Abyssal Atlantic Circulation during the Last Glacial Maximum

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Abstract

The ocean’s role in regulating atmospheric carbon dioxide on glacial-interglacial timescales remains a primary unresolved issue in paleoclimatology. Reduced mixing between deep waters may have aided storage of atmospheric CO₂ at the Last Glacial Maximum (LGM; 20,000 years ago), but data supporting this idea have remained elusive. Here we use published oxygen and carbon stable isotopic data in a new tracer balance to constrain the circulation and mixing in the deep ocean. We define the use of foraminiferal oxygen isotopes as a conservative tracer and infer the ratio of mass transport to vertical mixing in the abyssal Atlantic. By combining oxygen and carbon stable isotopic data, we also estimate the ratio of biological remineralization to vertical mixing. Our calculations suggest that both ratios were larger at the LGM. The simplest explanation is a reduction in vertical mixing between northern and southern component waters, driven by movement of this watermass boundary away from intense mixing near the seafloor. Our results support the hypothesis that changes in mixing between deep water masses played a role in sequestration of atmospheric carbon dioxide during glacial times.
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). Since greater than 90% of the combined oceanic, atmospheric and
terrestrial carbon resides in the deep ocean, it is believed to play a primary role in
regulating atmospheric CO₂ on glacial-interglacial time scales (Broecker, 1982; Sigman
and Boyle, 2000). The carbon isotopic ratio (δ¹³C) of benthic foraminifera from sediment
cores show that the deep ocean was more chemically stratified during the LGM (Boyle
and Keigwin, 1982; Curry and Lohmann, 1982; Kallel et al., 1988; Herguera et al., 1992;
Duplessy et al., 1988; Curry and Oppo, 2005). Toggweiler (1999) speculated that
reduced mixing between deep waters created the large vertical δ¹³C gradient and helped
sequester atmospheric CO₂ during the LGM. Alternatively, the LGM distribution of δ¹³C
in the Atlantic Ocean is explained by invoking changes in transport of North Atlantic
Deep Water (NADW) relative to Antarctic Bottom Water (AABW) (e.g., Duplessy et al.,
1988; Curry and Oppo, 2005). In this paper, we will use both stable carbon and oxygen
isotopic data to quantify the relative roles of transport and diffusion in the abyssal
circulation of the LGM.

Efforts to reconstruct the deep circulation using δ¹³C have been hindered by the fact
that δ¹³C is a non-conservative tracer – its distribution is affected not only by the physical
circulation of the ocean but also the biological remineralization of organic matter.
Organic carbon created by phytoplankton in the surface ocean is depleted in ¹³C relative
to ¹²C. As organic matter falls through the water column, remineralization at depth
produces ¹³C-depleted dissolved inorganic carbon that is incorporated in the shells of
benthic foraminifera. This process masks the effect of transport on the $\delta^{13}$C tracer field. We instead need a conservative tracer to reconstruct the deep circulation, or one which is influenced only by transport and mixing.

The stable oxygen isotopic ratio ($\delta^{18}$O) of foraminifera reflects the temperature and the $\delta^{18}$O of ambient seawater, which is directly related to salinity. Because both temperature and salinity are conservative tracers, so too is foraminiferal $\delta^{18}$O. Once a water parcel sinks from the surface ocean into the abyss, its temperature, salinity, and equilibrium $\delta^{18}$O value will change primarily by mixing with different watermasses. Here, for the first time, we exploit the conservative nature of $\delta^{18}$O to assess the LGM circulation of AABW. In combination with available $\delta^{13}$C data, we create a novel tracer balance to estimate the ratios of mass transport to vertical mixing and biological remineralization to mixing for the abyssal Atlantic. Our scaling calculations provide support for reduced vertical mixing between deep water masses during the LGM, evidence for which until now has been speculative and based on non-conservative tracers.

The strong vertical gradient in Atlantic Ocean benthic foraminiferal $\delta^{13}$C is a well-known feature of the LGM ocean. Less well-recognized is that LGM benthic foraminiferal $\delta^{18}$O profiles also have a larger gradient with water depth than today (Figure 1). At the Brazil Margin (30ºS), the change in *Cibicidoides* $\delta^{18}$O between 1 and 3 km was approximately 0.2 ‰ larger during the LGM (Figure 1a). The $\delta^{18}$O gradient is most easily observed in the difference between LGM and Holocene profiles (Figure 1b). LGM-Holocene $\delta^{18}$O values above 2 km water depth are generally less than 1.9 ‰, while
those below 2 km generally exceed 1.9 ‰. A least-square line fitting exercise indicates the LGM-Holocene δ\(^{18}\)O data are best described as two distinct groups separated in water depth at ~1800 m (SOM). The ‘step’ in the data is more pronounced at the Blake Ridge (30ºN), where it is observed in two genera of benthic foraminifera, *Cibicidoides* and *Uvigerina* (Figure 1d). At each location, the LGM-Holocene δ\(^{18}\)O pattern is driven by greater LGM δ\(^{18}\)O below ~1800 m. LGM δ\(^{18}\)O data from the Indian Ocean also show a clear watermass boundary near 1800 m water depth (Kallel et al., 1988) (SOM). The similar depth of this transition at the Brazil Margin and Blake Ridge, and the comparable absolute δ\(^{18}\)O value at which it occurs (~ 4.3 ‰) strongly suggests there was a watermass boundary near 2 km in the LGM Atlantic, consistent with inferences from benthic foraminiferal δ\(^{13}\)C (Oppo and Lehman, 1993).

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 along closed streamfunction contours (Figure 2). The water mass properties of AABW are set through to air-sea fluxes near Antarctica and mixing with overlying waters in the ocean interior. Air-sea fluxes are commonly used to estimate watermass transport in the present climate (e.g. Schlosser et al., 1991; Speer and Tziperman, 1992; Large and Nuser, 2001), but this type of data is lacking in the paleoclimate record. Instead, we estimate the circulation from tracer distributions in the ocean interior. In doing so, our intent is not to minimize the importance of air-sea fluxes, but to close the AABW circulation problem in the absence of surface flux information. AABW is
uniquely suited to this approach because unlike other watermasses it does not outcrop north of Drake Passage.

The derivation of the zonally-averaged equation for a tracer advected by the abyssal overturning circulation in the Atlantic Ocean is provided in the SOM, but Figure 2 provides a schematic of the key fluxes. The net amount of tracer flowing into the abyss with AABW is given by the zonally averaged overturning circulation, $\Psi$ (m$^3$ s$^{-1}$), multiplied by the difference in tracer concentration between the waters entering and leaving the South Atlantic, $\Delta \delta^{18}O$ (%$\epsilon$). For a conservative tracer like $\delta^{18}O$ the advection of tracer into the abyss must be balanced by mixing of tracer with overlying waters:

$$\Psi \Delta \delta^{18}O = \kappa \delta^{18}O_z A \quad (1)$$

where the right hand side represents the diffusion of tracer expressed as a turbulent diffusivity, $\kappa$ (m$^2$ s$^{-1}$), times the vertical tracer gradient at the upper boundary of AABW, $\delta^{18}O_z$ (%$\epsilon$ m$^{-1}$), and $A$ is the area across which mixing occurs (m$^2$). We neglect the contribution of along isopycnal tracer flux because it is an order of magnitude smaller than the diapycnal tracer flux today and even smaller at the LGM (SOM). Re-arranging so that measurable quantities are on the right and unknowns are on the left, we obtain:

$$\frac{\Psi}{\kappa} = \frac{\delta^{18}O_z}{\Delta \delta^{18}O} A \quad (2)$$

To calculate $\Psi/\kappa$ for the modern, we created a gridded dataset of $\delta^{18}O$ of calcite for the Atlantic based on available seawater $\delta^{18}O$ (LeGrande and Schmidt, 2006) and temperature (Gouretski and Koltermann, 2004) databases (SOM). We used a $\delta^{18}O$ value of 3.05%$\epsilon$ to define the upper boundary for AABW. We estimate that $\Psi/\kappa$ for AABW at
27°S is \(-1.3\pm0.2 \times 10^{10} \text{ m} (1\sigma)\). A range of \(\delta^{18}\text{O}\) values that encompass the realistic spatial extent of AABW produces similar \(\Psi/\kappa\) estimates. If we assume an average \(\kappa\) for the deep Atlantic of \(3\pm1.5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}\) (Ganachaud and Wunsch, 2000), then AABW transport at 27°S is \(4\pm2.0 \text{ Sv} (1 \text{ Sv} = 1\times10^6 \text{ m}^3 \text{ s}^{-1})\), consistent with geostrophic estimates in the Brazil Basin: \(6.7\pm0.5 \text{ Sv at } 31°\text{S and } 5\pm1.1 \text{ Sv at } 24°\text{S}\) (Speer and Zenk, 1993). Since foraminifera record in-situ rather than potential temperature, their \(\delta^{18}\text{O}\) values are not strictly conservative. If we adjust \(\delta^{18}\text{O}\) for the effects of adiabatic heating, the transport to vertical mixing estimates decrease by \(~15\%\) for both the modern and the LGM (SOM).

The budget describing the distribution of \(\delta^{13}\text{C}\) in the deep circulation cell has the additional contribution of a biological remineralization term, \(J (\%e \text{ s}^{-1})\), multiplied by the volume into which the remineralization occurs, \(V (\text{m}^3)\):

\[
\Psi \Delta \delta^{13}\text{C} = \kappa \delta^{13}\text{C}_z A + JV \tag{3}
\]

In this budget, \(\Psi\) acts as a sink of \(\text{^{13}C}/\text{^{12}C}\) for the abyssal cell, because inflow of AABW has lower \(\text{^{13}C}/\text{^{12}C}\) than the outflow. \(J\) is also a sink, as organic matter is strongly depleted in \(\text{^{13}C}\). The diffusive term is a source, since mixing with high \(\text{^{13}C}/\text{^{12}C}\) North Atlantic Deep Water (NADW) adds more \(\text{^{13}C}\) than \(\text{^{12}C}\) to the AABW cell. Dividing eqn. 3 through by \(\kappa \delta^{13}\text{C}_z A\), we obtain:

\[
Pe = \frac{\Psi \Delta \delta^{13}\text{C}}{\kappa \delta^{13}\text{C}_z A} \approx 1 + \frac{J V}{\kappa \delta^{13}\text{C}_z A} \tag{4}
\]

where 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. Using $\Delta \delta^{13}C$ and $\delta^{13}C$ from modern $\delta^{13}C$ tracer data (SOM) and the $\delta^{18}O$-based value of $\Psi/\kappa$ (eqn. 2), we estimate a bulk Pe for AABW of $2.2 \pm 0.6$ (Table 1). This results suggests that today the transport flux of $\delta^{13}C$ is roughly twice that of the diffusive flux.

The last term of eqn. 4 is the dimensionless ratio of the $\delta^{13}C$ flux due to remineralization and the $\delta^{13}C$ flux due to diffusion. For modern AABW, this ratio is approximately 1, indicating the remineralization and diffusive fluxes are similar (Table 1). In terms of the overall tracer balance, approximately 50% of the $\delta^{13}C$ transported by AABW is balanced by remineralization, with the remainder accounted for by diffusion. The non-conservative behavior of $\delta^{13}C$ therefore plays an important role in the tracer balance today. Assuming $\kappa$ of $3\pm1.5 \times 10^{-4}$ m$^2$s$^{-1}$ (Ganachaud and Wunsch, 2000), we estimate $J = -3\pm1.9 \times 10^{-3}$ ‰ yr$^{-1}$ (SOM). As a cross-check we calculated the remineralization rate for the deep Atlantic using a global compilation of sediment trap carbon flux data (Lutz et al., 2002). We estimate that remineralization of carbon causes $\delta^{13}C$ to change by $-1\pm0.6 \times 10^{-3}$ ‰ yr$^{-1}$ in the AABW circulation cell. This result is consistent with the $\delta^{18}O$- and $\delta^{13}C$-based value, indicating that the tracer budget provides a reasonable measure of remineralization for the modern abyssal Atlantic.

Our estimate of the bulk Peclet number for the LGM is based on depth transects of benthic foraminiferal stable isotopes from the Brazil Margin and the Blake Ridge. We assume that these few profiles characterize the Atlantic Ocean tracer field. As the number of measurements increases in the future, it will be possible to test whether the
data from the Brazil Margin and Blake Ridge are indeed representative. Using the data in hand, we estimate a Pe for the LGM of $6.8 \pm 2.9$, indicating that diffusion played a less important role than transport in controlling the distribution of $\delta^{13}C$ (Table 1). Since the LGM Pe is approximately 3 times the modern value, either the LGM transport was larger than today, the vertical mixing less, or some combination of the two.

The ratio of $\delta^{13}C$ remineralization to diffusion provides additional perspective on the relative role of transport and diffusion in the LGM abyssal Atlantic. We estimate that the ratio of remineralization to diffusive fluxes for the LGM was $5.8 \pm 2.9$, meaning that the remineralization of $\delta^{13}C$ was about 6 times larger than the diffusive flux (Table 1). Furthermore, the ratio of remineralization to diffusion at the LGM was about 5 times the modern value, suggesting that either remineralization in AABW was greater at the LGM or vertical diffusivity between AABW and northern source waters was reduced. The LGM tracer data also tell us that about 85% of the $\delta^{13}C$ transported by AABW is balanced by the remineralization of organic matter, compared to 50% today. This result provides quantitative support for the idea that the glacial $\delta^{13}C$ signal of the abyssal ocean is driven largely by accumulation of respired CO$_2$ (Toggweiler, 1999).

The simplest way to increase both the transport/diffusion and remineralization/diffusion ratios at the LGM is to reduce vertical mixing. We suggest this occurred due to upward movement of the watermass boundary at the Last Glacial Maximum. Today, the boundary between southern and northern source waters is near 4 km, 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 Basin, vertical mixing is approximately three times lower above 3.5 km than below 3.8 km (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 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.

It is possible that both $\Psi$ and $J$ were larger during the LGM, requiring no change in vertical diffusivity. What mechanisms could explain a factor of two or larger increase in $J$? Vertical migration of the watermass boundary could cause greater remineralization of carbon in southern component waters, but sediment trap data show that carbon flux per unit area is similar at 2 km and 4 km water depth (Lutz et al., 2002). Another way to change $J$ is to increase the export flux of carbon from surface waters between 30ºS and 60ºN. Given that surface nutrients are nearly completely consumed in the modern low-latitude Atlantic (Sigman and Boyle, 2000), it is unlikely that more efficient nutrient utilization could account for a glacial-interglacial contrast in export production.

Alternatively, the oceanic nutrient inventory may have been larger at the LGM (Broecker, 1982). However the tongue of high $\delta^{13}C$ Antarctic Intermediate Water that supplies nutrients to the thermocline in the modern Atlantic was apparently absent at the LGM (Curry and Oppo, 2005). Water equivalently depleted in $\delta^{13}C$ is found below 2 km at the LGM, too deep to affect productivity in the surface ocean. Cadmium data from the tropical and North Atlantic support the $\delta^{13}C$ results, showing increased nutrient levels in
the abyss, but reduced levels above 2 km (Boyle, 1992; Marchitto and Broecker, 2006). Finally, modest increases in low-latitude export production produce shifts in the ocean-average lysocline depth that are inconsistent with LGM observations (Sigman et al, 1998). Given these factors, it seems unlikely that J increased by a factor of two or more during the LGM.

Our results bear on recent questions concerning the deep Atlantic glacial-interglacial $\delta^{13}$C signal and atmospheric CO$_2$ levels. Contrary to conventional interpretations of the LGM $\delta^{13}$C distribution, the simplest explanation of the abyssal tracer data is a reduction in vertical mixing at the boundary between northern and southern component waters. Our scaling calculations provide support for the idea that less mixing between deep waters promotes sequestration of CO$_2$ during the LGM (Toggweiler, 1999). However for reduced mixing to be an effective trap, the carbon cannot leak out to the atmosphere in the Southern Ocean. This is a topic of ongoing debate (e.g. Sigman and Boyle, 2000; Keeling and Stephens, 2001) separate from the tracer balance presented here. Our approach demonstrates that the abyssal circulation can be constrained by combining passive paleoceanographic tracers with ocean dynamics. A clear next step is to add an active tracer like radiocarbon to our analysis. In doing so, we will be able to parse the distribution of abyssal ocean tracers during the LGM into separate contributions from transport, diffusion, and remineralization.
References


Monnin et al., Atmospheric CO$_2$ concentrations over the last glacial termination, Science, 291, 112-114, 2001.

Neftel et al., New measurements on ice cores samples to determine the CO$_2$ content of the atmosphere during the last 40,000 years, Nature, 295, 220-223, 1982.


Toggweiler, J. R., Variation of atmospheric CO$_2$ by ventilation of the ocean’s deepest water, Paleoceanography, 14, 571-588, 1999.

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<thead>
<tr>
<th></th>
<th>$Pe = \frac{\Psi \Delta \delta^{13}C}{\kappa \delta^{13}C_z A}$</th>
<th>$JV = \frac{J V}{\kappa \delta^{13}C_z A}$</th>
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<tbody>
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<td>Modern cell</td>
<td>2.2 ± 0.6</td>
<td>1.2 ± 0.6</td>
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<td>(δ$^{18}$O = 3.05)</td>
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<tr>
<td>LGM cell</td>
<td>6.8 ± 2.9</td>
<td>5.8 ± 2.9</td>
</tr>
<tr>
<td>(δ$^{18}$O = 4.30)</td>
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Table 1.

Key parameters of the Antarctic Bottom Water circulation during the modern and Last Glacial Maximum. The first column is the ratio of the transport flux to the diffusive flux of tracer, also known as a bulk Peclet number ($Pe$). $Pe$ is about 2 for the modern circulation, indicating transport moves roughly twice the amount of tracer as diffusion. The larger bulk $Pe$ at the LGM suggests diffusion played a smaller role than transport in controlling the distribution of $\delta^{13}C$. The second column is the ratio of remineralization flux to the diffusive flux. Today, the transport of $\delta^{13}C$ is balanced by approximately equal parts diffusion and remineralization. At the LGM, the transport of $\delta^{13}C$ is balanced primarily (~85%) by remineralization. Values for $A$, $V$, and $\delta^{18}O$ and $\delta^{13}C$ are given in Table 1s (SOM).
Figure captions:

Figure 1.  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 $\pm 0.3\%o$ of the mean defined by adjacent values in water depth. 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.

Figure 2. Schematic illustration of the LGM Atlantic Ocean abyssal circulation. Top) $\delta^{18}$O$_z$ is the gradient in $\delta^{18}$O at $\sim 2000$ m and $\Delta \delta^{18}$O is the difference in $\delta^{18}$O between 2000 m and the sea floor. $\delta^{18}$O is affected by advection ($\Psi$) from the Southern Ocean and vertical mixing ($\kappa$) with northern component water. Bottom) $\delta^{13}$C$_z$ is the gradient in $\delta^{13}$C at $\sim 2000$ m and $\Delta \delta^{13}$C is the difference in $\delta^{13}$C between 2000 m and the sea floor. $\delta^{13}$C is also affected by J, the remineralization of organic matter (dashed arrow). Rather than using air-sea fluxes near Antarctica, our approach depends on tracer distributions in the ocean interior distant from the outcropping region (grey area).