Southern Ocean Deep Circulation and Heat Uptake in a High-Resolution Climate Model

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Abstract

The dynamics of the lower cell of the meridional overturning circulation (MOC) in the Southern Ocean are compared in two versions of a global climate model: one with high-resolution (0.1°) ocean and sea ice and the other a lower resolution (1.0°) counterpart. In the high-resolution version, the lower cell circulation is stronger and extends further northward into the abyssal ocean. Using the water-mass-transformation framework, it is shown that the differences in the lower cell circulation between resolutions are explained by: greater rates of surface-water-mass transformation within the higher resolution Antarctic sea ice pack; and by differences in diapycnal-mixing-induced transformation in the abyssal ocean.

While both surface and interior transformation processes work in tandem to sustain the lower cell in the control climate, the circulation is far more *sensitive* to changes in surface transformation in response to atmospheric warming from raising carbon dioxide levels. The substantial reduction in overturning can be attributed primarily to reduced surface heat loss. At high resolution, the circulation slows more dramatically, with an anomaly that reaches deeper into the abyssal ocean and alters the distribution of Southern Ocean warming. The resolution dependence of associated heat uptake is particularly pronounced in the abyssal ocean (below 4000m), where the higher-resolution version of the model warms 4.5 times more than its lower-resolution counterpart.

1. Introduction

The vast majority of the energy gained by the climate system during periods of global radiative imbalance is stored within the global ocean (e.g., Levitus et al., 2001; Abraham et al., 2013). The effective volume of ocean available to warm at a given time is set by the rate at which heat can be moved from the surface mixed layer into the deep ocean (e.g., Hansen et al., 1985), redirecting heat that would otherwise warm the surface. Therefore, deep ocean heat uptake plays a critical role in slowing the pace of global surface warming in response to greenhouse gas forcing (e.g., Gregory 2000; Held et al., 2010; Raper et al., 2002; Kostov et al., 2014).

The deep and abyssal ocean is filled by water masses formed at the high-latitudes; deep ocean heat uptake proceeds through the warming and redistribution of these polar-sourced water masses. The deep water mass formed at the Southern Ocean surface, Antarctic Bottom Water (AABW), comprises a large fraction of the oldest waters found in the global abyssal ocean (Johnson, 2008; Gebbie and Huybers, 2012). Consequently, the heat content of the global abyssal ocean is directly influenced by Southern Ocean processes. Indeed, the estimated 0.1 Wm^{-2} rate of global ocean heat uptake below 2000 m over the last 30 years was driven primarily by the warming of AABW. This warming is likely linked to changes in the surface ocean and climate around Antarctica (Purkey and Johnson, 2010, 2013). Further, the impact of ocean heat uptake is particularly high within the Southern Ocean, where it acts to curb surface warming that would otherwise be strongly amplified by the combined effects of the sea-ice albedo feedback and a stably stratified atmosphere (Armour et al., 2013).

While the Southern Ocean plays a fundamental role in climate change, the unique influence of small-scale processes on its circulation presents modeling challenges. Processes such as mesoscale eddy mixing, internal wave breaking, turbulent overflows, and convection have first order effects on the overturning circulation. Because these processes are often sub-grid scale in modern general circulation models (GCMs), they are either parameterized or underresolved. The simplest and most ubiquitous parameterization is simply grid-scale diffusion and viscosity, which is often prescribed by numerical considerations rather than physical ones. Deep ocean circulation and stratification are quite sensitive to diapycnal diffusion (Bryan, 1987; Cummins et al., 1990), and diffusion (including spurious numerical diffusion) itself is resolution dependent (Griffies et al., 2000; Hill et al., 2012, Urakawa and Hasumi, 2014). Coarse resolution models employ mesoscale eddy transport schemes, comprised of isopycnal diffusion (Redi, 1982) and eddy-induced advection (Gent and McWilliams,1990, hereafter GM). These parameterizations involve tuning parameters that strongly influence the circulation (Danabasoglu and McWilliams, 1995; Gnanadesikan, 1999; Pradal and Gnanadesikan, 2014; Gnanadeskian et al., 2015). Overflow parameterizations have been employed with limited success (Danabasoglu et al., 2012; Snow et al., 2015), as have abyssal tidal mixing schemes (Jayne, 2009). The recent advent of global mesoscale resolving/permitting ocean models allows some such parameterizations and other scale-dependent processes to be evaluated against more direct explicit simulation, especially with respect to the mesoscale (McClean et al., 2011; Griffies et al. 2015).

The potential for inadequate model resolution to bias Southern Ocean dynamics has garnered substantial attention in the literature. The focus has been mostly directed towards the behavior of the upper cell of the meridional overturning circulation (MOC) (see e.g., Henning and Vallis, 2005; Hallberg and Gnandesikan, 2006; Farneti et al. 2010; Abernathey et al., 2011; Gent and Danabasoglu, 2011; Bryan et al, 2014; Farneti et al., 2014; Gent, 2016). In contrast, the impact of model resolution on lower cell of the MOC and its relationship to Southern Ocean abyssal heat uptake is relatively unexplored. This is the aim of our paper.

The lower MOC is sustained by processes that make surface water denser, “pushing” water into the abyssal ocean, and processes that reduce the density of dense water in the interior, “pulling” it upwards again. In the abyssal ocean, water is “pulled” up via diapycnal mixing, the magnitude of which is likely a key control on abyssal overturning strength (Nikurashin and Vallis, 2011, 2012).

Such mixing is accomplished by the breaking of locally generated internal waves over bathymetric features, geothermal heating, remote tidal dissipation and entrainment across strong density gradients in localized overflows (Naveira Garabato et al., 2004, 2007; Nikurashin et al., 2012, de Lavergne et al., 2015). Transient eddy fluxes, which redistribute potential energy and result in the northward transport of dense water in the abyss, have also been linked to mixing rates and overturning strength because of their influence on abyssal stratification (Ito and Marshall, 2008). A misrepresentation of these abyssal diapycnal mixing processes could introduce biases into the lower cell dynamics.

The descending branch of the lower cell is sustained by buoyancy loss at high latitudes via the interaction of cold, salty water that is rejected during sea ice growth (especially in coastal polynyas and leads); extremely cold, fresh water formed from sub ice-shelf melting; and comparatively warm, salty Lower Circumpolar Deep Water (LCDW) that upwells near the coast (Orsi et al., 1999; Jacobs, 2004; Gordon, 2001). Polynyas and leads, which are key moderators of heat loss and brine rejection, form on spatial and temporal scales below the resolution of the current generation of GCMs (Stössel et al., 2007, Wilmott et al., 2007). Additionally, recent studies have suggested that transient eddies are key to the transport of AABW across the continental shelf front (where the deformation radius is on order 1-10 km), as is the input of momentum from coastal easterly winds (Stewart and Thompson, 2012, 2015a; Thompson et al., 2014). In sum, sea ice formation, the formation of dense plumes, transient cross-shelf flow, and sub ice-shelf processes all occur on scales smaller than the grid spacing of most GCMs; as a result, most models misrepresent the volume and formation rate of AABW. Indeed, Heuzé et al. (2013) found that no model participating in phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al., 2012) formed AABW via the sinking of shelf waters. Instead the majority of models formed bottom waters by (possibly spurious) open ocean convection, leading to large biases in the abyssal ocean density simulated in most of the CMIP5 models analyzed in their study.

Here, we explore how increasing model resolution in the ocean and sea ice components influences the dynamics of the lower cell of the MOC. We are motivated by the questions: does resolution alter the lower cell in the mean state? And, does resolution affect the rate of Southern Ocean heat uptake under greenhouse warming? Ours is, to the best of our knowledge, the first study to investigate the resolution dependence of this circulation in a coupled model. We use a GCM configured at resolution high enough to capture coastal polynyas and to explicitly resolve transient eddies throughout much of the global ocean, and compare the results to a lower resolution counterpart, configured at a resolution typical of CMIP5 models.

2. Model setup and analysis

2a. Model and Experiments

We use the Community Climate System Model version 3.5 (CCSM3.5), with ocean and sea ice at two resolutions, while maintaining identically configured atmosphere and land components. The high-resolution version (HR) is run with 0.1° resolution in the sea ice and ocean components. The comparatively low-resolution version (LR) is run at 1° resolution in the ocean and sea ice. The atmospheric component has a finite-volume dynamical core and is run in both cases at identical resolution: 26 vertical levels, with horizontal resolution of 0.47° x 0.63° (as in Gent et al., 2010). The horizontal resolution of the land model is the same as for the atmosphere. The HR and LR setups are identical to those used in Kirtman et al. (2012), Bitz and Polvani (2012), and Bryan et al. (2014).

HR has a sufficiently fine ocean horizontal grid to be deemed eddy resolving at low and mid latitudes and eddy permitting at very high latitudes (south of ~50° S), following Smith et al.

(2000). Grid spacing near the Antarctic coast is ~3-5 km, which we will show improves the resolution of sea ice dynamics. In contrast, LR is non-eddy resolving, and relies on the GM eddy parameterization with a spatially and temporally varying GM coefficient that depends on the square of the local buoyancy frequency (see Danabasoglu and Marshall, 2007, and Gent and Danabasoglu, 2011). No overflow parameterizations were employed in this model, and neither of the resolutions examined here exhibits grid-scale-size, full-depth open-ocean convection (a problem endemic to some GCMs, i.e., Heuzé et al., 2013). Control integrations of HR and LR were run with 1990's carbon dioxide mixing ratios for 167 yr. Kirtman et al. (2012) describe in depth the configuration and climate of the control integrations.

Perturbed runs were branched from the control integrations of each resolution (LR and HR) at year 77. Each perturbed run was subject to a 1% increase in the carbon dioxide mixing ratio of the atmosphere until carbon dioxide doubling was reached at year 147; thereafter, carbon dioxide mixing ratios were held fixed for the next twenty years. To minimize the impact of possible climate drift, we define the "response" or "anomaly" of a given field in a perturbed run as the difference between its average during the 20-year period after carbon dioxide doubling and its average during the contemporaneous 20 years in the control integration. Because of the computational expense at high resolution, our control runs are necessarily short relative to the equilibration timescale of the deep ocean, and there is still a small drift in deep ocean temperatures from the branch point to the end of the control runs at either resolution (see Kirtman et al., 2012). We find this drift has a minimal effect on isopycnal inflation (explained subsequently) in our control simulations. However, we cannot conclusively rule out that the features we diagnose are transient.

One limitation of this model, of particular relevance to this study, is the absence of interactive ice shelves. Instead, the mass balance of Antarctica is enforced by uniformly freshening and cooling the ocean at the continental margin by any precipitation reaching the Antarctic continent that causes the snow depth to exceed a maximum value of 1m, termed “ice runoff”. Raising carbon dioxide results in an increase in ice runoff owing to higher snowfall rates over Antarctica from increased poleward moisture transport. However, in this model, the reduction in sea ice formation and increase in precipitation over the ocean near Antarctica are a much greater source of anomalous freshwater than Antarctic ice or meltwater runoff (see Kirkman and Bitz, 2011).

2b. Isopycnal Overturning and Water Mass Transformation

A challenge to understanding the impact of model resolution on circulation is the certainty that increasing resolution will affect multiple processes simultaneously. We try to assess the relative importance of different processes in altering the circulation by employing the water-mass-transformation framework, first introduced by Walin (1982) and further developed in many subsequent studies (e.g., Speer and Tziperman, 1992, Marsh et al., 2000, Bryan et al., 2006, Iudicone et al., 2008). We consider the ocean circulation most directly related to stratification by calculating an overturning streamfunction along surfaces of constant density (Döös and Webb, 1994). We refer to the time average of this field

$$(1)\ \psi^{\sigma}(\sigma, y) \equiv -\frac{1}{(t_1-t_0)}\int_{t_0}^{t_1}\int_{x_W}^{x_E}\int_{-B(x,y)}^{z(x,y,\sigma,t)} v(x,y,z^*,t)\,dz^*dxdt$$

as the *isopycnal MOC,* where *v* is the total meridional velocity (including the bolus velocity in LR), σ is potential density, x is longitude (positive eastward), y is latitude (positive northward), z is depth (positive upward), *B* is the ocean bottom depth, and *t* is time. The component of the MOC from Eq. 1 with v equal to the time-mean velocity and σ equal to the time-mean isopycnal depth is the *mean isopycnal MOC,* ψ_m^σ. The impact of transient eddies on the isopycnal MOC (or the *transient eddy-induced isopycnal MOC,* ψ_E^σ) can be expressed as the difference between the total and mean MOC components:

$$(2)\ \ \psi_E^\sigma = \psi^\sigma - \psi_m^\sigma.$$

The transient eddy-induced MOC emerges as a result of explicitly resolved eddies in HR, while in LR, it is computed from the GM parameterized bolus velocity field using Eq. 1 with $v = v_{bolus}$. At either resolution, the impact of standing eddies and gyres as well as any steady correlations between density and non-zonally uniform flow can be expressed as the difference between the mean isopycnal MOC and a *depth-space* MOC, ψ^z (Dufour et al., 2012), the latter of which is calculated in the more traditional method relative to surfaces of constant depth.

The isopycnal circulation ψ^σ persists on a global scale because of the continual redistribution of seawater between density classes. As illustrated in Fig. 1, the volume of a region of the ocean, V, below and southward of isopycnal σ and latitude y varies with the volume flux across its boundaries. Considering the volume, V, south of any given σ and y, the volume inflation can be expressed as:

$$(3)\ \ \frac{d}{dt}V(\sigma, y, t) = \psi^\sigma(\sigma, y, t) - [F(\sigma, y, t) + Dmix(\sigma, y, t)],$$

where the latter two terms together represent water mass transformation, with $F(\sigma, y, t)$ induced by air-sea fluxes at the sea surface (*surface transformation)* and $Dmix(\sigma, y, t)$ induced by mixing across density surfaces in the interior (*interior transformation)*. A general definition of surface transformation, F_{gen}, is:

$$(4)\ F_{gen}(\sigma, t) = -\frac{\partial}{\partial\sigma}\iint_{A[\sigma^* > \sigma]} f_{surf}(x, y, t)dA.$$

where f_{surf} is the spatial distribution of surface density flux, a function of heat and freshwater fluxes (f_{heat} and f_{water}, defined positive downward):

$$(5)\ f_{surf}(x,y,t) = -\frac{\alpha}{C_p} f_{heat}(x,y,t) - \frac{\rho_o}{\rho_{fw}} \beta S_o f_{water}(x,y,t).$$

Here, α and β are the coefficients of thermal expansion and saline contraction, respectively, c_p is the specific heat of seawater, ρ_0 is a reference seawater density, ρ_{fw} is the density of freshwater, and S_o is a reference salinity. To directly compare surface transformation to the circulation at a given latitude, *y,* as in Eq. 3, we must consider only the net surface transformation occurring south of *y*, which we call simply *F*. This can be calculated as:

$$(6)\ F(\sigma,y,t) = F_{gen}(\sigma,t)\mathcal{H}\left(\sigma - \sigma_{min}(y,t)\right),$$

where $\mathcal{H}$ is the Heaviside function and σ_{min} is the lowest density to outcrop at latitude *y*. The term *F* is positive toward lighter densities for direct comparison with the MOC in the Southern Hemisphere. The interior transformation, $Dmix$, can be defined similarly to *F*, however doing so requires knowledge of time-varying, three-dimensional diffusive fluxes, which were not saved for these simulations due to data storage limitations. Instead, we follow Marsh et al. (2000) and calculate *Dmix* at latitude y as a residual of the other terms in Eq. 3. This effectively measures the path integrated interior transformation at each density between the surface and *y***.** With this method, we cannot identify the specific mixing processes responsible for interior diapycnal volume fluxes, though we can gain insight into the relative importance of surface and interior transformation on MOC strength.

It is important to note two caveats to our analysis. First, we use a reference pressure of 2000 dbar to calculate potential density, known as σ_2, for both our surface transformation and MOC calculations, since our focus is on the deep ocean; however, the choice of reference pressure may affect our results (Iudicone et al., 2008, Stewart and Thompson, 2015b). Secondly, due to prohibitive amounts of data generated in HR, we were restricted to saving monthly data, meaning that higher frequency transient behavior isn't captured in our analysis. Ballarota et al. (2013) explored the temporal and spatial scales at which transient behavior was most influential: though daily timescales were important to the upper MOC cell, their impact was much smaller in the lower MOC cell. Furthermore, in a spectral analysis of eddy heat fluxes in the same class of eddy-resolving model, Abernathey and Wortham (2015) found that sub-monthly variability makes a negligible contribution to the total heat flux. Thus, we expect to capture the majority of transient fluctuations most relevant for our analysis.

3. Control State Results

It is essential to understand the mean state of the Southern Ocean at each resolution to interpret how resolution affects the response to carbon dioxide forcing. Compared to LR, HR produces consistently saltier and colder waters on the continental shelf and throughout the deep Southern Ocean (Fig. 2a-f). In the abyssal ocean (below 4000 meters) waters are colder by approximately 0.45 °C and saltier by approximately 0.016 psu on average. Near the surface, HR forms a fresher branch of Antarctic Intermediate Water (AAIW) and warmer surface water. In comparison to World Ocean Circulation Experiment (WOCE: publically available at http://seahunt.ucsd.edu/find.html) hydrographic data, both model resolutions form deep waters that are too saline. This salinity bias is more pronounced in HR, which may be the result of non-local processes impacting the salinity of upwelling circumpolar deep water (CDW), such as by brine rejection from sea ice in the Northern Hemisphere (Kirtman et al., 2012). Bottom waters formed in HR are too cold in some regions, while bottom waters in LR are consistently too warm. An example of these large-scale properties is shown along an Indian Ocean transect (WOCE identification IO8 in Fig. 2g-l), though we note that the comparison between each model and observations varies significantly with region.

These zonally averaged properties arise from rich spatial structures. Irrespective of resolution, the densest waters in the abyssal ocean (Fig. 3c-d) outcrop in the coastal regions of the Ross and Weddell Seas (Fig. 3a-b). The waters extending from the high latitude surface into the abyssal ocean are notably denser in HR. The densest of these bottom waters emanate from the Ross and Weddell continental shelves, as is illustrated by the distribution of ocean bottom temperature and salinity (Fig. 4). Both the density and the density response to increased resolution are highest in these shelf regions, and both decrease with distance from the shelf following topographically driven AABW export pathways. These distributions imply that differences in surface properties propagate into the deep ocean from these locations.

To understand how differences in stratification and abyssal properties are manifested in the large-scale circulation, we examine the MOC south of the equator. In Fig. 5 we show the relative contributions towards the circulation ψ^{σ} from each of the components in Eq. 2. In our control simulation, the volume inflation contributes only a small term (on order 1 Sv) and thus we ignore it in this calculation. To visualize the spatial distribution of the large-scale isopycnal circulation, we have projected each component onto the depth of each mean isopycnal in Fig. 5.

We focus our analysis on differences in the lower cell, which emerge more robustly when the circulation is defined along isopycnals (ψ^{σ})**.** In light of the recent focus on model resolution in the upper cell, as mentioned in the introduction, it is noteworthy that in this model, features of the lower cell vary significantly with resolution. There are two distinct local minima in ψ^{σ}(see Fig. 5a-d). One minimum is associated with the export of dense surface waters into the abyssal ocean and their return flow, contained south of 55° S, which we define as the "sub-polar range" of the lower cell. This feature is weak in the depth-space overturning, likely because much of the export of dense water occurs in the Ross and Weddell Gyres, as is suggested by the distribution of bottom temperature and salinity. The second minimum, in what we refer to as the "abyssal range" of the lower cell to describe overturning north of 55° S and south of the equator, is associated with the circulation of dense waters north of the ACC. Across this latitude range, bottom waters must mix sufficiently with lighter waters to upwell at the polar surface. The apparent spatial separation of these circulation minima, seen in most climate models (Farneti et al., 2015), derives from the partial recirculation of Weddell and Ross Sea waters at high latitudes. This separation is much reduced by density transport achieved by eddy fluctuations (resolved or parameterized), evident in the transient eddy-induced streamfunctions shown in Fig. 5. This counterclockwise transient eddy-induced cell, ψ_E^σ, is strongest between 45°-60° S and reaches from the upper ocean to the full depth (Fig. 5 e and f). The lack of resolution dependence in the structure of ψ_E^σ suggests that the eddy parameterization in LR is well calibrated, though ψ_E^σ is stronger near the ocean bottom in HR, indicating that resolved eddies may alter the northward export of very dense AABW. Further, it is possible that we underestimate the strength of the HR eddy-induced cell by using monthly velocities in Eq. 1, particularly in the upper cell where daily correlations between velocity and temperature may become more important (Ballarotta et al, 2013).

In contrast to ψ_E^σ, there are striking resolution-dependent differences in the mean isopycnal overturning, ψ_m^σ (Fig 5 c versus d); these differences dominate the resolution-dependence of the total overturning, ψ^σ. Mean transport in the sub-polar range is stronger in HR (26 Sv versus 22 Sv in LR) and occurs on isopycnals that are deeper in the zonal average, corresponding to the production of more and relatively denser AABW in HR, consistent with the abyssal temperature and salinity fields. Mean overturning in the abyssal range extends further northward and remains stronger throughout the abyssal ocean in HR, with an overturning maximum of 20 Sv around 40° S. In contrast, the majority of lower cell overturning occurs south of 50° S in LR, north of which abyssal overturning becomes relatively weak (with a maximum of 6 Sv).

The influence of resolution on surface and interior processes driving diapycnal volume exchange can be compared via Eq. 3. Implicit to this analysis is the notion that both surface fluxes and interior mixing processes can change the density of seawater on a given isopycnal, and in turn, any change in density must drive a volume flux of seawater across isopycnals; the transformation (separated into components *F* and *Dmix*) describes these induced volume fluxes. The sign of the transformation denotes the direction of this volume flux, where a positive transformation is a volume flux towards lighter isopycnals. Any slope in the transformation rate indicates that the diapycnal volume flux differs across neighboring isopycnals, thus inducing a convergence or divergence of seawater volume into or out of a given density class. In steady state, the convergence [divergence] of volume into a density class will necessitate export [import] of water at that density, i.e. its formation [destruction].

The surface transformation function ($F(\sigma, y, t)$ in Eq. 3) at 30° S captures the surface regeneration of the major water masses of the Southern Ocean, specifically, the regions of net positive transformation and net negative transformation associated with the upper cell and lower cell, respectively (see Fig. 6). Waters in the small density range of the lower cell (which differs slightly between models: ~36.82 $\leq \sigma \leq$ 37.72 in HR and ~36.2 $\leq \sigma \leq$ 37.54 in LR) occupies ~70% of the volume of the Southern Ocean; we focus our analysis on a shared density range which encompasses the majority of this cell at both resolutions ($\sigma_{lower\ cell}$ defined as 36.72$\leq \sigma \leq$ 37.72). Downwelling AABW is formed across the range of net volume flux convergence (≥37.25 in HR and ≥37.12 in LR), so transformation across this density range is of particular importance to understanding what “drives” descent from the surface.

To diagnose the distribution of interior transformation, *Dmix,* we consider how ψ^{σ} varies with density at a given latitude y_o. We then compare ψ^{σ} to the other terms in Eq. 3 (recalling that volume inflation is negligible) at several latitudes over the density range of the lower cell; the degree to which ψ^{σ} and *F* differ will indicate the impact diapycnal mixing has had on the circulation at each latitude. Fig. 7 illustrates the relative strength of $\psi^{\sigma}(\sigma_{lower\ cell}, y_o)$, $F(\sigma_{lower\ cell}, y_o)$, and $Dmix(\sigma_{lower\ cell}, y_o)$ at y_o = 64° S, 38° S, and 30° S.

We first examine y_o= 64° S, chosen as a compromise between the northern extent of the Ross and Weddell shelves to capture how flow from the continental shelf into the abyssal ocean transforms water masses. Here, ψ^{σ} bears a close connection to F (Fig. 7a and b). Vigorous surface transformation leads to a peak in ψ^{σ} at σ~37.25, in HR and σ~37.15 in LR; this peak is modestly increased by the diapycnal mixing of very dense waters along their descent from the shelf into the abyss (i.e., $Dmix$ drives a volume flux of 6 Sv in HR and 8 Sv in LR from the densest waters into the peak flow in ψ^{σ}). Somewhat surprisingly, the magnitude and relative distribution across density classes of this $Dmix$ is similar between resolutions, though it is translated to slightly denser classes in HR because of the higher densities of waters formed at the surface. So, while a number of ocean processes may depend on resolution, together these processes don’t result in significant resolution dependence in diapycnal volume transport, and thus large-scale circulation patterns, south of this latitude. Though some export off the Weddell continental shelf may occur northward of 64° S, Fig. 5a and b indicates that much of the total descent has occurred by this latitude. We conclude that the flux of dense water into the high latitude abyssal ocean differs with resolution primarily because more of it is being made at higher densities at the surface in HR.

A similar breakdown of the streamfunction in the abyssal range of the lower cell can be made at 38° S (Fig. 7c and d). At either resolution, there is significant diapycnal mixing and associated adjustment of the isopycnal circulation between 64°-38° S. The mixing between 64°-38° S (dashed green line in Fig. 7c and d) destroys waters denser than σ~37.3 in HR and σ~37.2 in LR by 38 S. The impact of this mixing on the circulation differs markedly between models. By 38° S, mixing redistributes volume from very dense water (σ >~37.15) into the $37.15 \leq \sigma \leq 37.27$ range in HR, shifting the peak in ψ^{σ} towards σ~37.2 but maintaining substantial northward flow. In contrast, mixing in LR lightens waters across the entire density range of the lower cell by 38° S, inducing volume fluxes towards lighter isopycnals and greatly reducing the northward flow across all density levels evident in Fig. 5. At both resolutions, $Dmix$ overcomes F in waters lighter than σ~37.0, reversing the direction of flow and the sign of ψ^{σ}.

Examination of the streamfunction components at 30° S reveals that little additional mixing occurs from 38°-30° S at either resolution. There is a small volume flux towards lighter density classes across much of this density range at both resolutions, further reducing the magnitude of counterclockwise (negative) overturning in denser waters, and increasing the magnitude of clockwise (positive) overturning for lighter waters in $\sigma_{lower\ cell}$.

In sum, both surface transformation and interior transformation are key to sustaining the circulation across the latitudes sampled here, as is anticipated in a (approximate) steady state. However, the relative importance of surface and interior transformation processes, and resolution dependence therein, has a consistent spatial structure, one which may have important effects on the pattern and mechanisms of ocean heat uptake, as discussed in the next section.

To uncover the actual oceanic processes responsible for the distribution of water mass transformation, we first consider surface transformation in more depth. Fig. 6 illustrates the relative roles of heat and freshwater fluxes in transforming surface waters. Somewhat surprisingly, in light of the strong control salinity variations exert on the density of polar waters, heat loss contributes most significantly to buoyancy loss across $\sigma_{lower\ cell}$ at both resolutions (but especially so in HR). Salt input contributes significantly only at very high densities in this density range, again particularly so in HR, though sea ice melt and precipitation are important to the transformation of lighter waters. The spatial distribution of surface transformation per unit area, over several key density classes in $\sigma_{lower\ cell}$ elucidates why this is so (Fig. 8).

These "transformation maps," or distributions of diapycnal velocity, are constructed from the 20-year mean of monthly estimates and capture the covariance of isopycnal migration and surface fluxes. These distributions demonstrate the important control isopycnal surface area has on total transformation rates (recall Eq. 4). Both heat and salt fluxes induce strong density gain at very high densities in the coastal polynyas of the Weddell and Ross Seas (and the entire coastal Antarctic region at decreasing densities) because of the intensity of brine rejection and the reduced control of heat fluxes on the density of very cold water. However, the total area of these coastal polynyas is relatively small, especially in HR, so the associated transformation is not substantial. The greater impact of salt rejection in LR follows from the larger spatial scale over which sea ice is formed. Perhaps counter-intuitively, relatively lower surface density fluxes associated with heat loss over broader regions of the sea ice pack account for more total transformation because they act over a larger area. In LR, dense isopycnal outcrops migrate over a smaller area than in HR, such that even extreme heat losses near the coast drive less total transformation at equivalent densities. At either resolution, the Ross and Weddell Sea regions are of particular importance in part because of their southward extent: they host large areas of continental shelf and sea ice, through which dense isopycnals can maintain contact with the surface over much of the year.

Elevated rates of heat-loss induced transformation in dense waters in HR imply larger rates of diapycnal heat convergence into these density classes. Waters in the upper 1000 m are generally colder in HR under most of the ice pack in the Ross and Weddell Seas, aside from a small region of warmer sub-surface temperatures in the Ross Sea. The resolution dependency of heat transport likely results from changes in the regional circulation and diapycnal temperature gradients, though it is difficult to attribute such changes to one particular process. In both regions, the standard deviation of wintertime temperature is slightly greater in HR, which may be key to sustaining intermittent high heat loss. Variations in temperature may arise from transient features in the flow as well as fluctuations the sea ice cover above.

Sea ice is well known to mediate rates of surface heat loss (Maykut, 1982). In HR, the sea ice is thinner and less extensive than in LR, in better agreement with observations, as discussed in depth by Bryan et al. (2014). While ice thickness is likely sensitive to ocean heat transport into the ice-zone, the resolution of ice dynamics also plays a central role in regulating ice thickness, and thus heat loss. In HR, finely-spaced grid cells allow sea ice to respond to more localized atmosphere and ocean conditions, leading to more pervasive and smaller-scale coastal polynyas. This leads to higher rates of brine rejection hugging the coast, in better agreement with observations (Willmott et al., 2007). In the pack ice, higher resolution enables a greater magnitude of divergence/convergence and shear, which creates the “leads” that are endemic to the observed Antarctic ice pack (Willmott et al., 2007). Lead opening exposes the ocean surface to the cold atmosphere in winter, driving enormous heat loss and rapid new ice formation. The resulting “frazil ice” only forms in open water (in our model), and is thus a proxy for the continuous exposure of the ocean surface. The rate of frazil ice formation has a broad peak, spanning austral fall through spring in HR (Fig. 9a.). In contrast, frazil ice formation in LR is sharply peaked in austral fall and is relatively weak in winter. The timing in LR is consistent with the northward expansion of the sea ice extent in fall, while the prolonged frazil ice production in HR indicates the consistent opening of polynyas and leads throughout the winter. These dynamics maintain more thin ice throughout the ice-covered waters in winter in HR (not shown), which drives areas of intense heat loss, as illustrated in Fig. 8a and 9b. The extremity of heat loss experienced through the sea ice in HR contributes to the northerly transit of dense isopycnal outcrops in winter; the outcrops in LR migrate less due to the insulating effects of its thicker ice pack and less frequent leads and polynyas.

Diagnosing the processes inducing different rates of diapycnal-mixing driven transformation is more difficult, given our methods of calculation. Generally, models must homogenize properties like temperature and salinity over the grid scale, a consequence of discretizing a continuum; simply reducing grid-cell size reduces this spurious diffusion and enables the formation of smaller-scale density gradients. This likely improves the scale of dense overflows resolved in HR, especially because bathymetry is better resolved. However, as is evident in Fig. 7, these processes have a minimal effect on altering the circulation south of 64° S, which may in part be a consequence of the small volume of the high latitude ocean, since volume has a strong control on transformation rates.

North of 64° S, an increase in resolution significantly impacts the magnitude of mixing. The meridional flow of dense water towards the subtropics likely occurs via transient mass fluxes (e.g., see Ito and Marshall, 2008, Lozier, 2010), and in deep western boundary currents (DWBC) where bathymetry allows (Orsi et al., 1999, 2002; Fukamachi et al., 2010). These boundary currents are particularly susceptible to real and numerical mixing because of their association with strong density gradients and shearing rates (Griffies et al., 2000). Mesoscale turbulence, acting to either erode or intermittently increase density gradients, likely depends on resolution, even without altering ψ_E^σ. Further, there may be different levels of numerical mixing induced across these currents; this spurious diffusion likely decreases with increasing resolution (to a degree, akin to the numerics discussed by Griffies et al., 2000), though it persists in eddying models (Urakawa and Hasumi, 2014). A representative example of the differences in DWBC characteristics between resolutions is shown at 30° S (Fig. 10), at a latitude chosen due to the large contrast in abyssal circulation strength apparent in both the isopycnal and depth-space overturning in Fig. 5. In HR, deep isopycnals slope steeply up towards the western continental boundaries or ridges, coincident with regions of strong meridional flow. The corresponding isopycnal slopes are notably flatter in LR, and the flow is much weaker. The reduced strength of these currents in LR, and associated weak abyssal overturning, follows from the elevated destruction of dense waters south of 30° S. However, Fig. 10 reveals that density gradients in HR can form on scales smaller than a single LR grid cell, indicating that correspondingly strong geostrophic currents are unresolvable in LR, even given a similar flux of dense water into the abyssal ocean. While the formation of boundary currents, and the mixing that erodes them, are coupled, further unraveling their interactions is beyond the scope of this article.

As a final diagnostic of the Southern Ocean circulation state and its dependence on resolution, we consider the twenty-year average integrated heat content tendency for a control volume bounded at 30° S, below each depth z^*, and spanning all longitudes. For HR:

$$(7)\quad -\frac{c_p\rho_o}{(t_1-t_0)}\int_{t_0}^{t_1}\int_{90^\circ S}^{30^\circ S}\int_x \int_{-B(x,y)}^{z^*}\Big[\left\{\frac{\partial\theta}{dt}\right\}_1 + \left\{\nabla\cdot\left(\overline{\bar{u}\theta}\right) + \frac{\partial}{\partial z}\overline{(w\theta)}\right\}_2 + \left\{\nabla\cdot\overline{u'\theta'}+\frac{\partial}{\partial z}\overline{w'\theta'}\right\}_3 - \left\{\frac{\partial}{\partial z}\left(\kappa_{bg}\frac{\partial\bar{\theta}}{\partial z}\right)\right\}_4 - \left\{\frac{\partial}{\partial z}\left(\kappa_{ML}\frac{\partial\bar{\theta}}{\partial z}\right)\right\}_5\Big]dxdydzdt = 0;$$

and for LR:

$$(8)\quad -\frac{c_p\rho_o}{(t_1-t_0)}\int_{t_0}^{t_1}\int_{90^\circ S}^{30^\circ S}\int_x \int_{-B(x,y)}^{z^*}[\left\{\frac{\partial\theta}{dt}\right\}_1 + \left\{\nabla\cdot\overline{(u\theta)} + \frac{\partial}{\partial z}\overline{(w\theta)}\right\}_2 + \left\{\nabla\cdot\overline{u_b\theta}+\frac{\partial}{\partial z}\overline{w_b\theta}\right\}_3 -$$

$$\left\{\frac{\partial}{\partial z}\left(\kappa_{bg}\frac{\partial\bar{\theta}}{\partial z}\right)\right\}_4 - \left\{\frac{\partial}{\partial z}\left(\kappa_{ML}\frac{\partial\bar{\theta}}{\partial z}\right) - \nabla\cdot\boldsymbol{K}_{redi}\nabla\theta\right\}_5]dxdydzdt = 0.$$

Note that the sign convention is such that positive vertical fluxes across the upper bounding surface, z^*, are acting to increase the heat content of the control volume. Here $\theta = \theta(x, y, z, t)$ is potential temperature, $u = u(x, y, z, t)$ and $w = w(x, y, z, t)$ are the horizontal and vertical velocity, respectively, an overbar denotes a time-mean, the prime denotes a deviation from the time mean, the subscript "b" denotes a bolus field in LR, c_p is the specific heat capacity for seawater, and ρ_o is a reference density. Plotted in Fig. 11 is the heat content tendency (labeled $\frac{\partial H}{dt}$) associated with the temperature trend (term 1). This tendency will depend on: heat transport by the mean flow (term 2); the transient eddy-induced flow (term 3); the background vertical diffusion, dependent on the background diffusivity, κ_{bg}(term 4); and a residual term that captures the remaining mixing processes, like KPP mixed-layer processes and convection (term 5). In LR, this residual term also includes the parameterized along isopycnal diffusion (the Redi flux, dependent on the parameterized isopycnal diffusivity tensor, $\boldsymbol{K}_{redi}$). Thus, part of the impact of mesoscale eddies on the vertical heat budget is captured in this residual term in LR; the Redi fluxes couldn't be directly calculated because the time dependent diffusivity tensor was not saved in the monthly output files. While there is a small trend in heat content at either resolution, in the top 500 meters, advective warming nearly balances convective cooling. Below this depth, to first order at either resolution, eddy heat fluxes are sufficient to counter heat fluxes by the mean flow, emphasizing the importance of eddies in diffusing heat across strong temperature gradients. This breakdown of heat fluxes will be useful in the subsequent section to elucidate the physics responsible for deep ocean warming under greenhouse forcing.

4. Carbon Dioxide Doubling Response

We now show how resolution affects the response of the Southern Ocean in the 20 years of carbon dioxide stabilization after doubling. The Southern Hemisphere surface air temperatures increase at both resolutions: air temperatures in the high southern latitudes warm by up to 8 °C over some regions of the ocean and sea ice pack. Bryan et al. (2014) discuss how climate change depends on resolution more generally; here we focus on the deep Southern Ocean's response.

At both resolutions, the lower cell of the isopycnal MOC slows substantially in response to surface warming (Fig. 12), though this reduction is notably greater in HR. The response of the overturning, $\Delta\psi^{\sigma}$, is dominated by a reduction in the circulation strength in the sub-polar range of the lower cell, with smaller changes across the abyssal range. The resolution dependence of the response to carbon dioxide doubling is primarily a feature of the mean isopycnal flow $\Delta\psi_M^{\sigma}$; while the transient eddy-induced circulation response to carbon dioxide doubling, $\Delta\psi_E^{\sigma}$, depends somewhat on resolution, it is comparatively small.

Surface warming and the associated changes in freshwater fluxes alter the surface water mass transformation, ΔF (see Fig. 13). There is a significant (positive) anomaly in ΔF across the outcrops of the lower cell. To understand the impact of these surface changes, relative to changes in interior transformation, we compare changes in the strength of processes contributing to the circulation response throughout the density range of the lower cell. Because the carbon dioxide forcing was only stabilized for 20 years, the response is transient and the isopycnal volume inflation is non-negligible. Thus, the response to carbon dioxide doubling apparent in the overturning, $\Delta\psi^{\sigma}$, includes the contributions from changes in isopycnal volume, $\Delta\frac{\partial V}{\partial t}$, surface water mass transformation, ΔF, and implied interior mixing, $\Delta Dmix$ (see Fig. 14). In both LR and

HR, the significant $\Delta\psi^{\sigma}$ at 64° S can be primarily attributed to ΔF; this affirms the close connection between surface transformation and flow in the sub-polar range as diagnosed in the control climate. Further, $\Delta Dmix$ at 64° S is largely explained by the magnitude and pattern of ΔF. In other words, the reduction in shelf water mixing directly reflects the reduction in shelf water production. Further to the north, at 38° S, there is little $\Delta\psi^{\sigma}$ at either resolution. The large change in mixing-driven transformation between these latitudes (which is primarily a reduction in the effectiveness of mixing in the control simulations, i.e. of the same shape but of opposite sign) can again be explained as the result of a reduction in downwelling dense waters. Further north yet, at 30° S, $\Delta\psi^{\sigma}$ is increasingly small, implying a very small reduction in mixing between 38°-30° S. Greater changes at these latitudes may occur over time, as the influence of high latitude processes spread northward; however, such changes aren't captured in our simulations. There is some redistribution of isopycnal volume from very dense to slightly less dense water classes, though it is small compared with ΔF.

In the control, ψ^{σ} is sustained by both F and $Dmix$, though the influence of each varies with latitude. In contrast, these transient results suggest that the $\Delta\psi^{\sigma}$ is almost entirely explained by a reduction in surface transformation, particularly so in the sub-polar range. Further, the resolution dependence of ΔF primarily explains the resolution dependence of $\Delta\psi^{\sigma}$. Since the pattern of ΔF is both more peaked and confined to denser isopycnals in HR, $\Delta\psi^{\sigma}$ in HR is larger and extends deeper in the water column than in LR. In HR, this peak is ~15 Sv and is centered at an average depth of 3.8 km; in LR, the peak reduction is ~12 Sv and is centered at an average depth of 1.7 km.

The surface transformation function response to carbon dioxide doubling is largely attributed to a reduction in surface heat fluxes. We interpret this as follows: in a warmer climate, there is a reduction in the air-sea temperature contrast in the coastal ocean, especially in wintertime, inducing a reduction in heat lost through the sea ice pack. The reduction in heat loss in response to carbon dioxide doubling is strongest in the winter months, and the (positive) anomaly in surface transformation occurs almost entirely in winter. Surface warming also drives a thinning of the ice pack irrespective of resolution. Sea ice volume is reduced by 40% in LR and 26% in HR. While there is some decrease in ice extent at each resolution, this volume reduction is mainly caused by changes in ice thickness. This relatively greater thinning in LR is characteristic of thicker ice in the mean state (Bitz and Roe, 2004). Because thinner ice is less insulating, thinning provides a damping effect on what would otherwise be a larger reduction in heat loss in response to warming air temperatures. This damping effect is greater in LR because of the more substantial thinning in LR, revealing how sensitive the response to carbon dioxide doubling of the sea ice-atmosphere-ocean system can be to the mean state ice thickness. Lastly, larger changes in the meridional circulation in HR may reduce the rate of heat convergence into very dense waters this region. The shape of the anomaly in surface transformation is also impacted by a shift in the spatial distribution in surface densities. These combined changes drive the larger reduction in surface water mass transformation across a narrower range of denser waters in HR (note the differences in positive anomalies peaked around 37.25 in HR and 37.18 in LR). A shift in the density classes into which sea ice melts partially offsets the strong reduction in buoyancy loss from reduced heat loss in slightly lighter waters.

The circulation response to carbon dioxide doubling, and the sensitivity of each model to surface transformation changes, leads to a strikingly different distribution of ocean warming with resolution, particularly in the high latitude abyssal ocean. Fig. 15a-b illustrates the zonal average temperature change with depth. In HR, warming extends along the path of dense water formation, from the coast into the abyssal ocean. In LR, warming is confined to the surface and mid depths, with nearly no warming of the abyssal ocean below 3500 m. These warming patterns lead to important differences in the total heat uptake with depth in the Southern Ocean (Fig. 15c). Changes in heat content are primarily explained by anomalous advective heat fluxes, as illustrated Fig. 16. In HR, the total advective warming into the ocean volume south of 30° S is a result of anomalous positive vertical and horizontal eddy heat fluxes and anomalous vertical fluxes by the mean flow, which are partially compensated by anomalous negative horizontal fluxes by the mean flow (not shown). In LR, the total advective warming is primarily a result of anomalous vertical eddy fluxes, which are also partially compensated by anomalous negative horizontal fluxes by the mean flow. As can be seen in Fig. 15, the pattern of warming bears a close connection with circulation changes, which are in turn intrinsically related to the control state circulation. The relative magnitude of the control MOC strength (ψ^{σ}) versus the magnitude of its reduction ($\Delta\psi^{\sigma}$) is important in explaining why heat uptake differs across models (Banks and Gregory, 2006, Xie and Vallis, 2011, Rugenstein et al., 2013, Kostov et al., 2014). We attempt to address this by partitioning the total heat flux response to carbon dioxide doubling as:

$$9)\ \Delta(vT) = T\Delta v + v\Delta T + H.O.T.,$$

where H.O.T. stands for higher order terms. Because we cannot calculate sub-monthly correlations between velocity and temperature, our calculations of the relative roles of eddy heat flux response are imperfect. With this caveat, we find that the total advective heat flux response due to changes in MOC strength, via Δv, is responsible for 2-3 times (depending on the depth) more of the warming below 2000 meters than changes in the temperature field, ΔT. Thus, a redistribution of ocean heat by the circulation response to doubling carbon dioxide is a key component of Southern Ocean warming.

5. Discussion and conclusions

These results support the notion that model resolution fundamentally alters simulated Southern Ocean dynamics. This possibility has been explored in numerous studies (e.g. Henning and Vallis, 2005; Hallberg and Gnandesikan, 2006; Farneti et al. 2010, Abernathey et al., 2011; Bryan et al, 2014; Farneti et al., 2014), which have justifiably focused on the first order effects of transient eddies on the residual circulation. In our experiments, the transient eddy-induced meridional circulation varies little with resolution, reinforcing findings by Gent and Danabasoglu (2011) and Gent (2016), who argued that employing an unconstrained spatially-varying GM coefficient greatly improves the agreement between the resolved and parameterized contribution of transient eddies to the circulation.

Here, we call attention to the influence of increased model resolution on the behavior of other fundamental small-scale processes, in particular sea ice divergence and shear and the formation of small-scale ocean density gradients and flows. These processes are generally only resolved to the extent that grid spacing permits- without additional sub grid-scale parameterizations. The increased resolution of these processes has a major influence on the mean isopycnal flow of the lower cell in this model. To quantify the spatial distribution of processes sustaining this circulation, and contributing to circulation differences with resolution, we consider the evolution of water mass transformation (see, e.g., Walin, 1982) across latitudes of the Southern Ocean and focus on lower cell density classes. We conclude that downwelling near Antarctica is consistent with heat loss from isopycnal outcrops in the vicinity of leads in the sea ice pack, the simulation of which is greatly improved by increased resolution: these fine-scale, intermittent sea ice openings (polynyas and leads) are more frequent at higher resolution, enabling vast areas of thinner ice and more efficient surface heat loss over localized scales. Greater heat loss from very dense isopycnals at high resolution is sustained by a greater diapycnal convergence of heat into these density classes by the flow. The resolution dependence of transient eddy fluctuations across the shelf front (i.e., Stewart and Thompson, 2013) may play a role diapycnal heat fluxes. However, even our higher resolution experiment is too coarse to explicitly resolve their dominant scales at the shelf front.

As water flows northward from the shelf regions into the abyssal ocean, transformation from diapycnal mixing processes contributes increasingly to patterns of flow. Progressively greater interior transformation is expected as the volume of most density classes increases over the immediate domain northward from the continental shelves; correspondingly distinct patterns of interior transformation between resolutions emerge as different dynamics act on progressively larger volumes of water. Our higher resolution case forms more vigorous, small scale DWBCs, and experiences less diapycnal mixing of dense waters, sustaining a stronger abyssal circulation; vigorous mixing damps abyssal overturning strength dramatically at lower resolution. These differences may be tied in part to a reduction in numerical mixing, though such mixing can persist in eddying models (Urakawa and Hasumi, 2014).

In the approximate steady state of our control simulation, processes regulating the transformation of seawater at the surface and in the interior together sustain the lower cell circulation. In contrast, we attribute *changes* in flow in response to a doubling of carbon dioxide predominantly to reductions in surface transformation. While the reduction in surface transformation leads to a dramatic reduction in overturning irrespective of resolution (14 Sv at high resolution, a reduction of ~54%, and ~10 Sv at lower resolution, a reduction of ~45%), these overturning changes are greater and deeper on average at high resolution because of a larger reduction in heat loss-induced transformation. There is a decrease in diapycnal mixing throughout the abyssal Southern Ocean with carbon dioxide doubling, though the magnitude of this decrease directly reflects the reduction in upstream dense water production at the surface. These results emphasize an important aspect in the dynamics of the lower cell circulation omitted from several theoretical and idealized studies (e.g., Ito and Marshall, 2008, Nikarushin and Vallis, 2011, 2012). While these studies include a simplified form of buoyancy loss in the high latitudes, they don't include a theory for the sensitivity of the circulation to changes in surface fluxes. In the absence of any evolving surface buoyancy forcing, their results stress the lower cell's sensitivity to other processes: particularly, changes in southward eddy heat fluxes across the ACC, westerly wind strength, and abyssal mixing rates. These insights are crucial to understanding the system, but have limitations. Our results suggest that the lower cell circulation is most sensitive, at least in a "global warming" type of perturbation, to changes in surface heat loss with atmospheric warming, and that this heat loss is sensitive to surface processes. This is supported in the conceptual frameworks of Shakespeare and Hogg (2012) and Stewart et al. (2014), and in the eddy-permitting model discussed by Kuhlbrodt et al. (2015). Further, the relationship between southward eddy heat fluxes and MOC strength becomes more convoluted in a framework that includes evolving interactions with the atmosphere. In fact, southward eddy heat fluxes across the ACC are lower in our high-resolution control experiment (Bryan et al., 2014), while the lower cell is stronger. It is possible that the influence of abyssal mixing rates, as mediated by southward eddy heat fluxes, and westerly winds strength, are overemphasized, or misrepresented, in models that don't include realistically responsive surface fluxes.

The omission of evolving surface fluxes may alter mechanisms of heat uptake. Zhang and Vallis (2013) consider the impact of model resolution on ocean heat uptake in an idealized ocean-only model, which imposed surface heat fluxes and included neither sea ice nor the dependence of density on salinity. They attribute the greater abyssal heat uptake simulated in their higher resolution model to the greater advection of heat by the stronger mean state circulation, explained as a result of higher eddy heat fluxes across the ACC. In contrast, we find that circulation changes significantly redistribute the existing internal heat reservoir in our model in response to surface changes. This redistribution of heat is a larger component of abyssal Southern Ocean warming than the advection of anomalous heat taken up at the surface in these simulations, in agreement with the results of Xie and Vallis (2011).

Our results also suggest a possible shortcoming in the behavior of many standard resolution coupled models. Armour et al. (in prep) studied Southern Ocean heat uptake across a range of standard resolution CMIP5 models, and found that the dominant mechanism for Southern Ocean heat uptake in these models was the northward advection, and ultimate subduction, of anomalous heat taken up at the Southern Ocean surface and resulting in warming in the upper 1500-2000 meters (Armour et al., submitted, Marshall et al. 2014). This mechanism of heat uptake is certainly active in our model, as evident by the significant mode water warming around 65°-35° S (Fig. 15a-b). However, while the heat uptake of the Southern Ocean varies little with resolution (2.5×10^{23} J at high resolution and 2.4×10^{23} J at low resolution), more of this heat enters the lower cell in our simulations in our high-resolution experiment. This changes the distribution of heat with depth (Fig. 15c) as well as the circulation regime in which this anomalous heat resides. The models analyzed by Armour et al. (in prep) produce unrealistically small volumes of, or spuriously formed, AABW (Heuzé et al., 2013). Our results raise the possibility that the response of standard resolution models to surface warming may be biased because of their inability to (realistically) simulate abyssal ocean warming, and instead these models warm too vigorously in the mode waters closer to the ocean surface. In other words, models with weaker lower cells versus those with stronger lower cells may have intrinsically different capacities to suppress surface warming over long timescales, as has been suggested for the AMOC (Winton et al., 2014). While temperature is not a passive tracer, the timescales at which anomalously warm interior waters might be expected to eventually alter SSTs likely differs for abyssal and mode waters because of their different residence times (i.e., Gebbie and Huybers, 2012). Since radiative feedbacks are influenced by SST (Winton et al., 2010), a bias in the fate of heat taken up at the Southern Ocean surface could alter the evolution of transient climate change (Armour et al, 2013). We leave exploration of this possibility to future studies.

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Figures captions

Figure 1. Schematic to demonstrate processes controlling the volume V poleward of an isopycnal surface σ and latitude y in Eq. 3. Volume fluxes into V include contributions of surface transformation induced by surface fluxes south of the outcrop at $S(\sigma)$ (component F), contributions from diapycnal mixing across the isopycnal surface σ (component $Dmix$), and total southward flow across y and below σ (component ψ^{σ}). Color of lines denoting the isopycnal surface (green), ocean surface (black), and latitude transect (pink) are added to

orient the reader to the components of diapycnal volume flux illustrated in Fig. 6 and Fig. 13.

Figure 2. Zonally averaged properties in the Southern Ocean for potential temperature (A-C) and salinity (D-F) in HR (A and D) and LR (B and E), and HR minus LR (C and F). Potential temperature (G-I) and salinity (J-L) along the World Ocean Circulation Experiment (WOCE) transect IO8 (longitude varies from 82°-95° E) in: Observations (G and J), HR minus observations (H and K), and LR minus observations (I and L).

Figure 3. Potential density referenced to 2000 dbar (σ_2) in kg/m^3. Annual average surface σ_2 (A-B) and annual-average zonal-average σ_2 (C-D) in HR (A and C) and LR (B and D). Color scale is chosen to highlight density variations in the deep and polar surface ocean.

Figure 4. Ocean bottom properties south of 50° S for temperature (A-C) and salinity (D-F) in HR (A and D), LR (B and E), and HR minus LR (C and D). The ocean bottom is defined as the deepest ocean grid cell.

Figure 5. Components of the MOC: isopycnal overturning (A-F) remapped to depth-latitude space using the zonal and time mean isopycnal depths in the Southern Hemisphere and overturning calculated in level coordinates (G-H). The top row is for HR and bottom row is for LR. The total isopycnal overturning (A-B) is broken into components of the mean isopycnal overturning (C-D) and transient eddy-induced isopycnal overturning (E-F). Contour intervals of 5 Sv are overlaid in black.

Figure 6. Total surface water mass transformation rate, *F*, in the Southern Ocean. The contributions from heat fluxes (red) and freshwater fluxes (blue) add up to the total (black) for (A) HR and (B) LR. The density spacing is increased for waters denser than 36.82 $\mathrm{kg\ m^{-3}}$, which outcrop over a small area at the surface but occupy ~70% of the ocean volume south of 30° S.

Figure 7. Isopycnal overturning streamfunction (ψ^σ, in magenta) and its components from the total surface transformation (*F*, in black) and total implied mixing at each latitude

(*Dmix*, in green) at 65° S (top row), 38° S (middle row), and 30° S (bottom row) in HR (left) and LR (right). For convenience, the total mixing between latitude pairs (in dashed green) is included (i.e. between 90°-64° S, 64°-38° S and 38°-30° S).

Figure 8. Distribution of transformation per unit area (positive towards lower densities) across density surfaces (roughly spanning the density range of net volume convergence, and thus surface-forced downwelling). Shown are: heat-driven buoyancy flux in HR (top row) and LR (second row) and freshwater-induced buoyancy flux in HR (third row) and LR (bottom row).

Figure 9. Left: monthly rates of total frazil ice formation in HR (black) and LR (blue). Right: distribution of area-weighted winter surface heat flux (W/m^2) in HR (black) and LR (blue). This distribution is calculated from monthly values during the period in which the majority of negative surface transformation occurs (JJAS), within a region bounded at the north by the contour of 5% ice concentration on average for this period, though the relationship is insensitive to choice of domain. Note skew towards more negative (out of the ocean) values in HR.

Figure 10. Zonal transects across key deep western boundary currents at 30° S as a function of longitude and depth (below 3000m). The meridional velocity (in color) is overlaid with isopycnals (in black) in HR (A-D) and LR (E-F). Positive (red) is northward. Isopycnal spacing varies with stratification, but is on average .01 kg m^{-3}. Bathometric features are noted to orient the reader. Note the reduced color scale for LR.

Figure 11. Relative contributions to the heat content tendency of the ocean volume below depth z^*, south of 30° S in HR (left) and LR (right). As defined in the legend at right, the total heat content will depend on: total advective fluxes (solid blue); mean advective fluxes (dashed blue); eddy advective fluxes (short dashed blue); diffusive fluxes (red) mixed layer, convective, and (in LR) Redi fluxes (magenta); and result in a very small temperature tendency (black solid). Fluxes are defined as positive downwards.

Figure 12. As in Fig. 5, but the anomaly in response to doubling carbon dioxide (i.e., perturbed – control). Contour intervals of 3 Sv are overlaid in black. Note reduced color scale for the eddy-induced and depth-space circulation changes (E-H).

Figure 13. As in Fig. 6, but the anomaly in response to doubling carbon dioxide (i.e., perturbed – control).

Figure 14. As in Fig. 7, but the anomaly in response to doubling carbon dioxide (i.e. perturbed – control). In addition, the anomalous response of the isopycnal volume inflation is shown in gold.

Figure 15) a-b) Zonally averaged temperature change for the Southern Hemisphere in HR (A) and LR (B), overlaid with contours of the anomalous isopycnal MOC streamfunction at 3, 6, 9, and 12 Sv; and cumulative fraction of heat uptake south of the equator (C) in HR (red) and LR (blue).

Figure 16. As in Fig. 11, but the anomaly in response to doubling carbon dioxide (i.e. perturbed – control).