroy, Changes in atmospheric CO_2: Factors regulating the glacial to interglacial transition, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present, Geophys. Monogr. Ser.*, vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 154–162, AGU, Washington, D. C., 1985.

Fairbanks, R. G., A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, *Nature*, *342*, 637–642, 1989.

Franck, V. M., M. A. Brzezinski, K. H. Coale, and D. M. Nelson, Iron and silicic acid concentrations regulate Si uptake north and south of the Polar Front Zone in the Pacific sector of the Southern Ocean, *Deep Sea Res., Part II*, *47*, 3315–3338, 2000.

Francois, R.M. A. Altabet, E.-F. Yu, D. M. Sigman, M. Bacon, M. Frank, G. Bohrmann, G. Bareille, and L. D. Labeyrie, Contribution of Southern Ocean surface-water stratification to low atmospheric CO_2 concentrations during the last glacial period, *Nature*, *389*, 929–935, 1997.

Gille, S. T., Mean sea surface height of the Antarctic Circumpolar Current from Geosat data: Method and application, *J. Geophys. Res.*, *99*, 18,255–18,273, 1994.

Gille, S. T., The Southern Ocean momentum balance: Evidence for topographic effects from numerical model output and altimeter data, *J. Phys. Oceanogr.*, *27*, 2219–2232, 1997.

Gordon, A. L., Seasonality of Southern Ocean sea ice, *J. Geophys. Res.*, *86*, 4193–4197, 1981.

Gordon, A. L., E. Molinelli, and T. Baker, Large-scale relative dynamic topography of the Southern Ocean, *J. Geophys. Res.*, *83*, 3023–3032, 1978.

Graham, D. W., B. H. Corliss, M. L. Bender, and L. D. Keigwin, Carbon and oxygen isotopic disequilibria of recent deep-sea benthic foraminifera, *Mar. Micropaleontol.*, *6*, 483–497, 1981.

Grobe, H., and A. Mackensen, Late Quaternary climatic cycles as recorded in sediments from the Antarctic continental margin, in *The Antarctic Paleoenvironments: A Perspective on Global Change, Antarct. Res. Ser.*, vol. 60, edited by J. P. Kennett and D. A. Warnke, pp. 349–276, AGU, Washington, D. C., 1992.

Grobe, H., A. Mackensen, H.-W. Hubberten, V. Spiess, and D. K. Futterer, Stable isotope record and late Quaternary sedimentation rates at the Antarctic continental margin, in *Geological History of Polar Oceans: Arctic Versus Antarctic*, edited by U. Bleil and J. Thiede, pp. 539–572, Kluwer Acad., Norwell, Mass., 1990.

Hays, J. D., J. A. Lozano, N. J. Shackleton, and G. Irving, Reconstruction of the Atlantic and western Indian Ocean sectors of the 18,000 B.P. Antarctic Ocean, in *Investigation of Late Quaternary Paleoceanography and Paleoclimatology*, edited by R. M. Cline and J D. Hays, *Mem. Geol. Soc. Am.*, *145*, 337–372, 1976.

Herguera, J. C., E. Jansen, and W. H. Berger, Evidence for a bathayal front at 2000 m depth in the glacial Pacific, based on a depth transect on Ontong Java Plateau, *Paleoceanography*, *7*, 273–288, 1992.

Hodell, D. A., Late Pleistocene paleoceanography of the South Atlantic sector of the Southern Ocean: Ocean Drilling Program Hole 704A, *Paleoceanography*, *8*, 47–67, 1993.

Hodell, D. A., C. D. Charles, and U. S. Ninnemann, Comparison of interglacial stages in the South Atlantic sector of the Southern Ocean for the past 450 kyr: Implications for Marine Isotope Stage (MIS) 11, *Global Planet. Change*, *24*, 7–26, 2000.

Hughes, C. W., and P. D. Killworth, Effects of bottom topography in the large-scale circulation of the Southern Ocean, *J. Phys. Oceanogr.*, *25*, 2485–2497, 1995.

Hutchins, D. A., and K. W. Bruland, Iron-limited diatom growth and Si:N uptake ratios in coastal upwelling regime, *Nature*, *393*, 561–564, 1998.

Keigwin, L. D., and E. A. Boyle, Late Quaternary paleochemistry of high-latitude surface waters, *Paleogeogr. Paleoclimatol. Paleoecol.*, *73*, 85–106, 1989.

Kim, S.-T., and J. R. O'Neil, Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates, *Geochem. Cosmochim. Acta*, *61*, 3461–3475, 1997.

Klinck, J. M., and D. A. Smith, Effect of wind changes during the Last Glacial Maximum on the circulation in the Southern Ocean, *Paleoceanography*, *8*, 427–433, 1993.

Kohfeld, K. E., Geochemistry and ecology of polar planktonic foraminifera, and applications to paleoceanographic reconstructions, Ph.D. thesis, Columbia Univ., New York, 1998.

Kumar, N., R. F. Anderson, R. A. Mortlock, P. N. Froelich, P. Kubik, B. Dittrich-Hannen, and M. Suter, Increased biological productivity and export production in the glacial Southern Ocean, *Nature*, *378*, 675–680, 1995.

Labeyrie, L. D., J.-C. Duplessy, and P. L. Blanc, Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years, *Nature*, *327*, 477–482, 1987.

Levitus, S., and T. P. Boyer, *World Ocean Atlas 1994*, vol. 4, *Temperature*, Natl. Oceanic and Atmos. Admin., Silver Spring, Md., 1994.

Levitus, S., R. Burgett, and T. P. Boyer, *World Ocean Atlas 1994*, vol. 3, *Salinity*, Natl. Oceanic and Atmos. Admin., Silver Spring, Md., 1994.

Lutjeharms, J. R. E., N. M. Walters, and B. R. Allanson, Oceanic frontal systems and biological enhancement, in *Antarctic Nutrient Cycles and Food Webs*, edited by W. R. Siegfried et al., pp. 11–21, Springer-Verlag, New York, 1985.

Lynch-Stieglitz, J., W. B. Curry, and N. C. Slowey, A geostrophic transport estimate for the Florida Current from the oxygen isotope composition of benthic foraminifera, *Paleoceanography*, *14*, 360–373, 1999.

Mackensen, A., H. Grobe, H.-W. Hubberten, V. Spiess, and D. K. Futterer, Stable isotope stratigraphy from Antarctic continental margin during the last one million years, *Mar. Geol.*, *87*, 315–321, 1989.

Mackensen, A., H.-W. Hubberten, T. Bickert, G. Fischer, and D. K. Futterer, The δ^{13}C in benthic foraminiferal tests of *Fontbotia wuellerstorfi* (Schwager) relative to δ^{13}C of dissolved inorganic carbon in southern ocean deepwater: Implications for glacial ocean circulation models, *Paleoceanography*, *8*, 587–610, 1993.

Mackensen, A., H. Grobe, H.-W. Hubberten, and G. Kuhn, Benthic foraminiferal assemblages and the δ^{13}C signal in the Atlantic sector of the Southern Ocean: Glacial-to-interglacial contrasts, in *Carbon Cycling in the Glacial Ocean: Constraints on the Ocean's Role in Global Change, NATO ASI Ser., Ser. I*, vol. 17, edited by R. Zahn et al., pp. 105–144, North Atl. Treaty Organ., Brussels, Belgium, 1994.

Marshall, D., Topographic steering of the Antarctic Circumpolar Current, *J. Phys. Oceanogr.*, *25*, 1636–1650, 1995.

Martin, J. H., and S. E. Fitzwater, Iron deficiency limits phytoplankton growth in the north-east Pacific subarctic, *Nature*, *331*, 341–343, 1988.

Martinson, D. G., Ocean heat and seasonal sea ice thickness in the Southern Ocean, in *Ice in the Climate System*, edited by W. R. Peltier, pp. 597–609, Springer-Verlag, New York, 1993.

Matsumoto, K., and J. Lynch-Stieglitz, Similar glacial and Holocene deep water circulation inferred from southeast Pacific benthic foraminiferal carbon isotope composition, *Paleoceanography*, *14*, 149–163, 1999.

Moore, J. K., M. R. Abbot, and J. G. Richman, Location and dynamics of the Antarctic Polar Front from satellite sea surface temperature data, *J. Geophys. Res.*, *104*, 3059–3073, 1999.

Niebler, H. S., Reconstruction of paleo-environmental parameters using stable isotopes and faunal assemblages of planktonic foraminifera in the Southern Atlantic Ocean, *Ber. Polarforschung*, *167*, 1–198, 1995.

Ninnemann, U. S., and C. D. Charles, Regional differences in Quaternary subantarctic nutrient cycling: Link to intermediate and deep water ventilation, *Paleoceanography*, *12*, 560–567, 1997.

O'Neil, J. R., R. N. Clayton, and T. K. Mayeda, Oxygen isotope fractionation in divalent metal carbonates, *J. Chem. Phys.*, *51*, 5547–5558, 1969.

Oppo, D. W., and M. Horowitz, Glacial deep water geometry: South Atlantic benthic foraminiferal Cd/Ca and δ^{13}C evidence, *Paleoceanography*, *15*, 147–160, 2000.

Orsi, A. H., T. I. Whitworth, and W. D. J. Nowlin, On the meridional extent and fronts of the Antarctic Circumpolar Current, *Deep Sea Res., Part I*, *42*, 641–673, 1995.

Petit, J. R., et al., Climate and atmospheric history of the past 420,000 years from Vostok ice core, Antarctica, *Nature*, *399*, 429–436, 1999.

Sarmiento, J. L., T. M. C. Hughes, R. J. Stouffer, and S. Manabe, Simulated response of the ocean carbon cycle to anthropogenic climate warming, *Nature*, *393*, 245–249, 1998.

Sarnthein, M., K. Winn, J.-C. Duplessy, and M. R. Fontugne, Global variations of surface ocean productivity in low and mid latitudes: Influence on CO_2 reservoirs of the deep ocean and atmosphere during the last 21,000 years, *Paleoceanography, 3*, 361–399, 1988.

Sarnthein, M., K. Winn, S. J. A. Jung, J.-C. Duplessy, L. Labeyrie, H. Erlenkeuser, and G. Ganssen, Changes in east Atlantic deepwater circulation over the last 30,000 years: Eight time slice reconstructions, *Paleoceanography, 9*, 209–267, 1994.

Schrag, D. P., and D. J. DePaolo, Determination of $\delta^{18}O$ of seawater in the deep ocean during the Last Glacial Maximum, *Paleoceanography, 8*, 1–6, 1993.

Schrag, D. P., G. Hampt, and D. W. Murray, Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean, *Science, 272*, 1930–1932, 1996.

Shackleton, N. J., Attainment of isotopic equilibrium between ocean water and the benthonic foraminiferal genus *Uvigerina*: Isotopic changes in the ocean during the last glacial, *Colloq. Int. C. N. R. S., 219*, 203–219, 1974.

Shackleton, N. J., and N. D. Opdyke, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale, *Quat. Res., 3*, 39–55, 1973.

Sigman, D. M., M. A. Altabet, D. C. McCorkle, R. Francois, and G. Fischer, The $\delta^{15}N$ of nitrate in the Southern Ocean: Consumption of nitrate in surface waters, *Global Biogeochem. Cycles, 13*, 1149–1166, 1999.

Sigmon, D. E., D. M. Nelson, and M. A. Brze-zinski, The Si cycle in the Pacific sector of the Southern Ocean: Seasonal diatom production in the surface layer and export to the deep sea, *Deep Sea Res., Part II*, in press, 2001.

Stute, M., M. Forster, H. Frischkorn, A. Serejo, J. F. Clark, P. Schlosser, W. S. Broecker, and G. Bonani, Cooling of tropical Brazil (5°C) during the Last Glacial Maximum, *Science, 269*, 379–383, 1995.

Suman, D. O., and M. P. Bacon, Variations in Holocene sedimentation in the North American Basin determined by Th-230 measurements, *Deep Sea Res., 36*, 869–878, 1989.

Takeda, S., Influence of iron availability of nutrient consumption ratio of diatoms in oceanic waters, *Nature, 393*, 774–777, 1998.

Toggweiler, J. R., and J. L. Sarmiento, Glacial to interglacial changes in atmospheric carbon dioxide: The critical role of ocean surface water in high latitudes, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present, Geophys. Monogr. Ser.*, vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 163–184, AGU, Washington, D. C., 1985.

Verardo, S., Late Pleistocene sea-surface temperatures in the Agulhas Current region, Ph.D. thesis, City Univ. of New York, New York, 1995.

Wefer, G., E. Suess, W. Balzar, G. Liebezeit, P. J. Mueller, C. A. Ungerer, and W. Zenk, Fluxes of biogenic components from sediment trap deployment in circumpolar waters of the Drake Passage, *Nature, 299*, 145–157, 1982.

Weiss, R. F., W. S. Broecker, H. Craig, and D. Spencer, *GEOSECS Indian Ocean Expedition*, vol. 5, *Hydrographic Data 1977–1978*, 48 pp., Natl. Sci. Found., Washington, D. C., 1983.

Wenk, T., and U. Siegenthaler, The high-latitude ocean as a control of atmospheric CO_2, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present, Geophys. Monogr. Ser.*, vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 154–162, AGU, Washington, D. C., 1985.

Woodruff, F., S. M. Savin, and R. G. Douglas, Biological fractionation of oxygen and carbon isotopes by recent benthic foraminifera, *Mar. Micropaleontol., 5*, 3–11, 1980.

Woodruff, S. D., R. J. Slutz, R. L. Jenne, and P. M. Steurer, A comprehensive ocean-atmosphere data set, *Bull. Am. Meteorol. Soc., 68*, 1239–1250, 1987.

———————————

R. Anderson and J. Lynch-Stieglitz, Lamont-Doherty Earth Observatory and Department of Earth and Environmental Sciences, Columbia University, Route 9W, Palisades, NY 10964, USA. (boba@ldeo.columbia.edu; jean@ldeo.columbia.edu)

K. Matsumoto, Program in Atmospheric and Oceanic Sciences, Princeton University, P.O. Box CN710, Sayre Hall, Princeton, NJ 08544–0710, USA. (kmatsumo@splash.princeton.edu)