Supporting Information

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Vertical Profile of Mixing in the Deep Ocean

Vertical mixing is typically quantified in terms of a vertical diffusivity κ, which represents the rate at which mixing would spread a patch of tracer over time. Direct estimates show that vertical diffusivities are strongly enhanced a few hundred meters above rough bottom topography (e.g., ref. 1). Direct estimates suggest that the vertical diffusivities averaged over ocean basins are of $O(10^{-4})$ m^2 s^{-1} at 4 km, the average ocean depth, and decay to $O(10^{-5})$ m^2 s^{-1} at 2 km, the typical height of oceanic ridges (e.g., refs. 2, 3.). Similar results are found from inverse models that infer vertical diffusivities from tracer distributions (4, 5) and from theoretical calculations of the internal wave generation rates (6). A review of the extensive literature concludes that diffusivities of $O(10^{-4})$ m^2 s^{-1} drive an overturning of O(10) Sv, as observed in the real ocean, and diffusivities of $O(10^{-5})$ m^2 s^{-1} could only return a very weak flow of O(1) Sv (1).

The enhanced vertical mixing below 2 km is triggered when abyssal flows (geostrophic or tidal) intersect rough bottom topography and radiate internal waves that break (6). The magnitude of the mixing depends on the strength of the abyssal flows and the ocean stratification and may have changed between the modern and glacial climates. Its vertical profile, with large values below 2 km and low values above, is set by the distribution of ocean topography. Much like surface gravity waves break when they encounter a beach regardless of the strength of the swell, internal waves break primarily close to bottom topography regardless of their amplitude. Thus, the vertical profile of mixing must have changed little at the LGM, because the topography was essentially the same. A recent study has verified that the magnitude of deep mixing likely increased at the LGM, but its vertical profile remained the same as today (7). The changes in ocean circulation described in the main paper depend on the depth range where mixing is enhanced, not on the magnitude of the deep mixing.

Numerical Models of the Last Glacial Maximum (LGM)

Our claim that the rearrangement in deep water masses observed at the LGM is linked through simple dynamics to the expansion of sea ice around Antarctica may seem at odds with reports that numerical models of the LGM struggle to reproduce this rearrangement (8). Although the National Center for Atmospheric Research Community Climate System Model version 3 (CCSM3) used to generate Fig. 5 shows an expansion of the abyssal water mass, other models do not. We believe that the range in model responses stems from various inconsistencies in the physics. Most models do not include a profile of vertical mixing that decays with height above the bottom, thereby missing a key dynamical aspect of the LGM oceans. Models differ in their treatment of atmospheric radiation and sea ice with some models capturing the expansion in sea ice better than others. The representation of geostrophic eddies is deficient, because it overestimates changes in isopycnal slope due to wind changes at the LGM.

Based on our study, we believe that LGM simulations would likely become more consistent if all models represented the key dynamical features of the LGM oceans, namely the vertical profile of vertical mixing, an appropriate representation of geostrophic eddies, and consistent sea ice physics. We plan to run such a model to test to what extent our zonally averaged view of the changes in the ocean circulation at the LGM will stand the test of a 3D circulation.

Shoaling of North Atlantic Deep Water (NADW) at the Last Glacial Maximum

The shoaling of NADW is best understood if one considers the transient response to an expansion of Southern Ocean (SO) sea ice. Any water sinking in the North Atlantic below the isopycnal separating the two cells would be trapped in the abyssal cell, because mixing would no longer be able to lift it high enough in the water column to outcrop around Antarctica in areas experiencing a positive buoyancy flux. The waters would instead experience a negative buoyancy flux, become denser, and sink back into the abyss. This transient conversion of NADW into denser Antarctic Bottom Water (AABW) would continue until the whole abyss was filled with waters denser than the ones formed in the North Atlantic. At this point waters sinking in the North Atlantic would start flowing above the abyssal dense cell and the equilibrium solution sketched in Fig. 4 would be achieved.

We sketched the solution with two overturning cells relevant for the LGM in Fig. 4, *Lower*. However, a second solution is possible (9). If the Northern Hemisphere air–sea fluxes generated waters lighter than any of the waters found at the surface in the Southern Ocean, the upper overturning cells would disappear and a pool of stagnant waters would develop in the North Atlantic, a solution possibly relevant for Heinrich events. The circulation in the upper cell is, however, inconsequential for the abyssal cell, which is the focus of this study.

Deep Salinity Budget at the Last Glacial Maximum

In the present day ocean, North Atlantic Deep Water comes to the surface under sea ice in the Southern Ocean with temperatures between 2–3 °C. These warm waters drive strong melting of sea ice and ice shelves resulting in a freshening of the surface waters, visible in Fig. 1 as a light-density tongue right at the sea surface around Antarctica. This freshwater offsets much of the increase in salinity driven by the formation of new ice through brine rejection. This results in weak salinity variations and as a result temperature dominates the density stratification in the modern Southern Ocean.

At the Last Glacial Maximum, the deep waters upwelling to the surface around Antarctica were close to the freezing point and resulted in little melting of sea ice and ice shelves (10). The waters that flowed poleward under the permanent sea ice were therefore exposed to significant brine rejection, resulting in a strong increase in their salinities. We can derive a scaling law for the salinity changes experienced by the waters as they flowed under ice.

In steady state, the meridional salinity variations of waters flowing under sea ice can be computed from the simple equation (ref. 11):

$$\psi S_y \approx \mathcal{F},$$

where ψ is the mass transport in the surface layer of the ocean exposed to the salinity flux $\mathcal{F}$, and S_y is the meridional salinity gradient. This equation has been obtained averaging the equation for the conservation of salinity over the depth of the ocean mixed layer (the upper layer of the ocean where temperature and salinity are vertically well mixed) and along a latitude circle. If we further integrate the equation in latitude between the Antarctic continent and the permanent sea ice line, we obtain that the salinity difference ΔS between waters upwelling at the permanent sea ice edge and those sinking around the Antarctic continent,

$$\Delta S \approx \mathcal{F} \times \ell_{ice}/\psi,$$

where ℓ_{ice} is the meridional extent of sea ice.

The CCSM3 model used to generate Fig. 5 simulates a mass transport of $\psi = 20 \times 10^6$ m^3 s^{-1} in the LGM mixed layer under sea ice, close to the present-day value. The area-averaged salinity flux is instead much larger than today as a result of the expansion of the permanent sea ice and the reduction of ice melting giving $\mathcal{F} \times \ell_{ice} = 2 \times 10^7$ m^3 psu s^{-1}. Our scaling law for these values gives a salinity difference $\Delta S \approx 1$ psu between the SO upwelling deep waters and the sinking abyssal waters. The model salinity distributions confirm this scaling and are consistent with pore fluid estimates of LGM salinity, which suggest that the abyssal ocean was at least 1 psu saltier than overlying waters. This large salinity difference dominated the density stratification in the Southern Ocean. The deep LGM ocean was everywhere within a few degrees of freezing, while a $10°$ C temperature difference would be required to match the impact on density of a 1-psu salinity difference. Based on model output and observations, we can then conclude that the isolated overturning cell spanning the deep ocean at the LGM and depicted in the schematic in Fig. 4, *Lower* was associated with a strong salinity gradient.

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