Abyssal Atlantic circulation during the Last Glacial Maximum:  
Constraining the ratio between transport and vertical mixing

D. Lund, J. Adkins, and R. Ferrari

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Abstract

The ocean’s role in regulating atmospheric carbon dioxide on glacial-interglacial timescales remains an unresolved issue in paleoclimatology. Reduced mixing between deep watermasses may have aided oceanic storage of atmospheric CO₂ during the Last Glacial Maximum (LGM) but data supporting this idea have remained elusive. The δ¹³C of benthic foraminifera indicate the Atlantic Ocean was more chemically stratified during the LGM, but the non-conservative nature of δ¹³C may complicate interpretation of the LGM signal. Here we use benthic foraminiferal δ¹⁸O as a conservative tracer to constrain the ratio of meridional transport to vertical diffusivity in the deep Atlantic. Our calculations suggest that the ratio was at least twice as large at the LGM. We speculate that the primary cause was reduced mixing between northern and southern component waters, associated with movement of this watermass boundary away from the zone of intense mixing near the seafloor. The shallower watermass boundary yields an order of magnitude increase in the volume of southern component water, suggesting its residence time may have increased substantially. Our analysis supports the idea that an expanded volume of Antarctic Bottom Water and limited vertical mixing enhanced the abyssal ocean’s ability to trap carbon during glacial times.
I. Introduction

At the Last Glacial Maximum (LGM; ~20,000 years ago), atmospheric carbon dioxide levels were 80-90 ppm lower than pre-industrial values (Neftel et al., 1982; Monnin et al., 2001). Because more than 90% of the combined oceanic, atmospheric and terrestrial carbon resides in the deep ocean, this reservoir is believed to play a primary role in regulating atmospheric CO₂ on glacial-interglacial time scales (Broecker, 1982; Sigman and Boyle, 2000). Since publication of the ‘Harvardton Bears’ box models, we have realized that nutrient utilization and overturning rate in Southern Ocean deep water are two of the key controls on atmospheric CO₂ levels (Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984). Toggweiler (1999) refined this view when he speculated that reduced mixing between deep watermasses formed in the Southern Ocean and the North Atlantic also helped sequester atmospheric CO₂ during glacial times. Vertical mixing, overturning rate, nutrient status, and volume of the glacial southern source deep water are all thought to be important controls in regulating pCO₂.

Deep ocean nutrient tracers δ¹³C and Cd/Ca provide a relatively clear picture of the water mass distribution in the LGM Atlantic. Benthic δ¹³C shows that the deep ocean was more isotopically stratified during the LGM (Figure 1) (Boyle and Keigwin, 1982; Curry and Lohmann, 1982; Duplessy et al., 1988; Kallel et al., 1988; Herguera et al., 1988; Keiwin, 2004; Curry and Oppo, 2005). The distribution of δ¹³C in the Atlantic has been explained by invoking changes in transport of Antarctic Bottom Water (AABW) relative to North Atlantic Deep Water (NADW) (Duplessy et al., 1988; Curry and Oppo,
However, tracers like $\delta^{13}C$ also respond to variations in water mass mixing, variations in surface air-sea fluxes, and non-conservative behavior. In this paper, we use $\delta^{18}O$ and $\delta^{13}C$ data in an attempt to better quantify the relative roles of transport, vertical mixing, and remineralization in the abyssal Atlantic during the LGM.

Efforts to reconstruct the deep circulation using $\delta^{13}C$ have been hindered by the fact that $\delta^{13}C$ is a non-conservative tracer – its distribution is affected not only by the ocean circulation but also the biological remineralization of organic matter. Organic carbon created by phytoplankton in the surface ocean is depleted in its $^{13}C/^{12}C$ ratio relative to dissolved inorganic carbon (DIC). As organic matter falls through the water column, remineralization at depth produces $^{13}C$-depleted DIC that is incorporated in the shells of benthic foraminifera. This process masks the effect of circulation and mixing on the $\delta^{13}C$ tracer field. A conservative tracer, i.e. one which is influenced only by transport and mixing, is needed if one is to attempt a reconstruction of the deep ocean circulation from tracer profiles.

The stable oxygen isotopic ratio of foraminifera ($\delta^{18}O_c$) varies as a function temperature and the $\delta^{18}O$ of seawater, which is directly related to salinity. Because both temperature and salinity are conservative tracers, so too is $\delta^{18}O_c$. In the upper ocean, $\delta^{18}O_c$ can be used as a proxy for seawater density because both parameters have a linear dependence on temperature and salinity (Lynch-Stieglitz et al., 1999). In the deep ocean, however, this relationship breaks down because seawater density varies non-linearly with temperature. Furthermore, the $\delta^{18}O$-seawater relationship can vary between watermasses
and on glacial-interglacial timescales (Zahn and Mix, 1991). Instead of estimating paleo-
densities for the abyss, here we simply exploit the conservative nature of $\delta^{18}O_c$. Once
$\delta^{18}O_c$ is set at the surface, the value is conserved by a water parcel as it moves into the
ocean interior; $\delta^{18}O_c$ can only be modified by mixing and transport. A minor caveat is
that in situ temperature recorded by foraminifera is not truly conservative because it can
be changed through compressive effects. However this effect is negligible for the present
study as we show below.

$\delta^{18}O$ evidence for an abyssal watermass boundary

The strong vertical gradient in North Atlantic Ocean benthic foraminiferal $\delta^{13}C$ is
a well-known feature of the LGM ocean. Less well-recognized is that LGM benthic $\delta^{18}O$
(we use $\delta^{18}O$ as shorthand for $\delta^{18}O_c$ for the remainder of the paper) profiles also have a
larger gradient with water depth than today (Figure 2). At the Brazil Margin (30ºS), the
change in *Cibicidoides* $\delta^{18}O$ between 1 and 3 km was approximately 0.3 ‰ larger during
the LGM (Figure 2a) (Curry and Oppo, 2005). The $\delta^{18}O$ gradient is most easily observed
in the difference between LGM and Holocene profiles (Figure 2b). The larger LGM
gradient also exists in $\delta^{18}O$ data from the Blake Ridge (30ºN), where it is observed in two
genera of benthic foraminifera, *Cibicidoides* and *Uvigerina* (Figure 2d). At each
location, the LGM-Holocene $\delta^{18}O$ pattern is driven by a larger increase in LGM $\delta^{18}O$
below ~ 2 km. Similar patterns have been observed in the Indian (Kallel et al., 1988) and
Pacific Oceans (Herguera et al., 1992) suggesting the LGM ‘step’ in $\delta^{18}O$ may have been
a global feature.
Diagnosing the transport to mixing ratio from tracer distributions

A detailed analysis of the processes that create the δ\textsuperscript{18}O profiles is necessary to identify what generates the observed differences in the LGM and modern data. The distributions of tracers in the ocean are set through a balance between advection (e.g., by the overturning circulation), air-sea fluxes at the surface, and mixing in the ocean interior. The overturning circulation of AABW in the Atlantic Ocean is associated with sinking of water through convection around Antarctica and a return flow at shallower levels. The properties of AABW are set by surface fluxes near Antarctica and mixing with overlying waters in the ocean interior. Air-sea fluxes have been used to estimate watermass transport in the present climate (Schlosser et al., 1991; Speer and Tziperman, 1992; Large and Nurser, 2001), but this type of data is lacking in the paleoclimate record. Instead, we infer the ratio between meridional transport and vertical diffusivity in the abyss from tracer distributions. AABW is uniquely suited to this approach because it does not outcrop at the surface north of Drake Passage and it is only modified through mixing with other watermasses.

Reconstructing the ocean circulation from tracer distributions is often necessary when tracer data are more readily available than velocity measurements. Using the inverse approach, Wunsch (1996) shows how to combine tracer measurements to constrain ocean circulation with least square techniques: the larger the set of measurements, the smaller the uncertainty on the inferred circulations and fluxes. Inversions with available LGM data from the Atlantic have been used to evaluate whether the modern circulation is the same as the past (Legrand and Wunsch, 1995;
Gebbie and Huybers, 2006; Marchal and Curry, 2008). For example, South Atlantic LGM $\delta^{18}$O data may be consistent with an upper ocean circulation similar to today or a reduction in vertical shear (Gebbie and Huybers, 2006). Here we acknowledge that the data are too sparse to uniquely constrain the LGM circulation field and we set a more modest goal. We instead evaluate which transport to vertical mixing ratio for AABW is consistent with the available $\delta^{18}$O data. We assume that the few available profiles are representative of an entire basin and evaluate the implications. This exercise is educational because it highlights the key factors that influence $\delta^{18}$O and $\delta^{13}$C profiles in the modern and LGM Atlantic. The question of whether our results apply to the entire Atlantic Ocean will have to wait until more stable isotopic profiles are available. A recent inverse calculation for the LGM Atlantic by Marchal and Curry (2008) focused on estimating the consistency of LGM data with the modern circulation. Our analysis differs from this work in two important ways. First, we only attempt to constrain the ratio of $\Psi/\kappa$ for the glacial southern source water mass, not the circulation of the entire basin. Second, we explicitly consider vertical mixing in our analysis whereas the previous work treated this term implicitly as an error in the tracer budget. Overall our goal is less ambitious than a full tracer inversion, but our data set is also more limited. We see these two approaches as complimentary in the goal to better understand the role of the deep ocean in glacial-interglacial climate change.

II. Methods

Tracer budget for $\delta^{18}$O
In this section, we provide an abbreviated description of the tracer budget for $\delta^{18}O$ and $\delta^{13}C$. The full derivation is given in the Appendix. Our tracer budget encompasses the volume of water below a surface of constant tracer concentration, as shown in Figure 3, departing from the traditional approach that considers volumes below a fixed depth. The advantage is that the net amount of tracer flowing into such a volume from the south can only be balanced by mixing of tracer across the upper surface. By construction, the mean circulation advects tracer into the volume from the south, but it cannot flux tracer across a contour. Consider the conservative tracer $\delta^{18}O$ (Figure 3). At the order of magnitude level, the transport of $\delta^{18}O$ from the south is given by the maximum of the zonally averaged overturning circulation, $\Psi \text{ (m}^3\text{ s}^{-1})$, multiplied by the difference in tracer at the top and bottom of our control volume, $\Delta \delta^{18}O \text{ (‰)}$. The transport of tracer into the South Atlantic is balanced by mixing with overlying waters giving:

$$2/3 \Psi \Delta \delta^{18}O = \kappa \delta^{18}O_z A$$  \hspace{1cm} (1)

where the right hand side represents the diffusion of tracer expressed as a turbulent diffusivity, $\kappa \text{ (m}^2\text{ s}^{-1})$, times the vertical tracer gradient at the upper boundary of AABW, $\delta^{18}O_z \text{ (‰ m}^{-1})$, and $A$ is the area across which mixing occurs \text{(m}^2). The scaling factor of 2/3 accounts for the fact that the top of our control volume corresponds to the maximum overturning streamfunction and therefore our budget does not include southward advection out of the Atlantic (see Appendix). It is useful to re-arrange the budget so that quantities that can be estimated from observations are on the right hand side and unknowns are on the left:

$$\Psi \kappa \approx \frac{3}{2} \delta^{18}O_z A$$  \hspace{1cm} (2)
In other words, we can describe the ratio of tracer transport ($\Psi$) to vertical mixing of tracer ($\kappa$) using vertical gradients in the $\delta^{18}O$ profiles, the vertical difference $\Delta\delta^{18}O$, and the area of the $\delta^{18}O$ iso-surface. By comparing the modern and LGM estimates of this ratio, we can evaluate the relative importance of transport to vertical mixing through time.

Our tracer budget approach is similar to W. Munk’s local balance between upwelling and mixing in his classic paper ‘Abyssal Recipes’ (Munk, 1966). We differ from this approach by balancing the total transport of tracer within a water mass against mixing with other waters. In essence, we swap local upwelling ($w$, m/sec) and mixing ($\kappa$, m$^2$/sec) for the integrated volume transport ($\Psi$, m$^3$/sec) and the mixing integrated over the whole upper 'lid' of the water mass ($A\kappa$, m$^4$/sec). This difference is crucial, because a local vertical balance rarely holds in the ocean.

Tracer budget for $\delta^{13}C$

The budget describing the distribution of $\delta^{13}C$ in the abyssal Atlantic (Figure 3) differs from that of $\delta^{18}O$ in that it has the additional contribution of a biological remineralization term, $J$ (%o s$^{-1}$), multiplied by the volume into which the remineralization occurs, $V$ (m$^3$):

$$2/3 \, \Psi \Delta \delta^{3}C + JV \approx \kappa \delta^{3}C z \, A$$  \hspace{1cm} (3)$$

where $\Delta\delta^{13}C$ is the difference in $\delta^{13}C$ between water entering the South Atlantic and $\delta^{13}C$ of the upper ‘lid’ of AABW, defined as the maximum in the overturning streamfunction. $\delta^{13}C z$ is the gradient in $\delta^{13}C$ at the upper boundary of AABW in the Atlantic. In this
equation, we place the sink terms on the left-hand side of the equation, and the source
term on the right. \( \Psi \) is positive and acts as a sink of \(^{13}\text{C}/^{12}\text{C} \) for the abyssal cell, because
incoming AABW has a low \(^{13}\text{C}/^{12}\text{C} \) ratio. \( J \) is also a sink, as organic matter is strongly
depleted in \(^{13}\text{C} \). The diffusive term is a source, since mixing with northern component
water adds dissolved inorganic carbon with a higher \(^{13}\text{C}/^{12}\text{C} \) ratio than AABW. Dividing
through by the diffusive flux and re-arranging the terms, we obtain:

\[
Pe \equiv \frac{2 \Psi \Delta \delta^{13}C}{3 \kappa \delta^{13}C_z A} = \frac{\delta^{18}O \Delta \delta^{13}C}{\Delta \delta^{18}O \delta^{13}C_z} = 1 - \frac{Jh}{\kappa \delta^{13}C_z} \quad (4)
\]

where \( h \) is the vertical extent of the volume considered, i.e. \( V=Ah \). The first term is a bulk
Peclet number, \( Pe \), the non-dimensional ratio of tracer transport to tracer diffusion. We
use a bulk \( Pe \) because we focus on the zonally integrated advection (i.e. transport) of
tracer. \( Pe \) can be estimated using \( \Psi/\kappa \) from the \( ^{18}\text{O} \) tracer budget and \( \Delta \delta^{13}C \) and \( \delta^{13}C_z \)
derived from the \( ^{13}\text{C} \) profiles. With substitution we see that \( Pe \) is only dependent on the
tracer distribution and not the volume or area. The non-conservative term \( J \) can only
make \( Pe \) smaller. In the following Results section, we evaluate modern and LGM \( ^{18}\text{O} \)
and \( ^{13}\text{C} \) profiles that sample southern source water. We also check that our tracer
balance for the modern ocean gives a value of \( \Psi \) that is consistent with other estimates.
In the Discussion we evaluate the \( \Psi/\kappa \) and \( Pe \) values implied by these results and
comment on their implications.

\section*{II. Results}

\subsection*{Modern \( ^{18}\text{O} \) budget}
Does the $\delta^{18}O$ tracer budget described above yield reasonable $\Psi/\kappa$ values for the modern AABW circulation in the Atlantic? To answer this question, we first need to outline the $\delta^{18}O$ domain for our analysis. Ideally, the $\delta^{18}O$ surface that defines our control volume should coincide with the upper boundary of AABW. Here we define the upper boundary for AABW as the neutral density surface $\gamma = 28.15$ kg/m$^3$, which ranges from 3000 to 4000 m water depth in the South Atlantic and corresponds to the maximum vertical salinity gradient (Figure 4). The $\gamma = 28.15$ kg/m$^3$ surface coincides with the maximum overturning streamfunction in the abyssal South Atlantic; the net flow of water below $\gamma = 28.15$ kg/m$^3$ is northward, while flow above $\gamma = 28.15$ kg/m$^3$ is primarily southward (Lumpkin and Speer, 2007). For consistency with the LGM budget, we set the southern edge of our domain at 27°S, the approximate latitude of the Brazil Margin cores (Curry and Oppo, 2005).

Today, AABW sits below the deepest core sites at Brazil Margin. As a result, our modern $\delta^{18}O$ budget must rely on estimates of the equilibrium $\delta^{18}O$ of calcite calculated from temperature and seawater $\delta^{18}O$ data. We created a dataset of gridded calcite $\delta^{18}O$ based on the $\delta^{18}O$ of sea water database (LeGrande and Schmidt, 2006), in-situ temperatures from the WOCE Global Hydrographic Climatology (Gouretski and Koltermann, 2004) and a paleotemperature equation (Kim and O'Neil, 1997) modified for an acid-fractionation factor of 1.0105. Sections of equilibrium $\delta^{18}O$ of calcite for the western and eastern Atlantic are shown in Figure 5. The Atlantic AABW endmember has $\delta^{18}O$ values $\geq 3.12$ ‰ while the NADW endmember has $\delta^{18}O$ values $\leq 3.00$ ‰.
Because $\delta^{18}O$ is conservative, the maximum vertical gradient in $\delta^{18}O (\delta^{18}O_z)$ should mark the approximate boundary between AABW and NADW.

We calculated $\delta^{18}O_z$ for all grid points north of 27°S using $\delta^{18}O$ estimates from immediately above and below a given $\delta^{18}O$ horizon, over a span of 500 m water depth. A plot of $\delta^{18}O_z$ across $\delta^{18}O$ surfaces from 3.00‰ to 3.20‰ is shown in Figure 6. The maximum in $\delta^{18}O_z$ occurs for values between 3.06‰ to 3.11‰. These $\delta^{18}O$ surfaces track very closely the $\gamma = 28.15$ kg/m$^3$ surface and provide an objective method to identify the upper boundary of AABW in the Atlantic from $\delta^{18}O$ profiles. Figure 6 also shows the vertical gradient in potential temperature and salinity across the $\delta^{18}O$ iso-surfaces, calculated using the WOCE Global Hydrographic Climatology (Gouretski and Koltermann, 2004). We take vertical differences across 250 m, the resolution of the climatology. The maximum vertical gradients in T and S correspond to the maximum in $\delta^{18}O_z$, confirming that $\delta^{18}O$ acts as a conservative tracer that tracks the upper boundary of AABW. We choose $\delta^{18}O = 3.08$‰ to define our modern control volume for AABW. This $\delta^{18}O$ surface has an average $\delta^{18}O_z$ of $(3.1\pm0.8) \times 10^{-4}$ ‰ m$^{-1}$ ($\pm1\sigma$) (Figure 7). $\delta^{18}O$ surfaces ranging from 3.06‰ to 3.11‰ have similar $\delta^{18}O_z$ values (Figure 6).

We calculated $\Delta \delta^{18}O$ for the modern by subtracting the $\delta^{18}O$ values nearest the seafloor at 27°S from the $\delta^{18}O$ value = 3.08 ‰ chosen for the upper boundary of our control volume. A cross-section of modern $\delta^{18}O$ at 27°S, the southern limit of our circulation cell, is shown in Figure 8. At this latitude, there are 21 separate 1-degree
longitude bins across the western Atlantic where the δ^{18}O value at the sea-floor exceeds 3.08‰. We calculated Δδ^{18}O for each bin separately, and then took the average to calculate a Δδ^{18}O of -0.09±0.01 ‰. If instead we focus on the intense flow along the western boundary where most of the tracer flux occurs, Δδ^{18}O = -0.15±0.01 ‰. To cover these two different options, we use an intermediate Δδ^{18}O value with an error that encompasses both (-0.12±0.04 ‰).

Using our estimates of δ^{18}O_2 and Δδ^{18}O and the tracer budget for δ^{18}O (equation 2), we estimate Ψ/κ = (1.8±0.8) x 10^{10} m for the modern δ^{18}O surface = 3.08‰ (Table 1). We obtain similar Ψ/κ estimates for δ^{18}O surfaces ranging from 3.06‰ to 3.11‰. Assuming a κ for the deep Atlantic of (3±1.5) x10^{-4} m^2s^{-1} (Ganachaud and Wunsch, 2000), then the tracer-derived AABW transport at 27°S is 5.5±3.7 Sv (1Sv = 1x10^6 m^3s^{-1}). We obtain a similar Ψ if we use an estimate of κ for the Brazil Basin at the approximate depth of the γ = 28.15 kg/m^3 surface [(3.3±2.4) x 10^{-4} m^2s^{-1}; Morris et al., 2001]. Our tracer-based estimates of Ψ agree with geostrophic estimates Ψ: 6.7±0.5 Sv at 31°S and 5.0±1.1 Sv at 24°S (Speer and Zenk, 1993). Thus, the δ^{18}O tracer budget provides a reasonable approximation of the modern Ψ for AABW in the Atlantic.

**LGM δ^{18}O budget**

The first step in outlining the LGM tracer budget involves defining the domain for AABW in the Atlantic. We use three vertical profiles of δ^{18}O for this task, one from the Brazil Margin (Curry and Oppo, 2005) and two from the Blake Ridge (Keigwin, 2004).
Similar to the modern, we can use $\delta^{18}O$ to identify the upper boundary of AABW at the LGM. *Cibicidoides* data from above 2000 m water depth at the Blake Ridge suggest that the northern component endmember had a $\delta^{18}O$ signature of $\leq 4.0$ per mil (Figure 2a). Data from below 2000 m water depth at the Brazil Margin indicate that the southern component endmember had a $\delta^{18}O$ signature of $\geq 4.5$ per mil (Figure 2c). Given that $\delta^{18}O$ is a conservative tracer, the maximum in $\delta^{18}O_z$ should mark the boundary between northern and southern component waters during the LGM.

Our estimates of $\delta^{18}O_z$ are based on a Monte Carlo approach. We first assign a 1σ calibration error of ±0.18‰ to each LGM data point to account for the scatter observed in regressions between coretop foraminiferal $\delta^{18}O$ and predicted $\delta^{18}O$ based on in-situ temperature and seawater $\delta^{18}O$ (Marchal and Curry, 2008). This is likely a conservative error given that field-based calibrations reflect not only vital effects, but also the use of non-modern coretops and spatial and temporal variability in water column properties. We generate 1000 surrogate vertical profiles where every $\delta^{18}O$ value is sampled from a Gaussian distribution with the observed mean value and a standard deviation of 0.18. For each profile, we calculate $\delta^{18}O_z$ by fitting a line to data spanning an 800 m water depth range, and vary the center point of the range from 1500 m to 2700 m. The depth spacing of deep cores at the Blake Ridge precludes use of a water depth range less than 800 m.

The maximum $\delta^{18}O_z$ occurs at $\delta^{18}O$ values of 4.4‰ for the *Cibicidoides sp.* data at both the Brazil Margin and the Blake Ridge, suggesting that the upper boundary of AABW in the Atlantic was defined by this value (Figure 9). A similar structure is
observed for the *Uvigerina* sp. data at Blake Margin, but with an offset in absolute $\delta^{18}O$
which is likely due to vital effects (Keigwin, 2004). Thus, the basic structure of the LGM
water column profiles is resilient to the large scatter induced by a ±0.18‰ calibration
error for $\delta^{18}O$. This result strongly suggests there was a watermass boundary near 2 km
water depth at the LGM, and that this boundary was defined by an $\delta^{18}O$ of 4.4±0.1‰.
The medians of each $\delta^{18}O_z$ distribution overlap at 1σ: (-0.9±0.2) x 10^{-3}, (-1.0±0.3) x 10^{-3},
and (-1.2±0.3) x 10^{-3}‰/m for Brazil *Cibicidoides*, Blake *Cibicidoides*, and Blake
*Uvigerina*, respectively. We use an average of all three (Figure 10, top) in our
calculation of $\Psi/\kappa$.

The difference in $\delta^{18}O$ between water entering the South Atlantic and the upper
boundary of AABW at 27ºS ($\Delta\delta^{18}O$) is the largest source of uncertainty in calculating
$\Psi/\kappa$ for the LGM. Although the upper value is well constrained, there are no $\delta^{18}O$ data
below 3000 m water depth at the Brazil Margin. To estimate $\delta^{18}O$ of AABW entering the
Atlantic, we extrapolate the linear trend in $\delta^{18}O$ from 2000-3000 m water depth to 4500
m, yielding a bottom value of 5.0±0.2‰. In the modern South Atlantic, $\delta^{18}O$ increases
linearly all the way to seafloor, supporting our assumption for the LGM. If instead
abyssal $\delta^{18}O$ was nearly constant below 3000 m ($\delta^{18}O \sim 4.6$ ‰), it increases our estimate
of $\Psi/\kappa$.

We calculated $\Delta\delta^{18}O$ as linear regressions between the mid-point of the maximum
$\delta^{18}O_z$ (i.e., the watermass boundary) and the bottom of the ocean at 4500 m for each of
the 1000 randomly generated profiles. The median $\Delta \delta^{18}\text{O}$ is -0.46±0.23 ‰ (Figure 10, middle panel), and the $\Delta \delta^{18}\text{O}/\delta^{18}\text{O}_z$ ratio peaks at 440±230 meters (Figure 10, bottom). The LGM $\Delta \delta^{18}\text{O}/\delta^{18}\text{O}_z$ ratio is statistically indistinguishable from the modern ratio (390±160 meters; Table 1). We plot $\Delta \delta^{18}\text{O}/\delta^{18}\text{O}_z$ instead of $\delta^{18}\text{O}_z/\Delta \delta^{18}\text{O}$ because division by $\Delta \delta^{18}\text{O}$ values close to zero creates infinitely large ratios.

Modern $\delta^{13}\text{C}$ budget

To evaluate $Pe$ for the modern Atlantic it is necessary to calculate $\delta^{13}\text{C}_z$ and $\Delta \delta^{13}\text{C}$. Our estimates of these parameters are derived from WOCE and Ocean-Atmosphere Carbon Exchange Study (OACES) $\delta^{13}\text{C}$ profiles (Figure 11). We estimate the budget for $\delta^{13}\text{C} = 0.70\%$ which approximately follows the $\delta^{18}\text{O} = 3.08\%$ surface. We calculated $\Delta \delta^{13}\text{C}$ at 30ºS using $\delta^{13}\text{C}$ of the iso-surface minus $\delta^{13}\text{C}$ near the seafloor. Because there were no water column $\delta^{13}\text{C}$ data below 4000 m near the Brazil site, data from 45ºS to 35ºS were used to estimate an abyssal $\delta^{13}\text{C}$ value of 0.46 ‰ (water depths > 4200 m). Together, these data yield a $\Delta \delta^{13}\text{C}$ of 0.24 ± 0.05 ‰ (Table 1).

We calculated the gradient in $\delta^{13}\text{C}_z (\delta^{13}\text{C}_z)$ across each iso-surface for the latitude bins presented in Figure 11, using data from 300 m above and below the designated $\delta^{13}\text{C}$ surface, similar to the water depth range used to calculate $\delta^{18}\text{O}_z$ for modern AABW. We estimate a $\delta^{13}\text{C}_z$ for the upper ‘lid’ of AABW of (4.8±1.0) x 10⁻⁴ ‰ m⁻¹ (Table 1). The $\delta^{13}\text{C}_z$ estimates are largest in the South Atlantic where the presence of nutrient-enriched (13C depleted) AABW is the most obvious. Our estimate of the
$\Delta \delta^{13}C/\delta^{13}C_2$ ratio (510±150 meters) is within one sigma error of the $\Delta \delta^{18}O/\delta^{18}O_2$ ratio for the modern Atlantic (390±160 meters).

LGM $\delta^{13}C$ budget

Our constraints for the LGM $\delta^{13}C$ budget come from the *Cibicidoides* $\delta^{13}C$ profiles at the Brazil Margin and Blake Ridge. As with the $\delta^{18}O$ data, we use a Monte Carlo technique to test the robustness of our tracer budget calculation to uncertainty in the $\delta^{13}C$ data. We estimate an RMS error of 0.14‰ for the regression between Holocene *Cibicidoides* $\delta^{13}C$ and $\delta^{13}C$ of dissolved inorganic carbon from Duplessy et al. (1984) (Figure 12). This calibration error is likely a maximum estimate, because: 1) the sediment cores used in the regression lack good chronological control, meaning that the regression between dissolved inorganic carbon $\delta^{13}C$ and benthic foraminiferal $\delta^{13}C$ includes error associated with the unknown time offset between sediment and hydrographic data, and 2) the hydrographic data used by Duplessy et al. (1984) were very sparse and required interpolation to match core locations.

Assuming a 0.14‰ uncertainty for individual $\delta^{13}C$ points, we randomly generate 1000 $\delta^{13}C$ profiles from these distributions. We calculate $\delta^{13}C$ by fitting a line to data spanning an 800 m water depth range, and vary the center point of the range from 1500 m to 2700 m. Figure 13 shows a 2-d histogram of the vertical $\delta^{13}C$ gradient ($\delta^{13}C_2$) and its corresponding $\delta^{13}C$ for these 1000 profiles. The largest gradient occurs at similar $\delta^{13}C$ values for each site, although the pdf is shifted towards lower $\delta^{13}C$ at the Brazil Margin. The two sites also have similar $\delta^{13}C_2$ values, but the pdf is shifted towards slightly higher
δ¹³C at Blake Ridge. The steeper tracer gradient at Blake Ridge may reflect the greater
influence of ¹³C-enriched northern component water at this latitude. To obtain the area
averaged δ¹³C for our budget, we averaged the δ¹³C estimates from each location. The
resulting values have a wide distribution with a median δ¹³C of 1.2 x 10⁻³‰ m⁻¹ (Figure
14, top panel).

The broad histogram of mean δ¹³C values can be combined with the Δδ¹³C data
from Brazil Margin (defined as before for Δδ¹⁸O) to calculate the Δδ¹³C / δ¹³C ratio. We
estimate a δ¹³C at 4500 m of -0.8±0.2‰, based on linear extrapolation of the δ¹³C trend
in shallower water. This value is in good agreement with δ¹³C for the abyssal southwest
Atlantic (Ninnemann and Charles, 2002). The median of the difference between δ¹³C at
the seafloor and the top of our control volume (Δδ¹³C) for the LGM is 1.2‰ (Figure 14,
middle panel). In combination with δ¹³Cz, this yields a median Δδ¹³C / δ¹³C of 930±250
m (Figure 14, bottom panel). This distribution overlaps with the distribution of Δδ¹⁸O
/ δ¹⁸Oz for the LGM (440±230 m; Figure 10).

III. Discussion

Modern vs. LGM Ψ/κ and Pe

Our estimates of δ¹⁸Oz and Δδ¹⁸O for the LGM yield a Δδ¹⁸O/δ¹⁸Oz distribution
that overlaps with modern values. However, to calculate Ψ/κ we must also account for
the larger surface area over which tracer diffused into and out of AABW during the
LGM. The maximum in δ¹⁸O at δ¹⁸O ~ 4.4‰ indicates the upper ‘lid’ of AABW spanned 27°S to at least 30°N and was located at a water depth of ~ 2000 m. The δ¹³C tracer field suggests that AABW reached 60°N in the Atlantic and encompassed both the western and eastern basins (Duplessy et al., 1988; Oppo and Lehman, 1993; Curry and Oppo, 2005). The area of the horizontal surface in the Atlantic Ocean from 27°S to 60°N along the 2000 m isobath is approximately 4.5 x 10¹³ m² (Table 2). If AABW reached to only 50°N during the LGM, its area would have been 4.3 x 10¹³ m², about 10% lower. By comparison, the area of the upper ‘lid’ of AABW in the modern Atlantic is (4.7±0.5) x 10¹² m². Therefore, the area of AABW was about an order of magnitude larger at the LGM. Accounting for the difference in surface area, we estimate that Ψ/κ was a factor of

8 larger at the LGM.

The considerable errors for both our Ψ/κ estimates make it difficult to precisely determine the change in Ψ/κ from LGM to modern. We can, however, estimate the minimum difference between the two climate states. The smallest possible Ψ/κ for the LGM is set by the largest possible Δδ¹⁸O/δ¹⁸Oz ratio, or 1200 m (Figure 10, bottom). Using this value yields a minimum Ψ/κ for the LGM of 5.7 x 10¹⁰ m. Because the minimum estimate of the LGM Ψ/κ is at least 2x the modern value, either the LGM transport was larger than today, the vertical mixing less, or some combination of the two. We explore the likelihood of these scenarios below. Given that the Δδ¹⁸O/δ¹⁸Oz histogram also includes values near zero, the upper limit for Ψ/κ during the LGM is unconstrained by the available data.
By adding $\delta^{13}C$ to the tracer budget we can assess how the larger $\Psi/\kappa$ ratio affected the bulk $Pe$ defined in equation (4). One of the long-standing issues in paleoceanography is whether the glacial $\delta^{13}C$ tracer field primarily reflects the biological remineralization of organic matter or the physical circulation of the ocean. In principle, we should be able to partition the influences of remineralization and circulation on $\delta^{13}C$ using the tracer budget. As outlined in the methods, we can estimate a Peclet number for $\delta^{13}C$ (the ratio of $\delta^{13}C$ transport to $\delta^{13}C$ diffusion) using parameters from the $\delta^{13}C$ profiles ($\Delta\delta^{13}C$ and $\delta^{13}C_z$) and $\Psi/\kappa$ derived from the $\delta^{18}O$ tracer budget. If $Pe = 1$, then the diffusive and transport fluxes balance, and the remineralization flux is negligible. When $Pe < 1$, the tracer displays non-conservative behavior because remineralization influences the budget. For example, a $Pe$ of 0.5 would indicate that ~50% of the diffusive flux is balanced by remineralization. $Pe$ values above 1 indicate the tracer inventory is out of balance.

Using our estimates of $\delta^{13}C_z$, $\Delta\delta^{13}C$, and the $\Psi/\kappa$ from the $\delta^{18}O$ budget, we calculate a $Pe$ of $1.3 \pm 0.7$ for modern AABW (Table 1). The advective flux of $\delta^{13}C$ appears to be largely balanced by the diffusive flux. If this is correct, $\delta^{13}C$ is essentially conservative and it can be used to independently estimate $\Psi/\kappa$. Our $\delta^{13}C$-based $\Psi/\kappa$ estimate of $1.4\pm0.4 \times 10^{10}$ m is very similar to the $\delta^{18}O$-derived value (Table 1). However, at one-sigma the minimum $Pe$ also suggests that remineralization may account for up to 40% of the $\delta^{13}C$ sink for the abyssal Atlantic. Our results also includes $Pe$ values > 1, which would indicate the tracer inventory is out of balance and that we have not accounted for all of the important fluxes.
For the LGM, our estimates of Δδ^{18}O/δ^{18}O_z and Δδ^{13}C/δ^{13}C_z yield a Pe of 2.1±1.3 (Table 2). Similar to the modern, Pe spans a range of values above and below 1. Assuming a Pe ~ 1, the δ^{13}C-derived Ψ/κ would be (7.3±2.5) x 10^{10} m, within error of the δ^{18}O-derived value of (16±9) x 10^{10} m. The agreement is not as good as for the modern, however. Overall, the LGM Pe estimates span over a higher range of values, suggesting the LGM budget may be imbalanced. Given that Pe values > 1 imply we have not fully accounted for all the fluxes of tracer, it is important to ask where there are likely sources of error in our analysis. In the next section we consider the uncertainties in the tracer balances.

Sources of Error in the Tracer Budgets

Along Isopycnal Mixing Into the Control Volumes

The full budget for δ^{18}O outlined in the Appendix includes contributions from mixing both across (κ) and along isopycnal surfaces (κ_{iso}). Our tracer budget includes only cross-isopycnal mixing in the estimate of the Ψ/κ ratio, an acceptable approximation if isolines of δ^{18}O are parallel to lines of constant density. The additional along isopycnal mixing that arises if the isosurfaces are not parallel is negligible only if:

\[ \kappa \gg \kappa_{iso} |s_c - s_\rho|^2 \]  

where \( s_c \) is the slope of the tracer surface (\( s_c = \) horizontal δ^{18}O gradient / vertical δ^{18}O gradient) and \( s_\rho \) is the slope of the density surface at the same point. The difference
is approximately equal to the along isopycnal tracer gradient divided by the vertical tracer gradient. In the modern Atlantic, we define the upper boundary of AABW using the neutral density surface $\gamma \sim 28.15 \text{ kg/m}^3$ or equivalently the $\delta^{18}O = 3.08\%_o$ surface. The median $\delta^{18}O_x$ across this surface is $-2.9 \times 10^{-4} \%_o \text{ m}^{-1}$ (Figure 15, top). We estimate $\delta^{18}O_y$ at each grid point using $\delta^{18}O$ values immediately to the north and south (given a grid spacing of $1^\circ$, $\delta^{18}O_y$ is calculated over $2^\circ$ latitude). The median along isopycnal gradient is $1.5 \times 10^{-8} \%_o \text{ m}^{-1}$ (Figure 15, bottom). Together these yield $|s_c - s_{\rho}| \approx 5 \times 10^{-5}$. Using eqn. 5 and typical diffusivities, $\kappa_{\rho} = O(1000) \text{ m}^2/\text{s}$ and $\kappa = O(10^{-4}) \text{ m}^2/\text{s}$ (Stammer, 1998; Ledwell et al., 1998), we find that the along-density diffusion of $\delta^{18}O$ is ~40 times smaller than the cross-density diffusion and can be neglected for the modern case.

Given that we have no direct constraints on the abyssal density field in the geologic past, we can only speculate about along isopycnal mixing of $\delta^{18}O$ and $\delta^{13}C$ at the LGM. We assume that isopyncals in the deep Atlantic were nearly horizontal north of $30^\circ$S, as they are today (Figure 4). Available stable isotope data for the LGM suggest the $\delta^{18}O$ isolines were nearly horizontal as well. The LGM boundary between southern and northern component waters occurred at $4.4\pm0.1\%_o$, or $2000\pm200 \text{ m}$ water depth at the Brazil Margin ($\sim30^\circ$S) and Blake Ridge ($\sim30^\circ$N). Today, $\delta^{18}O$ surfaces that bound the upper lid of AABW ($3.05\text{-}3.10\%_o$) vary by up to 1 km over the same latitude range (Figure 5). It is therefore plausible that both $s_c$ and $s_{\rho}$ were small at the LGM and along isopycnal diffusion of $\delta^{18}O$ played a negligible role in the tracer budget. For both today
and the LGM, along-isopycnal diffusion of $\delta^{18}\text{O}$ into AABW appears to be important only in the Southern Ocean, where AABW comes to the surface and tracer and density slopes become very steep. The Southern Ocean, however, is south of our tracer budget domain.

While along isopycnal mixing is not important for $\delta^{18}\text{O}$, it may be an issue for $\delta^{13}\text{C}$ because it has a different tracer field. Using $\delta^{13}\text{C}$ values at the approximate depths of the neutral density surface $\gamma \sim 28.15$ in the modern ocean, we estimate a $\delta^{13}\text{C}_z$ of $\sim 5 \times 10^{-4}$‰ m$^{-1}$ (Table 1) and an along-isopycnal $\delta^{13}\text{C}$ gradient of $\sim 5 \times 10^{-8}$‰ m$^{-1}$ (a 0.2‰ increase over 60° latitude; Figure 16). Together, this yields $|s_e - s_\rho| \approx 1 \times 10^{-4}$. These values suggest the along-isopycnal diffusion of $\delta^{13}\text{C}$ is $\sim 10\%$ of the diapycnal diffusion in the modern Atlantic. Therefore, along-isopycnal mixing seems to be a minor term in the modern $\delta^{13}\text{C}$ tracer budget, only slightly larger than for the $\delta^{18}\text{O}$ budget.

Along isopycnal mixing may play a more significant role in the budget for $\delta^{13}\text{C}$ at the LGM. The contrast in $\delta^{13}\text{C}$ between northern and southern component waters at the LGM (~2.5 per mil) was more than twice the $\delta^{18}\text{O}$ contrast (~1 per mil) (Figure 17). As a result, along isopycnal mixing likely played a more important role in the tracer budget for $\delta^{13}\text{C}$ than $\delta^{18}\text{O}$. Indeed, our estimate of $Pe$ for the LGM includes values greater than 1. If in reality $Pe$ was this large, it would indicate that we underestimated the $^{13}\text{C}$ flux into our control volume. A likely candidate for this missing flux is along isopycnal mixing.
**Other Possible Errors in the $\delta^{18}O$ Budget**

In the preceding discussion, we referred to $\delta^{18}O$ as a conservative tracer. Because benthic foraminiferal $\delta^{18}O$ reflects in-situ rather than potential temperature, it is not strictly conservative. The effect of compressive heating on the $\Psi/\kappa$ estimates is small, however. In the case of the modern AABW, the pressure effect makes water entering the South Atlantic at 4500 m about 0.1°C warmer than water leaving the South Atlantic at 3500 m. Given a $\delta^{18}O$-T relationship of 0.2 ‰ per °C, the implication is that the deep $\delta^{18}O$ value is reduced by about 0.02 ‰. If we adjust $\Delta\delta^{18}O$ for this effect, the modern $\Psi/\kappa$ estimate decreases by 11%. During the LGM, the correction is larger (0.05 ‰) because the upper boundary of southern component water was near 2000 m water depth. If we increase the LGM $\Delta\delta^{18}O$ by 0.05 ‰, $\Psi/\kappa$ decreases by 13%. While significant, the $\Delta\delta^{18}O$ adjustments do not change our result that $\Psi/\kappa$ was larger at the LGM.

Our LGM budget is based on three vertical $\delta^{18}O$ profiles, one from the Brazil Margin and two from the Blake Ridge. This is a small data set for analyzing the LGM southern source water mass, especially given that much more $\delta^{13}C$ data has been used to make tracer maps like Figure 1. We use these locations because they are the only LGM $\delta^{18}O$ profiles in the Atlantic which meet the following criteria: 1) the cores used to create each depth profile are from tightly constrained geographic regions, 2) the sedimentation rates in the cores are high (generally > 5 cm kyr$^{-1}$), 3) the core depths span most of the water depth range relevant to AABW, and 4) the stable isotope data used to create each profile was run within an individual laboratory.
Whether the data were run in a single laboratory is particularly important for creating high fidelity profiles of $\delta^{18}O$ vs. water depth. Inter-laboratory offsets of $\sim 0.3 \%$ for benthic $\delta^{18}O$ (Ostermann and Curry, 2000; Hodell et al., 2003) are similar to the total range of $\delta^{18}O$ variability in the abyssal Atlantic during the LGM. As a result, it is very difficult to make coherent maps of benthic $\delta^{18}O$ similar to those that exist for $\delta^{13}C$ (Zahn and Mix, 1991). As laboratory inter-calibration improves and additional depth transects from high sedimentation rate sites become available, we will be able to test whether the few available profiles provide a reliable picture of the abyssal Atlantic $\delta^{18}O$ tracer field at the LGM.

Comparing the $\delta^{13}C$ and $\delta^{18}O$ results

The discussion above makes two important points. First, the ratio of $\Psi/\kappa$ at the LGM was $\sim 8x$ larger than it is today. Second, the difference in $Pe$ values was not nearly as large, though uncertainties in the $\delta^{13}C$ budget make it difficult to quantify the ratio of $\delta^{13}C$ transport to tracer diffusion. Given this analysis, it is instructive to consider the LGM stable isotope profiles (Figure 17). At $27^\circ S$, $\delta^{18}O$ increases monotonically below the boundary between northern and southern source waters. This shape is to be expected if dense, cold, and salty waters are transported into the Brazil Basin at the LGM. By the time this watermass reaches the Blake Ridge in the Northern Hemisphere vertical diffusion has worked to homogenize the vertical $\delta^{18}O$ profile.
The structure of the $\delta^{13}C$ profiles is fundamentally different than those for $\delta^{18}O$. First, the slope of the $\delta^{13}C$ profiles is opposite that for the $\delta^{18}O$ profiles, because the southern endmember has lower $\delta^{13}C$ than the northern endmember. Second, the slope of the Blake and Brazil profiles is essentially identical (Figure 17). An extra flux of low $\delta^{13}C$ works against the tendency of vertical mixing to homogenize the profile. A likely candidate for this extra flux is biological remineralization. By adding isotopically depleted DIC to the water mass continuously along its flow path, this non-conservative term has the opposite effect than diffusive mixing of isotopically enriched northern source water. That these two processes would leave the $\delta^{13}C$ profiles roughly parallel seems like a coincidence as there is no a priori reason that mixing and remineralization should be of nearly the same size.

The Combined Roles of Transport and Vertical Mixing

Our analysis of the available LGM $\delta^{18}O$ profiles in the Atlantic shows that $\Psi/\kappa$ for AABW was larger than today. Similar to the local balance between upwelling and diffusion postulated for the modern North Pacific by W. Munk (his ‘$z^*$’ is similar to our $\Psi/\kappa$), a conservative tracer can tell you the ratio of advection to diffusion, but it can not constrain the two independently. However, we speculate that higher $\Psi/\kappa$ during the LGM could have been due to reduced mixing between northern and southern component waters, associated with movement of this watermass boundary away from the seafloor. Today, the upper boundary of AABW sits near 4 km water depth, whereas at the LGM it was at approximately 2 km. In the modern ocean, vertical diffusivities decrease away from bottom topography (Kunze et al., 2006). Averaged over the Atlantic, vertical
mixing is approximately three times lower above 4 km water depth than below (Ganachaud and Wunsch, 2000). In the Brazil Basin, vertical diffusivities over rough topography are an order of magnitude lower at 2 km than at the seafloor (Polzin et al., 1997). Reduced mixing due to upward migration of the watermass boundary at the LGM is therefore consistent with the vertical profile of $\kappa$ in the modern ocean, in both sign and magnitude. Our budget results suggest that reduced vertical mixing across the upper boundary of AABW may play a primary role in setting the Atlantic $\delta^{18}O$ tracer field during the LGM.

Upward movement of the AABW 'lid' may have been caused by an increase in the total area of deepwater formation during the LGM. We speculate that AABW formation occurred over a greater proportion of the Southern Ocean and therefore isopycnals exposed at the surface occupied a larger depth range in the abyss. Today, AABW is formed by a combination of brine rejection during sea ice formation and the export of freshwater by wind-driven Ekman transport (Foldvik et al., 2004; Orsi et al., 1999). Although greater total brine rejection may have enhanced density flux at the LGM (Shin et al., 2003), the density flux would also be sensitive to changes in freshwater export or melting of the Antarctic Ice Sheet. Thus, it is plausible that the AABW formation rate ($\Psi$) remained similar to today, while $\kappa$ decreased due to shoaling of AABW.

In contrast to our mixing-based interpretation of the $\delta^{18}O$ profiles, the large vertical gradient in $\delta^{13}C$ at the LGM has been interpreted in terms of watermass transport alone (Duplessy et al., 1988; Curry and Oppo, 2005). Curry and Oppo (2005) argued that
steep gradients in δ^{13}C are likely maintained by robust deep water flow, if vertical mixing was similar to or greater than today. If the increase in Ψ/κ during the LGM was entirely a function of transport, it would imply that Ψ increased by a factor of 8, a highly unlikely scenario. If instead we use the minimum Ψ/κ for the LGM, it would require a factor of two increase in transport. Although more feasible, this would also be a large change in the rate of AABW formation. By comparison, a factor of 2-8 reduction in vertical mixing is reasonable considering the profile of κ in the modern ocean. In all likelihood, both vertical mixing and transport likely changed, but we believe that most of the increase in Ψ/κ at the LGM was due to decreased vertical mixing.

It has been suggested that oceanic vertical mixing was enhanced at the LGM, due in part to exposure of continental shelves and greater tidal dissipation in the deep sea (Wunsch, 2003). If tidal dissipation was confined primarily to water depths below 2 km, however, its effect on mixing across the upper boundary of AABW would have been minimal. Rather than speculate on how κ changed in the past, our scenario calls on the variability in κ with depth in the ocean. The overall shape of this profile must be similar in alternative circulation scenarios, due simply to the presence of rough topography at the bottom of the ocean.

Our constraints on Ψ/κ have implications for the residence time of AABW in the Atlantic during the LGM. The ~ 2 km depth of the boundary between southern and northern component waters indicates the volume of AABW in the Atlantic was very large. Assuming that AABW spanned from 27ºS to 60ºN and across the width of the
Atlantic, we estimate its volume was \((1\pm0.2) \times 10^{17} \text{ m}^3\). For the modern \(\delta^{18}O\) surface = 3.08‰, the approximate volume of AABW is \((4\pm0.4) \times 10^{15} \text{ m}^3\), ~20x lower than our LGM estimate. If transport during the LGM was similar to the modern value, the residence time would have increased 20x. Even if transport was twice that of today, the residence time for AABW would have increased by an order of magnitude. A greater residence time for AABW is consistent with glacial reconstructions of highly depleted radiocarbon values (Keigwin, 2004; Robinson et al., 2005) and low Pa/Th ratios in the abyssal North Atlantic (Gherardi et al., 2009).

While our analysis of the \(\delta^{18}O\) tracer budget places new quantitative constraints on the LGM \(\Psi/\kappa\) ratio, we would also like to place the results in the context of glacial-interglacial CO₂ variability. The ‘Harvardton Bears’ box models highlighted the interdependence of nutrient utilization in the Southern Ocean and deep overturning rate in setting atmospheric pCO₂ (Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984). Less vigorous circulation and lower preformed nutrient concentrations for southern source deep water both work to lower atmospheric carbon dioxide levels. Toggweiler (1999) refined the ‘Harvardton Bears’ by parsing the deep ocean’s role into overturning rate and the efficiency of mixing between southern and northern component deep waters. Slower overturning works to lower pCO₂, but carbon and nutrients are brought into contact with the atmosphere if the two deep waters mix efficiently because northern component waters flow back to the North Atlantic via the shallow circulation (in his model at least). Reducing mixing between the deep water masses makes the southern cell a more efficient carbon trap. A lower \(\kappa\) allows for
smaller changes in overturning strength and/or nutrient utilization to produce the same lowering of atmospheric pCO₂. In this sense, the results from our δ¹⁸O budget may provide an important clue as to how the glacial deep ocean was involved in lowering atmospheric carbon dioxide levels.

IV. Conclusions

The δ¹⁸O of benthic foraminifera in the deep Atlantic show the presence of a watermass boundary at a depth of 2 km during the LGM, defined by a δ¹⁸O value of approximately 4.4 ‰. This boundary outlines the spatial domain of AABW and allows for the creation of a conservative tracer budget based on δ¹⁸O. Our approach accounts for transport of δ¹⁸O into the South Atlantic and vertical mixing of δ¹⁸O across the upper boundary of AABW. The δ¹⁸O budget yields reasonable estimates of AABW flow for the modern, giving us confidence that the budget captures the key factors that influence δ¹⁸O in the ocean interior, namely tracer transport and tracer diffusion. For the LGM, we find that the ratio of transport to vertical mixing (Ψ/κ) was at least twice as large as today. Either the transport was greater, the vertical mixing less, or some combination of these two factors. We suggest that the increase in Ψ/κ is most easily explained by reduced vertical mixing. Today, the watermass boundary between northern and southern component waters is located at approximately 4000 m water depth, whereas at the LGM the boundary was near 2000 m. The decrease in vertical mixing over this water depth range in the modern Atlantic (3-10x) (Polzin et al., 1997; Ganachaud and Wunsch, 2000) is broadly consistent with the implied decrease during the LGM.
Box-model experiments invoke limited AABW ventilation and vertical mixing to sequester atmospheric carbon dioxide during the LGM (Toggweiler, 1999). In this scenario, reduced exchange between abyssal and deep water promotes the accumulation of respired CO$_2$. If reduced vertical mixing at the upper ‘lid’ of AABW sequestered significant quantities of carbon during glacial times, it should be observed outside the Atlantic Ocean as well. Preliminary $\delta^{18}$O evidence from the Indian (Kallel et al., 1988) and Pacific (Herguera et al., 1992) show an apparent watermass boundary at 2 km water depth, suggesting the vertical diffusivity may have been reduced on global basis.

Sequestration in the abyssal ocean during glacial times also requires minimal leakage of carbon dioxide from the Southern Ocean to the atmosphere. Sea-ice coverage and stratification are other key factors that control Southern Ocean out-gassing, yet there is little agreement on how the processes change on glacial-interglacial timescales (Sigman and Boyle, 2000; Keeling and Stephens, 2001).

Our approach demonstrates that the abyssal $\Psi/\kappa$ ratio can be constrained using a simple tracer budget and traditional oxygen stable isotope data. Because transport and vertical mixing also influence $\delta^{13}$C, it is likely that the strong $\delta^{13}$C gradient in the glacial Atlantic was driven in part by reduced mixing between northern and southern component waters. The $\delta^{13}$C data also suggest that remineralization may influence the evolution of the $\delta^{13}$C profiles from south to north. In theory, we can combine the $\delta^{13}$C and $\delta^{18}$O data to quantify the relative influence of transport, mixing, and remineralization in the abyssal Atlantic. Understanding these parameters is key to understanding the ocean’s role in
glacial-interglacial CO$_2$ cycles, and will be a priority for future refinements to the tracer budget approach. In addition, it will be essential to construct new high-fidelity profiles of $\delta^{13}C$ and $\delta^{18}O$ to test whether the results we present here are representative of the Atlantic basin as a whole.
Appendix

Derivation of the zonally-averaged tracer equation

The evolution of the concentration of a generic tracer \( c \) released into the ocean is governed by the advection-diffusion equation,

\[
\frac{\partial c}{\partial t} + (u + u_{\text{eddy}}) \cdot \nabla c = \nabla \cdot \left( K_{\text{iso}} \nabla_{//} c \right) + \nabla \cdot (\kappa \nabla c) + \frac{\partial F_{\text{air-sea}}}{\partial z} + J \tag{1}
\]

where \( t \) is time, \( \nabla c \) is the three dimensional tracer gradient, and \( \nabla_{//} c \) is the along density surface tracer gradient. This equation describes the evolution of \( c \) averaged over the space and timescales characteristic of ocean turbulence, i.e. O(100 km) and O(1 week). At these large scales the tracer is advected by the averaged ocean circulation \( u \) and by an eddy-driven circulation \( u_{\text{eddy}} \): turbulent eddies result from instabilities of the mean ocean currents and their interactions produce a velocity equivalent to the Stokes drift resulting from surface gravity waves (Plumb and Ferrari, 2005). In addition to a velocity, turbulent eddies drive diffusion of tracers along density surfaces, represented by an isopycnal eddy diffusivity \( \kappa_{\text{iso}} \), and diffusion across density surfaces, represented by the diapycnal diffusivity \( \kappa \). In the abyssal ocean, the isopycnal diffusivity is typically 7-8 orders of magnitude larger than the diapycnal diffusivity \( (10^3 \text{ m}^2/\text{s} \text{ versus } 10^{-4} \text{ m}^2/\text{s}) \), because eddy fluxes are directed primarily along density surfaces. \( F_{\text{air-sea}} \) is the flux of tracer at the ocean surface and \( J \) includes any interior sources and sinks of non-conservative tracers. The tracer equation must be supplemented by boundary conditions. For the tracers considered in this paper the tracer flux across the solid boundaries is zero – the surface flux is represented explicitly by \( F_{\text{air-sea}} \) term. Knowledge of a tracer concentration \( c \), be it \( \delta^{18} \text{O} \) or \( \delta^{13} \text{C} \), is generally not sufficient to infer the full ocean
velocity field. But it can be used to infer some bulk properties of the ocean circulation, like the zonally-averaged overturning.

We can write an expression for the overturning circulation as a function of latitude and tracer concentration, much alike oceanographers do when they compute the overturning circulation as a function of latitude and density (Andrews, 1983; Nurser and Lee, 2004). Consider a zonal slice of the ocean at a fixed latitude \( y \), that is an \( x - z \) section through the ocean. We define \( A(y, C, t) \) as the area of that section below a certain tracer concentration value \( C \) and we assume that the tracer is increasing towards the surface (otherwise the inequalities must be reversed),

\[
A(y, C, t) = \iint_{c \leq C} dx \, dz \quad (2)
\]

The area \( A(y, C, t) \) can be changed either by horizontal advection of water with tracer concentration larger than \( C \), or by non-conservative processes like diffusion, air-sea fluxes or interior sources and sinks,

\[
\frac{\partial A}{\partial t} = -\frac{\partial}{\partial y} \left( \iint_{c \leq C} (v + v_{eddy}) \, dx \, dz \right) - \frac{\partial}{\partial x} \left[ \nabla \cdot (K_{iso} \nabla C) + \nabla \cdot (\kappa \nabla C) + \frac{\partial F_{air-sea}}{\partial z} + J \right] \, dx \, dz \quad (3)
\]

and \( v \) and \( v_{eddy} \) are the meridional velocity and eddy-induced velocity. Equation (3) can be simplified if we integrate between any latitude \( y \) and the northern or southern continental boundaries and assume steady state,

\[
\Psi(y, C) = \frac{\partial}{\partial x} \left[ \iiint_{c \leq C} \left( \nabla \cdot (K_{iso} \nabla C) + \nabla \cdot (\kappa \nabla C) + \frac{\partial F_{air-sea}}{\partial z} + J \right) \, dx \, dy \, dz \right] \quad (4)
\]

where \( \Psi(y, C) \) represents the mass meridional transport across the latitude \( y \),

\[
\Psi(y, C) = \iint_{c \leq C} (v + v_{eddy}) \, dx \, dz \quad (5)
\]
In deriving (4), we assumed that \( \Psi(y, C) \) vanishes at the southern boundary around Antarctica.

Integrating between the surface defined by the \( C \)-contour and the ocean bottom, we obtain the fundamental equation for our analysis,

\[
\int \Psi(y, c) \, dc \approx \int_{c=c} K_{iso} \nabla c \cdot n \, dS + \int_{c=c} \kappa \nabla c \cdot n \, dS + \int_{c=C} F_{air-sea} \, dxdy + \int_{c=c} J \, dxdydz
\]

(6)

where \( n \) is unit vector normal to the bounding surface \( S \), defined by the \( C \)-contour. The equation can be further simplified because, in the ocean, horizontal tracer gradients are many orders of magnitude smaller than vertical gradients, i.e. surfaces of constant tracer concentration are approximately flat. Expanding equation 6 in the limit of small tracer slopes, one obtains,

\[
\int \Psi(y, c) \, dc \approx \int_{c=c} K_{iso} \left| s_c - s_\rho \right|^2 \frac{\partial c}{\partial z} \, dxdy + \int_{c=c} \kappa \frac{\partial c}{\partial z} \, dxdy + \int_{c=C} F_{air-sea} \, dxdy + \int_{c=C} J \, dxdydz
\]

(7)

where \( s_c \) is the slope of the tracer surface and \( s_\rho \) is the slope of the density surface at the same point. Mixing of tracer across the \( C \)-contour is achieved both by along- and across density diffusion. Diapycnal diffusion dominates if:

\[
\kappa \gg K_{iso} \left| s_c - s_\rho \right|^2
\]

(8)

For the conservative tracer \( \delta^{18}O \), the \( s_c - s_\rho \) term is very small, as shown in the paper, and isopycnal mixing can be ignored. The situation is not as clear for \( \delta^{13}C \).
Consider now the circulation associated with AABW, i.e. a circulation that does not outcrop at latitudes north of 60°S. Then the integral (7) for latitudes north of 60°S reduces to,

$$\int \Psi(y, c) \, dc = \int\int_{c=c} K_{iso} \left| s - s_0 \right|^2 \frac{\partial c}{\partial z} \, dx \, dy + \int\int_{c=c} \kappa \frac{\partial c}{\partial z} \, dx \, dy + \int\int\int_{c=c} J \, dx \, dy \, dz$$  \hspace{1cm} (9)

For conservative tracers $J = 0$ and the equation has a simple interpretation. The left-hand side of (9) represents the tracer transport across a latitude $y$, $\Psi \Delta C$. Tracer transport can only be achieved for $\Delta C \neq 0$, i.e., if the tracer concentration going north is different from the concentration returning south. In order for the tracer concentration to change as it goes around the overturning loop, there must be diffusion of tracer from above, the term in the right-hand side of (9). At the order of magnitude level, equation (9) states,

$$\Psi(y, C) \Delta C \propto \kappa \frac{\partial C}{\partial z} \, A$$  \hspace{1cm} (10)

where $A$ is the horizontal area span by the tracer surface $C$ north of $y$. We dropped the along-density surface diffusion term, because in the region of interest diapycnal diffusion dominates (see Discussion section in main text). The simple balance (10) can be compared with the expression derived by W. Munk in the famous paper on ‘Abyssal Recipes’ (Munk, 1966):

$$w \Delta C \propto \kappa \frac{\partial C}{\partial z}$$  \hspace{1cm} (11)

This budget is consistent with (10) if vertical upwelling is uniform and can be estimated as $\Psi/A$, i.e. the mass flux across the surface $A$. However our approach shows that the budget is not local. The balance is between lateral transports of tracer and global vertical fluxes, not between local vertical upwelling and local vertical fluxes.
The budget described by eqn. (9) includes tracer transport into and out of the deep South Atlantic. However, our tracer domain is specifically limited to water flowing into the South Atlantic. Today, we define this domain using the depth of the maximum overturning streamfunction, which corresponds to the neutral density surface $\gamma = 28.15$ kg/m$^3$ (Lumpkin and Speer, 2007) and it is well tracked by a maximum in salinity vertical gradient (Figure 4). Paleoceanographic proxies are too sparse to fully evaluate the terms in (9). We therefore make some simplifying assumptions about the vertical structure of $\Psi$ and $C$. We assume $C$ has a linear vertical gradient within the volume considered, i.e.

$$\frac{\partial C}{\partial z} = \frac{\Delta C}{D}$$ (12)

where $D$ is the thickness of AABW. The streamfunction is instead assumed to be quadratic in $z$:

$$\Psi(z) = -\Psi \frac{z(z - 2D)}{D^2}$$ (13)

where $\Psi$ is the magnitude of the streamfunction (and hence of the transport) at the top of AABW and the ocean bottom is set at $z=0$. This is the simplest vertical structure that gives a northward flow of AABW and a southward flow of NADW. Plugging these expressions in the first integral of eq.(9), we obtain,

$$\frac{2}{3} \Psi \Delta C \approx \kappa \frac{\partial C}{\partial z} A$$ (14)
Figure captions:

Figure 1. Contour map of benthic foraminiferal $\delta^{13}C$ for the Western Atlantic during the Last Glacial Maximum (Curry and Oppo, 2005). The strong vertical gradient in $\delta^{13}C$ was likely due to 1) enhanced remineralization of organic carbon at depth, 2) greater transport of AABW relative to northern component water, 3) reduced mixing between AABW and northern component water, or 4) some combination of these effects.

Figure 2. A) Brazil Margin Cibicidoides spp. $\delta^{18}O$ (solid circles) (Curry and Oppo, 2005), outliers (open circles) and interpolated data points (+). Outliers are defined as those values which exceed the mean $\delta^{18}O$ from adjacent water depths by more than $\pm 0.3\%$. Interpolated values are based on linear interpolation between data points at adjacent water depths. B) Brazil Margin LGM-Holocene $\delta^{18}O$. C) Blake Ridge Cibicidoides spp. $\delta^{18}O$ (solid circles) (Keigwin, 2004) outliers (open circles), and interpolated data points (+). Also included are Uvigerina spp. $\delta^{18}O$ (grey triangles) (Keigwin 2004). D) Blake Ridge LGM-Holocene $\delta^{18}O$. Using a Student t-test and assuming a threshold 2-sided p value of 0.05, we are able to reject the null hypothesis that the shallow and deep LGM-Holocene $\delta^{18}O$ data are sampled from the same population ($p \approx 0.005$ for Brazil Margin Cibicidoides sp., $p \approx 0.0005$ for Blake Ridge Cibicidoides sp., and $p \approx 0.02$ for Blake Ridge Uvigerina sp.). This result is not surprising given that LGM-Holocene $\delta^{18}O$ data above 2 km fall in the upper left-hand quadrant of each plot, while the deeper data fall in the lower right-hand quadrant.

Figure 3. Two dimensional representation of the transport and mixing of a generic tracer (C) in the abyssal Atlantic. In reality, the tracer surface varies in three dimensions. The tracer budget is computed for volumes north of 27ºS and below a constant tracer value. $C_z$ is the vertical tracer gradient at the upper ‘lid’ of AABW and $\Delta C$ is the difference in C between the tracer surface and the sea floor. A conservative tracer like $\delta^{18}O$ is affected by transport ($\Psi$) from the Southern Ocean (solid arrow) and vertical mixing ($\kappa$) with overlying waters. Horizontal dashed arrow represents transport of northern component water. A non-conservative tracer like $\delta^{13}C$ is also affected by $J$, the remineralization of organic matter (dashed vertical arrow). Our approach depends on tracer distributions in the ocean interior (grey areas) distant from the watermass outcropping region near Antarctica.

Figure 4. Top) Salinity section along 30ºW in the Atlantic Ocean. The contour interval is 0.02 salinity units. The largest vertical gradient in salinity in the deep Atlantic between 30ºS and the equator exists from 3800-4300 m water depth, the approximate location of the upper boundary for AABW. Bottom) Neutral density section along 30ºW in the Atlantic Ocean. The maximum vertical salinity gradient corresponds to a neutral density surface $\gamma = 28.15$ kg/m³ (dashed line), which also marks the maximum overturning streamfunction for the South Atlantic (Lumpkin and Speer, 2007). Salinity and neutral density data are from the WOCE Global Hydrographic Climatology (Gouretski and Koltermann, 2004).
Figure 5. Contoured map of zonally averaged $\delta^{18}O$ for the eastern Atlantic (top) and western Atlantic (bottom). The gridded dataset was created using an $\delta^{18}O$ of sea water database (LeGrande and Schmidt, 2006), WOCE Global Hydrographic Climatology temperatures (Gouretski and Koltermann, 2004), and a paleotemperature equation (Kim and ONeil, 1997) adjusted for an acid fractionation factor of 1.0105. $\delta^{18}O$ values exceeding 3.08‰ exist primarily in the southwestern Atlantic where the influence of AABW is most pronounced.

Figure 6. The vertical gradient in $\delta^{18}O$ (top), potential temperature (middle), and salinity (bottom) across $\delta^{18}O$ iso-surfaces ranging from 3.00‰ to 3.20‰. One-sigma standard deviations are given as dashed lines. Estimates of $\delta^{18}O_z$ are based on the gridded $\delta^{18}O$ dataset (see text). Temperature and salinity data are from the WOCE Global Hydrographic Climatology (Gouretski and Koltermann, 2004). The maximum $\delta^{18}O_z$ occurs between $\delta^{18}O$ values of 3.06‰ and 3.10‰.

Figure 7. Top) Contour map of the vertical $\delta^{18}O$ gradient ($\delta^{18}O_z$) at $\delta^{18}O = 3.08$ ‰. This $\delta^{18}O$ surface spans the southwestern Atlantic. Bottom) Histogram of $\delta^{15}O_z$ values along the $\delta^{18}O = 3.08$ ‰ surface in the southwestern Atlantic. The mean $\delta^{18}O_z$ value for this $\delta^{18}O$ surface is $-3.1 \times 10^{-4}$ ‰ m$^{-1}$.

Figure 8. Contoured cross-section of $\delta^{18}O$ at 27.5ºS in the southwest Atlantic. $\delta^{18}O$ values are contoured in 0.02‰ increments from 3.00 to 3.20‰. We calculated $\Delta\delta^{18}O$ for the $\delta^{18}O$ iso-surface by subtracting the $\delta^{18}O$ values nearest the seafloor from 3.08‰. $\Delta\delta^{18}O$ is larger if the calculation is limited to the western boundary, where the majority of tracer fluxes into our control volume (see text for details).

Figure 9. Probability density functions from the Monte Carlo results of $\delta^{18}O_z$ as a function of $\delta^{18}O$ for the LGM data. These contour plots show that the water mass boundary at ~4.4‰ between northern and southern source waters at the LGM is a robust feature of the data even when an error of 0.18‰ is added to each data point. The top panels are from Cibicidoides sp. data at the Brazil Margin and the Blake Ridge and each peak at ~4.4‰. The bottom panel is from Uvigerina sp. data at the Blake Ridge and peaks at ~4.9‰. The offset in $\delta^{18}O$ between the species is most likely due to vital effects.

Figure 10. Top) Probability density function of the average $\delta^{18}O_z$ for the three LGM $\delta^{18}O$ profiles. Middle) Probability density function of $\Delta \delta^{18}O$ at the Brazil Margin. Bottom) Probability density function of $\Delta \delta^{18}O/\delta^{18}O_z$ for the LGM data. We plot $\Delta \delta^{18}O/\delta^{18}O_z$ instead of $\delta^{18}O_z/\Delta \delta^{18}O$ because division by $\delta^{18}O_z$ values close to zero creates infinitely large ratios. The peak in the $\Delta \delta^{18}O/\delta^{18}O_z$ PDF overlaps with the average of $\Delta \delta^{18}O/\delta^{18}O_z$ for the modern $\delta^{18}O$ surface of 3.08‰ (solid vertical line; 390±160 m).

Figure 11. Water column $\delta^{13}C$ of DIC values for the western Atlantic Ocean, partitioned into 10-degree latitude bins. Data from 35ºS to 5ºN are from the Ocean Atmosphere Carbon Exchange Study (OACES; http://cdiac3.ornl.gov/waves/) and data from 5ºN to
45°N are from World Ocean Circulation Experiment (WOCE) lines AR01, A20, and A22 (http://cchdo.ucsd.edu/atlantic.htm). The influence of AABW is most prominent in the South Atlantic, where δ\(^{18}\)C values below 4000 m approach -0.5 per mil. North of 15°N, the influence of AABW is barely perceptible.

Figure 12. Scatter plot of dissolved inorganic carbon δ\(^{13}\)C vs. Holocene Cibicidoides δ\(^{13}\)C data from Duplessy et al. (1984). We estimate an RMS error of 0.14‰ for the regression between the two parameters. Inset) Histogram of residual values (i.e., foraminiferal δ\(^{13}\)C values predicted by the regression minus observed foraminiferal δ\(^{13}\)C).

Figure 13. Probability density function of the vertical gradient in δ\(^{13}\)C (δ\(^{13}\)C\(_z\)) vs. δ\(^{13}\)C where the maximum in δ\(^{13}\)C\(_z\) occurs. At Brazil Margin, the maximum in δ\(^{13}\)C\(_z\) occurs at δ\(^{13}\)C values between 0.3‰ and 0.8‰. At Blake Ridge, the range is from 0.5‰ to 1.0‰. The maximum δ\(^{13}\)C\(_z\) values are generally higher at Blake Ridge, with a median of 1.4x10\(^{-3}\) ‰/m, while at Brazil Margin the median is 1.0x10\(^{-3}\) ‰/m.

Figure 14. Top) Average LGM δ\(^{13}\)C\(_z\) for the Brazil Margin and Blake Ridge from Figure 13. Middle) LGM Δδ\(^{13}\)C for the Brazil Margin based on 1000 randomly generated profiles, a calibration error of ±0.14‰, and a value at 4500 m water depth of -0.8‰ (see text). Bottom) Probability density function of Δδ\(^{13}\)C/δ\(^{13}\)C\(_z\) for the LGM data. The median for the LGM is 930±250 m, larger than that for the modern (510±150 m).

Figure 15. Top) Histogram of the vertical gradient in δ\(^{18}\)O (δ\(^{18}\)O\(_z\)) across the neutral density surface γ = 28.15 kg/m\(^3\) in the modern Atlantic. Nearly all of the values are negative, meaning δ\(^{18}\)O increases with depth. The median δ\(^{18}\)O\(_z\) is 2.9 x 10\(^{-4}\) ‰ m\(^{-1}\). Bottom) Histogram of the along isopycnal gradient in δ\(^{18}\)O (δ\(^{18}\)O\(_x\)) where γ = 28.15 kg/m\(^3\). Values are both positive and negative, meaning fluxes are into and out of the control volume. We use the median (1.5 x 10\(^{-8}\) ‰ m\(^{-1}\)) to yield the largest tracer slope (s\(_r\)) for evaluating along versus across isopycnal diffusion of δ\(^{18}\)O (see text).

Figure 16. Plot of δ\(^{18}\)O (o) and δ\(^{13}\)C (x) vs. latitude along the neutral density surface γ = 28.15. To estimate the δ\(^{13}\)C values, we determined the approximate depth of γ = 28.15 in each latitude bin of Figure 11 and used the δ\(^{13}\)C at that water depth. We estimate an along isopycnal gradient of 5 x 10\(^{-8}\) ‰ m\(^{-1}\) for δ\(^{13}\)C, approximately twice the horizontal gradient for δ\(^{18}\)O.

Figure 17. Left) Water depth vs. benthic foraminiferal δ\(^{13}\)C at the LGM for the Brazil Margin (solid line) (Curry and Oppo, 2005) and Blake Ridge (dashed blue) (Keigwin, 2004). At the Brazil Margin, we estimated δ\(^{13}\)C below 3000 m by linear extrapolation of the 2000-3000 m trend to 4500 m. The resulting value of -0.8 ‰ is consistent with values of -0.8‰ from the southeast Atlantic (Ninnemann and Charles, 2002). The maximum vertical gradient in δ\(^{13}\)C (δ\(^{13}\)C\(_z\)) occurs between 2500 m and 3000 m water depth. Right) Water depth vs. benthic foraminiferal δ\(^{18}\)O at the LGM for the Brazil Margin (solid line) and Blake Ridge (dashed line). At the Brazil Margin, we estimated δ\(^{18}\)O below 3000 m by linear extrapolation of the 2000-3000 m trend to 4500 m. A lower
$\delta^{18}O$ value at 4500 m decreases $\Delta \delta^{18}O$ and therefore increases $\Psi/\kappa$ for the LGM. The maximum vertical gradient in $\delta^{18}O$ ($\delta^{18}O_z$) occurs at ~2000 m water depth (horizontal grey bar).
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### Table 1. Parameters for modern tracer budget ($\delta^{18}$O = 3.08‰, $\delta^{13}$C = 0.7‰, 27°S to equator)

<table>
<thead>
<tr>
<th>$\delta^{18}$O</th>
<th>$\delta^{18}$O$_z$ (x10$^{-4}$‰ m$^{-1}$)</th>
<th>$\Delta\delta^{18}$O (%o)</th>
<th>$\Delta\delta^{18}$O / $\delta^{18}$O$_z$ (m)</th>
<th>Area (x10$^{12}$ m$^2$)</th>
<th>$\Psi/k$ (x10$^{10}$ m)</th>
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</thead>
<tbody>
<tr>
<td>AABW average</td>
<td>-3.1 ± 0.8</td>
<td>-0.12 ± 0.04</td>
<td>390 ± 160</td>
<td>4.7 ± 0.5</td>
<td>1.8 ± 0.8</td>
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<table>
<thead>
<tr>
<th>$\delta^{3}$C</th>
<th>$\delta^{3}$C$_z$ (x10$^{-4}$‰ m$^{-1}$)</th>
<th>$\Delta\delta^{3}$C (%o)</th>
<th>$\Delta\delta^{3}$C / $\delta^{3}$C$_z$ (m)</th>
<th>Area (x10$^{12}$ m$^2$)</th>
<th>$\Psi/k$ (x10$^{10}$ m)</th>
<th>Pe</th>
</tr>
</thead>
<tbody>
<tr>
<td>AABW average</td>
<td>4.8 ± 1.0</td>
<td>0.24 ± 0.05</td>
<td>510 ± 150</td>
<td>4.7 ± 0.5</td>
<td>1.4 ± 0.4</td>
<td>1.3 ± 0.7</td>
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### Table 2. Parameters for LGM tracer budget ($\delta^{18}$O = 4.4‰, $\delta^{13}$C = 0.6‰, 27°S to 60°N)*

<table>
<thead>
<tr>
<th>$\delta^{18}$O</th>
<th>$\delta^{18}$O$_z$ (x10$^{-4}$‰ m$^{-1}$)</th>
<th>$\Delta\delta^{18}$O (%o)</th>
<th>$\Delta\delta^{18}$O / $\delta^{18}$O$_z$ (m)</th>
<th>Area (x10$^{12}$ m$^2$)</th>
<th>$\Psi/k$ (x10$^{10}$ m)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Brazil (Cib)</strong></td>
<td>-8.7 ± 2.4</td>
<td>-0.46 ± 0.23</td>
<td>530 ± 250</td>
<td>45 ± 9</td>
<td>13 ± 7</td>
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<tr>
<td><strong>Blake (Cib)</strong></td>
<td>-9.9 ± 3.1</td>
<td>-0.46 ± 0.23</td>
<td>440 ± 230</td>
<td>45 ± 9</td>
<td>16 ± 9</td>
</tr>
<tr>
<td><strong>Blake (Uvi)</strong></td>
<td>-10.2 ± 3.1</td>
<td>-0.46 ± 0.23</td>
<td>440 ± 230</td>
<td>45 ± 9</td>
<td>16 ± 9</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td>-10.0 ± 1.7</td>
<td>-0.46 ± 0.23</td>
<td>440 ± 230</td>
<td>45 ± 9</td>
<td>16 ± 9</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>$\delta^{3}$C</th>
<th>$\delta^{3}$C$_z$ (x10$^{-4}$‰ m$^{-1}$)</th>
<th>$\Delta\delta^{3}$C (%o)</th>
<th>$\Delta\delta^{3}$C / $\delta^{3}$C$_z$ (m)</th>
<th>Area (x10$^{12}$ m$^2$)</th>
<th>$\Psi/k$ (x10$^{10}$ m)</th>
<th>Pe</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Brazil (Cib)</strong></td>
<td>10.0 ± 2.0</td>
<td>1.14 ± 0.25</td>
<td>1140 ± 340</td>
<td>45 ± 9</td>
<td>6 ± 2</td>
<td>2.2 ± 1.2</td>
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<tr>
<td><strong>Blake (Cib)</strong></td>
<td>14.0 ± 3.2</td>
<td>1.14 ± 0.25</td>
<td>930 ± 250</td>
<td>45 ± 9</td>
<td>7 ± 3</td>
<td>2.1 ± 1.3</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td>12.0 ± 2.0</td>
<td>1.14 ± 0.25</td>
<td>930 ± 250</td>
<td>45 ± 9</td>
<td>7 ± 3</td>
<td>2.1 ± 1.3</td>
</tr>
</tbody>
</table>

* Domain of AABW at LGM spanned from 27°S to 60°N and from the 2000 m isobath to the seafloor, a volume of $\sim$1 x 10$^{17}$ m$^3$. For the modern, AABW spans from 27°S to the equator and from 4000 m to the seafloor, a volume of $\sim$5 x 10$^{15}$ m$^3$. 
Figure 1
Figure 2
Figure 3
Figure 4
Figure 5
Figure 6
Figure 7
Figure 8
Figure 9
Figure 10
Figure 11
Cibicidoides spp. δ¹³C (per mil, VPDB)

Figure 12
Figure 13
Figure 14
Figure 15
Figure 16

\[ \delta^{18}O \text{ for } \gamma \text{ surface (‰)} \]

\[ \delta^{13}C \text{ for } \gamma \text{ surface (‰)} \]
Figure 17