Geochemistry Geophysics Geosystems G3
AN ELECTRONIC JOURNAL OF THE EARTH SCIENCES
Published by AGU and the Geochemical Society

Article
Volume 7, Number 10
3 October 2006
Q10N03, doi:10.1029/2005GC001226
ISSN: 1525-2027

Meridional overturning circulation in the South Atlantic at the last glacial maximum

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[1] The geostrophic shear associated with the meridional overturning circulation is reflected in the difference in density between the eastern and western margins of the ocean basin. Here we examine how the density difference across 30°S in the upper 2 km of the Atlantic Ocean (and thus the magnitude of the shear associated with the overturning circulation) has changed between the last glacial maximum and the present. We use oxygen isotope measurements on benthic foraminifera to reconstruct density. Today, the density in upper and intermediate waters along the eastern margin in the South Atlantic is greater than along the western margin, reflecting the vertical shear associated with the northward flow of surface and intermediate waters and the southward flowing North Atlantic Deep Waters below. The greater density along the eastern margin is reflected in the higher δ^{18}O values for surface sediment benthic foraminifera than those found on the western margin for the upper 2 km. For the last glacial maximum the available data indicate that the eastern margin foraminifera had similar δ^{18}O to those on the western margin between 1 and 2 km and that the gradient was reversed relative to today with the higher δ^{18}O values in the western margin benthic foraminifera above 1 km. If this reversal in benthic foraminifera δ^{18}O gradient reflects a reversal in seawater density gradient, these data are not consistent with a vigorous but shallower overturning cell in which surface waters entering the Atlantic basin are balanced by the southward export of Glacial North Atlantic Intermediate Water.

Components: 6680 words, 6 figures, 1 table.

Keywords: Last glacial maximum; South Atlantic; meridional overturning circulation; oxygen isotopes; benthic foraminifera.

Index Terms: 4532 Oceanography: Physical: General circulation (1218, 1222); 4926 Paleoceanography: Glacial; 4962 Paleoceanography: Thermohaline.

Received 16 December 2005; **Revised** 25 May 2006; **Accepted** 23 June 2006; **Published** 3 October 2006.

Lynch-Stieglitz, J., W. B. Curry, D. W. Oppo, U. S. Ninneman, C. D. Charles, and J. Munson (2006), Meridional overturning circulation in the South Atlantic at the last glacial maximum, *Geochem. Geophys. Geosyst.*, *7*, Q10N03, doi:10.1029/2005GC001226.

Theme: Past Ocean Circulation
Guest Editors: Jean Lynch-Stieglitz, Catherine Kissel, and Olivier Marchal

1. Introduction

[2] Today there is a relatively strong meridional overturning circulation in the Atlantic Ocean basin, reflecting the export of North Atlantic Deep water and the compensating inflow of surface and intermediate waters. Most data-based reconstructions [e.g., *Ganachaud and Wunsch*, 2000; *Roemmich and Wunsch*, 1985; *Schmitz*, 1995; *Schmitz and McCartney*, 1993; *Talley et al.*, 2003] of this overturning cell show that the strength of the shallow inflow of thermocline and intermediate waters is relatively constant throughout the subtropics and tropics from the tip of Africa in the south (30°S) to just south of the polar seas in the north (Figure 1). The crossover between northward flowing upper waters and southward flowing North Atlantic Deepwater occurs at approximately 1 km in both the North and South Atlantic. It is still unclear whether a similar surface to deep overturning cell in the Atlantic, along with the associated northward transport of heat, existed during the last glacial maximum (LGM).

[3] Deepwater tracer distributions inferred from chemical and isotopic measurements on the tests of benthic foraminifera suggest a strong stratification in the North Atlantic Ocean, with a low nutrient, high δ^{13}C water mass (often called Glacial North Atlantic Intermediate Water, GNAIW) occupying depths down to about 2 km depth, and a high nutrient, low δ^{13}C water mass underneath [e.g., *Boyle and Keigwin*, 1987; *Curry and Oppo*, 2005; *Duplessy et al.*, 1988; *Oppo and Lehman*, 1993; *Sarnthein et al.*, 1994]. These data are generally interpreted as supporting a shallower overturning circulation in the glacial Atlantic, similar to that observed in some models of LGM ocean circulation. *Curry and Oppo* [2005] argue that a vigorous circulation (with respect to the strength of the vertical mixing) is required in order to maintain the sharp property gradient between the shallower and deeper water masses and the large bathymetric gradient in δ^{13}C below 2 km.

[4] The deep water below 2 km had higher nutrient concentrations than today, and almost certainly contained predominantly waters with a southern (Antarctic) source [e.g., *Marchitto et al.*, 2002]. However, it is clear from many tracers that there was also a contribution of low nutrient waters from the north to the deep (below 2 km) Atlantic. *Boyle and Keigwin* [1982] reconstructed Cd in the deep North Atlantic and showed that these waters were more nutrient depleted than average ocean values, arguing against a complete shut-down in North Atlantic deepwater production. More recently, *Rickaby et al.* [2000] find that reconstructed glacial Cd concentrations were lower in the western basin than the eastern basin in the North Atlantic below 2.5 km water depth, supporting a North Atlantic source of deep waters. The reconstructed seawater δ^{13}C are also higher in the North Atlantic than in the South Atlantic during the LGM, suggesting continued contribution of deep waters from the north [e.g., *Broecker*, 2002; *Matsumoto and Lynch-Stieglitz*, 1999]. *Lea and Boyle* [1990] found that the barium/calcium ratios in benthic foraminifera demand continued contributions of low nutrient waters from the north as well. Radiocarbon reconstructions for the deep North Atlantic show older ventilation ages than today [*Broecker et al.*, 1990; *Keigwin*, 2004; *Keigwin and Schlegel*, 2002], but still younger than waters in the South Atlantic [*Goldstein et al.*, 2001] suggesting a continued contribution of high radiocarbon water to deep waters in the North Atlantic.

[5] While the processes controlling the formation rates and properties of the upper (GNAIW) and lower deep waters in the North Atlantic were probably different for each water mass and are not well understood at this time, it is clear that the renewal of these water masses did not cease entirely during the last ice age (perhaps except

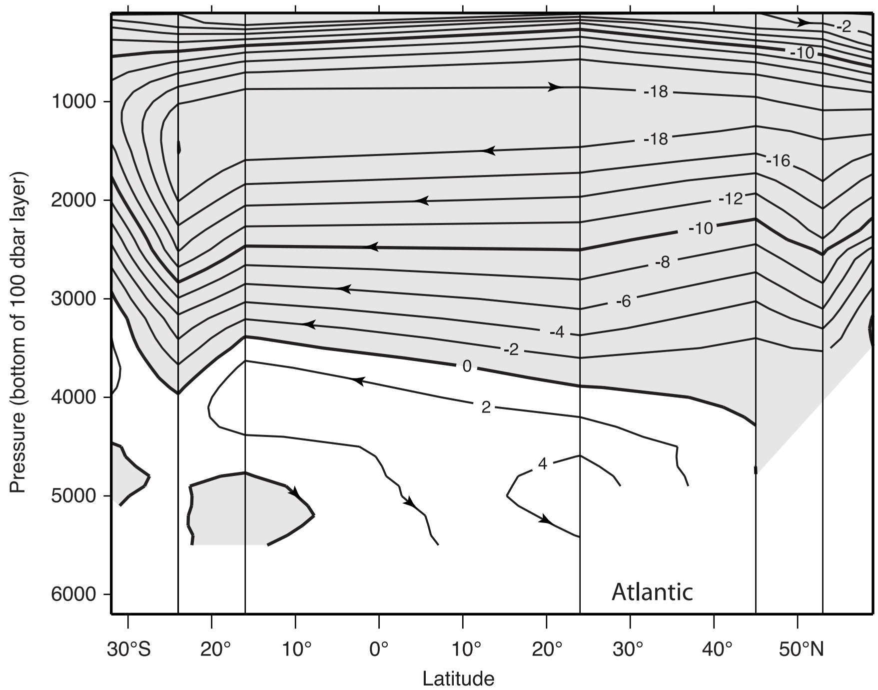


Figure 1. A data-based meridional overturning stream function (Sv = 10^6 m^3 s^{-1}) for the modern Atlantic from *Talley et al.* [2003]. Note that there is little change in the strength of the overturning from north to south within the Atlantic Basin.

during Heinrich event 1 [*McManus et al.*, 2004]). The waters above 2 km may have even been renewed faster than they are today. Today, deep-water formation involves the cooling and sinking of surface waters in the North Atlantic, and is therefore associated with a strong northward transport of heat. It is not clear whether the LGM Atlantic had an overturning cell involving the northward transport of surface waters (and thus heat), or whether the ventilation was accomplished differently.

[6] *Yu et al.* [1996] find the Pa-231/Th-230 ratio in the LGM sediments in the Atlantic is lower than the production ratio, implying relatively short residence time for waters in the LGM Atlantic. *McManus et al.* [2004] come to a similar conclusion based on a high-resolution Pa/Th record from the North Atlantic. This data could reflect the continued presence of an overturning cell involving the northward transport of surface waters, but a short residence time could also be accomplished by rapid flushing of deep or intermediate waters either from the north or the south. *Lynch-Stieglitz et al.* [1999b] find a reduced vertical shear in the geostrophic velocity in the Florida Straits, consistent with a weaker surface branch of the overturning circulation. This does not contradict the idea of continued production of deep and intermediate water masses in the North Atlantic during the LGM, but does suggest that either the overturning cell was weaker, or that the upper and lower deep water masses were formed by mechanisms which did not draw large quantities of surface water northward through the Florida Straits.

[7] In this study, we attempt to directly measure the strength of the meridional overturning cell at 30°S in the LGM Atlantic by reconstructing the cross-basin density gradient associated with the geostrophic shear in the overturning circulation.

2. Methods and Materials

[8] The large-scale surface-to-deep meridional overturning circulation is reflected in a density contrast across the upper North Atlantic (denser waters on the western margin for a given water depth), and a similar density contrast across the upper South Atlantic (denser waters on the eastern margin for a given water depth) (Figures 2 and 3). Specifically, the vertical shear in the geostrophic transport integrated across the basin is proportional to the east-west density difference at any given depth, $[(f\rho_0)/g]\partial_z(L_x\overline{v}) = \rho_e - \rho_w$, where f is the

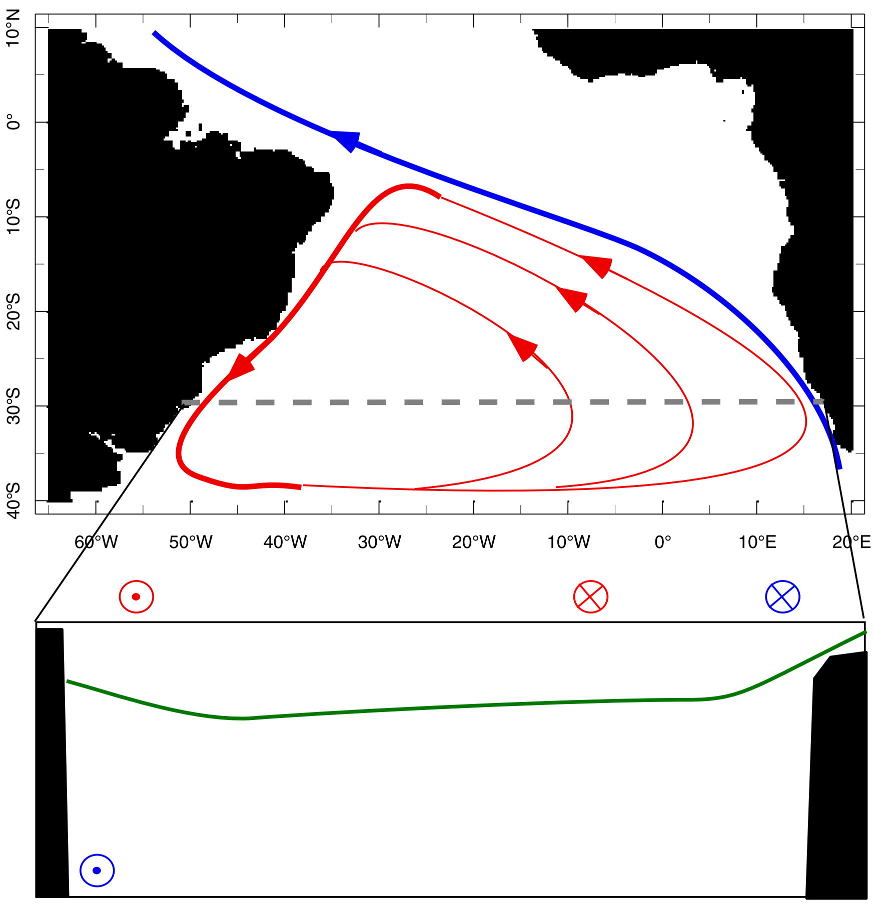


Figure 2. Schematic diagram showing the wind-driven (red) and the surface branch of the overturning circulation (blue) which result in a net tilt of the thermocline (green line) across the South Atlantic at 30°S. A cross indicates flow into the page, and a dot indicates flow out of the page. At 30°S, where the net Ekman transport is small, the time-averaged wind-driven circulation in the upper ocean is in geostrophic balance. Over long timescales, the wind-driven circulation is closed in the upper ocean (the amount of water moving southward in the western boundary current is balanced by the northward flow in the interior). In a simple two-layer ocean, the wind-driven circulation is not associated with a net tilt in the thermocline. However, there is a net tilt of the thermocline across the ocean basin which reflects the net northward flow of surface waters which are the surface branch of the meridional overturning circulation.

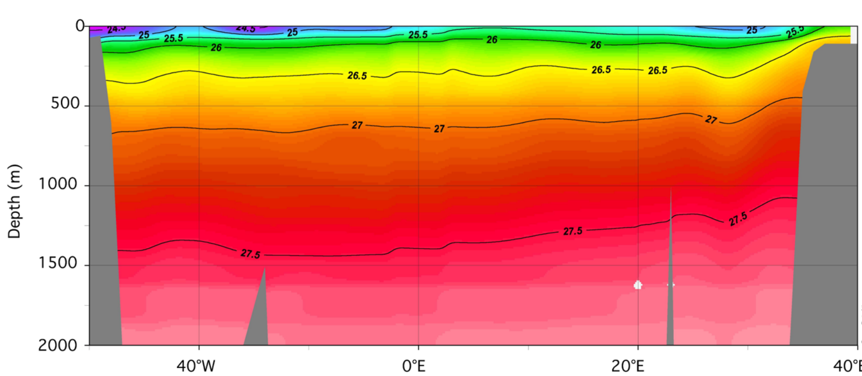


Figure 3. A section of potential density (referenced to the surface) across the South Atlantic at 30°S [*Olbers et al.*, 1992; *Schlitzer*, 2006] which illustrates the east-west density contrast (or tilt in the thermocline in the simple view of Figure 2), associated with the meridional overturning circulation.

Coriolis parameter, ρ is the density of seawater, L_x is the width of the basin and $\bar{v}$ is the average velocity across the basin [*Marotzke et al.*, 1999]. The geostrophic shear associated with the subsurface branch of the shallower wind-driven overturning cells [*McCreary and Lu*, 1994] is also reflected in the density contrast across the ocean basins [*Lynch-Stieglitz*, 2001; *Veronis*, 1981]. The shear associated with the shallow wind-driven cells disappears where the average wind stress over the ocean basin is zero (generally around 30°N and 30°S where the zonal wind direction reverses). For this reason, if we are primarily interested in the shear associated with the large scale surface to deep overturning, it is best to monitor the density contrast near 30°N or 30°S. However, as one moves from north to south within the Atlantic basin the changes in east-west density contrast, which reflect the changing geostrophic shear and Coriolis parameter, are reflected almost exclusively by changes in density along the western margin [*Lynch-Stieglitz*, 2001]. The flows near the eastern boundary of the Atlantic are weak, and are in geostrophic balance. The requirement that water not flow into or out of the ocean margin then requires that the density not change in a north-south direction along the eastern boundary [*Veronis*, 1973].

[9] We can use the δ^{18}O from the calcite tests of benthic foraminifera preserved in ocean sediments to estimate density on the ocean margins because both the δ^{18}O of calcite and density increase as a result of increasing salinity or decreasing temperature [*Lynch-Stieglitz et al.*, 1999a]. For times in the geologic past, our ability to reconstruct density from the δ^{18}O of foraminiferal calcite is most limited by our knowledge of the relationships between the δ^{18}O of seawater and salinity, as well as the relationship between temperature and salinity.

[10] Here we use oxygen isotope measurements on benthic foraminifera to examine the density contrast across the South Atlantic at 27°S for the Holocene and last glacial maximum (LGM). We focus on sediment cores above 2 km water depth because the subtle Holocene gradient below 2 km cannot be resolved using the foraminifera data. In addition, the relationship between δ^{18}O of seawater and salinity is different in the deep sea than in the main thermocline due to the impact of sea ice formation on the δ^{18}O and salinity of these coldest deep waters. It would be difficult to assess how this relationship may have changed in the past.

[11] A depth transect of sediment cores on the western margin between 26–28°S off Brazil were collected and oxygen isotope measurements on benthic foraminifera were reported by *Curry and Oppo* [2005]. The *Curry and Oppo* [2005] isotope measurements were made on *Cibicidoides* and *Planulina* species which have been shown to accurately record seawater δ^{18}O and calcification temperature [*Duplessy et al.*, 2002; *Lynch-Stieglitz et al.*, 1999a], including *C. wuellerstorfi*, *C. kullenbergi*, *C. pachyderma*, *C. floridanus*, and *P. ariminensis*. These data were supplemented by oxygen isotope data on *C. lobatulus* from an additional piston core (RC12-279) at 35.35°S for western margin vertical transect over a water depth of 441 to >2000 meters. We use the published Holocene and LGM time slices which were chosen on the basis of benthic and planktonic oxygen isotope stratigraphy as well as radiocarbon dates on selected cores [*Curry and Oppo*, 2005]. Because changes in the cross basin density contrast must be accommodated along the western margin of the ocean, a reduced shear in the overturning circulation will, all else being equal, be associated with denser waters (higher δ^{18}O in foraminifera) along the western margin. Because the LGM values in the *Curry and Oppo* [2005] transect were chosen to be the δ^{18}O maximum, it is possible that these values reflect periods of low overturning within the LGM rather than average LGM values. However, most of the oxygen isotope maxima appear to cover a broad period of time, suggesting a relatively constant profile during the LGM.

[12] Because the density does not change much in the north-south direction along the eastern margin, we can use cores from a wider latitude range to reconstruct the vertical density structure at the corresponding latitude (27°S) along the eastern (African) margin. We have investigated all of the piston cores from the Lamont core collection in the depth range 200–1000 m between the equator and the southern tip of Africa. While many of these cores are unsuitable (no Holocene to LGM section within the core, insufficient foraminifera, etc.), there were a number of cores which were identified as covering the transition between LGM and present on the basis of planktonic and/or benthic foraminifera oxygen isotope stratigraphy (Table 1, Figure 4). Most measurements on benthic foraminifera were, like the western margin cores, from specimens of the genera *Planulina* or *Cibicidoides*. For two of the cores (V19-248, V19-249) that had very low abundances of *Cibicidoides* and *Planulina*, we also analyzed *Bolivina*, applying an offset of

Table 1. South Atlantic Sediment Cores Contributing to the Profiles Shown in Figure 5

Core	Latitude, °N	Longitude, °E	Depth, m	Holocene $\delta^{18}O$	n	sd	Glacial $\delta^{18}O$	n	sd	Species	Source
V19-236	−33.9	17.6	280	1.75	2	0.12	3.07	6	0.13	*C. pachyderma, P. wuellerstorfi*	this study
ODP 1078	−11.9	13.4	426	1.44	32	0.07	2.91	21	0.10	*B. dilatata*	*Rühlemann et al.* [2004]
V12-70	−6.5	11.4	450	1.68	1		2.83	4	0.10	*P. ariminensis*	this study
V19-257	−21.0	12.4	651	2.17	11	0.05				*C. pachyderma*	this study
V29-140	−3.1	9.3	719	2.26	4	0.04	3.48	1		*C. pachyderma*	this study
ODP 1079	−11.9	13.3	755	1.98	16	0.08	3.20	4	0.08	*C. pachyderma*, *Planulina*	this study
V16-51	−33.5	17.0	898	2.47	5	0.11	3.47	2	0.21	*C. pachyderma, P. wuellerstorfi*	this study
V19-258	−20.4	11.6	965	2.37	14	0.10				*C. pachyderma, P. wuellerstorfi*	this study
							3.67	4	0.05	*Bolivina*	this study
BT4	−4.3	10.4	1000	2.57	1		3.83	1		*Cib. sp.*	*Curry et al.* [1988]
MG-237	−5.2	11.3	1000	2.70	12	0.09	3.90	4	0.08	*Cib. sp*	*Sarnthein et al.* [1994]
V19-259	−19.9	11.0	1170	2.53	10	0.08	3.39	5	0.12	*C. pachyderma, P. wuellerstorfi*	this study
							3.93	6	0.03	*Bolivina*	
ODP 1087	−31.5	15.3	1372	2.49	3	0.04	3.84	4	0.19	*C. wuellerstorfi*	*Pierre et al.* [2001]
GeoB 1711-4	−23.3	12.4	1967	2.50	11	0.09	4.50	9	0.08	*P. wuellerstorfi*	*Kirst et al.* [1999]; *Little et al.* [1997]
ODP 1084	−25.5	13.0	1992	2.71	6	0.06	4.56	3	0.03	*P. wuellerstorfi*	this study

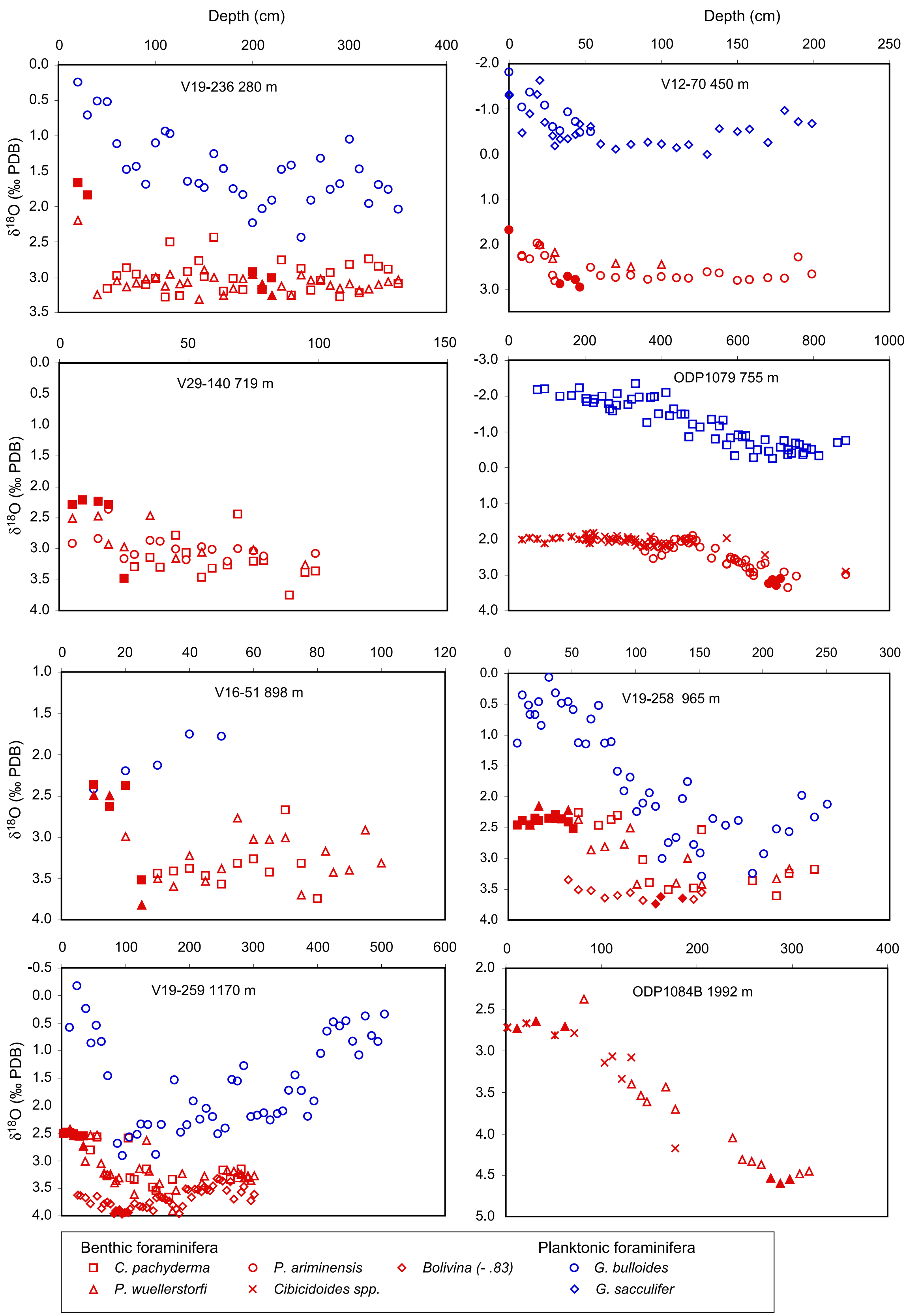


Figure 4. Previously unpublished down core oxygen isotope data from planktonic (blue) and benthic (red) foraminifera. Darkened symbols indicate data that contributed to the Holocene and glacial averages shown in Figure 5 and Table 1.

0.83‰ for comparison with the other benthic foraminifera measurements [*Herguera et al.*, 1992; *McCorkle et al.*, 1997].

[13] All oxygen isotope measurements from the eastern margin cores in this study except for ODP 1079 were made at Lamont-Doherty Earth Observatory on a Micromass Optima with Multiprep individual acid bath carbonate preparation device. Data were calibrated to PDB using NBS-19, NBS-18 and an in-house standard. Sample intercalibration yields no systematic differences between measurements made on this system and those for the western margin which were made primarily at Woods Hole Oceanographic Institution [*Curry and Oppo*, 2005]. Measurements on ODP 1079 were made at Scripps Institution of Oceanography on a Finnigan MAT252 equipped with a common acid bath carbonate preparation device were calibrated to PDB using NBS-19 and an in-house coralline carbonate standard. Replicate measurements at selected depths in ODP 1079 were made at Woods Hole Oceanographic Institution, and were found to be 0.18‰ heavier than those made at Scripps. A correction of 0.18‰ was applied to the data from ODP 1079 for better consistency with the data measured at WHOI and LDEO. The new eastern margin data were supplemented with published benthic δ^{18}O records from the literature to form the eastern margin profile (Table 1).

[14] Average profiles for the Holocene and LGM benthic δ^{18}O for the eastern margin were constructed. The time intervals were chosen on the basis of the planktonic and benthic δ^{18}O records (Figure 4). By choosing the maximum δ^{18}O values for the eastern margin cores, we will reconstruct the maximum possible LGM density gradient across the basin. If, in fact, during the LGM the benthic δ^{18}O on the eastern margin were lower, our reconstruction would overestimate the magnitude of the shear in the overturning circulation. Similarly, if there was a divergence in values between the species, we chose to use the species which gave the most positive δ^{18}O values which also will produce a maximum reconstruction of the density gradient and shear in the overturning circulation.

3. Results and Discussion

[15] On the African margin sedimentation rates were low and benthic foraminiferal abundances showed a dramatic shift from a Holocene dominance of *Planulina* and *Cibicidoides* to a LGM dominance of *Bolivina* and other benthic foraminifera common in areas of high overlying productivity and low oxygen concentration. The combination of abundance changes, low sedimentation rates, bioturbation, and the possibility of down slope transport on the continental margins leads to down core records which show significant differences depending on the species analyzed, and large amounts of scatter when species become rare (Figure 4).

[16] However, the Holocene profiles do show consistently higher δ^{18}O values on the eastern margin than on the western margin in the upper 1 km, consistent with the values predicted from modern hydrographic data (Figure 5). For the LGM, despite our efforts to choose the most positive δ^{18}O values for the eastern margin profile, the eastern margin profile shows δ^{18}O that is either the same as, or even lower than, the δ^{18}O in the western margin profile.

[17] While the relationship between δ^{18}O, T and S in the South Atlantic today associates higher δ^{18}O in the foraminifera with waters of higher density, was this necessarily true in the past? Presuming a stable water column, the fact that there is a general increase in δ^{18}O of the benthic foraminifera with depth for both profiles suggests that this was indeed the case. Could a change in relationship between δ^{18}O in foraminifera and density account for a collapse in the cross-basin δ^{18}O gradient with no reduction in cross basin density gradient? Because the density structure in the main thermocline is dominated by temperature, a complete flattening of the δ^{18}O/Salinity relationship (no change in δ^{18}O for a large change in salinity) due to a very isotopically heavy fresh end-member (from sea ice formation for example) would yield a reduction in the cross basin δ^{18}O of the benthic foraminifera at 750 m water depth from 0.3‰ to 0.25‰. There is no reason to suspect that the main thermocline would not be temperature dominated during the LGM, especially given the continued presence of strong sea surface temperature gradients in the subpolar North and South Atlantic where the main thermocline ventilation occurs. This assumption could be checked using an independent proxy for temperature in these profiles such as the Mg/Ca or Sr/Ca ratios in benthic foraminifera [*Lear et al.*, 2002; *Rosenthal et al.*, 2006].

[18] Today the T-S-δ^{18}O relationship in the main thermocline is relatively constant across the narrow South Atlantic basin due to the presence of mixing along isopycnals and the lack a proximal water

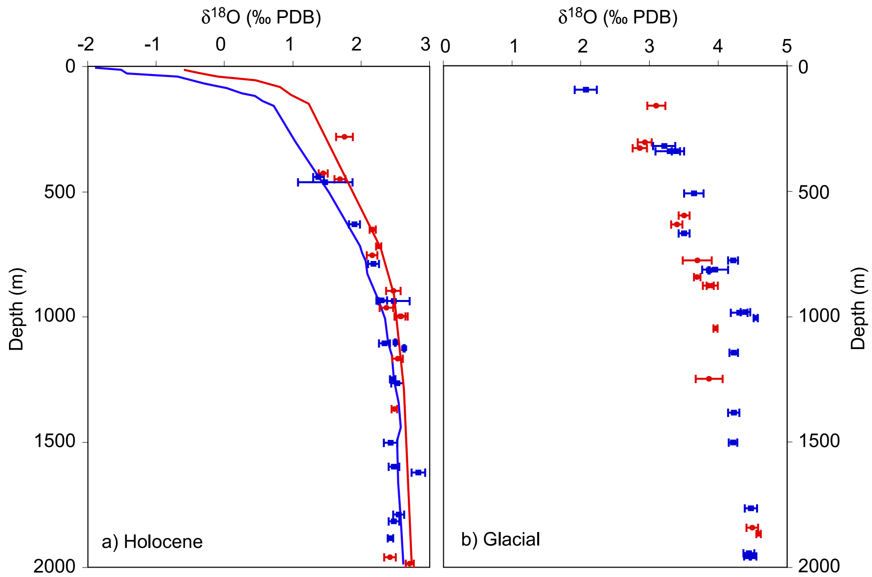
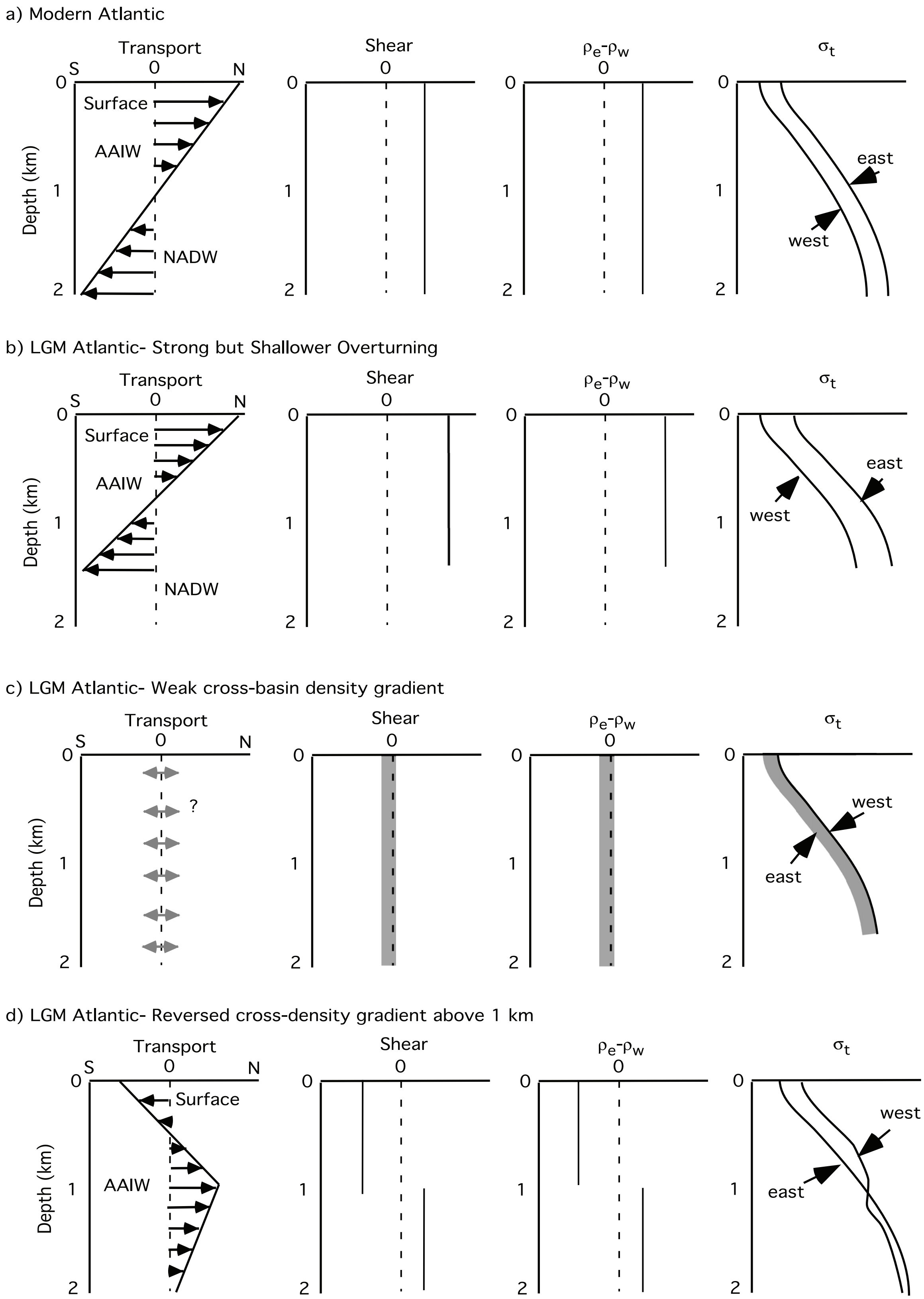


Figure 5. South Atlantic Ocean margin δ^{18}O from benthic foraminifera for (a) the Holocene and (b) the last glacial maximum. Data from the eastern (African) margin are in red, and the data from the western (Brazilian) margin are in blue. The error bars are the 1-sigma standard deviation of the analyses contributing to the average for each core for the time period in question (Table 1). Where only one analysis was used, an error bar of 0.08 (typical of the 1 sigma standard deviation of replicate analyses of analyses of δ^{18}O in carbonates) was used. The modern predicted δ^{18}O for benthic foraminifera at 30°S are indicated by the solid lines using the relationship between T and δ^{18}O for *Cibicidoides* species of *Lynch-Stieglitz et al.* [1999a].

mass source. However, if there were a strong source of intermediate water production on one side of the South Atlantic, this could decouple the T-S-δ^{18}O properties from one side to the next. Depending on the source of the water masses on either side, it is possible in this scenario that differences in the δ^{18}O of seawater from one side to the next could lead to a reduced cross basin gradient in the δ^{18}O of foraminifera in the presence of a cross basin density gradient similar to today. G. A. Gebbie and P. Huybers (Meridional circulation during the Last Glacial Maximum explored through a combination of South Atlantic δ^{18}O observations and a geostrophic inverse model, submitted to *Geochemistry, Geophysics, Geosystems*, 2006; hereinafter referred to as Gebbie and Huybers, submitted manuscript, 2006) demonstrate that some very different temperature and salinity combinations could produce glacial δ^{18}O values along the eastern margin that are lower despite being from waters of higher density. However, the required horizontal gradients in temperature and salinity are much larger than today's, and could be checked with an independent paleo-temperature proxy. Clearly further constraints on the T-S-δ^{18}O relationship in the LGM ocean from pore waters and paleo-temperature proxies will be quite useful in better interpreting the foraminifera data.

[19] In the discussion that follows we will presume that the LGM δ^{18}O data represents a collapse or reversal in the cross-basin density gradient. A northward shift of the wind systems could reduce the northward geostrophic transport in the upper 500 meters or so, which would reduce the magnitude of the shear in the meridional flow. However, all else being equal, this reduction would not be enough to change the sense of the shear associated with the meridional overturning circulation and could not account for the entire reduction in density contrast especially at deeper levels [*Lynch-Stieglitz*, 2001]. The elimination or reversal of the cross basin density gradient would seem to require a different sort of circulation altogether, one in which warm surface waters do not enter the Atlantic basin to compensate the export of cold deep waters (NADW) or intermediate waters (GNAIW).

[20] Simply shoaling the NADW overturning cell during the glacial (Glacial North Atlantic Intermediate Water) would tend to increase the shear and the density gradient between the eastern and western margins (Figure 6b), in contrast with our observations. Even if the flow were somewhat weaker, the compression to shallower depths would tend to increase the shear and thus the



a) Modern Atlantic
Transport
S
0
N
Surface
AAIW
NADW
Depth (km)
0
1
2
Shear
ρ_e-ρ_w
σ_t
east
west
b) LGM Atlantic- Strong but Shallower Overturning
c) LGM Atlantic- Weak cross-basin density gradient
?
d) LGM Atlantic- Reversed cross-density gradient above 1 km

Figure 6

density gradient across the basin. If the shear associated with the upper limb of NADW were compressed from 0–3 km depth to 0–2 km depth (a 50% increase) the density gradient, which is directly proportional to the shear should also increase by 50%. Some general circulation and intermediate complexity models show a strong, shallow overturning which is closed within the Atlantic basin for the LGM, implying that the return circulation is not derived from waters outside of the Atlantic. If this were the case, we might not see a density gradient associated with the maximum in the overturning this far south. Many of these low-resolution models also show many of the streamlines of the modern NADW overturning cell rising from the deep ocean into the warm surface waters. This behavior in the relatively low resolution general circulation and intermediate complexity models is due to fact that they are more diffusive than the real ocean, which shows relatively little mixing across isopycnals and the associated upwelling of deep waters at low latitudes (Figure 1). If the closed deep overturning cells observed in these low-resolution models do not accurately reflect the real ocean circulation, we do not necessarily expect that the closed shallow overturning cells simulated for glacial conditions are realistic either.

[21] On the other hand, it is difficult to envision that GNAIW could gain their high $\delta^{13}C$, low nutrient signal without a significant contribution from warm surface waters. Perhaps while today the large-scale MOC in the Atlantic is not closed at 30° (surface and intermediate waters are imported into the Atlantic and deep waters are exported), a shallower GNAIW cell was closed to the north of 30°S. This could happen if the wind systems and oceanographic regimes in the Southern Ocean shifted northward. However, subtropical fauna existed at these latitudes during the LGM as they do today, and the reconstructed sea surface temperature patterns do not suggest a migration of the Subtropical Front northward of 30°S [*Niebler et al.*, 2003]. Even if there were a GNAIW circulation which was closed within the Atlantic Basin, the weaker flow through the Florida Straits during the LGM suggests that it would have probably been considerably weaker than the overturning associated with NADW today [*Lynch-Stieglitz et al.*, 1999b]. In short, we find the ocean margin $\delta^{18}O$ data presented here very difficult to reconcile with the inferences based on other proxies of a shallower version of today's Atlantic overturning circulation during the LGM.

[22] While a collapse in the density gradient in the upper 2 km of the South Atlantic is consistent with a scenario in which there is no surface to deep overturning cell extending to the South Atlantic (Figure 6c), a reversal of the density gradient would require a circulation completely unlike the modern Atlantic. Some models with a completely collapsed overturning circulation show a vigorous northward flow of Antarctic Intermediate Water balanced by a southward flow of surface waters in the South Atlantic [e.g., *Weaver et al.*, 2003]. This would imply a shear in the meridional circulation that is consistent with the ocean margin $\delta^{18}O$ data that suggest a density reversal above 1 km (Figure 6d). A vigorous import of AAIW would, however, seem to require a shear in the opposite sense below the core of AAIW (1 km). This is not seen in the deeper data, but if the shear were relatively weak, it could be hidden within the errors and large scatter of the data.

[23] Can these scenarios without a surface to deep overturning extending to the South Atlantic be consistent with the Pa/Th data which support a weakened but significant flushing of the Atlantic deep waters during the LGM [*McManus et al.*, 2004; *Yu et al.*, 1996]? Are they consistent with the argument that the sharp nutrient boundary in the deep North Atlantic and the large bathymetric gradients in $\delta^{13}C$ require vigorous renewal of both water masses [*Curry and Oppo*, 2005]? We believe that it is possible to have a significant ventilation of

Figure 6. (a) Schematic illustration of the vertical structure of the net meridional transport (overturning circulation) across the South Atlantic, the shear in the overturning circulation, the east-west density contrast, and the density at the ocean margins. The shear in the overturning circulation is reflected in the higher density on the eastern margin relative to the western margin. (b) The same schematic for a meridional overturning circulation of equal strength, but compressed to shallower depth in the glacial Atlantic. In this case we would expect an even greater difference between the eastern and western margin density. (c) A schematic consistent with our LGM profiles, assuming little difference between the eastern and western density profiles. This implies little shear in the overturning circulation in the upper 2 km of the South Atlantic. (d) Another schematic consistent with our LGM profiles, assuming that the apparently higher density on the western margin is robust and reflects the shear associated with an inflow of AAIW.

deep waters in the North Atlantic during glacial times accomplished without an overturning cell involving the northward transport of surface waters. Our data allows the possibility that the high nutrient deep waters below 2 km in the glacial Atlantic could be well ventilated from the south (increased production of AABW). The shear associated with a vigorous AABW cell would be below 2 km and not be reflected in a density difference between the profiles above this depth. A more vigorous AABW cell would be consistent with the bathymetric changes in δ^{13}C in both the North and South Atlantic, which appear to decrease by 0.7 to 1‰ per km below 2 km [*Curry and Oppo*, 2005]. Bathymetric gradients like this are difficult to maintain without active advection because vertical mixing is so strong near the mid-Atlantic Ridge [*Ledwell et al.*, 2000; *Mauritzen et al.*, 2002; *Polzin et al.*, 1997]. As discussed in the introduction, deep waters formed in the North Atlantic would have contributed to this deeper water mass as well.

[24] It is also difficult, but not impossible, to envision vigorous export of GNAIW without a strong northward transport of surface waters. Perhaps this water mass is ventilated in the well developed cyclonic subpolar gyre north of the polar front (which lay significantly farther to the south during the LGM), where upwelled cold water is made more dense by cooling and/or sea-ice formation at the sea surface. This mechanism could also be associated with little shear in the upper 2 km at lower latitudes. Or, as discussed above, a GNAIW overturning cell was confined in the Atlantic basin, north of 30°S.

[25] Clearly we still have no scenario for the circulation of the glacial Atlantic that is completely consistent with our understanding of all of the existing data. However, since it is the temperature contrast between the northward moving warm surface waters and cold deep waters which is associated with the large amounts northward heat transport in the Atlantic Ocean, a severe reduction of shear in the upper 2 km of the glacial South Atlantic would imply a large reduction in northward ocean heat transport in the South Atlantic regardless of the specific scenario.

[26] The quality of the records do not give us a high degree of confidence in the LGM density profile along the eastern margin, and it is possible that better records on the eastern margin could yield higher δ^{18}O on the eastern margin during the LGM, consistent with a vigorous surface to deep overturning. It is also possible that local formation of intermediate water masses in the South Atlantic could lead to different relationships between δ^{18}O of foraminifera and density on the eastern and western margins, confounding the use of δ^{18}O to infer cross-basin density gradients (e.g., Gebbie and Huybers, submitted manuscript, 2006). It is also likely that there were large changes in the strength of the overturning on millennial timescales during glacial times and on the deglaciation, which we could never resolve with these data. For example our ''LGM'' profiles may include data for H1 and H2, times when deepwater circulation may have been dramatically different than during the LGM proper.

4. Conclusions

[27] We examined oxygen isotope data from benthic foraminifera on both sides of the South Atlantic, and find that the cross-basin gradient seen in the Holocene is absent or even reversed during the last glacial maximum. If the reduction in the cross basin gradient in the δ^{18}O of benthic foraminifera reflects a reduction in cross basin density, these data imply a reduction in the shear associated with the Atlantic meridional overturning circulation. We feel that the quality of the materials we analyzed here do not provide a definitive assessment of the cross basin density gradient during glacial times, but rather show the potential of this approach. Hopefully further work will serve to confirm/refute and quantify the scenario we outline here with well dated, high-resolution isotope records from the margins of the Atlantic Ocean, coupled with independent paleo-temperature estimates. However, it is very hard to reconcile the ocean margin δ^{18}O data we present in this paper with a strong overturning cell involving the northward transport of surface waters in the glacial Atlantic.

Acknowledgments

[28] We acknowledge the technical assistance of Pat Malone, Linda Baker, Heather Griffiths, Christa Farmer, and Jennifer Mayer. This work was supported by NSF award OCE-9984989/OCE-0428803 to J.L.-S., NSF award OCE-9986748 to D.W.O. and W.B.C., NSF OCE-0222111 to C.D.C., and SEGRF fellowship at LLNL to J.M. Sample material used in this project provided by the Lamont-Doherty Earth Observatory Deep-Sea Sample Repository. Support for the curation of the core collection is provided by the National Science Foundation through OCE-0350504. Comments from

Claire Waelbroeck and an anonymous reviewer improved this manuscript.

References

Boyle, E. A., and L. D. Keigwin (1982), Deep circulation of the North Atlantic over the last 200,000 years: Geochemical evidence, *Science*, *218*, 784–787.

Boyle, E. A., and L. D. Keigwin (1987), North Atlantic thermohaline circulation during the last 20,000 years linked to high latitude surface temperature, *Nature*, *330*, 35–40.

Broecker, W. S. (2002), Constraints on the glacial operation of the Atlantic Ocean's conveyor circulation, *Isr. J. Chem.*, *42*, 1–14.

Broecker, W. S., M. Klas, E. Clark, S. Trumbore, G. Bonani, W. Wolfli, and S. Ivy (1990), Accelerator mass-spectrometric radiocarbon measurements on foraminifera shells from deep-sea cores, *Radiocarbon*, *32*, 119–133.

Curry, W. B., and D. W. Oppo (2005), Glacial water mass geometry and the distribution of $\delta^{13}C$ of ΣCO_2 in the western Atlantic Ocean, *Paleoceanography*, *20*, PA1017, doi:10.1029/2004PA001021.

Curry, W. B., J.-C. Duplessy, L. D. Labeyrie, and N. J. Shackleton (1988), Changes in the distribution of $\delta^{13}C$ of deep water ΣCO_2 between the last glaciation and the Holocene, *Paleoceanography*, *3*, 317–341.

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

Duplessy, J. C., L. Labeyrie, and C. Waelbroeck (2002), Constraints on the ocean oxygen isotopic enrichment between the Last Glacial Maximum and the Holocene: Paleoceanographic implications, *Quat. Sci. Rev.*, *21*, 315–330.

Ganachaud, A., and C. Wunsch (2000), Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data, *Nature*, *408*, 453–457.

Goldstein, S. J., D. W. Lea, S. Chakraborty, M. Kashgarian, and M. T. Murrell (2001), Uranium-series and radiocarbon geochronology of deep-sea corals: Implications for Southern Ocean ventilation rates and the oceanic carbon cycle, *Earth Planet. Sci. Lett.*, *193*, 167–182.

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

Keigwin, L. D. (2004), Radiocarbon and stable isotope constraints on Last Glacial Maximum and Younger Dryas ventilation in the western North Atlantic, *Paleoceanography*, *19*, PA4012, doi:10.1029/2004PA001029.

Keigwin, L. D., and M. A. Schlegel (2002), Ocean ventilation and sedimentation since the glacial maximum at 3 km in the western North Atlantic, *Geochem. Geophys. Geosyst.*, *3*(6), 1034, doi:10.1029/2001GC000283.

Kirst, G. J., R. R. Schneider, P. J. Muller, I. von Storch, and G. Wefer (1999), Late Quaternary temperature variability in the Benguela Current System derived from alkenones, *Quat. Res.*, *52*, 92–103.

Lea, D. W., and E. A. Boyle (1990), A 210,000-year record of barium variability in the deep northwest Atlantic-Ocean, *Nature*, *347*, 269–272.

Lear, C. H., Y. Rosenthal, and N. Slowey (2002), Benthic foraminiferal Mg/Ca-paleothermometry: A revised core-top calibration, *Geochim. Cosmochim. Acta*, *66*, 3375–3387.

Ledwell, J. R., E. T. Montgomery, K. L. Polzin, L. C. St Laurent, R. W. Schmitt, and J. M. Toole (2000), Evidence for enhanced mixing over rough topography in the abyssal ocean, *Nature*, *403*, 179–182.

Little, M. G., R. R. Schneider, D. Kroon, B. Price, C. P. Summerhayes, and M. Segl (1997), Trade wind forcing of upwelling, seasonality, and Heinrich events as a response to sub-Milankovitch climate variability, *Paleoceanography*, *12*, 568–576.

Lynch-Stieglitz, J. (2001), Using ocean margin density to constrain ocean circulation and surface wind strength in the past, *Geochem. Geophys. Geosyst.*, *2*(12), doi:10.1029/2001GC000208.

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

Lynch-Stieglitz, J., W. B. Curry, and N. Slowey (1999b), Weaker Gulf Stream in the Florida Straits during the Last Glacial Maximum, *Nature*, *402*, 644–648.

Marchitto, T. M., D. W. Oppo, and W. B. Curry (2002), Paired benthic foraminiferal Cd/Ca and Zn/Ca evidence for a greatly increased presence of Southern Ocean Water in the glacial North Atlantic, *Paleoceanography*, *17*(3), 1038, doi:10.1029/2000PA000598.

Marotzke, J., R. Giering, K. Zhang, D. Stammer, C. Hill, and T. Lee (1999), Construction of the adjoint MIT ocean general circulation model and application to Atlantic heat transport sensitivity, *J. Geophys. Res.*, *104*, 29,529–29,547.

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

Mauritzen, C., K. L. Polzin, M. S. McCartney, R. C. Millard, and D. E. West-Mack (2002), Evidence in hydrography and density fine structure for enhanced vertical mixing over the Mid-Atlantic Ridge in the western Atlantic, *J. Geophys. Res.*, *107*(C10), 3147, doi:10.1029/2001JC001114.

McCorkle, D. C., B. H. Corliss, and C. A. Farnham (1997), Vertical distributions and stable isotopic compositions of live (stained) benthic foraminifera from the North Carolina and California continental margins, *Deep Sea Res., Part I*, *44*, 983–1024.

McCreary, J. P., and P. Lu (1994), Interaction between the subtropical and equatorial ocean circulations: The subtropical cell, *J. Phys. Oceanogr.*, *24*, 466–497.

McManus, J. F., R. Francois, J. M. Gherardi, L. D. Keigwin, and S. Brown-Leger (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, *428*, 834–837.

Niebler, H. S., H. W. Arz, B. Donner, S. Mulitza, J. Patzold, and G. Wefer (2003), Sea surface temperatures in the equatorial and South Atlantic Ocean during the last glacial maximum (23–19 ka), *Paleoceanography*, *18*(3), 1069, doi:10.1029/2003PA000902.

Olbers, D., V. Gouretski, G. Seiß, and J. Schröter (1992), *Hydrographic Atlas of the Southern Ocean*, vol. 17, 82 plates, Alfred Wegener Inst. for Polar and Mar. Res., Bremerhaven, Germany.

Oppo, D. W., and S. J. Lehman (1993), Mid-Depth Circulation of the Subpolar North-Atlantic During the Last Glacial Maximum, *Science*, *259*, 1148–1152.

Pierre, C., J. F. Saliege, M. J. Urrutiaguer, and J. Giraudeau (2001), Stable isotope record of the last 500 k.y. at Site 1087 (Southern Cape Basin) [online], *Proc. Ocean Drill. Program Sci. Results*, *175*. (Available at http://www-odp.tamu.edu/publications/175_SR/chap_12/chap_12.htm)

Polzin, K. L., J. M. Toole, J. R. Ledwell, and R. W. Schmitt (1997), Spatial variability of turbulent mixing in the abyssal ocean, *Science*, *276*, 93–96.

Rickaby, R. E. M., M. J. Greaves, and H. Elderfield (2000), Cd in planktonic and benthic foraminiferal shells determined by thermal ionisation mass spectrometry, *Geochim. Cosmochim. Acta*, *64*, 1229–1236.

Roemmich, D., and C. Wunsch (1985), Two transatlantic sections: Meridional circulation and heat flux in the subtropical North Atlantic Ocean, *Deep Sea Res.*, *32*, 619–664.

Rosenthal, Y., C. H. Lear, D. W. Oppo, and B. K. Linsley (2006), Temperature and carbonate ion effects on Mg/Ca and Sr/Ca ratios in benthic foraminifera: Aragonitic species Hoeglundina elegans, *Paleoceanography*, *21*, PA1007, doi:10.1029/2005PA001158.

Rühlemann, C., S. Mulitza, G. Lohmann, A. Paul, M. Prange, and G. Wefer (2004), Intermediate depth warming in the tropical Atlantic related to weakened thermohaline circulation: Combining paleoclimate data and modeling results for the last deglaciation, *Paleoceanography*, *19*, PA1025, doi:10.1029/2003PA000948.

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

Schlitzer, R. (2006), Ocean Data View, software package, Alfred Wegener Inst. for Polar and Mar. Res., Bremerhaven, Germany.

Schmitz, W. (1995), On the Interbasin-scale thermohaline circulation, *Rev. Geophys.*, *33*, 151–173.

Schmitz, W. J., and M. S. McCartney (1993), On the North Atlantic circulation, *Rev. Geophys.*, *31*, 29–49.

Talley, L. D., J. L. Reid, and P. E. Robbins (2003), Data-based meridional overturning streamfunctions for the global ocean, *J. Clim.*, *16*, 3213–3226.

Veronis, G. (1973), Model of the world ocean circulation: I. Wind-driven, two layer, *J. Mar. Res.*, *31*, 228–288.

Veronis, G. (1981), Dynamics of large-scale ocean circulation, in *Evolution of Physical Oceanography: Scientific Surveys in Honor of Henry Stommel*, edited by B. A. Warren and C. Wunsch, pp. 140–183, MIT Press, Cambridge, Mass.

Weaver, A. J., O. A. Saenko, P. U. Clark, and J. X. Mitrovica (2003), Meltwater pulse 1A from Antarctica as a trigger of the Bolling-Allerod warm interval, *Science*, *299*, 1709–1713.

Yu, E. F., R. Francois, and M. P. Bacon (1996), Similar rates of modern and last-glacial ocean thermohaline circulation inferred from radiochemical data, *Nature*, *379*, 689–694.