Water Mass Exchange in the Southern Ocean in Coupled Climate Models

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ABSTRACT

The authors estimate water mass transformation rates resulting from surface buoyancy fluxes and interior diapycnal fluxes in the region south of 30°S in the Estimating the Circulation and Climate of the Ocean (ECCO) model-based state estimation and three free-running coupled climate models. The meridional transport of deep and intermediate waters across 30°S agrees well between models and observationally based estimates in the Atlantic Ocean but not in the Indian and Pacific, where the model-based estimates are much smaller. Associated with this, in the models about half the southward-flowing deep water is converted into lighter waters and half is converted to denser bottom waters, whereas the observationally based estimates convert most of the inflowing deep water to bottom waters. In the models, both Antarctic Intermediate Water (AAIW) and Antarctic Bottom Water (AABW) are formed primarily via an interior diapycnal transformation rather than being transformed at the surface via heat or freshwater fluxes. Given the small vertical diffusivity specified in the models in this region, the authors conclude that other processes such as cabbeling and thermobaricity must be playing an important role in water mass transformation. Finally, in the models, the largest contribution of the surface buoyancy fluxes in the Southern Ocean is to convert Upper Circumpolar Deep Water (UCDW) and AAIW into lighter Subantarctic Mode Water (SAMW).

1. Introduction

The zonal-mean view of Southern Ocean water mass circulation can be summarized as a balance between the southward-flowing thermocline waters and North Atlantic Deep Water (NADW) and the northward-flowing Subantarctic Mode Water (SAMW), Antarctic Intermediate Water (AAIW), and Antarctic Bottom Water (AABW) (e.g., Speer et al. 2000; Kuhlbrodt et al. 2007; Talley 2008). These water masses circulate throughout most of the global ocean and ventilate the ocean interior. South of the Antarctic Circumpolar Current (ACC), the formation of bottom water occurs via a transformation of upwelled salty Lower Circumpolar Deep Water (LCDW) as it is cooled during the wintertime and made more salty through brine rejection from sea ice formation (Rintoul 1998; Rintoul et al. 2001; Bindoff et al. 2001). Near the Polar and Subantarctic Fronts, warming combined with an overlying strong precipitation band results in a buoyancy gain, transforming Upper Circumpolar Deep Water (UCDW) that has upwelled to the surface into lighter AAIW and SAMW.

Significant uncertainty exists as to the magnitude of and processes responsible for Southern Ocean water mass overturning circulation south of 30°S. Numerous studies have inferred the net water mass exchange from velocity estimates in inverse models (e.g., Ganachaud and Wunsch 2000; Talley et al. 2003; Talley 2008), and each come to different conclusions regarding water mass formation. A few studies have estimated basin-averaged water mass transformation rates via inverse models. Sloyan and Rintoul (2001) estimated that, south of 30°S, 34 Sv (1 Sv = 10⁶ m³ s⁻¹) of upwelled UCDW is converted at the surface
into the lighter AAIW and SAMW layers. However, 31 Sv of AAIW/SAMW recirculates into the UCDW layer in the ocean interior, resulting in little net transformation between the UCDW and AAIW/SAMW layers. Lumpkin and Speer (2007) also estimated water mass conversion at the surface and within the ocean interior and found a larger net transformation of UCDW to the lighter AAIW/SAMW layers. Differences between the Lumpkin and Speer (2007) and Sloyan and Rintoul (2001) results were attributed to differences in the inverse model, rather than in prescribed surface forcing. However, Badin and Williams (2010) found differing transformation rates (though a similar zonally averaged pattern) when comparing the transformation resulting from surface forcing in two observational datasets and a reanalysis product.

The picture of water mass conversion in general circulation models (GCMs) can differ fundamentally from the one suggested by observationally based estimates. We use results from a GCM with 4° horizontal ocean resolution (Gnanadesikan et al. 2004) that has been shown to capture the distribution of transient tracers and the observationally based transport estimates of Talley et al. (2003) to illustrate this difference (Fig. 1). The GCM estimates that about half of the incoming NADW is converted in the Southern Ocean to the denser bottom water export, and the rest is converted to lighter water mass classes. However, observationally based estimates propose two distinct water mass conversion cells that are almost independent of each other; the lower cell is composed of deep water being converted almost entirely to denser bottom waters, with little exchange with lighter densities. This implies that bottom waters flow out of the Southern Ocean and upwell at low latitudes to the lighter deep-water mass classes. In the observationally based upper cell, southward-flowing thermocline waters are converted primarily into SAMW and AAIW ($\sigma_\theta =$ 26.4–27.4 kg m$^{-3}$), which is also in contrast to the GCM.

It remains unclear whether the differences between the observationally based and GCM pictures of the Southern Ocean diapycnal circulation arise from differences in atmospheric reanalysis fluxes used in inverse models, ocean model physics, or the way in which the flow is broken down into water mass classes. The goal of this study is to shed further light on these issues by considering how different buoyancy forcing across the ocean surface and physical mixing in the interior shape the overturning circulation of water masses in the Southern Ocean in widely used coarse-resolution climate models. We rely on the Walin (1982) formulation for estimating the transformation (i.e., conversion) between water mass layers that results from the influence of surface air–sea fluxes and nonadvective diapycnal fluxes within the ocean interior. The original Walin (1982) analysis was carried out considering water masses defined by isothermal layers but has since been extended to density coordinates (e.g., Tziperman 1986; Speer and Tziperman 1992; Marshall et al. 1999). The majority of the literature incorporating this water mass transformation method has been confined to regions of rich surface flux data, such as the North Atlantic (e.g., Marshall et al. 1999; Nurser et al. 1999; Maze et al. 2009). In contrast, the literature for Southern Ocean water mass transformation, with its incomplete spatial and temporal observations of heat and freshwater (FW) surface fluxes, is quite limited.

The large range of water mass transformation rates diagnosed in the few studies quantifying water mass transformation around the Southern Ocean using the Walin (1982) transformation method demonstrates the sensitivity of water mass formation to model grid, resolution, and parameterizations. In addition, water mass transformation is sensitive to different surface buoyancy flux reanalysis products (based on few observations in the Southern Ocean) used to force the model bulk layer atmosphere see discussions in, for example, Josey et al.
(1999), Large and Yeager (2008), and Griffies et al. (2009). Using an ocean–ice z-level model, forced by satellite and National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Global Reanalysis 1 (NCEP-1) climatological fluxes and analyzed in neutral density space, Iudicone et al. (2008b) estimated a formation of approximately 30 Sv in the neutral density class 26.0–27.4 kg m\(^{-3}\) layer via external buoyancy forcing. Examining the same density class in an isopycnal coordinate model forced by different climatological fluxes for atmospheric fields than Iudicone et al. (2008b), Marsh et al. (2000) calculated less than 10 Sv of water mass formation at the ocean–atmosphere interface. In addition, Iudicone et al. (2008b) and Marsh et al. (2000) propose different contributions of heat and freshwater fluxes to the net surface transformation. Similarly, Spence et al. (2009) concluded that four different grid-resolution runs based on the same model of intermediate complexity produced different magnitudes of heat and freshwater contributions for intermediate water formation and higher-density classes. Not only do the surface forcing differences affect water mass formation, but the relative contribution of the surface, mixed layer, and mixing in the main thermocline to water mass formation, particularly for high densities (Spence et al. 2009), remains unclear.

Here, we provide insights into the roles of buoyancy and mixing processes driving Southern Ocean water mass formation and identify model features that impact the water mass formation rates. We use three coupled models from the Geophysical Fluid Dynamics Laboratory (GFDL) and the Estimating the Circulation and Climate of the Ocean (ECCO; Wunsch et al. 2009) assimilated solution to provide insights into the buoyancy and mixing processes driving Southern Ocean water mass formation, and we identify model features that impact the water mass formation rates. We begin with an overview of the climate models and the Walin (1982) transformation method in section 2. We then diagnose the volume transport across 30°S, which we regard here as the northern boundary of the Southern Ocean, for observationally based estimates and the four models in section 3. In section 3, we also quantify transformation due to air–sea and ocean–ice fluxes, which we combine with the layer export at 30°S to infer the diapycnal flux resulting from interior mixing. Our transformation results are discussed in section 4.

2. Methods

a. Transformation framework

To separate the impacts of surface fluxes from interior processes, we use the transformation methods first proposed by Walin (1982) and more recently extended to density coordinates by other studies (e.g., Tziperman 1986; Marshall et al. 1999; Nurser et al. 1999). We present our results in neutral density coordinates, denoted by \(\gamma'^{n}\) (Jackett and McDougall 1997). Figure 2 shows the processes and transports that can inflate or deflate the volume of a density layer \(k\), bounded by density surfaces \(\gamma_{k}^{n}\) and \(\gamma_{k+1}^{n}\), with a meridional boundary defined at 30°S. The volume flux from the density layer above or below is due to the following:

1. ocean–atmosphere and ocean–ice buoyancy fluxes, which change the surface density of the layer resulting in a dianeutral flux within the surface layer \([F(\gamma_{k}^{n})]\); in Fig. 2, buoyancy loss results in a southward flux [and
2. processes that give rise to dianeutral transport in the ocean interior below the ocean surface \([D(\gamma_{k}^{n})]\), including dianeutral mixing, diapycnal eddy transport in the mixed layer, cabbeling, and thermobaricity; where the flux is positive toward higher densities].

The sum of the resultant fluxes across the density layer gives the water mass formation \(\Delta F + \Delta D\), where we define the difference of given a variable \(x\) across a density layer as \(\Delta x = x(\gamma_{k}^{n}) - x(\gamma_{k+1}^{n})\). The water mass formation of the layer is balanced by the change in the volume of layer \(k\) over time \(dV_{k}/dt\) and the transport in or out of the layer at the meridional boundary (in this case, 30°S) \(-\Delta \Psi\) according to

\[
\frac{dV_{k}}{dt} - \Delta \Psi = \Delta F + \Delta D, \tag{1}
\]

where each term has units of Sverdrups. The transport across 30°S, the layer volume change with time, and the
surface buoyancy fluxes are easily computed from regularly archived model output, and we use the layer balance described above to diagnose the formation of water masses in the ocean interior ($\Delta D$); that is,

$$\Delta D = \frac{dV_k}{dt} - \Delta \Psi - \Delta F. \quad (2)$$

We further define the net transformation across a neutral surface as $G(\gamma_k^n)$ (positive values indicating transformation to higher densities) given by

$$G(\gamma_k^n) = F(\gamma_k^n) + D(\gamma_k^n). \quad (3)$$

We find $G(\gamma_k^n)$ by summing Eq. (1) from layer $n$ to the index of the interface at the bottom of the densest layer $N$ (neglecting geothermal and frictional heat), so

$$G(\gamma_k^n) = \sum_{j=k}^{N} \frac{dV_j}{dt} - \Psi(\gamma_k^n) = F(\gamma_k^n) + D(\gamma_k^n). \quad (4)$$

According to Eq. (4), the net transformation is balanced by the net volume change, minus the export of the density layer, and is driven by the surface and interior contributions. Rewriting the above equation gives the equation for the interior transformation term,

$$D(\gamma_k^n) = \sum_{j=k}^{N} \frac{dV_j}{dt} - \Psi(\gamma_k^n) - F(\gamma_k^n). \quad (5)$$

We calculate the water mass transformation and transports, as well as the volume layer change in time, in depth coordinates using monthly model output. We then bin the monthly transformation diagnostics into neutral density coordinates (0.1 kg m$^{-3}$ bins), using the software described in Jackett and McDougall (1997) to calculate neutral density, and then take the annual mean of the transformation rates. The transformation due to surface (ocean–atmosphere and ocean–ice) buoyancy fluxes $F(\gamma^n)$ is given by

$$F(\gamma_k^n) = -\frac{1}{(\gamma_{k+1}^n - \gamma_k^n)} \int_A \left( \frac{Q_{net}}{C_p} \right) dA$$

$$+ \frac{1}{(\gamma_{k+1}^n - \gamma_k^n)} \int_A \left[ \rho_0 \beta S(FW_{net}) \right] dA. \quad (6)$$

The buoyancy fluxes are integrated over the area of the density outcrop $dA$. The first term in Eq. (6) is the transformation due to surface heat fluxes $Q_{net}$, where $\alpha$ is the heat expansion coefficient and $C_p$ is the specific heat capacity. The net heat flux includes the shortwave and longwave radiative forcing, latent and sensible heat fluxes. We are unable to distinguish the penetrative shortwave heating component in all four models, as is done in Iudicone et al. (2008b), and hence exclude the shortwave radiation that penetrates below the surface layer. The second term in Eq. (6) is the transformation due to freshwater fluxes $FW_{net}$, where $\beta$ is the coefficient of saline contraction, $\rho_0$ is the mean ocean density, and $S$ is the surface salinity. The net freshwater flux is calculated from precipitation, evaporation, river runoff, and the exchange of freshwater between ocean and ice.

### b. Observationally based estimates

We take a moment here to describe the observationally based inverse estimates with which we later compare the transport across 30$^\circ$S in the four climate models. The Talley et al. (2003) and Talley (2008) studies use geostrophic velocity data (Reid 1994, 1997; Robbins and Toole 1997) with empirically chosen reference levels and NCEP–NCAR reanalysis winds (Kalnay et al. 1996) to construct the volume transports across 30$^\circ$S. Sloyan and Rintoul (2001) use an inverse box model with 23 neutral density layers, forced at the surface with Comprehensive Ocean–Atmosphere Data Set (COADS) heat fluxes (da Silva et al. 1994), Global Assimilation and Prediction Scheme (GASP) freshwater fluxes (Budd et al. 1995), and Hellerman and Rosenstein (1983) wind stress. The inverse model used in Lumpkin and Speer (2007) has almost double the number of neutral density layers as in Sloyan and Rintoul (2001), and it uses an average of five surface flux products, including NCEP–NCAR and 15-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15) wind stress. The observationally based estimates differ in magnitude because of the number of density layers, resolution, model physics, water mass constraints, velocity reference levels, and surface forcing. For example, in Talley et al. (2003) and Talley (2008), the Indonesian Throughflow (ITF) transport is assumed to be ~10 Sv, whereas it is 13.2 Sv in Lumpkin and Speer (2007), resulting in a stronger southward flow in the Indian Ocean upper layer.

### c. Climate models

Our transformation analysis is performed with three free-running (i.e., no assimilation) GFDL coupled atmosphere–ocean–ice models, which all have the same horizontal and vertical grid resolution (Table 1) and a free surface that enables the exchange of real freshwater fluxes. For all three coupled climate models, we use the second century average of monthly means from a control run with constant 1990 radiative forcing (also used in Gnanadesikan et al. 2006). In the GFDL Climate Model version 2.1 (CM2.1) (Griffies et al. 2005; Delworth et al. 2006; Gnanadesikan et al. 2006), the representation of
Table 1. Model features of the three GFDL models and ECCO model. Shown are the horizontal resolution (x, y), number of depth levels z, the along-isopycnal diffusivity $A_y$, the vertical background diffusivity $K_v$, and the dry snow albedo (note that ECCO does not include an ice model). The latitude in the GFDL models has increased resolution moving toward the equator.

<table>
<thead>
<tr>
<th>Model</th>
<th>Ocean (reference)</th>
<th>x</th>
<th>y</th>
<th>z</th>
<th>$A_y$ (m$^2$ s$^{-1}$)</th>
<th>$K_v$ (m$^2$ s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GFDL-CM2.1</td>
<td>MOM4.0 (Gnanadesikan et al. 2006)</td>
<td>1°</td>
<td>1/3°-1°</td>
<td>50</td>
<td>600</td>
<td>1.5 × 10$^{-5}$ to 1.2 × 10$^{-4}$</td>
</tr>
<tr>
<td>GFDL-CM3</td>
<td>MOM4p1 (Griffies 2009)</td>
<td>1°</td>
<td>1/3°-1°</td>
<td>50</td>
<td>600</td>
<td>1.5 × 10$^{-5}$ to 1.2 × 10$^{-4}$</td>
</tr>
<tr>
<td>GFDL-CM2M</td>
<td>MOM4p1 (Griffies 2009)</td>
<td>1°</td>
<td>1/3°-1°</td>
<td>50</td>
<td>600</td>
<td>1.0 × 10$^{-5}$ to 1.5 × 10$^{-5}$</td>
</tr>
<tr>
<td>ECCO</td>
<td>MITgcm (Marshall et al. 1997)</td>
<td>1°</td>
<td>1°</td>
<td>50</td>
<td>1000</td>
<td>1.0 × 10$^{-5}$</td>
</tr>
</tbody>
</table>

The distribution of water mass density, ACC transport, ocean circulation, and the Southern Hemisphere westerlies is consistent with observations (e.g., Russell et al. 2006; Sloyan and Kamenkovich 2007; Sen Gupta et al. 2009). The second model, CM3, has a similar ocean component to CM2.1 but has a reformulated atmosphere with interactive aerosols, which significantly alters the radiative budget (Donner et al. 2011; Griffies et al. 2011). The third GFDL model, CM2M, has a similar atmosphere component as CM2.1 coupled to version 4p1 of the Modular Ocean Model (MOM4p1), using updated ocean model parameterizations such as the Gent and McWilliams (1990, hereafter GM90) eddy flux parameterization (see description later in this section). Comparing these three GFDL models gives us an indication of how a model with a different atmosphere or updated ocean physics might affect the water mass transformation rates and surface buoyancy fluxes.

The CM2.1 model includes a Boussinesq (i.e., volume conserving) ocean component, the MOM4.0 (Griffies et al. 2003). The vertical mixing profile is based on Bryan and Lewis (1979), which has a background vertical diffusivity of $1.5 \times 10^{-5}$ m$^2$ s$^{-1}$ in the low-latitude pycnocline, $3 \times 10^{-5}$ m$^2$ s$^{-1}$ in the pycnocline poleward of 40°S, and $1.2 \times 10^{-4}$ m$^2$ s$^{-1}$ in the deep ocean, as well as the K-profile parameterization (KPP) boundary layer mixing scheme of Large et al. (1994) (Table 1). For CM3, the ocean model has been upgraded to MOM4p1 (Griffies 2009), which uses a slightly different vertical coordinate but is still Boussinesq. The ocean model parameterizations remain the same as in CM2.1. The CM2M uses the MOM4p1 with updated ocean model parameterizations. The three models use the GFDL Sea Ice Simulator for the ice component (Winton 2000). CM3 and CM2M use canonical ice albedos of 0.68, instead of dark ice albedos of 0.58 (which were included in CM2.1 to reduce the temperature and sea ice biases in the Northern Hemisphere). The dry snow albedo has been increased from 0.8 in CM2.1 to 0.85 in CM3 and CM2M.

The parameterization of subgrid scale processes (e.g., eddies) can significantly influence the physical circulation of water masses. The three coarse-resolution GFDL models analyzed here parameterize eddy-induced transports and other small-scale processes using the flux parameterization of GM90. The GM90 coefficient is proportional to the slope of the isopycnals, except at slopes greater than the set slope maximum. The slope maximum is used to combat numerical instabilities arising from steep isopycnals and is set to 0.002 in CM2.1 and CM3. This slope tapering was included to allow parameterizations to deal with the mixed layer where isopycnal slopes go to infinity and can reduce the eddy-induced transport component of Southern Ocean overturning circulation, particularly at high latitudes (Large et al. 1997; Döös and Webb 1994; Griffies et al. 2005; Gnanadesikan et al. 2007; Farneti and Delworth 2010). By contrast, in CM2M there is no slope maximum, because a new formulation permitting increasing eddy-induced transport, even at steep isopycnal slopes, is modeled (Ferrari et al. 2010). In addition, CM2M includes the Simmons et al. (2004) turbulence parameterization scheme, which represents the spatial variability due to tidal generation of internal waves.

We compare the coupled model results to those of an ocean GCM from the ECCO consortium. A brief description of the ECCO model version used in this analysis is given below; further details can be found in Wunsch et al. (2009) and Forget (2010). The model we refer to as ECCO in this paper is the global state estimate from the data-rich Argo period 2004-06 and is known as the Ocean Comprehensible Atlas (OCCA; Forget 2010), based on the ocean–ice Massachusetts Institute of Technology GCM (MITgcm) (Marshall et al. 1997). ECCO assimilates global datasets from altimetry, hydrographic sections, Argo floats, and data collected from instruments attached to elephant seals (in the Southern Ocean; see online at http://biology.st-andrews.ac.uk/seaos/). The ocean model is forced by surface fluxes from the NCEP–NCAR reanalysis product (Kalnay et al. 1996). At each time step, information regarding past observations is used in an iterative technique to reduce the least squares misfit between the models and data. The ECCO model has a horizontal grid resolution of $1° \times 1°$ and 50 vertical levels, making it almost identical to the Southern Ocean resolution of the three GFDL models. A uniform vertical diffusivity of $1 \times 10^{-5}$ m$^2$ s$^{-1}$ is applied throughout the ocean. The vertical diffusivities in the both the GFDL and ECCO models are smaller...
than those observed for the Southern Ocean pycnocline (Cisewskia et al. 2005) and for the deep Drake Passage (Naveira Garabato et al. 2007). Because high levels of mixing in the ocean interior are expected to inflate interior layers at the expense of surface layers, neglect of such mixing should make it easier to form dense Antarctic Bottom Water from Circumpolar Deep Water. Our models thus represent a lower bound for the surface fluxes and interior transformations required to produce AABW export from the Southern Ocean.

3. Results

a. Volume transport across 30°S

The estimates of transport across 30°S are qualitatively similar among the three GFDL models, ECCO, and the observationally based studies. For example, the UCDW and LCDW layers flowing equatorward in the Pacific are balanced by southward transport of AAIW and Pacific Deep Water (PDW) in the subtropical gyre. Quantitatively, the models are in best agreement with the observationally based estimates in the Atlantic sector. Around 15 Sv of NADW enters the Southern Ocean between 27.5 and 28.1 kg m$^{-3}$, balanced by about 10 Sv of upper thermocline waters and 5 Sv of intermediate water. Both the magnitude and direction of the AABW transport in the Atlantic varies among the models.

We highlight three differences in the volume transports between the GCMs and observationally based estimates. First, in the Indian Ocean (Fig. 3, bottom left), the observationally based estimates of CDW transports and transformation of this water to lighter waters
(SAMW/AAIW/IDW) north of 30°S are larger than those produced by any of the climate models. The GCMs simulate a southward Agulhas Current similar to the 49.5 Sv assumed in Talley (2008), with ECCO simulating the weakest circulation (Table 2). Although the balancing northward flow in the Indian subtropical gyre in Talley (2008) is made up of both the upper and CDW layers, it is mainly the upper-layer northward flow that balances the southward-flowing Agulhas Current in the GFDL and ECCO models. This difference may stem from the mixing parameterization in the models or from the inability for inverse models to represent a mean circulation state from snapshots of boundary conditions from different seasons and years.

Second, the Pacific Ekman transport (included in the upper layer; Fig. 3) and the ITF transport (Table 2) in the GFDL and ECCO models are slightly stronger than in Talley (2008). In CM2M, there is a stronger export of thermocline waters from the Southern Ocean into the Pacific basin and hence a larger ITF transport (13.8 Sv). In Talley (2008), the northward-flowing deep-water masses in the Pacific (PDW/LCDW) upwell into the upper layer before feeding the Indian Ocean basin through the ITF. In contrast, there is little upwelling of intermediate and deep-water masses in the GFDL and ECCO models in the Pacific basins, with the northward-flowing Pacific upper layer dominating the ITF transport into the Indian Ocean.

Finally, the greatest difference in the net transport estimates can be seen in the deepest water mass layers (AABW/CDW/LCDW) in all basins in Fig. 3. The observationally based estimates diagnose a total of 20–45 Sv of deep water entering the Southern Ocean between 27.1 and 28.1 kg m⁻³, with the majority being transformed into the bottom water layer (Fig. 3, bottom right). The GFDL and ECCO models produce a weak LCDW/AABW transport across 30°S; instead, the incoming modeled NADW is transformed to lighter mode and intermediate water. The weaker net bottom water export in CM2.1 and CM3 (less than 3 Sv) than in CM2M is partially due to the restrictions imposed on the GM90 parameterization, and the removal of the maximum neutral slope parameter in CM2M likely contributes to its stronger bottom water export of 7 Sv (e.g., Gnanadesikan et al. 2007).

The net water mass transformation is calculated by vertically integrating the transport across 30°S [$\Psi(y^g)$; Eq. (2)]. Figure 4 illustrates the cumulative sum of transport across 30°S, binned into 0.1 kg m⁻³ bins. We use Talley (2008) (blue curve) as an example for interpretation of the figure: at the LCDW/AABW and NADW interface ($y^g = 28.1$ kg m⁻³), there is 27 Sv of incoming NADW that is converted to higher densities. In addition, 3 Sv of NADW is converted to lighter SAMW/AAIW at the $y^g = 27.2$ kg m⁻³ interface, giving a total of 30 Sv of NADW entering the Southern Ocean. The upper layer is transformed to the denser SAMW/AAIW layer at the $y^g = 26.6$ kg m⁻³ interface. The estimates of Talley et al. (2003) and Talley (2008) (black and dark blue curves, respectively)

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**Table 2.** Volume transport across 30°S (Sv) between the surface and 27.0 kg m⁻³ (Indian Ocean), 27.2 kg m⁻³ (Pacific Ocean), and 26.5 kg m⁻³ (Atlantic Ocean). Density boundaries are chosen for comparison with Talley (2008). Transport values listed are the westward-flowing ITF (110°–140°E) and the southward-flowing Agulhas Current (Indian Ocean; net southward flow between 30° and 45°E), East Australian Current (EAC; Pacific Ocean; net southward flow between 150° and 160°E), and Brazil Current (Atlantic Ocean; net southward flow between 35° and 50°W).

<table>
<thead>
<tr>
<th>Model/obs</th>
<th>Agulhas Current ($y^g &lt; 27.0$ kg m⁻³)</th>
<th>EAC ($y^g &lt; 27.2$ kg m⁻³)</th>
<th>Brazil Current ($y^g &lt; 26.5$ kg m⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Talley (2008)</td>
<td>10.0</td>
<td>49.5</td>
<td>24.0</td>
</tr>
<tr>
<td>ECCO</td>
<td>10.9</td>
<td>34.8</td>
<td>13.6</td>
</tr>
<tr>
<td>CM2.1</td>
<td>11.7</td>
<td>48.7</td>
<td>21.2</td>
</tr>
<tr>
<td>CM3</td>
<td>10.8</td>
<td>53.5</td>
<td>23.3</td>
</tr>
<tr>
<td>CM2M</td>
<td>13.8</td>
<td>55.3</td>
<td>17.6</td>
</tr>
</tbody>
</table>
indicate that the SAMW/AAIW layer is fed primarily from lighter thermocline waters. Lumpkin and Speer (2007), Sloyan and Rintoul (2001), and the GFDL and ECCO models (pale colors) indicate that the SAMW/AAIW layer receives little or no flow from thermocline waters and instead is formed through the conversion of the NADW layer.

The GFDL and ECCO models and observationally based estimates have similar rates of deep water entering the Southern Ocean (Fig. 3, bottom right). In the GFDL and ECCO models, this deep water (NADW; Fig. 4) is converted mostly into lighter SAMW/AAIW. In contrast, the observationally based estimates have most of NADW transformed into denser LCDW/AABW. Among the four GCMs, ECCO shows a stronger transformation of NADW to LCDW/AABW than the GFDL models.

Differences between the observationally based estimates and the GCMs in the definition of the water mass density classes can result in smaller conversion rates. The GFDL models simulate a warmer and fresher upper ocean and a warmer and saltier abyssal Atlantic basin compared with observations and ECCO (Fig. 5). These differences in the temperature and salinity fields and hence density affect the density at which water masses are transformed (Fig. 4). For example, the NADW–LCDW/AABW boundary in Talley (2008) is 28.1 kg m$^{-3}$, showing that 27 Sv of NADW is converted to higher densities. At 28.1 kg m$^{-3}$, ECCO shows that 9 Sv of NADW is converted to LCDW/AABW, whereas this rate increases to 14 Sv at the ECCO turning point (28.2 kg m$^{-3}$).

b. Surface heat and freshwater fluxes south of 30°S

We have analyzed the layer export [$\Psi(x)$; Fig. 4] in the previous section and now move onto quantifying how surface buoyancy fluxes influence water mass transformation [$F$; Eq. (6)]. The water mass layers that
will be used hereafter are defined according to the models’ temperature and salinity properties, and the bounding densities that are applied to all basins are listed in Table 3. For example, a low potential vorticity layer is used to define SAMW, and a salinity minimum is used to define AAIW. These classes differ from those defined in Figs. 3 and 4 (in which we use definitions from Talley 2008), because the GFDL models are warmer and fresher than observed (see Fig. 5). Figure 6 illustrates the net transformation due to surface fluxes in the Southern Ocean (south of 30°S), with the resulting surface formation rates given in Table 4.

Upwelled UCDW is destroyed at the surface primarily via freshening. It is converted partly to higher densities (LCDW and AABW) and also combines with destroyed AAIW and upper-layer water masses (because of surface warming) to form SAMW. Note that the AAIW layer is destroyed at the surface, where there is a net buoyancy gain, but the net formation of AAIW (i.e.,

<table>
<thead>
<tr>
<th>Water mass (acronym)</th>
<th>Density class (kg m⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Surface</td>
<td>26.0–26.5</td>
</tr>
<tr>
<td>Subantarctic Mode Water (SAMW)</td>
<td>26.0–27.0</td>
</tr>
<tr>
<td>Antarctic Intermediate Water (AAIW)</td>
<td>27.0–27.5</td>
</tr>
<tr>
<td>Upper Circumpolar Deep Water (UCDW)</td>
<td>27.5–28.0</td>
</tr>
<tr>
<td>Lower Circumpolar Deep Water (LCDW)</td>
<td>28.0–28.2</td>
</tr>
<tr>
<td>Antarctic Bottom Water (AABW)</td>
<td>28.2–bottom</td>
</tr>
</tbody>
</table>

| TABLE 3. The density class definitions for each water mass from the top to the bottom of the water column (kg m⁻³). |
\( \Delta F + \Delta D \) is indeed a positive rate. This implies AAIW is primarily formed via mixing in the mixed layer and main thermocline. Although the net water mass transformation due to air–sea fluxes (green curve) is similar in the four models, the heat (red curve) and freshwater (blue curve) contributions differ greatly (Fig. 6). None of the four models are in quantitative agreement for the net water mass transformation into AABW because significant differences in interaction between the models’ ocean and ice components. The heat flux contribution to the surface transformation is negligible in the AABW layer, indicating that LCDW is transformed into AABW via freshwater fluxes at the ocean–ice interface (Table 4).

Overall, the formation rates in ECCO are smaller than in the GFDL models, as are the ECCO transports listed in Table 2. Because this pattern is displayed throughout the water column, it is likely not a consequence of our water mass definitions and is more likely due to the ECCO’s ocean–atmosphere and ocean–ice heat and freshwater exchange. In the upper layer, warming (for densities less than 25.7 kg m\(^{-3}\)) and cooling (25.7–26.0 kg m\(^{-3}\)) dominate the transformation processes. In the GFDL models, we find that this large heat contribution is strongest in the Indo-Pacific region; however, it is strongest in the Atlantic for the ECCO model. The weak role of evaporation in ECCO continues to 26.6 kg m\(^{-3}\), followed by strong freshening in the denser half of the SAMW layer and throughout the AAIW layer. In ECCO, the transformation driven by surface heat fluxes oscillates over a 12-Sv range, resulting in a small net change in the transformation. Conversely, there is no net evaporation over the SAMW layer in the GFDL models and a stronger gradient in the heat flux curve.

The CM3 and CM2.1 models have the same ocean model component but differ in their atmospheric model component (see model description in section 2). In both models, the destruction (negative formation rate) of the upper layer is mainly due to surface warming and the surface evaporation plays a secondary role. However, in CM3 we find the net surface transformation via freshwater fluxes (2.3 Sv) is almost 5 times that of CM2.1 (0.5 Sv). This stronger freshwater flux in CM3 is found in the Pacific, the same basin in which the CM3 AABW net freshwater flux exceeds that of CM2.1.

We briefly focus on the mode and intermediate water mass layers, for which we can compare our results to previous studies. Across the SAMW outcrop (\( \gamma^g = 26–27 \) kg m\(^{-3}\)), the net surface formation results from a net

<table>
<thead>
<tr>
<th>Water mass</th>
<th>Model</th>
<th>Heat</th>
<th>FW</th>
<th>( F )</th>
<th>( \Delta \phi + \frac{dV}{dt} )</th>
<th>( D )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper</td>
<td>ECCO</td>
<td>-9.1</td>
<td>-3.4</td>
<td>-12.4</td>
<td>-8.2</td>
<td>4.2</td>
</tr>
<tr>
<td></td>
<td>CM2.1</td>
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<td>0.5</td>
<td>-14.9</td>
<td>-4.2</td>
<td>10.6</td>
</tr>
<tr>
<td></td>
<td>CM3</td>
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<td>-13.2</td>
<td>-10.0</td>
<td>3.2</td>
</tr>
<tr>
<td></td>
<td>CM2M</td>
<td>-14.7</td>
<td>2.7</td>
<td>-12.0</td>
<td>-8.7</td>
<td>3.3</td>
</tr>
<tr>
<td>SAMW</td>
<td>ECCO</td>
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<td>22.1</td>
<td>25.6</td>
<td>14.7</td>
<td>-10.9</td>
</tr>
<tr>
<td></td>
<td>CM2.1</td>
<td>33.2</td>
<td>16.1</td>
<td>49.3</td>
<td>18.2</td>
<td>-31.0</td>
</tr>
<tr>
<td></td>
<td>CM3</td>
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<td>19.0</td>
<td>50.8</td>
<td>21.6</td>
<td>-29.2</td>
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<tr>
<td></td>
<td>CM2M</td>
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<td>11.9</td>
<td>34.7</td>
<td>18.3</td>
<td>-16.3</td>
</tr>
<tr>
<td>AAIQ</td>
<td>ECCO</td>
<td>-5.5</td>
<td>-4.4</td>
<td>-9.9</td>
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<td>16.6</td>
</tr>
<tr>
<td></td>
<td>CM2.1</td>
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<td>-5.2</td>
<td>-21.2</td>
<td>7.3</td>
<td>28.4</td>
</tr>
<tr>
<td></td>
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<td>6.8</td>
<td>-19.6</td>
<td>4.8</td>
<td>24.4</td>
</tr>
<tr>
<td></td>
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<td>-9.4</td>
<td>4.2</td>
<td>13.5</td>
</tr>
<tr>
<td>UCDW</td>
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<td>-15.8</td>
<td>-5.4</td>
<td>-4.1</td>
<td>1.3</td>
</tr>
<tr>
<td></td>
<td>CM2.1</td>
<td>3.5</td>
<td>-26.8</td>
<td>-23.3</td>
<td>-3.9</td>
<td>19.4</td>
</tr>
<tr>
<td></td>
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<td>-28.7</td>
<td>-1.9</td>
<td>26.8</td>
</tr>
<tr>
<td></td>
<td>CM2M</td>
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<td>-27.5</td>
<td>-23.8</td>
<td>-3.9</td>
<td>19.9</td>
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<tr>
<td>LCDW</td>
<td>ECCO</td>
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<td>1.5</td>
<td>2.1</td>
<td>-18.1</td>
<td>-20.2</td>
</tr>
<tr>
<td></td>
<td>CM2.1</td>
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<td>4.2</td>
<td>9.3</td>
<td>-12.0</td>
<td>-21.3</td>
</tr>
<tr>
<td></td>
<td>CM3</td>
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<td>7.4</td>
<td>10.8</td>
<td>-12.0</td>
<td>-22.8</td>
</tr>
<tr>
<td></td>
<td>CM2M</td>
<td>3.8</td>
<td>5.2</td>
<td>9.0</td>
<td>-18.0</td>
<td>-27.0</td>
</tr>
<tr>
<td>AABW</td>
<td>ECCO</td>
<td>8 x 10^{-4}</td>
<td>1.8 x 10^{-3}</td>
<td>2.6 x 10^{-3}</td>
<td>7.7</td>
<td>7.7</td>
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<tr>
<td></td>
<td>CM2.1</td>
<td>0.09</td>
<td>-1.8 x 10^{-3}</td>
<td>0.09</td>
<td>-0.6</td>
<td>-0.7</td>
</tr>
<tr>
<td></td>
<td>CM3</td>
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<td>-0.9</td>
<td>0.9</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>CM2M</td>
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<td>0.4</td>
<td>0.5</td>
<td>7.2</td>
<td>6.7</td>
</tr>
</tbody>
</table>

Table 4. Table of water mass formation rates for each water mass, as defined in Table 3, for the ECCO, CM2.1, CM3, and CM2M models. The formation rate is the negative difference of the transformation rate (Sv). The bold values display formation rates for the transformation due to surface heat and FW fluxes, which combine to give the net transformation due to air–sea and ocean–ice fluxes \( F \). Also listed are the net transformation diagnosed from the layer export and layer volume changes (\( \phi + \frac{dV}{dt} \)) and the inferred transformation due to interior mixing [\( D \); see Eq. (2)]. Positive rates indicate formation, whereas negative rates indicate destruction (i.e., conversion to another density class).
ocean buoyancy loss via cooling in the GFDL models and net evaporation in the ECCO model and a buoyancy gain due to freshening (Fig. 6). The buoyancy loss makes the 26.0 kg m\(^{-3}\) surface migrate equatorward, and the buoyancy gain makes the 27.0 kg m\(^{-3}\) surface migrate poleward. The SAMW is thus inflated at the surface, resulting in a net formation. Between \(\gamma_0 = 26\) and 27 kg m\(^{-3}\), the four models used here form between 26 and 51 Sv of SAMW via surface buoyancy fluxes (Table 4), whereas Iudicone et al. (2008b) found 33 Sv for the same range. Using three observational and reanalysis datasets, Badin and Williams (2010) diagnosed 45–60 Sv of water mass formation between \(\sigma_0 = 26\) and 27 kg m\(^{-3}\), dominated by surface heat fluxes. Using a combination of reanalysis and observational datasets, Spence et al. (2009) estimated around 40 Sv of water mass formation via surface cooling between the buoyancy peaks that bound mode and intermediate waters in high- and coarse-resolution models. Although these estimates vary widely (26–60 Sv), they all depict SAMW formation dominated by surface heat fluxes and partially fed by the destruction of neighboring AAIW.

c. **Inferred interior transformation south of 30°S**

The transformation due to interior mixing processes can be inferred from the transformation due to surface buoyancy fluxes, the layer export, and the layer volume change with time [Eq. (5)]. Figure 7 illustrates the net water mass transformation, divided into the surface transformation and the inferred interior transformation. Water masses lighter than LCDW show a similar net transformation curve, with the two transformation components (i.e., surface and inferred interior) acting in opposition. As the surface forcing acts to form a water mass, the interior transformation processes convert it to other layers and vice versa. Weak surface buoyancy fluxes in the AABW layer indicate that the formation of AABW occurs primarily via interior transformation of LCDW and that the magnitude of this interior transformation is similar to that of the AABW export across 30°S.

We pause here to explore this result of opposing transformation components in the GFDL and ECCO models by comparison with other water mass conversion studies. Except between 25.5 and 26.0 kg m\(^{-3}\), we find that the surface transformation cannot be used to solely represent Southern Ocean water mass transformation. This result is supported by Iudicone et al. (2008b, Fig. 11a) and by Spence et al. (2009), who show that the transformation occurring at the surface in models of varying grid resolution is of a similar magnitude to the flux across the mixed layer base (which is included in our inferred interior transformation term). Separating the interior transformation into mixed layer and main thermocline processes is beyond the scope of this analysis. However, we do wish to note that, in recent observationally based studies, the water mass formation resulting from surface heat and freshwater forcing (e.g., Badin and Williams 2010) for mode and intermediate waters is at least one-third larger than the subduction across the mixed layer (e.g., Sallée et al. 2010). The model-based and observationally based studies indicate that mixed layer processes are not the only diapycnal processes transforming water masses beneath the ocean surface.

For the LCDW and AABW layers, the roles of the interior and surface transformation in this study and others remain inconclusive. Heat loss and salt gain at the ocean–ice interface has been observed as the primary mechanism for AABW formation (e.g., Rintoul et al. 2001; Williams et al. 2008, and references therein), but there is only a small buoyancy flux in the models in our study at these densities. In addition, Iudicone et al. (2008b) show that interior mixing dominates the formation of LCDW, but the surface and interior transformation terms are similar for the formation of denser AABW. A weak ocean–ice buoyancy flux is to be expected in the ECCO simulation, because there is no coupled ice model. In the GFDL models, as with many climate models, the small buoyancy flux at the ocean ice interface may be due to a failure to resolve or parameterize polynyas, ice shelves, icebergs, or overflows. Grid resolution is not necessarily linked with increased surface fluxes at the ocean ice interface. For example, Spence et al. (2009) show large differences between their four model surface buoyancy fluxes for abyssal waters, with no surface transformation in their high-resolution simulation.

How do the net water mass transformation \(\psi + \frac{dV}{dt}\), the transformation due to surface \(F\), and the transformation due to interior diameutral mixing process \(D\) compare among the four models? The CM3, CM2M, and ECCO models show a net formation of AABW from interior mixing, and CM2.1 shows a small destruction of AABW (−0.6 Sv). This destruction of AABW does not indicate that AABW flows poleward in CM2.1 but rather is a consequence of significant volume change in the AABW layer. The most significant difference in the formation rates between the four models is due to differing surface fluxes, and this highlights the importance of accurately simulating ocean–atmosphere and ocean–ice interaction for water mass formation. For example, the net formation rate due to surface fluxes in the ECCO model in the UCDW and LCDW layers is −5.4 and 2.1 Sv, respectively, which is less than a quarter of the GFDL models’ formation rates.

Our diagnostics from the last three columns of Table 4 are schematized in Fig. 8. The models show a net formation of SAMW, AAIW, and AABW and a net destruction of the upper, UCDW, and LCDW layers. We
summarize the transformation of water masses via surface buoyancy fluxes and interior mixing as follows:

- We find that 15–20 Sv of combined UCDW and LCDW enters the Southern Ocean at 30°S. About a third of the LCDW is converted to denser AABW via interior mixing, and this 1–8 Sv of AABW then flows out of the Southern Ocean. The remaining two-thirds of LCDW is transformed in the ocean interior to lighter UCDW. Less than 1 Sv of LCDW reaches the surface and is converted to AABW.
- The majority of UCDW upwells to the surface, where freshwater fluxes convert about a third of UCDW to LCDW and the rest to AAIW. Less than 10 Sv of UCDW is transformed to lighter AAIW via interior mixing.
- AAIW is destroyed at its outcrop via surface warming, where 13–39 Sv of AAIW is transformed to lighter SAMW. The remaining 12–15 Sv of SAMW formed at the surface is supplied by the destruction of the lighter upper layer (via warming). SAMW subducts into the ocean interior, where about half of it is converted into
adjacent water masses, and the remaining 15–22 Sv flows out of the Southern Ocean.

- The 4–7 Sv of exported AAIW at 30°S is of similar magnitude to the southward-flowing upper layer.

4. Summary and conclusions

This paper provides a multimodel analysis of the transformation of Southern Ocean water masses in a neutral density framework. We have used the ECCO assimilation solution, which is strongly constrained to observations, to diagnose Southern Ocean rates of water mass transformation and circulation. The three coarse-resolution GFDL models compare reasonably well with ECCO in their transport estimates but tend to have slightly higher transformation rates. As found in previous studies (e.g., Marsh et al. 2000; Sloyan and Rintoul 2001; Iudicone et al.

Fig. 8. Schematic of the transformation results simulated by the GFDL and ECCO models. The gray box represents a zonal ocean view between Antarctica and 30°S and within are the six water masses and their bounding densities (kg m⁻³). The net buoyancy gain (red text) or loss (blue text) and the largest contributing term to that buoyancy gain/loss are shown above each water mass layer outcrop. Directly below is shown the direction of the resulting range of net surface transformation rates (green text) in the first ocean model layer across the four models. In the interior, the purple squiggly arrows indicate the inferred direction of the interior water mass transformation process, with the range of model transformation rates indicated (note that “interior” includes mixed layer processes). The volume transport across 30°S in each layer is indicated by black arrows (northward transport indicated by arrows directed right), and the corresponding ranges of model transports are boxed. Water masses whose net transformation is dominated by the surface transformation are highlighted in green, and those where the interior conversion is the dominant process are highlighted in purple. Three water masses of the six have a net formation (as indicated by an asterisk): SAMW, AAIW, and AABW.

Fig. 9. Schematic of the inferred net water mass transformation for the region south of 30°S based on estimates of the net transport across 30°S in (a) Sloyan and Rintoul (2001), (b) ECCO, and (c) the three GFDL models. Transports are in units of Sverdrups. The four transport layers in each figure are chosen to maximize the northward (to the right) and poleward (to the left) flows. The inferred conversions between layers the net result of interior and surface water mass transformation.
2008b,a), the interior processes play a significant role in the net transformation of Southern Ocean water masses. Our principal findings for water mass transport and transformation are as follows:

1) The model estimates of meridional transport of intermediate and deep waters across 30°S are similar to the observationally based estimates for the Atlantic Ocean but disagree significantly in the Indian and Pacific Ocean basins.

2) In the GFDL models and ECCO, at most half of the southward-flowing deep water entering the Southern Ocean is converted to denser bottom waters, with the remainder being transformed to lighter mode and intermediate waters (Figs. 9b,c). In contrast, observationally based estimates (e.g., Sloyan and Rintoul 2001) suggest that most of the inflowing deep waters are converted to denser layers, with little interaction with the upper ocean (Fig. 9a).

3) In the GFDL and ECCO models, LCDW that flows into the Southern Ocean contributes to the formation of both lighter UCDW and denser AABW (and also AAIW in the ECCO simulation), primarily as a result of interior transformation (Table 4).

4) The magnitude and sign of the heat and freshwater fluxes at the AABW and LCDW outcrops differ among the four models, leading to different surface transformation into the AABW layer.

We draw the following major conclusions from our analysis. First, it is interesting that the various estimates of meridional transport across 30°S of intermediate and deep waters agree so well with each other in the Atlantic Ocean but disagree in the Indo-Pacific. We believe that this difference may be due at least in part to tuning of models to well-known metrics of NADW formation. By contrast, there are no comparable well-known measures of Indo-Pacific intermediate and deep-water formation and transport against which models test their performance. We conclude that greater attention needs to be given to estimating how important these deep and intermediate flows are to properly simulating the oceanic uptake of anthropogenic carbon and heat and to devising corresponding transport metrics for these water masses.

Second, we infer very large water mass conversion rates in the ocean interior south of 30°S in the GFDL models. Given the relatively small vertical diffusivity in this region, we conclude that other processes such as cabbeling and thermobaricity and/or lateral dianeutral fluxes in the mixed layer must be important in this transformation. Both thermobaricity and cabbeling have been noted as key mechanisms in balancing the formation of AAIW (Iudicone et al. 2008a; Klocker and McDougall 2010). We are presently implementing a more detailed analysis package in the GFDL models, which will enable us to explicitly calculate the magnitude of the various mixing processes contributing to interior water mass transformation across density surfaces.

Finally, all transformation estimates are currently missing key physical processes. For example, observationally based estimates (inverse models) of water mass transformation do not represent ocean–ice fluxes. Although GCMs do represent such fluxes, they do not adequately resolve small-scale atmospheric forcing that leads to phenomena such as polynyas and leads. Additionally, the models analyzed here poorly represent interior transformations due to overflows (Winton et al. 1998). This problem will be addressed in ongoing studies with a coupled model with an identical atmosphere as CM2M but an isopycnal ocean component (R. W. Hallberg 2010, personal communication). An isopycnal coordinate system allows newly formed bottom water to descend down the continental shelf without spurious mixing and numerical entrainment into adjacent grid cells, resulting in a more realistic representation of AABW formation and circulation.

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