A simple mixing explanation for late Pleistocene changes in the Pacific-South Atlantic benthic δ¹³C gradient

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Abstract. The fact that the deep-ocean benthic δ¹³C minimum shifted from the North Pacific to the South Atlantic during the Last Glacial Maximum is often interpreted as evidence of a change in deep water circulation, such as the development of deep water ventilation in the North Pacific or a decrease in Southern Ocean overturning. This study re-evaluates the implications of changes in benthic δ¹³C gradients by comparing Pacific Deep Water (PDW) δ¹³C measurements with the values expected for the null hypothesis that PDW ventilation sources remained unchanged throughout the Late Pleistocene. The δ¹³C compositions of PDW, Northern Component Water (NCW) and Southern Component Water (SCW) are estimated from regional benthic δ¹³C stacks of 3–6 sites. Changes in PDW δ¹³C and PDW-SCW δ¹³C gradients over the past 800 kyr are found to be well described by a constant mixture of 60% NCW and 40% SCW plus a constant Pacific remineralization offset of −0.5‰. Thus, a change in PDW ventilation cannot be inferred solely on the basis of changes in the Pacific-South Atlantic benthic δ¹³C gradient.

1 Introduction

An important unanswered question about glacial ocean circulation is whether the shift of minimum δ¹³C values from the North Pacific to the South Atlantic at the Last Glacial Maximum (LGM) implies a major change in Pacific circulation. Benthic δ¹³C values are often used as a proxy for deep water ventilation because deep water δ¹³C becomes progressively more negative with age due to the accumulation of remineralized low-δ¹³C organic carbon. Today deep waters formed in the North Atlantic and Southern Ocean mix in approximately equal proportions to form Circumpolar Deep Water (CDW) (Broecker et al., 1998), which flows northward into the South Pacific at depth and returns southward at mid-depths in the form of Pacific Deep Water (PDW) (Ganachaud et al., 2003; Talley et al., 2003). Because southward flowing PDW progressively mixes with younger water during its return trip, today the oldest deep water with the most negative δ¹³C value is found at mid-depth in the North Pacific (Kroopnick, 1985; Matsumoto et al., 2002).

However, at the LGM the most negative benthic δ¹³C measurements are found in the deep South Atlantic. Does this shift imply that at the LGM Pacific deep water was younger than South Atlantic deep water? Previous studies have suggested that the change in benthic δ¹³C gradients was caused by additional mixing between PDW and North Pacific Intermediate Water (NPIW) (Duplessy et al., 1988; Herguera et al., 1992; Keigwin, 1998; Oppo and Horowitz, 2000). Alternatively, North Atlantic water may have been the primary ventilation source for the deep Pacific with intermediate water providing only a weak, secondary ventilation source (Toggweiler et al., 2006). In this study a simple mixing model is used to test the null hypothesis that PDW δ¹³C values result from a constant mixing ratio of water from the North and South Atlantic throughout the Late Pleistocene. This model reproduces most of the changes in PDW δ¹³C, including glacial reversals of the Pacific-South Atlantic δ¹³C gradient.

Section 2 reviews the causes of δ¹³C variability and describes δ¹³C gradients in the ocean today and at the LGM. Section 3 estimates the δ¹³C values of PDW, Northern Component Water (NCW) and Southern Component Water (SCW) for the last 800 kyr using regional δ¹³C stacks. Section 4 compares PDW δ¹³C with the results of a simple mixing model based on the assumption that PDW is composed of a constant proportion of NCW and SCW with a constant

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remineralization offset. Section 5 discusses model and data limitations and the implications of the model results, including alternative interpretations. Finally, Sect. 6 summarizes my conclusions.

2 Background

Modern $\delta^{13}$C gradients are explained by differences in water masses’ age and surface water processes (biological productivity and air-sea gas exchange). Here, age is defined as the length of time since water last had contact with the surface ocean. As $^{13}$C-depleted organic carbon sinks below the thermocline and remineralizes at depth, it increases the $\delta^{13}$C of surface water and decreases the $\delta^{13}$C of deep water. The flux of low-$\delta^{13}$C organic carbon from the surface to deep ocean produces a remineralization (or “aging”) offset because the longer that deep water remains out of contact with the surface, the more negative its $\delta^{13}$C value becomes. Another important source of carbon isotope fractionation is temperature-dependent air-sea gas exchange. The change in the $\delta^{13}$C of the ocean is thought to change by $-0.3%e$ (Duplessy et al., 1988) during glacial periods as the result of a net transfer of low-$\delta^{13}$C organic carbon from the continents to the ocean (Shackleton, 1977).

Today North Atlantic Deep Water (NADW), which fills most of the deep Atlantic above 4000 m, has the highest $\delta^{13}$C of all deep water ($\sim 1%e$, Kroopnick, 1985) because it is recently formed from surface water that experiences high biological productivity. Antarctic Bottom Water (AABW), which is formed from water with incomplete nutrient utilization, has lower $\delta^{13}$C values ($\sim 0.2$; Kroopnick, 1985) and is found below 4000 m in the South Atlantic. Although PDW is formed from approximately equal parts NADW and AABW (Broecker et al., 1998), remineralization in the deep Pacific makes PDW $\delta^{13}$C values more negative than AABW, with a minimum value of approximately $-0.2%e$ in the oldest water in the North Pacific (Kroopnick, 1985; Keigwin, 1998; Matsumoto et al., 2002). Alternatively, a study by Johnson (2008) suggests that PDW may be approximately 60% AABW, 20% NADW and 20% Antarctic Intermediate Water (AAIW).

The $\delta^{13}$C of carbonate from benthic foraminifera is used to estimate past changes in a water mass’s $\delta^{13}$C composition and spatial extent. Most studies agree that epibenthic *Cibicidoides* species accurately record the $\delta^{13}$C of deep water (e.g., Graham et al., 1981; McCorkle et al., 1990, 1997). However, in areas of high productivity the $\delta^{13}$C of benthic foraminifera may be affected by a phytodetrital effect caused by localized remineralization of organic carbon in or at the surface of marine sediments (Mackensen et al., 1993).

Because water mass properties and formation processes differ between glacial and interglacial, I use the general term Northern Component Water (NCW) to describe intermediate/deep water formed in the North Atlantic and the term Southern Component Water (SCW) to describe deep/bottom water formed in the Southern Ocean. LGM NCW, often called Glacial North Atlantic Intermediate Water (GNAIW), shoals to 2000 m and has higher $\delta^{13}$C values than NCW today (Oppo and Lehman, 1993). Large changes in Atlantic $\delta^{13}$C values below 2000 m result from changes in the mixing ratio of NCW and SCW (i.e., more SCW during glacials) and the $\delta^{13}$C values of the two water masses (Curry and Oppo, 2005).

The $\delta^{13}$C value of SCW may have been as low as $-0.9%e$ at the LGM (Ninnemann and Charles, 2002; Curry and Oppo, 2005). Repeated $\delta^{13}$C measurements of LGM benthic foraminifera from different locations throughout the South Atlantic are all consistent with a low-$\delta^{13}$C glacial SCW endmember, indicating that these values are not an artifact of the phytodetrital effect (Ninnemann and Charles, 2002; Curry and Oppo, 2005). The large change in SCW $\delta^{13}$C is considered evidence for glacial reductions in air-sea exchange and overturning rates around Antarctica (e.g., Marchitto and Broecker, 2006; Bouttes et al., 2009), which could result from a northward shift in westerly winds over the Southern Ocean, increased sea ice cover, and/or increased surface stratification (Toggweiler et al., 2006). Increases in the biological productivity of the Southern Ocean may also contribute to the decrease in $\delta^{13}$C (Brovkin et al., 2007; Martinez-Garcia et al., 2009).

Because PDW is derived from a mixture of NCW and SCW that progressively accumulates low-$\delta^{13}$C organic carbon during its transit, one might expect that minimum $\delta^{13}$C values would always occur in the deep Pacific. However, LGM Pacific $\delta^{13}$C measurements (although somewhat sparse) indicate that the lowest LGM Pacific $\delta^{13}$C values ($-0.4%e$, Matsumoto et al., 2002) were much less negative than the $-0.9%e$ values found in the deep South Atlantic. LGM reconstructions show a slight decrease in PDW $\delta^{13}$C below 2000 m and a slight increase above 2000 m (Herguera et al., 1992; Matsumoto et al., 2002). The lowest values remain in the north but shift from 2000 m today to 3000 m at the LGM (Matsumoto et al., 2002). The only observations of lower Pacific $\delta^{13}$C values ($-0.8%e$) are in the western South Pacific at 2000–3000 m where SCW enters the Pacific (McCave et al., 2008).

Thus, during the LGM the lowest $\delta^{13}$C values are found in the South Atlantic rather than the Pacific. Does reversal of the PDW-SCW $\delta^{13}$C gradient at the LGM imply an additional source of PDW ventilation in the glacial Pacific? Some studies have proposed that NPIW ventilates glacial PDW (e.g., Duplessy et al., 1988; Herguera et al., 1992; Keigwin, 1998; Oppo and Horowitz, 2000). The flow of modified GNAIW into the Pacific may also contribute to glacial PDW ventilation (Lynch-Stieglitz and Fairbanks, 2005).
Table 1. Core locations.

| Core      | Lat  | Long | Depth (m) | Reference                  |
|-----------|------|------|-----------|----------------------------|
| North Atlantic |
| DSDP 552  | 56.0 | -23.2| 2301      | Shackleton and Hall (1984)  |
| ODP 658   | 20.8 | -18.7| 2264      | Tiedemann (1991)            |
| ODP 980   | 55.5 | -14.7| 2169      | Oppo et al. (1998); McManus et al. (1999); Flower et al. (2000) |
| ODP 982   | 57.5 | -15.9| 1145      | Venz et al. (1999); Venz and Hodell (2002) |
| ODP 983   | 60.4 | -23.6| 1983      | McIntyre et al. (1999); Raymo et al. (2004) |
| ODP 984   | 61   | -24  | 1650      | Raymo et al. (2004)         |
| South Atlantic |
| ODP 1089  | -40.9| 9.9  | 4621      | Hodell et al. (2001)        |
| ODP 1090  | -42.9| 8.9  | 3702      | Venz and Hodell (2002)      |
| GeoB 1211 | -24.5| 7.5  | 4085      | Bickert and Wefer (1996)    |
| Pacific   |
| ODP 677   | 4.2  | -83.7| 3461      | Shackleton et al. (1990)    |
| ODP 806B  | 0.3  | 159.4| 2520      | Berger et al. (1996); Bickert et al. (1993) |
| ODP 846   | -3.1 | -90.8| 3307      | Mix et al. (1995a); Shackleton et al. (1995) |
| ODP 849   | 0.2  | -110.5| 3851      | Mix et al. (1995b)          |

1994; Lynch-Stieglitz et al., 1996; Matsumoto et al., 2002; Toggweiler et al., 2006). One ocean circulation model reproduces the glacial PDW-SCW δ13C gradient as the result of weak ventilation primarily from NCW with no increase in mixing between PDW and Pacific intermediate waters (Toggweiler et al., 2006). It is easy to understand how mixing with intermediate waters could ventilate the upper 2500 m of the glacial Pacific. Here I focus on the PDW-SCW δ13C reversal below ~2500 m where the contributions of NPIW and GNAIW to PDW are most uncertain. My approach to investigating PDW ventilation sources is first to estimate the δ13C of the different water masses using regional δ13C stacks and then to compare the PDW stack with the δ13C values that would be expected from the null hypothesis that PDW ventilation sources and the Pacific remineralization offset remained constant throughout the late Pleistocene. This allows for better quantification of how the relationships between NCW, SCW, and PDW may have changed during glacial cycles.

3 Regional δ13C stacks

Changes in the δ13C of NCW, SCW, and PDW over the last 800 kyr are estimated by stacking (averaging) δ13C records of Cibicidoides benthic foraminifera from each region. The stacks include all of the continuous records for each water mass spanning 0–800 ka with a temporal resolution of ≤5 kyr (Table 1; Fig. 1). The age model for each δ13C record is produced by aligning the core’s benthic δ18O record to the LR04 benthic δ18O stack (Lisiecki and Raymo, 2005). The age model uncertainty associated with this alignment is ~2 kyr within each ocean basin, but Pacific δ18O may briefly lag Atlantic δ18O by as much as 4 kyr during terminations (Skinner and Shackleton, 2005; Lisiecki and Raymo, 2009).

The δ13C of NCW (Fig. 2a) is estimated by stacking six North Atlantic records from water depths of ≤2300 m (Table 1). These benthic δ13C records are all similar with little glacial-interglacial change (Raymo et al., 2004; Lisiecki et al., 2008) but with low-frequency power that may represent long-term cyclicity in the global carbon cycle (Raymo et al., 1997; Wang et al., 2004). The δ13C of SCW is estimated using three South Atlantic cores from ≥3700 m water depth: ODP 1089, ODP 1090, and GeoB1211 (Fig. 1a). This stack may not represent a pure SCW signal because ODP Site 1090 is estimated to be 60% SCW and 40% NCW today (Venz and Hodell, 2002). However, the NCW contribution is likely smaller during glacial periods. The effects of NCW intrusions at SCW sites are discussed in Sect. 5.1. The Pacific stack is composed of three records from 3300–3850 m depth in the Eastern Equatorial Pacific (ODP 677, ODP 846, and ODP 849) and one record from 2520 m in the Western Equatorial Pacific (ODP 806B). The δ13C records from these sites are similar to one another (Fig. 1b) and to shorter records from the North Pacific (Keigwin, 1998), the southeast Pacific (Matsumoto and Lynch-Stieglitz, 1999), and other western equatorial Pacific sites (Herguera et al., 1992; Mix et al., 1991). The PDW and SCW δ13C stacks (Fig. 2a) are well correlated (r=0.78), but the glacial-interglacial amplitude of PDW δ13C is only about half that of SCW δ13C. PDW δ13C is slightly more negative than SCW δ13C during interglacials but much more positive than SCW during glacial periods.
Fig. 1. Component records for (a) the SCW $\delta^{13}$C stack (black): ODP 1090 (blue), ODP 1089 (red), and GeoB1211 (green) and (b) the PDW $\delta^{13}$C stack (black): ODP 677 (blue), ODP 806 (gray), ODP 846 (green), and ODP 849 (red). See Table 1 for citations.

4 A simple mixing model

Reversal of the PDW-SCW $\delta^{13}$C gradient during glacial has sometimes been interpreted as evidence of a change in PDW ventilation. However, the implications of past changes in the PDW-SCW $\delta^{13}$C gradient are best evaluated in comparison with the null hypothesis that PDW ventilation remained unchanged throughout the Late Pleistocene. The PDW $\delta^{13}$C values expected for the null hypothesis are simulated using a simple mixing model that assumes PDW is derived from a constant mixing ratio of NCW and SCW plus a constant remineralization (age) offset. This is given by the equation

$$\delta^{13}C_{PDW} = f \delta^{13}C_{NCW} + (1-f) \delta^{13}C_{SCW} + c$$  (1)

where $f$ is the fraction of PDW originating from the North Atlantic and $c$ is the (negative) $\delta^{13}$C offset produced by the remineralization of organic carbon in the deep Pacific. The mixing ratio $f$ and remineralization offset $c$ are estimated empirically by minimizing the root mean square error between the model and the PDW stack for 800–0 ka, yielding values of 0.60 for $f$ and $-0.51\%o$ for $c$. In other words, the best fit between the model and data is produced by a mixture of 60% NCW and 40% SCW with a $-0.5\%o$ remineralization offset between the Atlantic and Pacific. The model estimates a greater percentage of NCW than in PDW today (Broecker et al., 1998; Johnson, 2008), but this discrepancy may be an artifact of interglacial intrusions of NCW at SCW sites (see Sect. 5.1).

Given the simplicity of the model, the overall fit between model and data is quite good for the entire Late Pleistocene (Fig. 2b), with a correlation coefficient of 0.87 and a root mean square error of 0.13‰. The model also reproduces the reversal of the PDW-SCW $\delta^{13}$C gradient during glacial without any change in the mixing ratio of NCW and SCW or the Pacific remineralization offset. During interglacials when the $\delta^{13}$C gradient between NCW and SCW is relatively small, the Pacific remineralization offset is sufficient large to make PDW $\delta^{13}$C more negative than SCW $\delta^{13}$C. However, during glacial when the NCW-SCW gradient is large, the (constant) remineralization offset is not large enough to make PDW $\delta^{13}$C more negative than SCW $\delta^{13}$C. Thus, the null hypothesis cannot be rejected solely on the basis of $\delta^{13}$C gradients. Other possible constraints on PDW ventilation are discussed in Sects. 5.2 and 5.4.

5 Discussion

5.1 Data limitations

One important limitation to evaluating the null hypothesis of constant PDW ventilation is the spatial and temporal resolution of Late Pleistocene benthic $\delta^{13}$C data. The $\delta^{13}$C of NCW is fairly well constrained by data from six sites above 2300 m in the North Atlantic. However, there is some evidence for $\delta^{13}$C gradients between different NCW components (Matsumoto and Lynch-Stieglitz, 1999; Millo et al., 2006), particularly before 600 kyr ago (Raymo et al., 2004). In the Pacific and South Atlantic, the availability of $\delta^{13}$C data is limited by poor carbonate preservation in these regions. This creates uncertainty about how well the PDW and SCW $\delta^{13}$C stacks describe their respective water masses.

The strong similarity of all four Pacific $\delta^{13}$C records (Fig. 1a) suggests that the PDW stack is probably a good representation of $\delta^{13}$C values from 2500–3850 m in the equatorial Pacific. The lack of strong $\delta^{13}$C gradients in the Pacific below 2000 m today and during the LGM (Herguera et al., 1992; Matsumoto et al., 2002) suggests that the PDW stack may also be representative of a large portion of the
Fig. 2. (a) Regional stacks of δ^{13}C for NCW (red), SCW (blue), and PDW (black). (b) PDW δ^{13}C stack (black) and the δ^{13}C predicted by the null hypothesis (purple), simulated by a constant mixture of 60% NADW and 40% CDW plus a remineralization offset of −0.5‰. This simple mixing model produces a good fit with changes in PDW δ^{13}C and the PDW-SCW δ^{13}C gradient.

5.2 Model limitations and alternative interpretations

An additional concern is that the simple mixing model presented here cannot adequately test for all possible changes in PDW ventilation even if the δ^{13}C values of NCW, SCW, and PDW are known exactly. A unique solution can be found only if no other water masses contribute to PDW and if either f or c is assumed to be constant. Changes in the relative flux of NCW into the Pacific could be disguised by compensating variations in the Pacific remineralization offset, potentially resulting from changes in overturning rates, the δ^{13}C of sinking organic matter, or Pacific surface productivity (e.g., Robinson et al., 2005). More research is needed to constrain the extent to which the Pacific remineralization offset may have varied over the last 800 kyr.

The ventilation of PDW by water masses other than NCW and SCW would also produce errors in the model’s estimates of the PDW mixing ratio and remineralization offset. NPIW is sometimes cited as a possible additional ventilation source for glacial PDW (e.g., Duplessy et al., 1988; Herguera et al., 1992; Keigwin, 1998). Increased production and deeper mixing of NPIW may explain higher δ^{13}C values at 2000 m in the Pacific, but sharp gradients in δ^{13}C, δ^{18}O, and apparent ventilation age between 2000 m and 3000 m at the LGM suggest strong density stratification and little mixing with NPIW below 3000 m (Matsumoto et al., 2002). Latitudinal δ^{13}C gradients are also inconsistent with NPIW ventilation. LGM δ^{13}C values at 3000 m in the North Pacific (near the possible ventilation source) are more negative than LGM δ^{13}C values at 3000 m in the equatorial Pacific (Matsumoto et al., 2002).
Could additional PDW ventilation occur in the South Pacific? One study estimates that AAIW constitutes up to 20% of modern PDW at 3000 m (Johnson, 2008). Because glacial $\delta^{13}C$ values for AAIW are approximately 0‰ (Pahnke and Zahn, 2005), AAIW contributions to PDW could also help explain why glacial PDW has a more positive $\delta^{13}C$ value than glacial SCW. Therefore, the estimated contributions of NCW and SCW from the simple mixing model should not be considered definitive. However, a simulation of glacial ocean circulation reproduces the PDW-SCW $\delta^{13}C$ gradient without any increase in mixing between AAIW and PDW at the LGM (Toggweiler et al., 2006). Additionally, if mixing with AAIW were the primary source of PDW ventilation, one would expect PDW $\delta^{13}C$ values to be more positive near the source of AAIW ventilation and more negative at greater distances from the ventilation source. Thus, the deepest Pacific site would be expected to have the most negative $\delta^{13}C$ values. Instead, ODP Site 849 at 3851 m is more positive than the two shallower EEP sites (Fig. 1b).

Another possible southern ventilation source is suggested by an LGM depth profile of $\delta^{13}C$ in the Southwest Pacific, which detects an additional water mass below 3500 m with a $\delta^{13}C$ value of $-0.3‰$ (Cibicidoides-equivalent as measured in Uvigerina) (McCave et al., 2008). McCave et al. (2008) propose that this water mass represents deep water formed in the Ross Sea and that topographic barriers at 3500 m in the Southern Ocean prevent dense, low-$\delta^{13}C$ glacial deep water formed in the Weddell Sea from entering the Pacific below 3500 m. This does not necessarily imply a change in PDW ventilation; instead, the apparent shift may arise because Ross and Weddell Sea waters are indistinguishable during interglacials and only develop different $\delta^{13}C$ and Cd/Ca signatures during glacials (Rosenthal et al., 1997; Ninnemann and Charles, 2002; McCave et al., 2008). However, the paleo-extent of Ross Sea Water is difficult to assess because LGM $\delta^{13}C$ transects in the low and mid-latitude Pacific show no $\delta^{13}C$ gradients below 2000 m (Herguera et al., 1992; Matsumoto et al., 2002; McCave et al., 2008). Additional study is needed to determine the extent to which Ross Sea Water may have influenced PDW $\delta^{13}C$.

5.3 Model-data residuals

The simple mixing model presented here provides no evidence for glacial changes in the mixing ratio of NCW and SCW or for a long-term trend in their relative contributions to the Pacific over the last 800 kyr. Separately analyzing the time intervals 800–400 ka and 400–0 ka yields estimates of 60% and 62% NCW, respectively. However, the Pacific remineralization offset shifts from $-0.47‰$ to $-0.58‰$ between the two intervals. Possible mechanisms for a long-term trend in the remineralization offset are changes in the rate of deep water export to the Pacific or changes in the flux or $\delta^{13}C$ composition of carbon falling from the surface ocean.

For the 800–0 ka model, model-data residuals are all less than 0.4‰ (Fig. 2b). Several short intervals with average residuals of $>0.1‰$ occur at: 635–575 ka ($-0.17‰$), 530–475 ka (0.13‰), 325–275 ka (0.14‰) and 80–0 ka (0.11‰). The durations of these model-data misfits are insufficient to determine whether they are caused by changes in the ratio of NCW or the Pacific remineralization offset. For example, the best fit for 80–0 ka is obtained from a mixture of 66% NCW and a 0.69‰ Pacific remineralization offset, but the fit for 60% NCW and a 0.62‰ Pacific remineralization offset is nearly as good. The two scenarios have less than a 5% difference in root mean square error. Alternatively, some of these model-data residuals may result from temporary intrusions of NCW at SCW sites, which would cause the model to overestimate PDW $\delta^{13}C$. This is not a likely explanation for residuals during glacial conditions when NCW is less extensive, but NCW intrusions could explain residuals during interglacial intervals, such as 530–475 ka and 325–275 ka.

5.4 Additional tests of the null hypothesis

The null hypothesis (a constant mixing ratio of NCW and SCW with a constant Pacific remineralization offset) reproduces the reversal of the PDW-SCW $\delta^{13}C$ gradient during glacial periods without any changes in PDW ventilation. However, this requires that the relative contribution of NCW to PDW at 2500–3850 m remain approximately constant despite dramatic shoaling of NCW in the glacial Atlantic. One can further test the null hypothesis by assessing whether glacial NCW reached the deep Pacific and constraining the relative export rates of NCW and SCW.

Because NCW must pass through the South Atlantic to reach the deep Pacific, one possible test of the null hypothesis is identification of NCW in the South Atlantic. Glacial $\delta^{13}C$ values were approximately 0.2‰ at $\sim 3000$ m in the western South Atlantic (Curry and Oppo, 2005; Lisiecki et al., 2008), which suggests an approximately equal mixture of NCW (1.2‰) and SCW (−0.8‰). Additionally, a vertical $\delta^{13}C$ profile at $\sim 45^\circ S$ in the eastern Atlantic has an LGM $\delta^{13}C$ value of $\sim 0‰$ at 2500 m, indicative of substantial mixing between NCW and SCW (Hodell et al., 2003). Could NCW at $\sim 2500$ m in the South Atlantic reach the glacial deep Pacific? Possible routes of GNAIW through the Southern Ocean are highly speculative because glacial benthic $\delta^{13}C$ measurements south of 50° S are nearly nonexistent. The Antarctic Circumpolar Current may have been highly effective at mixing GNAIW to sill depths of 3500 m en route to the Pacific or GNAIW may have reached the deep Pacific via deep water formation in the Ross Sea (McCave et al., 2008). Alternatively, some studies have suggested that NADW formation continued throughout the glaciation (Matsumoto and Lynch-Stieglitz, 1999; Millo et al., 2006).

If NCW reached the glacial deep Pacific, could its export rate have been comparable to that of SCW? Estimates of the glacial overturning rates of NCW and SCW vary widely (e.g.,
Toggweiler et al., 2006; Weber et al., 2007; Huybers et al., 2007; Govin et al., 2009; Bouttes et al., 2009), and their flux into the Pacific is even more difficult to constrain. Pa/Th data suggest that the rate of North Atlantic overturning at the LGM was nearly as high as in the Holocene (McManus et al., 2004). Export of at least some GNAIW into the Pacific is supported by identification of the air-sea exchange signature of GNAIW in the glacial Pacific (Lynch-Stieglitz et al., 1996) and analysis of δ13C gradients between the open Pacific and Tasman Sea (Russon et al., 2009). Pore water estimates of deep water temperature and salinity during the LGM suggest that PDW may have been more similar to NCW than SCW (Adkins et al., 2002). Additionally, the simulation of glacial ocean circulation by Toggweiler et al. (2006) finds that GNAIW was the primary ventilation source for glacial PDW. More research is needed to evaluate whether the ratio of NCW and SCW fluxes into the Pacific actually remained constant.

5.5 Water mass ages

Finally, I consider the mixing model’s implications for the relative ages of PDW and SCW at the LGM. The simple model presented here is actually consistent with either young or old glacial SCW, where age is defined as the length of time since deep water last had contact with the surface ocean. At one extreme, SCW may have remained relatively young with rapid overturning (e.g., Ninnemann and Charles, 2002; Curry and Oppo, 2005) and obtained low-δ13C values through other processes, such as reduced air-sea gas exchange (e.g., Piotrowski et al., 2008) or increased Southern Ocean productivity (e.g., Brovkin et al., 2007; Martinez-Garcia et al., 2009) which would increase the accumulation of isotopically depleted carbon in the deep South Atlantic. In this case, water mass ages, overturning rates, and fluxes into the Pacific could have been quite similar to modern circulation (Huybers et al., 2007; Marchal and Curry, 2008), but benthic δ13C would might not correlate with ventilation age because changes in productivity or air-sea gas exchange could overwhelm the remineralization effect (e.g., Charles et al., 1993).

At the other extreme, the model results could also be consistent with a stagnant layer of old SCW in the glacial deep Atlantic (e.g., Sikes et al., 2000; Toggweiler et al., 2006; Brovkin et al., 2007; Marchitto et al., 2007). A large density gradient (Adkins et al., 2002) could have impeded vertical mixing within the Atlantic, and the 3500-m sill in the Southern Ocean would have blocked SCW flow into the deep Pacific (McCave et al., 2008). The glacial ocean circulation model of Toggweiler et al. (2006) found that SCW was ventilated in only one grid cell in the Pacific sector of the Southern Ocean. In such a scenario, the fluxes of both SCW and NCW into the Pacific would likely be reduced. A decrease in the rate of remineralization during glaciations could potentially explain why no change in the Pacific remineralization (age) offset is apparent in my simple mixing model. Glacial PDW could be younger than SCW if its age is the average of young NCW and old SCW plus an age offset that is not large enough to make it older than SCW. Radiocarbon estimates of deep water ages (e.g., Broecker et al., 2007; Marchitto et al., 2007) may provide additional constraints on changes in overturning rates.

6 Conclusions

Late Pleistocene changes in the δ13C values of NCW, SCW, and PDW are estimated from regional benthic δ13C stacks. Orbital-scale changes in PDW δ13C and the PDW-SCW δ13C gradient are found to be well described by a constant mixture of 60% NCW and 40% SCW with a constant −0.5‰. Pacific remineralization offset for the last 800 kyr. Inter-glacial intrusions of NCW at SCW sites may explain the slight discrepancy between the model’s mixing ratio estimate and the modern mixing ratio of 50% NCW (Broecker et al., 1998). Model-data comparison also yields no evidence for a long-term trend in the NCW contribution to PDW over the last 800-kyr, but the Pacific remineralization offset may have gradually shifted from −0.47‰ to −0.62‰.

A constant ratio in the flux of NCW and SCW to the deep Pacific provides the simplest explanation of glacial changes in the PDW-SCW δ13C gradient changes. During interglacials when the δ13C gradient between NCW and SCW is relatively small, the Pacific remineralization offset is sufficiently large to make PDW δ13C more negative than SCW δ13C. However, during glacials when the NCW-SCW gradient is large, the (constant) remineralization offset is not large enough to make PDW δ13C more negative than SCW δ13C. Thus, the null hypothesis of constant Pacific ventilation cannot be rejected on the basis of changes in the PDW-SCW δ13C gradient. Additionally, the pattern of δ13C gradients within the Pacific at the LGM does not appear to support mixing between PDW and NPIW below 3000 m (Matsumoto et al., 2002).

The results presented here suggest that glacial boundary conditions may not have greatly altered the relative contributions of NCW and SCW to PDW. However, a simple mixing model cannot provide conclusive results because changes in the mixing ratio of the two water masses could be disguised by changes in the Pacific remineralization rates or by contributions from other water masses such as AAIW or Ross Sea Water. More research is needed to constrain the δ13C composition and spatial extent of glacial water masses and changes in the Pacific remineralization offset throughout the Late Pleistocene.
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