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Sensitivity of the Southern Ocean overturning circulation to surface buoyancy forcing

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[1] The sensitivity of the Southern Ocean overturning to altered surface buoyancy forcing is investigated in a series of eddy-permitting, idealised simulations. The modelled response indicates that heat and freshwater fluxes in the Southern Hemisphere mid-latitudes may play a significant role in setting the strength of the overturning circulation. Enhanced buoyancy fluxes act to increase the meridional overturning up to a limit approaching the wind-driven Ekman transport. The sensitivity of the overturning to surface buoyancy forcing is strongly dependent on the relative locations of the wind stress profile, buoyancy forcing and upwelling region. The numerical simulations provide support for the hypothesis that changes in upwelling during deglaciations may have been driven by changes in heat and freshwater fluxes, instead of, or in addition to, changes in wind stress. **Citation:** Morrison, A. K., A. M. Hogg, and M. L. Ward (2011), Sensitivity of the Southern Ocean overturning circulation to surface buoyancy forcing, *Geophys. Res. Lett.*, *38*, L14602, doi:10.1029/2011GL048031.

1. Introduction

[2] The southern limb of the oceanic meridional overturning circulation plays a key role in the Earth's response to climate change. The rise in atmospheric CO_2 at glacial-interglacial transitions has been attributed to outgassing of enhanced upwelling water masses in the Southern Ocean [*Skinner et al.*, 2010], while subducting water masses currently represent a significant fraction of the global anthropogenic CO_2 sink [*Gruber et al.*, 2009]. Despite the significance of the Southern Ocean overturning to past and future climate change, the dynamics and sensitivity to forcing changes are poorly understood.

[3] The upper cell of the Southern Ocean meridional overturning is commonly considered to be a wind-driven circulation, with dense waters upwelling along steeply tilted isopycnals, driven by divergent Ekman transport at the surface [*Toggweiler and Samuels*, 1993]. Coarse resolution modelling studies support this theory, showing overturning increasing in proportion to westerly wind stress [*Saenko et al.*, 2005]. More recently, investigations into the response of the overturning to changing wind stress in eddy-permitting models have revealed the significance of the eddy field to the overturning dynamics; increases in Ekman transport are at least partially compensated by increases in opposing eddy-induced transport, implying only a weak sensitivity of the net residual overturning transport to wind stress [*Hallberg and Gnanadesikan*, 2006; *Viebahn and Eden*, 2010; *Farneti et al.*, 2010]. The focus of these modelling studies has been on the effect of wind stress forcing on the overturning, while very little modelling work has examined the response of the overturning to changes in surface buoyancy forcing.

[4] Air-sea fluxes of heat and precipitation over the Antarctic Circumpolar Current (ACC) region are necessary for the conversion of dense upwelled water masses into lighter waters at the surface [*Speer et al.*, 2000]. Theoretical frameworks, based on zonally averaged representations of the meridional overturning circulation [*Marshall and Radko*, 2003; *Radko and Kamenkovich*, 2011], show the magnitude of the overturning increasing linearly with the net buoyancy flux supplied to the surface mixed layer. While the effect of mid-latitude buoyancy fluxes on the overturning has not been extensively tested in an eddy-permitting numerical model, coarse resolution simulations have highlighted the link between southern hemisphere mid-latitude freshwater forcing and the formation of deep water in the North Atlantic [*Saenko et al.*, 2003]. It has also been proposed that the increase in atmospheric CO_2, observed at the end of the last glacial maximum, was a result of increased upwelling in the Southern Ocean, driven by an increased surface buoyancy flux [*Watson and Naveira Garabato*, 2006]. In addition, observational studies have shown that heat and freshwater fluxes into the Southern Ocean have increased over the past few decades [*Böning et al.*, 2008] and modelling studies suggest that this trend will continue with future climate change [*Bracegirdle et al.*, 2008]. However, due to the sparse coverage of flux data in the Southern Ocean and the difficulties of investigating buoyancy forcing in coupled climate models, the dependence of the meridional overturning on surface buoyancy forcing has not been fully explored.

[5] In this paper we investigate the dependence of the overturning on surface buoyancy forcing in the mid-latitudes through a series of eddy-permitting, idealised simulations of the Southern Ocean.

2. Model

[6] We use GOLD, the primitive equation, isopycnal layered ocean model of *Hallberg and Gnanadesikan* [2006], which is a recent adaptation of the Hallberg Isopycnal Model [*Hallberg*, 1995]. The idealised domain is a zonally reentrant, 40° wide sector of the Southern Ocean with a simple Drake Passage-like sill (Figure 1a). A Mercator grid, with grid size decreasing towards the southern boundary, is employed at two different horizontal resolutions ($\frac{1}{4}$° and $\frac{1}{8}$°), resulting in square grid cells of size 14 km ($\frac{1}{4}$°) and 7 km ($\frac{1}{8}$°) at 60°S. The latitudinal extent is from 70°S to the equator. A sponge of width 2° at the equator relaxes the density stratification to observational data [*Levitus*, 2010] at the northern boundary with a decay timescale of 1 day. This allows an effective

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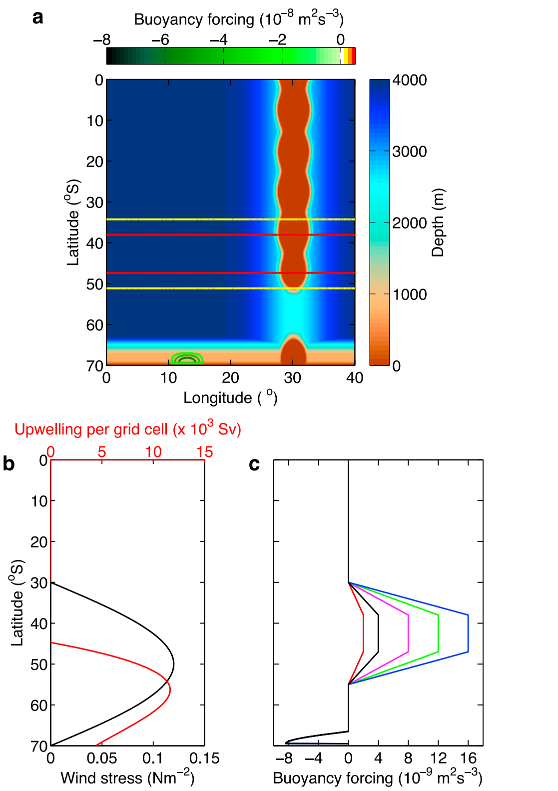

Figure 1. (a) Bathymetry, with spatial distribution of buoyancy forcing shown in contours. (b) Zonally averaged wind stress (black) and theoretical Ekman-induced upwelling (red). (c) Zonally averaged buoyancy forcing (control in black, perturbations in colour).

parameterisation of North Atlantic Deep Water (NADW) formation, without constraining incoming velocities. Bathymetry consists of an Antarctic shelf and Drake Passage-like sill, providing an unblocked circumpolar passage down to 2500 m below the surface, with a maximum ocean depth of 4000 m. A three layer bulk mixed layer is used, in addition to nine interior constant density layers (shown in Figure 2a). Biharmonic viscosity with an additional Smagorinsky component provides numerical closure. Biharmonic coefficients are set to 2×10^{11} m^4 s^{-1} ($\frac{1}{4}°$) and 2×10^{10} m^4 s^{-1} ($\frac{1}{8}°$). A weak diapycnal diffusivity of 10^{-6}m^2 s^{-1} in the interior ensures a largely adiabatic circulation. The control cases are spun up for ~120 years (Figure 2c). Perturbations from the end of the control spin-up are allowed to reach equilibrium (~30 years), and 25 year averages after these spin-up times were used for the results presented in this paper.

[7] The modelled residual overturning is forced by zonally and temporally invariant, sinusoidal winds, with a maximum of 0.12 Nm^{-2} at 50°S (Figure 1b). The idealised buoyancy forcing used in the model (Figures 1a and 1c) is based on net heat and freshwater surface fluxes derived from several observational and reanalysis products. Sparse data coverage (and loosely constrained reanalyses) over the Southern Ocean is no impediment to this study, as our framework tests the sensitivity of the overturning to forcing changes, rather than simulating the exact structure and magnitude of the overturning. The region of mid-latititude positive buoyancy forcing (i.e., acting to decrease the density of surface waters) in the control run has a magnitude of 4×10^{-9} m^2 s^{-3}, incorporating both heat and freshwater fluxes. The region of negative buoyancy forcing has a zonal average peaking at -8×10^{-9} m^2 s^{-3}, but is spatially localised (a necessary condition for the formation of an Antarctic Bottom Water cell in this model). We use a fixed, non-interactive buoyancy flux, to allow precise control of surface buoyancy forcing and an increased ability to isolate dynamical processes.

3. Results

[8] The surface forcing generates a Southern Ocean-like state (Figure 2) with tilted isopycnals, an energetic eddy field with associated fronts and jets and two overturning cells which broadly resemble observations [*Speer et al.*, 2000]. The lower Antarctic Bottom Water cell is driven principally by the negative buoyancy forcing near the southern boundary. The ACC transport in the model is 108 Sv in the $\frac{1}{4}°$ runs and 91 Sv in the $\frac{1}{8}°$ runs. The upper meridional overturning cell simulates the upwelling of NADW in the latitudes unblocked by bathymetry. Scaled up to the full width of the Southern Ocean, the transports of the upper and lower cells in Figure 2d would be 23 Sv and 9 Sv respectively. The meridional overturning in the model is largely adiabatic in the interior, with diapycnal transport occurring predominantly in the mixed layers, as allowed by the surface buoyancy forcing and diapycnal eddy transports.

[9] As the mid-latitude surface buoyancy forcing is increased, so does the overturning in the upper cell (Figure 3), while the lower overturning cell is unaffected. Due to the adiabatic nature of the overturning, the changes in a single profile of the streamfunction (as shown in Figure 3a) are representative of the changes in the structure of the entire overturning cell. When surface buoyancy forcing is weak, the modelled overturning is substantially less than the magnitude of the wind-driven Ekman transport (Figure 3b). At progressively larger values of buoyancy forcing, the overturning approaches a limit close to the value of the Ekman transport (offset by the opposing eddy-induced transport). The simulated $\frac{1}{8}°$ overturning is less sensitive to buoyancy forcing than the overturning in the $\frac{1}{4}°$ model. Though it may appear that the reduced sensitivity in the $\frac{1}{8}°$ model is a result of eddy compensation, this is not the case; the eddy kinetic energy (EKE) is nearly insensitive to buoyancy forcing. In the $\frac{1}{8}°$ model, a doubling of buoyancy forcing increases EKE by less than 5%, whereas a doubling of wind stress increases EKE by more than 100%.

[10] Increased buoyancy forcing results in an expected lightening of surface waters and an associated southward shift in the location of isopycnal layer outcropping (Figure 4a). The structure of the overturning in the interior remains largely unchanged and therefore the upwelling region also shifts southward, following the movement of the outcropping. Note that the data shown in Figure 4a indicates roughly the latitude of the northern edge of the upwelling, which is spread over a 5°–8° latitude band. The limiting behaviour seen in Figure 4a arises due to the restricted range of the mid-latitude buoyancy forcing. There is nearly no change in the latitude of outcropping south of 50°–55°S (Figure 4b). Therefore the upwelling shifts southward only so far as the mid-latitude buoyancy perturbations extend. The reduced surface density at high latitudes in the $\frac{1}{8}°$ model (as indicated by Figure 4 and

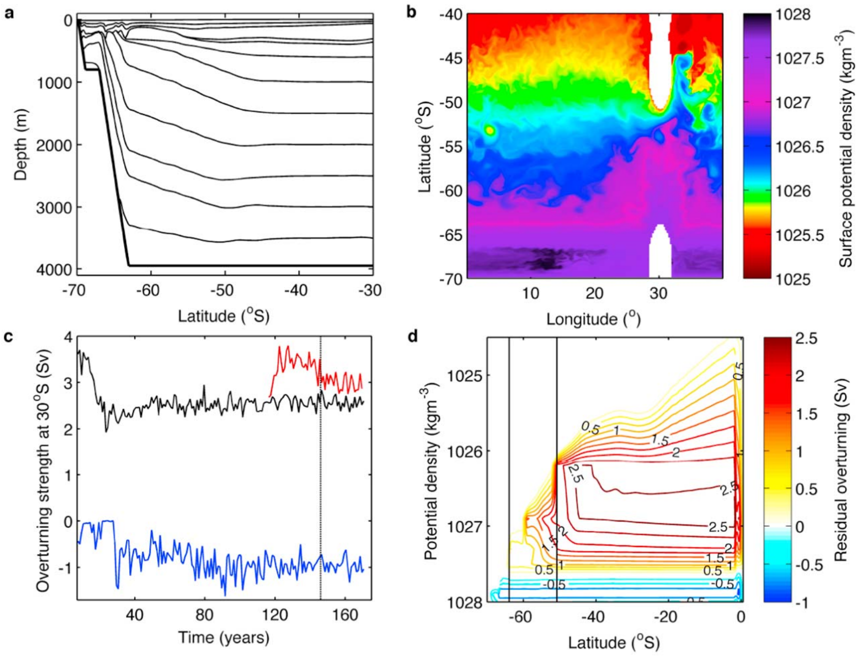

Figure 2. (a) Zonally and temporally averaged isopycnal layers for the $\frac{1}{8}°$ control simulation. (b) Snapshot of surface potential density for the $\frac{1}{8}°$ control simulation. (c) Timeseries of $\frac{1}{8}°$ overturning spinup. The maximum of the upper (lower) cell in the control run at 30°S is shown in black (blue), with the buoyancy forcing perturbation 16×10^{-9} m^2 s^{-3} shown in red. The vertical line indicates the point after which the 25 year averages were calculated. (d) Residual overturning streamfunction in density space for the $\frac{1}{8}°$ control simulation. Positive values (red) indicate clockwise transports, while negative values (blue) indicate anti-clockwise transports. The vertical black lines show the extent of the unblocked circumpolar passage (above the sill). Horizontal sections of the streamfunction imply adiabatic transport along isopycnal layers. Since the isopycnals are strongly tilted in the south, this also indicates upwelling.

the 25% reduction in ACC transport compared with the $\frac{1}{4}°$ model) may be linked to an enhancement of the eddy field at higher resolution. EKE is around 40% higher in the $\frac{1}{8}°$ model and the increased eddy field acts to reduce the tilt of isopycnal layers near the surface, resulting in more southerly outcropping in the higher resolution case.

4. Discussion

[11] These idealised model results indicate that the overturning circulation is sensitive to mid-latitude buoyancy forcing. Increases in buoyancy forcing act to shift the location of upwelling southward, while enhancing the overturning circulation. The enhanced upwelling is qualitatively consistent with theoretical predictions [*Marshall and Radko*, 2003; *Radko and Kamenkovich*, 2011] which indicate the need for water mass transformation to occur if upwelled water masses are to travel northwards in the surface layers without being immediately subducted. The change in upwelling location helps to control the sensitivity of the overturning to buoyancy forcing.

[12] The upwelling region in the model is determined by a number of factors, including bathymetry, the eddy field, wind and buoyancy forcing. The modelled upwelling therefore does not necessarily coincide with the location of the maximum potential Ekman-induced upwelling (profile shown in Figure 1b). As the upwelling location is forced southward by the mid-latitude buoyancy forcing perturbations, the modelled upwelling shifts from a region of low Ekman-induced upwelling, southward to a region of higher Ekman-induced upwelling. The Ekman-induced upwelling transport ($\nabla \times (\frac{\tau}{\rho_0 f})\ \Delta x \Delta y$) increases rapidly from zero at 45°S to a maximum at 56°S. The more northerly location of the upwelling in the $\frac{1}{4}°$ runs with low buoyancy forcing (Figure 4a) results in a strong sensitivity of the overturning to buoyancy forcing perturbations, as the upwelling moves through a region with rapidly increasing Ekman-induced upwelling. The reduced sensitivity observed in the $\frac{1}{8}°$ simulations may be traced to the increased stratification and the location of the upwelling region on the boundary of the applied mid-latitude buoyancy forcing. However, the profile of the Ekman-induced upwelling, which is nearly constant (<10% variation) over the range 52°–61°S, adds a further limit to the increase of the overturning circulation. Buoyancy forcing perturbations extending further south would not significantly alter the overturning sensitivity.

[13] In our model, the sensitivity depends on the relative locations of the wind stress, buoyancy forcing and upwelling. The greatest sensitivity occurred for low buoyancy forcing at $\frac{1}{4}°$ resolution, as the upwelling region was located furthest from the latitude of maximum Ekman-induced upwelling. The sensitivity for other models and the real ocean may vary widely from our results, depending on the stratification and the exact location and details of topography, wind stress and buoyancy forcing. However, we have highlighted that the Southern Ocean overturning may be strongly dependent on mid-latitude buoyancy forcing under certain conditions. Given the importance of the meridional overturning strength

for the future of the oceanic CO_2 sink, our results indicate an urgent need to improve monitoring of Southern Ocean buoyancy forcing and overturning.

[14] The sensitivity of the overturning to changes in buoyancy forcing presented here may play a role in solving the mystery of glacial-interglacial transitions. Radiocarbon evidence supports the hypothesis that the increased atmospheric CO_2 observed in Antarctic ice cores following the Last Glacial Maximum was released from deep water masses upwelling in the Southern Ocean [*Skinner et al.*, 2010], though the trigger and mechanism for the increased upwelling remains unclear. Current theories rely predominantly on strengthening and southward shifting winds as the driving force for increased upwelling [*Toggweiler et al.*, 2006]. However, both observations and models have cast doubt on this wind driven theory. Paleoreconstructions of wind using pollen and dust sources are sparse and inconsistent [*Fischer et al.*, 2010], while global coarse resolution modelling studies have shown that even a doubling in wind stress produces only ~35 ppm increase in atmospheric CO_2 [*d'Orgeville et al.*, 2010]. Higher resolution models are likely to show even less sensitivity to wind stress changes as a result of eddy compensation. On the contrary, it is plausible that large changes in buoyancy forcing occurred at the end of glacial periods, due to sudden changes in air temperature and sea ice extent. The dependence we have found of the overturning on surface buoyancy forcing provides evidence for the hypothesis that buoyancy forcing (in addition to changes in wind stress) may have played a major role in the enhanced upwelling and subsequent outgassing of CO_2 from the Southern Ocean during glacial-interglacial transitions.

[15] We have shown, using an idealised, but high resolution, model of the Southern Ocean, that surface buoyancy forcing, in addition to wind stress, may also be significant in setting the strength of the meridional overturning and that it may play an important role in past and future climate change. Increased mid-latitude buoyancy forcing may have led to enhanced Southern Ocean upwelling and therefore the rise in deglacial atmospheric CO_2 at glacial-interglacial transitions. Given the significance of the Southern Ocean CO_2 sink to future climate change, we have outlined the need for improved observations and an increased understanding of the dynamics of the meridional overturning circulation.

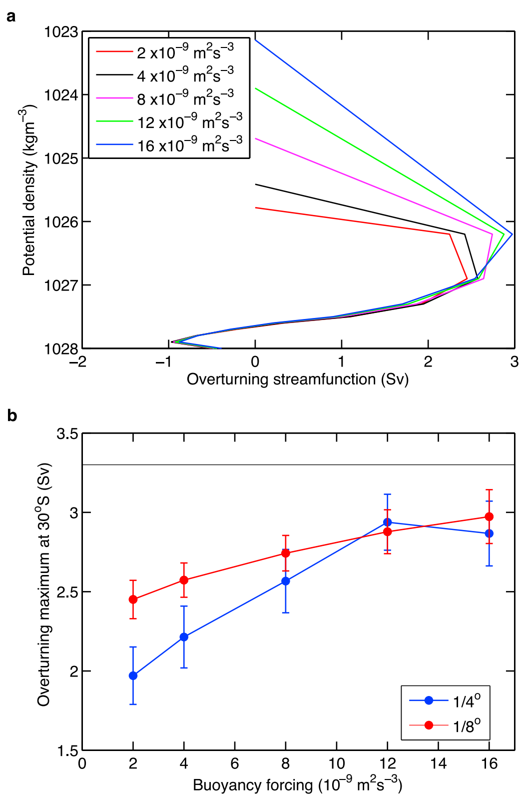

Figure 3. (a) Transects of the overturning streamfunction (such as shown in Figure 2d) at 30°S for a range of buoyancy forcing perturbations, run at $\frac{1}{8}°$ resolution. The upper (lower) overturning cell is represented by positive (negative) streamfunction values. (b) Maximum of the overturning streamfunction at 30°S for varied surface buoyancy forcing and resolution. The error bars show the standard deviation of 1 year overturning averages from the 25 year mean. The horizontal black line indicates the maximum of the analytically calculated Ekman transport.

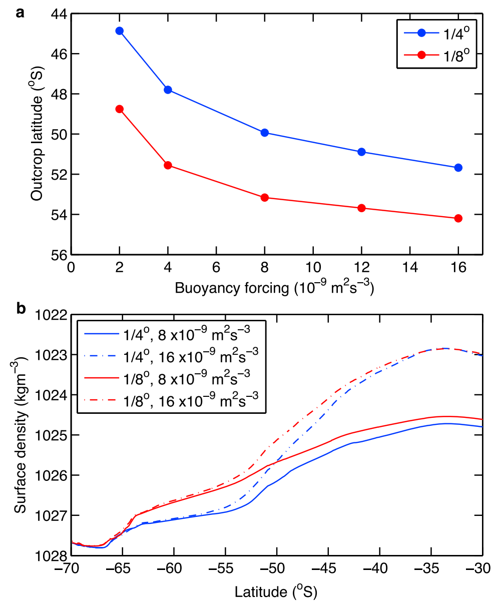

Figure 4. (a) The variation, with applied buoyancy forcing, of the latitude at which the 1 Sv overturning streamfunction outcrops into the upper isopycnal layer. (b) Mixed layer density for a range of simulations with varied resolution and buoyancy forcing. All values in this figure are zonal and temporal (25 year) averages.

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