AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER

10.1002/2016GL070058

Key Points:
- Southern Ocean surface forcing controls the global deep ocean density stratification over a larger depth range than previously thought
- Fixed Southern Ocean surface buoyancy fluxes rather than fixed surface density controls the deep density stratification
- But the temperature and salinity components of the deep stratification are strongly influenced by Northern Hemisphere surface forcing

Supporting Information:
- Supporting Information S1

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Citation:
Sun, S., I. Eisenman, and A. Stewart (2016), The influence of Southern Ocean surface buoyancy forcing on glacial-interglacial changes in the global deep ocean stratification, *Geophys. Res. Lett.*, *43*, doi:10.1002/2016GL070058.

Received 17 JUN 2016
Accepted 14 JUL 2016
Accepted article online 21 JUL 2016

The influence of Southern Ocean surface buoyancy forcing on glacial-interglacial changes in the global deep ocean stratification

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Abstract Previous studies have suggested that the global ocean density stratification below ∼3000 m is approximately set by its direct connection to the Southern Ocean surface density, which in turn is constrained by the atmosphere. Here the role of Southern Ocean surface forcing in glacial-interglacial stratification changes is investigated using a comprehensive climate model and an idealized conceptual model. Southern Ocean surface forcing is found to control the global deep ocean stratification up to ∼2000 m, which is much shallower than previously thought and contrary to the expectation that the North Atlantic surface forcing should strongly influence the ocean at intermediate depths. We show that this is due to the approximately fixed surface freshwater fluxes, rather than a fixed surface density distribution in the Southern Ocean as was previously assumed. These results suggest that Southern Ocean surface freshwater forcing controls glacial-interglacial stratification changes in much of the deep ocean.

1. Introduction

During the Last Glacial Maximum (LGM), the climate was characterized by a colder global-mean temperature and lower atmospheric CO_2 concentration compared with today [e.g., *Clark et al.*, 2009]. An enhanced stratification of the deep ocean (below ∼1000 m depth) has been proposed as a key contributor to the lower atmospheric CO_2 concentration at the LGM by acting as a more effective carbon trap [*Bouttes et al.*, 2009; *Adkins*, 2013]. The deep ocean stratification also influences the strength of the abyssal overturning circulation, which has been invoked to explain reduced CO_2 outgassing and hence a lower CO_2 concentration at the LGM [*Sarmiento and Toggweiler*, 1984; *Anderson et al.*, 2009; *Sigman et al.*, 2010].

As a large-scale feature that is closely tied to the global ocean overturning circulation, the processes that maintain the stratification of the deep ocean (including both abyssal and middepth regions) have attracted substantial attention for many years. Studies by *Munk* [1966] and *Munk and Wunsch* [1998] proposed that the deep stratification and overturning circulation are controlled to first order by a balance between the vertical advection and diffusion of buoyancy. More recent studies have suggested that Southern Ocean processes play a key role in closing the global overturning circulation and setting the deep ocean stratification [*Marshall and Speer*, 2012; *Wolfe and Cessi*, 2010].

Nikurashin and Vallis [2011, 2012] combined these ideas in a conceptual model, in which the surface density was specified in the Southern Ocean. In this model, the abyssal stratification associated with the abyssal overturning circulation, i.e., the lower cell that spreads Antarctic Bottom Water (AABW) throughout the global ocean below ∼3000 m, is essentially set by the Southern Ocean surface density profile with some modulation by the competing wind-driven and eddy-driven overturning circulations in the Southern Ocean.

Above the abyssal overturning circulation and below the main thermocline (typically from 3000 m to 1000 m depth in the Atlantic Ocean), diapycnal mixing is relatively weak [*Kunze et al.*, 2006]. The stratification in this middepth region is associated with the nearly adiabatic pole-to-pole overturning circulation (i.e., the upper cell) [*Wolfe and Cessi*, 2011] that spreads North Atlantic Deep Water (NADW) southward from the North Atlantic and spreads Antarctic Intermediate Water (AAIW) northward from the Southern Ocean [*Talley*, 2013; *Lozier*, 2012]. The stratification at this depth is modulated by surface buoyancy and momentum forcing conditions in both the Southern Ocean and the North Atlantic [*Wolfe and Cessi*, 2011].

Though these idealized modeling studies are conceptually illuminating, the applicability of their predictions to the real ocean is limited. Most of these studies employ an idealized topography, a single ocean basin, and a single thermodynamic variable (rather than including both temperature and salinity), which leads to an overturning circulation that is split into two isolated cells [e.g., *Wolfe and Cessi*, 2010, 2011; *Munday et al.*, 2013]. However, a property-based reconstruction of the overturning circulation suggests that the upper and lower cells are in fact actively coupled and follow a three-dimensional pathway through all of the major ocean basins [*Talley*, 2013]. Additionally, idealized modeling studies typically employ restoring to a fixed buoyancy profile over a prescribed time scale at the ocean surface, which may not accurately reflect the surface buoyancy fluxes in regions where they are dominated by freshwater fluxes, such as the Southern Ocean [*Cerovecki et al.*, 2011; *Stewart et al.*, 2014].

The present study is the first (as far as the authors are aware) to investigate the influence of the Southern Ocean surface forcing on the global deep ocean stratification in the relatively realistic setting of a comprehensive climate model. In section 2, we describe the experimental setup, which consists of three ocean-only climate model simulations that are designed to isolate the influence of the Southern Ocean surface forcing on the changes in the global deep ocean stratification between the LGM and the preindustrial (PI) climate. In section 3, we present the model simulation results and discuss the relative roles of the Southern Ocean and the Northern Hemisphere surface forcing in setting the global deep ocean stratification. In section 4, we use a conceptual model to interpret the results from the climate model simulations. Concluding remarks are provided in section 5.

2. Experimental Design

We use a state-of-the-art climate model, the National Center for Atmospheric Research (NCAR) Community Earth System Model version 1.1.2 (CESM1.1.2), which we run in a configuration with only the ocean component active and the atmosphere, sea ice, and land runoff specified from two previous coupled simulations. One coupled simulation represents the PI climate [*Gent et al.*, 2011], and the other coupled simulation represents the LGM climate [*Brady et al.*, 2013]. Further information about the model setup and forcing is included in the supporting information.

We perform three experiments that share the same model configuration (including the same PI ocean bathymetry) but have different ocean surface forcing: one control run (PI) is forced by PI surface conditions, a second control run (LGM) is forced by LGM surface conditions, and a test run (Test) is forced by LGM surface conditions in the Southern Ocean and PI surface conditions elsewhere. More precisely,

$$F_{\mathrm{Test}} = \gamma F_{\mathrm{PI}} + (1-\gamma) F_{\mathrm{LGM}},$$

where γ is 0 to the south of 40°S, 1 to the north of 30°S, and increases linearly from 0 to 1 between 40°S and 30°S. Here F_{PI} and F_{LGM} denote the surface forcing fields derived from the PI and LGM coupled runs, respectively, and F_{Test} denotes the surface forcing fields used for the Test run (see supporting information for further details). In each case, the coupled model output is used to construct surface forcing fields that repeat every 30 years.

All three runs are initialized from the same initial conditions obtained from the PI coupled run. The length of each integration is listed in Table S1 of the supporting information along with the trend during the last 120 years of the global volume-average temperature, ideal age, and Atlantic Meridional Overturning Circulation (AMOC) maximum (defined as the maximum total overturning circulation streamfunction below 500 m in the Atlantic Ocean including contributions from both the mean flow and the parameterized eddies). Although the trends are nonzero, Table S1 indicates that all three runs are close to equilibration (see also supporting information Figures S3 and S6). Note that all simulations are initiated from the PI coupled run, so the PI ocean-only run equilibrates more rapidly than the Test and LGM runs. Unless otherwise noted, the results presented in this study are averaged over the last 20 years of each model run.

Figure 1 shows the zonal-mean surface buoyancy flux (B) in each simulation, along with its heat (B_{HF}) and freshwater (B_{FW}) components defined as

$$B = B_{\mathrm{HF}} + B_{\mathrm{FW}} \equiv \alpha g \frac{Q_{\mathrm{HF}}}{c_p \rho_0} + \beta S g Q_{\mathrm{FW}}, \quad (1)$$

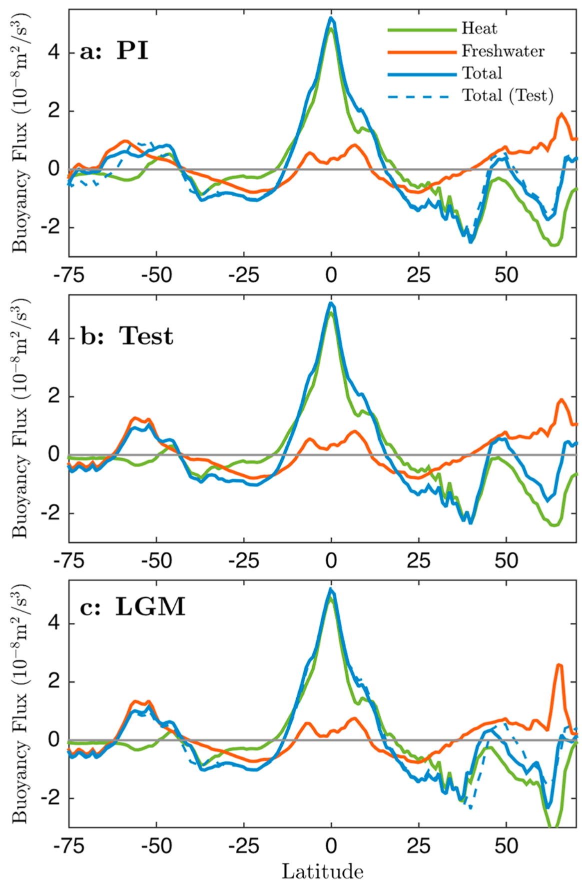

Figure 1. (a–c) Zonal-mean surface buoyancy flux for the three model runs along with its heat and freshwater components. The total buoyancy flux from the Test run is plotted for comparison in Figures 1a and 1c as a blue dashed line.

where g is the gravitational acceleration, ρ_0 is a reference density, c_p is the specific heat of seawater, S is the ocean surface salinity, α is the thermal expansion coefficient, β is the saline contraction coefficient, Q_{HF} is the net air-sea heat flux (positive for ocean heat gain), and Q_{FW} is the net freshwater flux (positive for ocean freshwater gain). The freshwater flux is approximately fixed by the prescribed forcing in each run. It is mainly associated with sea ice melting and freezing, river runoff, precipitation, and evaporation, and all but the last of these fields are fully specified in the simulations. The surface heat flux in these simulations, on the other hand, more closely resembles a restoring boundary condition [cf. *Haney*, 1971]. The freshwater flux does include a "weak restoring" component to avoid unbounded local salinity trends under mixed boundary conditions [*Griffies et al.*, 2009], but this component does not appear to substantially influence the results presented here, as discussed in the supporting information. Figure 1 shows that south of 45°S, the buoyancy flux is mostly dominated by the freshwater flux, implying that the Southern Ocean is subject to a surface buoyancy flux that is approximately fixed, i.e., independent of ocean state.

Note that the form of the surface buoyancy flux (restoring boundary condition or fixed buoyancy flux) has been shown to strongly influence the response of the deep ocean to surface forcing perturbations. In an eddy-resolving channel model, *Abernathey et al.* [2011] found different sensitivities of the overturning circulation to surface wind stress between simulations with fixed buoyancy flux and those with restoring boundary conditions, as was similarly found in a conceptual model by *Stewart et al.* [2014].

3. CESM Simulation Results

3.1. Stratification and Overturning Circulation

We first discuss the mean stratification and overturning circulation in the Test simulation, introducing a conceptual decomposition of the domain into three dynamically distinct regions in order to facilitate interpretation of the results. We focus our analysis on the Atlantic basin because, due to the formation of the NADW, the Northern Hemisphere surface forcing is expected to have more influence on the deep ocean stratification in the Atlantic basin than in the Pacific and Indian basins. A meridional section of σ_2 (i.e., potential density referenced to 2000 dbar) that is zonally averaged between 25°W and 35°W in the Test run is presented in Figure 2b, with the residual overturning circulation streamfunction in the Atlantic Ocean included as black contours.

By comparing the overturning circulation streamfunction to the potential density, we identify three distinct isopycnal regions in the Atlantic Basin which are separated by isopycnal surfaces ρ_1 and ρ_2. This is shown schematically in Figure 2c. Here ρ_2 is defined as the density of the isopycnal that separates the upper and lower overturning circulation cells. As shown in Figure 2a, it also coincides with the border between the regions of

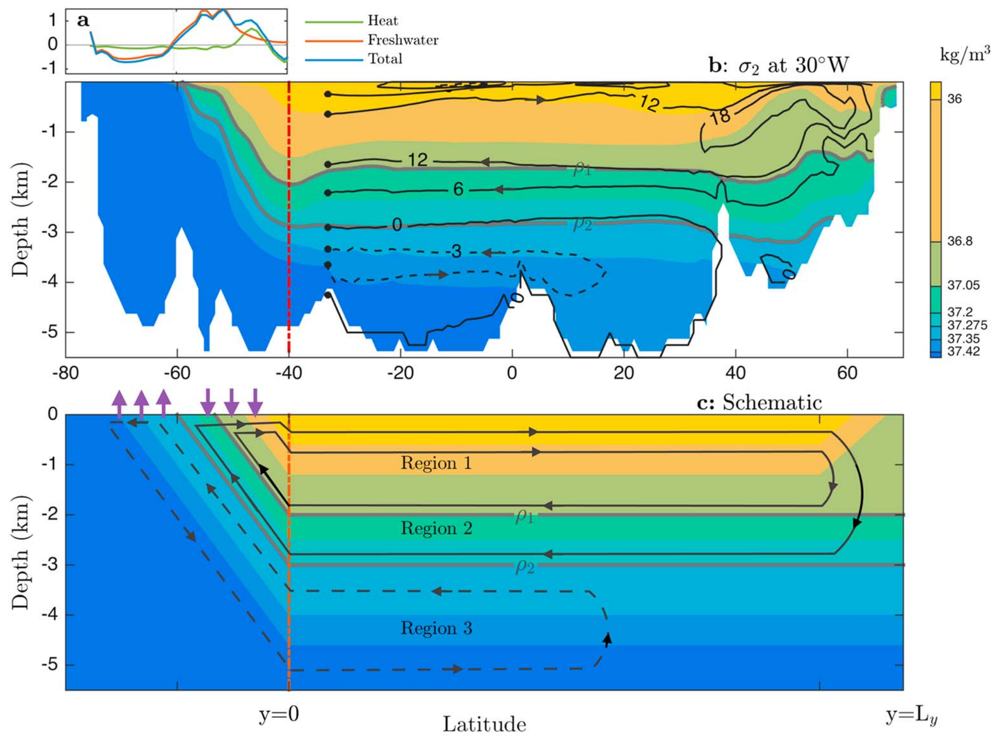

Figure 2. (a) Buoyancy forcing averaged between 25°W and 35°W in the Southern Ocean in the Test run, plotted in units of 10^{-8} m^2/s^3. (b) Meridional section of σ_2 (shading) from the Test run, averaged zonally between 25°W and 35°W. The residual overturning circulation streamfunction in the Atlantic Ocean, calculated on σ_2 surfaces and then mapped back to depth coordinates, is included as black contours with arrows indicating the direction of flow. (c) Schematic of the isopycnals (shading) and overturning circulation (black lines with arrows). Purple arrows in the Southern Ocean indicate the direction of buoyancy flux, with ocean buoyancy loss indicated by upward arrows. The northern boundary of the Antarctic Circumpolar Current (ACC) is indicated in panels b and c by a red dash dotted line. The thick gray lines represent the isopycnals that separate these 3 regions (ρ_1 and ρ_2). Here ρ_2 is defined as the density of the isopycnal surface that separates the upper and lower overturning circulation cells, and ρ_1 is defined more approximately as the density of the isopycnal below which the isopycnal surfaces are approximately flat and hence are not substantially affected by the near-surface wind-driven circulation.

buoyancy loss and gain in the Southern Ocean in the long-term mean, which is approximately 10° south of the westerly wind maximum in the Atlantic Sector of the Southern Ocean. We define ρ_1 as the uppermost isopycnal surface that outcrops in the Southern Ocean but not in the Northern Hemisphere in the long-term mean. Below ρ_1, the isopycnal surfaces are nearly flat in Figure 2b, implying that they are not substantially affected by the surface wind-driven circulation.

In Region 3, isopycnals outcrop only in regions in the Southern Ocean where the ocean loses buoyancy at the surface (Figure 2b). Region 3 coincides with the depths spanned by the counterclockwise lower overturning circulation cell (ψ<0). In Region 2, which represents the middepth ocean, isopycnals outcrop only in regions in the Southern Ocean where the ocean gains buoyancy in the long-term mean (Figure 2b), although they occasionally outcrop in the high-latitude North Atlantic during the winter season. In both Region 2 and Region 3, isopycnals are approximately flat except in the Southern Ocean, and hence, they do not appear to be affected by the wind-driven circulation except in the Southern Ocean.

Region 1 spans from the top of Region 2 to the surface. Here isopycnals outcrop in both the Southern Ocean and the high-latitude North Atlantic in the long-term mean, and the influence of the wind-driven circulation becomes apparent particularly in the subpolar gyre of the North Atlantic (40–60°N). In the PI and LGM model runs, we identify analogous regions and adjust the potential densities ρ_1 and ρ_2 to match the isopycnals that separate them (see Figure S8 for the PI and LGM).

3.2. Role of Southern Ocean Surface Forcing

We now compare the basin-average stratification in the Atlantic basin between 20°S and 20°N. The result is presented in Figure 3 as the squared Brunt-Väisälä frequency N^2, which is reported in CESM as $N^2 \equiv -\frac{g}{\rho_0}\frac{\partial \sigma_{p_r}}{\partial z}$,

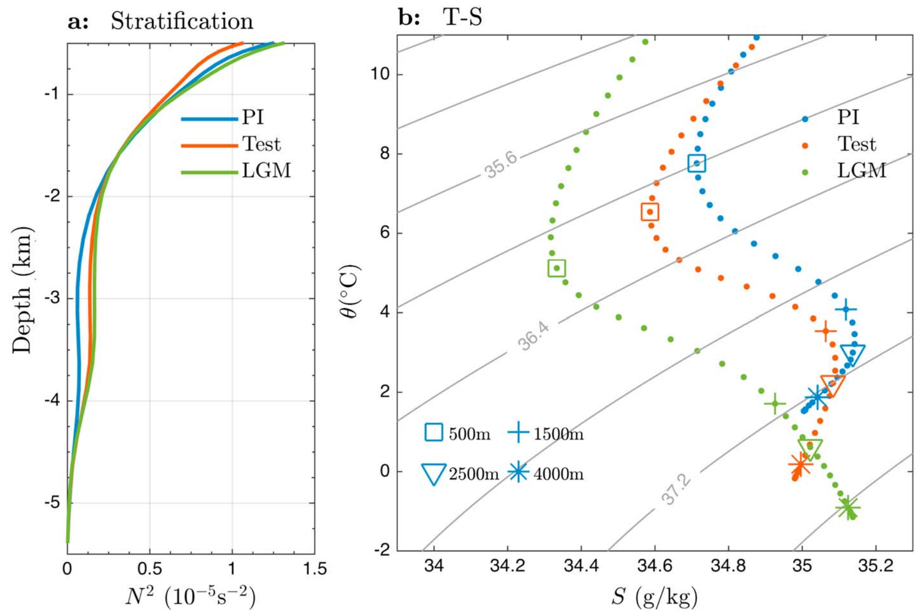

Figure 3. (a) Stratification (represented by the squared Brunt-Väisälä frequency N^2) averaged laterally between 20°S and 20°N in the Atlantic Ocean. (b) Temperature-salinity diagram for the three runs averaged between 20°S and 20°N in the Atlantic Ocean. Contours of σ_2 are indicated. The differences in σ_2 between 500 m and 1500 m are 0.99, 0.89, and 1.01 kg/m^3 in the PI, Test, and LGM runs, respectively; the differences between 2500 and 4000 m are 0.09, 0.20, and 0.27 kg/m^3, respectively.

where σ_{p_r} is the potential density referenced to the local pressure. The Test run closely reproduces the deep ocean stratification of the LGM run below approximately 2000 m, but not between 500 m and 1500 m. This indicates that the influence of the Southern Ocean on the deep stratification extends much higher in depth level than previously thought [e.g., *Nikurashin and Vallis*, 2012], approximately 1000 m above the boundary between the upper and lower overturning cells in the Atlantic. The stratification in the other major ocean basins largely supports this conclusion, suggesting that the surface forcing in the Southern Ocean is responsible for the enhanced global deep ocean stratification during the LGM in CESM (see supporting information for details).

Next, we examine the thermal and haline components of the deep ocean density stratification. Figure 3b shows a temperature-salinity (T-S) diagram, averaged laterally over the Atlantic basin between 20°S and 20°N. This figure indicates that the density difference between 2500 m (triangles) and 4000 m (stars) is much smaller in the PI run than in the Test and LGM runs, consistent with Figure 3a. However, the deep ocean temperature and salinity stratification in the Test run are strikingly different from the LGM run, having a negative rather than positive deep salinity stratification that more closely resembles the PI run. Though the density stratification is more dynamically relevant, the temperature and salinity stratifications are also important because they influence the stored heat and solubility of the abyssal waters, thereby affecting the capacity for carbon storage in the ocean.

Hence, Figure 3 implies that although North Atlantic surface forcing does not substantially affect the deep ocean density stratification, it does strongly influence the global deep ocean temperature and salinity profiles. This occurs in such a way that the deep ocean temperature and salinity differences between the simulations have canceling contributions to the deep ocean density stratification. This may be because both isopycnal advection and diffusion can influence the temperature and salinity along isopycnals between the Southern Ocean surface and the abyssal ocean, whereas the deep ocean density stratification is constrained by the Southern Ocean surface forcing. Consequently, there is a degree of freedom in how temperature and salinity vary with depth in the deep ocean.

4. Conceptual Model

Previous idealized studies [e.g., *Nikurashin and Vallis*, 2012] suggested that the density stratification in what we identify as Region 3 is constrained by the surface buoyancy forcing in the Southern Ocean. This is because surface buoyancy restoring essentially fixes the density gradient at the surface, and the approximately constant isopycnal slope in the Southern Ocean maps this surface density gradient to the abyssal ocean density stratification. In Region 2, however, they suggested that the stratification is substantially influenced by North Atlantic surface forcing as well, in contrast with the result presented in section 3.

In this section we adapt the zonally integrated conceptual model of *Nikurashin and Vallis* [2011] to investigate why the stratification in Region 2 in the CESM simulations appears to be largely controlled by the Southern Ocean alone. As discussed below, we find that the approximately fixed surface buoyancy flux in the Southern Ocean exerts a strong control over the density stratification in both Region 3 and Region 2, even though Region 2 contains the southward flow of the NADW.

As derived in the supporting information Text S3.i, the deep ocean stratification (N^2) can be related to the Southern Ocean surface buoyancy forcing in the conceptual model via a buoyancy budget equation:

$$\frac{\kappa L_x L_y}{N^2(z)} \frac{\partial}{\partial z} N^2(z) = \psi^*(z) + L_x \frac{B(-z/s)}{s N^2(z)}. \tag{2}$$

This states that the net diffusively driven upwelling across a given depth (or isopycnal surface) in the interior basin (left-hand side of equation (2)) is equal to the net export of NADW below that depth at the northern end of the basin (ψ^*) plus the net transformation from lighter to denser water at the Southern Ocean surface due to the zonal-mean surface buoyancy flux (B). Here κ is the diapycnal diffusivity, L_y and L_x are the meridional and zonal length scales of the basin (as in Figure 2b), ψ^* is the residual overturning circulation streamfunction at the northern boundary of the basin (i.e., at $y=L_y$), and s is the isopycnal slope in the Southern Ocean. Note that positive values of B here correspond to positive buoyancy input to the ocean and that the isopycnal slope (s) is negative.

Motivated by Figure 1, we model the surface buoyancy forcing as a fixed flux that varies with latitude, $B=B(y)$. In equation (2), B is evaluated at the location at which an isopycnal lying at depth z north of the Southern Ocean outcrops at the Southern Ocean surface, $y=-z/s$ (cf. Figure 2c). In order to simplify the conceptual model, we assume that both κ and s are constant. We find that s is approximately identical among the three simulations discussed above [cf. *Gent and Danabasoglu*, 2011], which is consistent with the assumption of constant isopycnal slope in the conceptual model. In supporting information Text S3.ii, we present a more general analysis that allows the isopycnal slope to change in response to the strength of overturning circulation. Note that the depth dependence of κ has been shown to be important for aspects of the deep ocean stratification, especially close to the depth of bottom topography [*Mashayek et al.*, 2015].

Region 3 is defined to lie below the southward flow of NADW, so ψ^* vanishes in this region (see Figure S8). Equation (2) in Region 3 thus can be written as

$$\frac{\partial}{\partial z} N^2(z) = \frac{B(-z/s)}{\kappa s L_y}. \tag{3}$$

Figure 3a shows that the stratification at the ocean bottom (N^2_{bot}) is close to zero in all three simulations, i.e., $N^2_{\mathrm{bot}} \approx 0$. Therefore, the stratification N^2 at any depth z within Region 3 is equal to the vertical integral of the right-hand side of equation (3) from the ocean bottom up to that depth, and hence, it is solely determined by the Southern Ocean surface buoyancy forcing. Because $B_{\mathrm{LGM}} \approx B_{\mathrm{Test}}$ in Figure 1c, it follows that $N^2_{\mathrm{LGM}} \approx N^2_{\mathrm{Test}}$ throughout Region 3 in Figure 3, where the subscripts indicate the model run. It should be noted that this is true only because the buoyancy forcing takes the form of a fixed flux in equation (2): if a relaxation boundary condition were applied as in previous idealized modeling studies [e.g., *Wolfe and Cessi*, 2011; *Nikurashin and Vallis*, 2012], then the stratification in Region 3 would be at least slightly impacted by interhemispheric effects, as shown by *Fučkar and Vallis* [2007] and in equation (S7) in the supporting information.

This argument does not extend to Region 2, because the southward flow of NADW is nonzero there, so the ψ^* term in equation (2) does not vanish. Instead, it can be shown that in order to produce a substantial difference between the Test and LGM stratification in Region 2, a very large change in ψ^* would be required, which is much larger than the difference in ψ^* between the LGM and PI simulations. Rearranging equation (2) and taking the difference between the LGM and Test simulations, we obtain

$$\kappa L_y L_x \frac{\partial}{\partial z}\left(N^2_{\mathrm{LGM}}(z) - N^2_{\mathrm{Test}}(z)\right) = N^2_{\mathrm{LGM}}(z)\psi^*_{\mathrm{LGM}}(z) - N^2_{\mathrm{Test}}(z)\psi^*_{\mathrm{Test}}(z) \tag{4}$$

in Region 2. Here we have neglected the difference between the Test and LGM fixed surface buoyancy fluxes in the Southern Ocean, $B_{\mathrm{LGM}} - B_{\mathrm{Test}}$, because Figure 1c shows this term to be small.

At the boundary between Region 2 and Region 3 ($\sim$3000 m depth), Figure 3a indicates that the stratification at this depth is approximately equivalent between the LGM and Test simulations, i.e., $\Delta N^2 \equiv N^2_{\mathrm{LGM}} - N^2_{\mathrm{Test}} \approx 0$. Qualitatively, in order for the terms on the right-hand side of equation (4) to produce a vertical change in ΔN^2 of order N^2, the difference between the NADW transports $\left(\Delta\psi^* \equiv \psi^*_{\mathrm{LGM}} - \psi^*_{\mathrm{Test}}\right)$ in Region 2 must be large. Scaling arguments suggest that this requires $\Delta\psi^* \sim \kappa L_y L_x / H_2$, where $H_2 \approx 1000$ m is the vertical thickness of Region 2 (see supporting information for details). For typical oceanic parameter values, this requires a change in the NADW transport streamfunction $\Delta\psi^*$ of $\mathcal{O}$(10 sverdrup (Sv)). However, the strength of the streamfunction in this region is less than 10 Sv in the LGM and Test simulations, with the difference between the two being only $\Delta\psi^* \sim 2$ Sv. Thus, in the absence of extreme perturbations to the high-latitude Northern Hemisphere surface forcing, the Southern Ocean essentially controls the stratification throughout Region 2, consistent with the CESM result (Figure 3). This is also true when we relax the assumption of constant isopycnal slope (see supporting information Text S3.ii).

In Region 1, where the isopycnals outcrop in both the Southern Ocean and the North Atlantic, the ocean stratification is expected to be affected by a variety of processes, including the wind-driven gyre circulation and surface forcing in the high northern and southern latitudes [*Wolfe and Cessi*, 2011].

Conceptually, the analysis above suggests that the stratification in Region 3 is constrained by the requirement that all buoyancy loss by density classes at the surface in the Southern Ocean south of the outcrop position of ρ_2 must be balanced outside of the Southern Ocean by the net interior diffusive buoyancy flux across ρ_2. This argument can almost be extended to Region 2, except that the injection of NADW also contributes to the buoyancy budget in this region. However, because the southward NADW transport in Region 2 needs to change by much more than it does between the LGM and PI runs to substantially impact the stratification, this contribution from NADW can thus be thought of as essentially constant. Consequently, the surface buoyancy flux in the Southern Ocean provides a strong control of the stratification up to $\sim$2000 m depth, as the CESM simulations indicate. This stands in contrast with previous idealized modeling studies [e.g., *Nikurashin and Vallis*, 2012; *Wolfe and Cessi*, 2011], where the stratification in the depth range that we identify as Region 2 is affected by the Northern Hemisphere surface forcing as well.

We emphasize that this conceptual model provides only an approximate qualitative picture of the effect of Southern Ocean surface buoyancy forcing on the global deep ocean stratification. The simplifications involved in the conceptual model make it difficult to find direct quantitative points of contact with the CESM simulations. For example, as shown in Figures S4 and S5 of the supporting information, the stratification profiles in the Pacific and Indian Oceans look different from the Atlantic. Understanding of this difference would require knowledge of the three-dimensional global overturning circulation, which is not included in the zonal-mean representation of the conceptual model.

5. Summary

The CESM ocean-only simulations presented here suggest that surface buoyancy forcing in the Southern Ocean largely controls the response of the abyssal stratification to LGM conditions. This is superficially consistent with previous understanding [*Nikurashin and Vallis*, 2011, 2012]. However, we furthermore find that this control extends up to approximately 2000 m depth, which is close to the core of the upper overturning circulation cell in the Atlantic. This is much shallower than expectations based on previous idealized modeling studies, which found the stratification above the abyssal ocean (i.e., in the middepth) to be substantially affected by North Atlantic surface forcing [e.g., *Nikurashin and Vallis*, 2012; *Wolfe and Cessi*, 2011]. We interpret the simulation results using a zonally integrated conceptual model. The analysis suggests that the control of the Southern Ocean surface buoyancy forcing over the global deep ocean stratification relies crucially on the

Southern Ocean surface buoyancy flux being dominated by approximately fixed freshwater fluxes. This is in contrast with previous idealized modeling studies [e.g., *Nikurashin and Vallis*, 2012; *Wolfe and Cessi*, 2011], in which the control of the Southern Ocean surface buoyancy forcing over the global deep ocean stratification relies on restoring thermal fluxes. This change in the form of the surface buoyancy forcing extends the control of the Southern Ocean surface forcing up to the core of the NADW overturning circulation cell.

In contrast to deep ocean density stratification, however, we find that although North Atlantic surface forcing does not substantially affect the deep ocean stratification, it does strongly influence the global deep ocean temperature and salinity profiles. In other words, the North Atlantic forcing causes temperature and salinity changes which have canceling contributions to the density. The temperature and salinity stratifications are important because they influence the stored heat and solubility of the abyssal waters, thereby affecting the capacity for carbon storage in the ocean.

In this study we used the ocean component of a single comprehensive climate model, and it is possible that other models may exhibit different responses to similar changes in the surface forcing. For example, the response may depend on the choice of parameterization scheme for unresolved mesoscale eddies [e.g., *Munday et al.*, 2013] and gravity currents [e.g., *Legg et al.*, 2009]. Running CCSM3.5 at an eddy-permitting resolution, *Bryan et al.* [2014] found that the simulated Southern Ocean processes are substantially different than a standard-resolution simulation. The parametrization of diapycnal mixing induced by tidally generated internal waves may also need to be modified to accurately simulate the LGM ocean [*Green et al.*, 2009]. Furthermore, it should also be noted that we are unable to isolate the influence of the Southern Ocean surface wind forcing in the model as it is varied together with the surface buoyancy forcing.

In conclusion, these results suggest that Southern Ocean surface freshwater forcing is largely responsible for the global deep ocean stratification differences between the LGM and PI climates. Considering the influence of deep ocean stratification on CO_2 outgassing [e.g., *Bouttes et al.*, 2009; *Adkins*, 2013], this implies that Southern Ocean surface freshwater forcing plays a central role in glacial-interglacial changes in atmospheric CO_2 concentration. It also implies that Southern Ocean surface freshwater forcing may have a strong influence on the deep ocean stratification and CO_2 storage in future climate change scenarios.

Acknowledgments
This work was supported by National Science Foundation grant OCE-1357078 and OCE-1538702. Without implying their endorsement, we thank Till Wagner, Shang-Ping Xie, Paola Cessi, Maxim Nikurashin, and Tim Merlis for their helpful comments and discussions. We are also grateful to Frank Bryan and David Bailey for technical help with the CESM model setup. Files related to the setup of the CESM simulations, as well as processed model output, are available at http://eisenman.ucsd.edu/code.html. Unprocessed model output is available by request from the corresponding author.

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