On Challenges in Predicting Bottom Water Transport in the Southern Ocean

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ABSTRACT

Changes in the Southern Ocean lower-limb overturning circulation are analyzed using a set of climate models. In agreement with some recently developed theoretical models, it is found that the overturning can be strongly affected by winds. In particular, the simulated strengthening of large-scale southward transport in the abyss is explicitly driven by zonal wind stress. However, there is a considerable range among the climate models in their projected changes of Southern Ocean wind stress. Furthermore, the strengthening of large-scale southward transport tends to be compensated by eddy-induced northward flows in the abyss, particularly at eddy-permitting resolution. As a result, the net Antarctic Bottom Water (AABW) export may only be weakly affected. However, none of the models considered accounts for the possibility that a fraction of the eddy kinetic energy may be converted to diapycnal mixing. If this were the case, the presented energetic arguments suggest that stronger Southern Ocean winds would result in a stronger AABW transport.

1. Introduction

Antarctic Bottom Water (AABW) is a global-scale water mass, formed in several regions in the Southern Ocean. The flows associated with AABW fill the abyssal ocean with cold waters and play an important role in transporting carbon and other geochemical tracers. However, understanding the driving mechanisms and predicting these flows, at least in an integral sense, present a number of theoretical and numerical challenges. Some of these are considered here, focusing on the driving mechanisms within the Southern Ocean. Neither the theories nor the climate models are complete. They do, however, suggest that diapycnal mixing, mesoscale eddies, and winds can all be important in controlling the net transport in the Southern Ocean abyss.

Some theoretical challenges are considered in Ito and Marshall (2008). Based on the Southern Ocean budgets of momentum and buoyancy, they propose the following scaling relation for the lower-limb overturning:

$$\Psi = \frac{\overline{\Psi}}{2}(1 - \sqrt{1 + \phi}), \tag{1}$$

where Ψ is the net, or residual, overturning streamfunction that combines the Eulerian-mean circulation ($\overline{\Psi}$), which is set by surface wind stress, and eddy-induced circulation (Ψ^*); that is, $\Psi = \overline{\Psi} + \Psi^*$. The dimensionless quantity ϕ reflects the relative magnitude of the mixing-driven and wind-driven circulations, that is, $\phi \propto kK(\overline{\Psi})^{-2}$, with k and K being the eddy diffusivities due to small-scale turbulence and mesoscale processes, respectively.

Ito and Marshall (2008) (see also Kamenkovich and Goodman 2000) focus on the role of mixing and, in particular, on the limit where $\phi \gg 1$ (strong mixing and/or weak wind) so that $\Psi \sim -\sqrt{kK}$. Using an ocean general circulation model with an idealized basin geometry, Ito and Marshall (2008) show that this relation may hold reasonably well. A similar limit has been arrived at by Nikurashin and Vallis (2011) using a more general theoretical framework developed for a global, interhemispheric component of the lower-limb overturning circulation. The theoretical model of Nikurashin and Vallis (2011) is based on an assumption that, in a dynamical equilibrium, the diabatic dynamics of the

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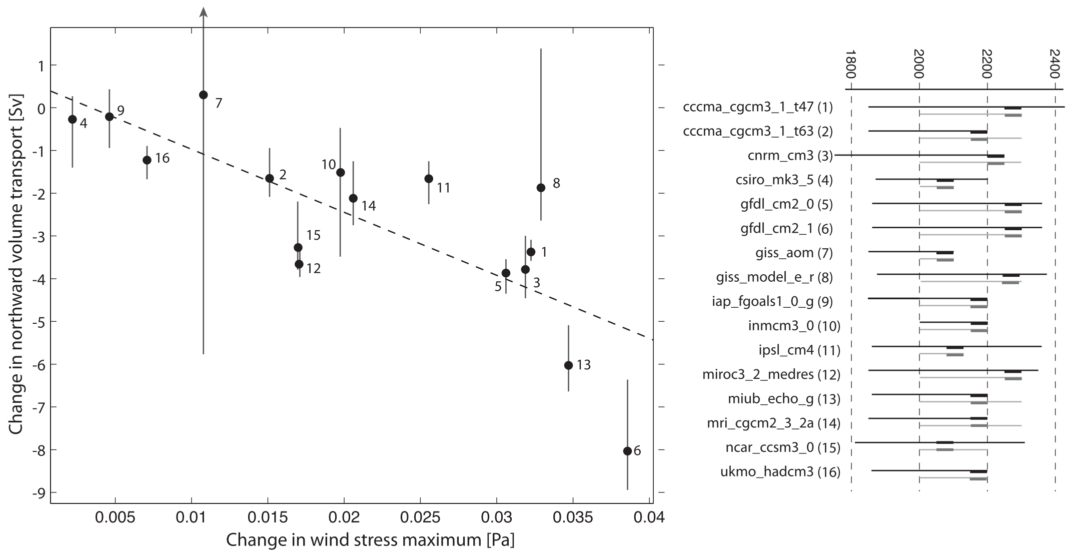

FIG. 1. (left) Changes in northward volume transport (Sv) averaged between the bottom and 1000 m above the bottom (for depths > 3000 m) and changes in the maximum zonally averaged zonal wind stress over the Southern Ocean. Transports are calculated as averages from 55° to 60°S (which correspond to Drake Passage latitudes in most models); bars indicate the range in velocities calculated at individual latitudes between 55° and 60°S. The dashed line is a regression. The Goddard Institute for Space Studies Atmosphere–Ocean Model [(GISS-AOM) (7)] is particularly sensitive to the latitude at which the velocity average is taken (upper bound above the *y*-axis limit, as shown). (right) The thin lines show the duration (yr) of the preindustrial (black) and A1B (gray) velocity and wind stress data in each of the 16 CMIP3 models, as provided to the PCMDI archive. The thick segments indicate the periods of time over which the averaging was done to obtain the anomalies presented in the left panel.

lower-limb overturning north of the Southern Ocean matches the nearly adiabatic dynamics within the Southern Ocean. They show, in particular, that in the limit of weak diapycnal mixing the rate of the lower-limb overturning is inversely proportional to the Southern Ocean wind stress.

Here, we use results from a set of global climate models, including a model with a relatively high resolution, and focus on the role of winds on the volume transport in the Southern Ocean abyss. We first illustrate that changes in the mean meridional flow in the abyss correlate with changes in the zonal wind stress. This is expected from the basic theoretical considerations (e.g., Nikurashin and Vallis 2011). An eddy-permitting model is then employed to show that the (explicitly) eddy-driven meridional volume flux in the Southern Ocean, including in the abyss, is relatively large. This flux can also change in response to plausible changes in the wind stress, counteracting the changes in the Eulerian-mean transport. As a result, the net AABW export may only be weakly affected. In addition, we discuss the situation where the residual circulation and its major components (the mean overturning, mesoscale eddies and diapycnal mixing) are all maintained by the wind: wind steepens isopycnals and generates eddies, and the eddy kinetic energy is converted to enhanced diapycnal mixing (Tandon and Garrett 1996). In this case, the presented energetic arguments suggest that stronger Southern Ocean winds would lead to a stronger AABW transport.

2. Coarse-resolution models

a. IPCC AR4 simulations

We first use the model data provided by the Program for Climate Model Diagnosis and Intercomparison (PCMDI) within the Coupled Model Intercomparison Project (CMIP3) [for details see Sen Gupta et al. (2009) and references therein]. Unlike in typical ocean-only simulations, wind stress is not imposed in these climate models. Rather, it can respond to plausible future changes in the climate, driven by the accumulation of greenhouse gases in the atmosphere. The latter follows here the A1B scenario within the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR4). The greenhouse gas concentrations increase to year 2100, with CO_2 levels reaching twice their present-day value; thereafter, the greenhouse gas concentrations are kept constant.

We analyze the data from relatively long, 50-yr periods (Fig. 1, right) to avoid aliasing by low-frequency natural variability. Climate drift is removed by considering the anomalies in the simulated fields of interest relative to their values in the corresponding concurrent control runs, where the concentrations of greenhouse gases are held at their preindustrial values. Only 16 CMIP3 models have overlapping meridional ocean velocity and zonal wind stress data available for their control and A1B scenario runs.

Output from PCMDI includes monthly averaged meridional velocity, but does not provide any information regarding eddy-induced velocities. The analysis is further complicated by the fact that different climate modeling groups follow different approaches for representing eddy-induced flows in their ocean models. In particular, most, but not all, groups employ the Gent and McWilliams (1990) parameterization. Some of them follow the ideas presented in Visbeck et al. (1997), whereas others simply set K to a constant value. As such, our examination of the climate models in this section is restricted to changes in the mean resolved transport (the eddy-induced circulation will be addressed in the next section and also in section 3). Furthermore, since much of the model output has been interpolated onto nonnative grids, it is difficult to compute an overturning streamfunction in density space, as the calculation, especially at higher densities, becomes very sensitive to the grid. Instead, we present the change in the mean northward transport (with negative values implying stronger southward flows) in the bottom 1000 m of the water column (for the ocean below 3000-m depth) as a function of the change in the maximum zonally averaged zonal wind stress over the Southern Ocean.

All the CMIP3 models predict an intensification of the zonal westerlies over the Southern Ocean, with a multimodel mean increase of about 12% (relative to preindustrial times). At the same time all but one of the models simulate a stronger mean southward flow between 55° and 60°S in the abyss (Fig. 1, left). This is not surprising. Westerly winds over the Southern Ocean drive the northward Ekman transport at the surface, which is balanced, on zonal mean, by the southward geostrophic flow below the sill depth of the highest ridge (Munk and Palmén 1951). Since only the mean resolved transport is used, Fig. 1 shows essentially the sensitivity of the mean southward return flow to the wind so that a linear relationship is expected between the two (see also Sen Gupta et al. 2009). This relationship is robust across the 16 models, with a correlation coefficient of 0.78. This suggests that despite major model differences in resolution and model parameterizations, about 60% of the variance in the southward return flow in the abyssal Southern Ocean is purely related to the differences in the local northward Ekman flux at the surface. The correlation is relatively insensitive to the depth range over which the transport values are computed. However, it is sensitive to latitude. North of the Southern Ocean and, in particular, between 30° and 35°S, only ⅔ of the models predict a significant decline in the large-scale northward transport in the abyss (not shown). This may suggest that in these models diabatic processes play a leading-order role in maintaining the lower-limb overturning (Nikurashin and Vallis 2011). Since different climate models use different representations of diapycnal mixing and values of k, it is difficult to arrive at a robust conclusion. It is also likely that the abyssal ocean is not close to a steady state on these time scales. An additional discussion of the changes in the AABW transport outside of the Southern Ocean is presented in the next subsection.

b. IPCC AR5 simulations based on the CanESM2

We next consider simulations based on the recently developed second-generation Canadian Earth System Model [CanESM2; see Arora et al. (2011) and references therein]. The purpose is twofold. First, we want to evaluate the contribution of the eddy-induced transport to the changes in the AABW outflow. Second, we consider the evolution of the lower-limb overturning, both within the Southern Ocean and outside of it, under the historical (1850–2005) and future (2006–2100) natural and anthropogenic forcing schemes that correspond to the newly developed representative concentration pathways of greenhouse gases and aerosols (RCPs; information online at http://www.pik-potsdam.de/~mmalte/rcps/index.htm). The RCPs have been developed for climate model simulations in support of the Fifth Assessment Report of the IPCC (IPCC AR5; due in 2013/14). We consider the AABW response under three RCPs (2.6, 4.5, and 8.5). Some other results, such as changes in the surface climate simulated by the CanESM2 under these RCPs, can be found in Arora et al. (2011).

The general structure of the residual overturning streamfunction simulated by the CanESM2 within the dense water classes and averaged for 1850–2005, is shown in Fig. 2c. The upper-limb overturning is associated with the formation of deep water in the North Atlantic, its transport to the south, and upwelling. The lower-limb overturning cell (which is our interest here) is associated with the outflow of AABW from the Southern Ocean, its transformation to lighter water, and its return to the south. Between 50° and 60°S, the net (residual) export of the densest waters from the Southern Ocean (Fig. 2c) is maintained by the eddy-induced transport (Fig. 2b; see also Fig. 3a), whereas the Eulerian-mean component (Fig. 2a) transports the dense water south, as expected. North of the Southern Ocean, the residual lower-limb overturning (Fig. 2c) is essentially represented by the mean (resolved) circulation (Fig. 2a).

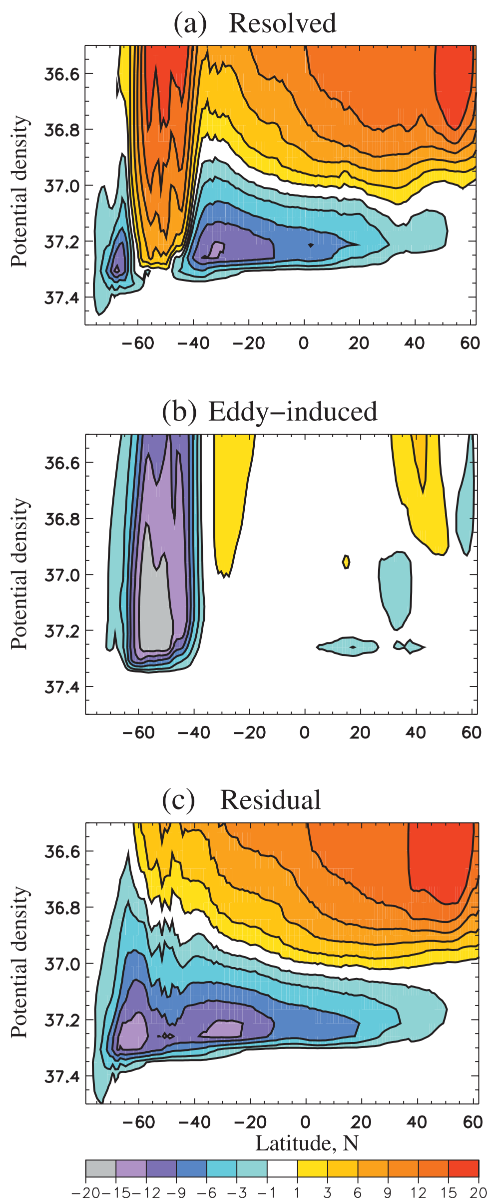

FIG. 2. Meridional overturning circulation (Sv) in the deep ocean in potential density (σ_2) coordinates averaged for the period of 1850–2005 in the CanESM2 historical run: (a) resolved, (b) eddy induced, and (c) the sum of the two, or the residual circulation. Negative values indicate counterclockwise circulation. The model employs the Gent and McWilliams (1990) parameterization with the layer thickness diffusivity of 10^3 m^2 s^{-1}.

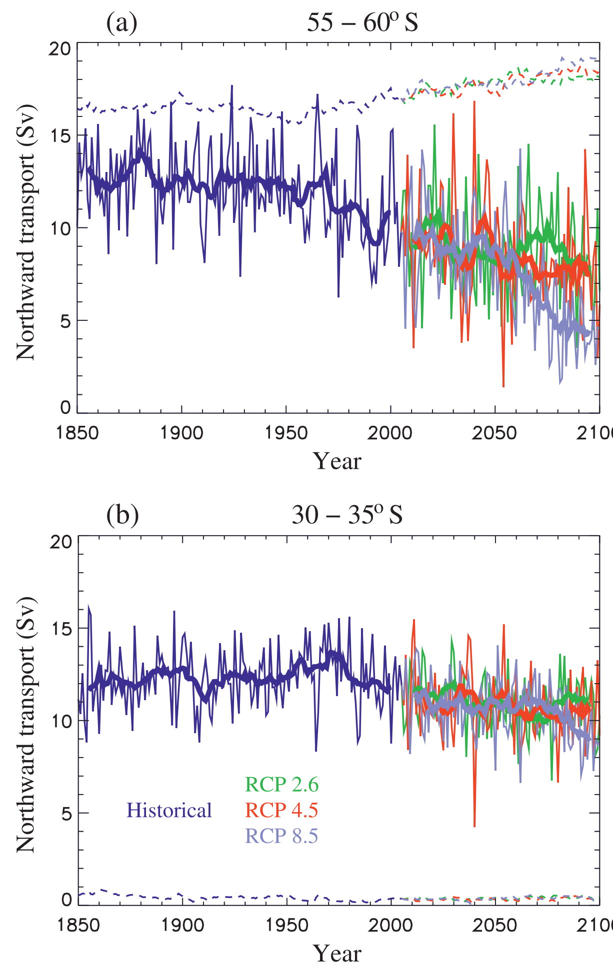

FIG. 3. Time series of (thin) annual-mean and (heavy) 11-yr running mean of northward transport of the densest waters, averaged between (a) 55°–60°S and (b) 30°–35°S for residual(solid) and eddy-induced (dashed) transports. Colors indicate the CanESM2 historical run, and the runs where the forcing follows the three RCPs.

Time series of the net densest water outflow from the Southern Ocean, averaged between 55° and 60°S, is shown in Fig. 3a. During the second half of the twentieth century, the simulated AABW export starts to decline. This general tendency for a decline continues into the twenty-first century. The rate of the AABW export decline is comparable between the RCPs during the first half of the twenty-first century (Fig. 3a), that is, when the corresponding concentrations of greenhouse gases are similar. [By the middle of the twenty-first century, the atmospheric CO_2 concentration is still between 450 and 550 ppm for all of the RCPs, whereas by the end of the century the estimates are approximately 420, 540, and 930 ppm in, respectively, RCP 2.6, 4.5, and 8.5 (Arora et al. 2011).] The decrease in the net residual northward transport of the densest waters at 55°–60°S is maintained in part by the stronger southward transport due to the Eulerian-mean component. By the end of the twenty-first century, the model predicts that the RCP-mean increase in the zonal wind stress maximum in the Southern Ocean will be about 8%, relative to the historical (1850–2005) period. The change in the Eulerian-mean component is compensated by the (parameterized) eddy-induced transport, including at the densest water classes (Fig. 3a). However, the compensation is not complete. In addition, since the lower-limb overturning cell is closed by the diabatic (cross isopycnal) flow, it can change in response to changes in the density structure [see Eq. (4) in section 4]. In particular, an increase in the slopes of isopycnals and/or a decrease of k would result in a weaker cross-isopycnal flow and, hence, in a weaker residual circulation (at steady state).

As noted above, north of the Southern Ocean the eddy-induced meridional circulation in the abyss is negligible (Fig. 2b) so that the mean resolved overturning essentially matches the residual circulation. This is one of the assumptions in the theoretical model of the global abyssal overturning circulation presented recently by Nikurashin and Vallis (2011). Their theory relies on the matching of the nearly adiabatic residual dynamics in the circumpolar region to the diabatic dynamics in the region north of the Southern Ocean, so that there is an interaction between the two regions. Nikurashin and Vallis (2011) argue, in particular, that in the limit of weak diapycnal mixing the rate of the interhemispheric lower-limb overturning is inversely proportional to the Southern Ocean wind stress, whereas in the case of strong diapycnal mixing in the abyss, such as in our model (Saenko and Merryfield 2005), the overturning in the abyss tends to be independent of the wind stress. Time series of the northward flux of the densest waters, averaged between 30° and 35°S, is shown in Fig. 3b. The flux declines, but not as much as that between 55° and 60°S. In general, the decline is not inconsistent with the results of Nikurashin and Vallis (2011), although their theory applies to a steady-state ocean under a prescribed surface climate, whereas these are still transient simulations.

So far we have used coarse-resolution models to diagnose the potential changes in the abyssal circulation. Motivated by the possibility for an important role of the eddy-driven circulation in controlling the net AABW export from the Southern Ocean, we next consider its response to a perturbed wind stress in a high-resolution model.

3. Eddy-permitting simulation

Hallberg and Gnanadesikan (2006) have argued that wind-driven changes in the Eulerian-mean transport could be largely compensated by changes in the eddy-driven transport. The ocean model they employed was configured to cover only the Southern Hemisphere. Here, we further test this idea using a global eddy-permitting model. In particular, we aim to evaluate the impacts of the plausible changes in the Southern Ocean wind stress on the (explicitly) eddy-driven flows in the abyss. The model is a version of the University of Victoria (UVic) climate model with horizontal resolution of $0.2° \times 0.4°$ (latitude $\times$ longitude) [for details on the model, see Spence et al. (2009, 2010). This corresponds to a resolution of about 25 km at 60°S, which is larger than typical local values of the first baroclinic Rossby radius. We note, however, that mesoscale eddies, at least those that can be identified from satellite altimetry including those in high latitudes, have characteristic diameters of the order of 100 km (Chelton et al. 2007).

For our purposes we consider two of the experiments discussed in Spence et al. (2010), which correspond to the model version with reduced viscosity. The focus is on the eddy-induced circulation within the densest water classes. One of the experiments is a standard run (Control), where wind stress is specified from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset. The second experiment (Wind) is identical to the control, except that the zonal wind stress is perturbed by imposing an anomaly. The anomaly (not shown; see Fig. 7 in Fyfe et al. 2007) is such that the perturbed stress increases south of 50°S. The maximum increase is about 20% of its value in the Control experiment. Further details on the design of the experiments and model spinup can be found in Spence et al. (2010).

The eddy-induced overturning, estimated as in Spence et al. (2009), is shown in Fig. 4a (for $\sigma_3 > 41$). Between 55° and 60°S, the eddies transport roughly 10 Sv (1 Sv $\equiv 10^6$ m^3 s^{-1}) northward for $\sigma_3 > 41.4$. Most of this flux takes place within the densest water classes, that is, around 41.8 σ_3 (in the model, the mean depth of 41.8 σ_3 around 55°S is 3124 m, reaching in many places more than 4000 m). This shows, consistent with the arguments presented in Ito and Marshall (2008) and with our results in section 2b, that mesoscale processes can be important in maintaining the lower-limb overturning in the Southern Ocean. Within the lighter water classes, between 41.6 and 41.7 σ_3, the eddy-induced northward volume flux tends to oppose the return of deep water to the Southern Ocean (Fig. 4b).

In response to the imposed stronger wind stress in the Southern Ocean, the eddy-driven overturning intensifies (Fig. 4c). Most of the intensification takes place for $\sigma_3 < 41.7$. In particular, between 41.5 and 41.7 σ_3 about 10 Sv is transported northward in the Wind experiment, whereas only about 3 Sv is transported in the Control experiment (in the model, the mean depth of 41.7 σ_3 around 55°S is about 2395 m, reaching locally more than 4000 m). As a result, the residual overturning in the Southern Ocean abyss is only weakly affected (Figs. 4b,d). This is in general agreement with Hallberg and Gnanadesikan (2006).

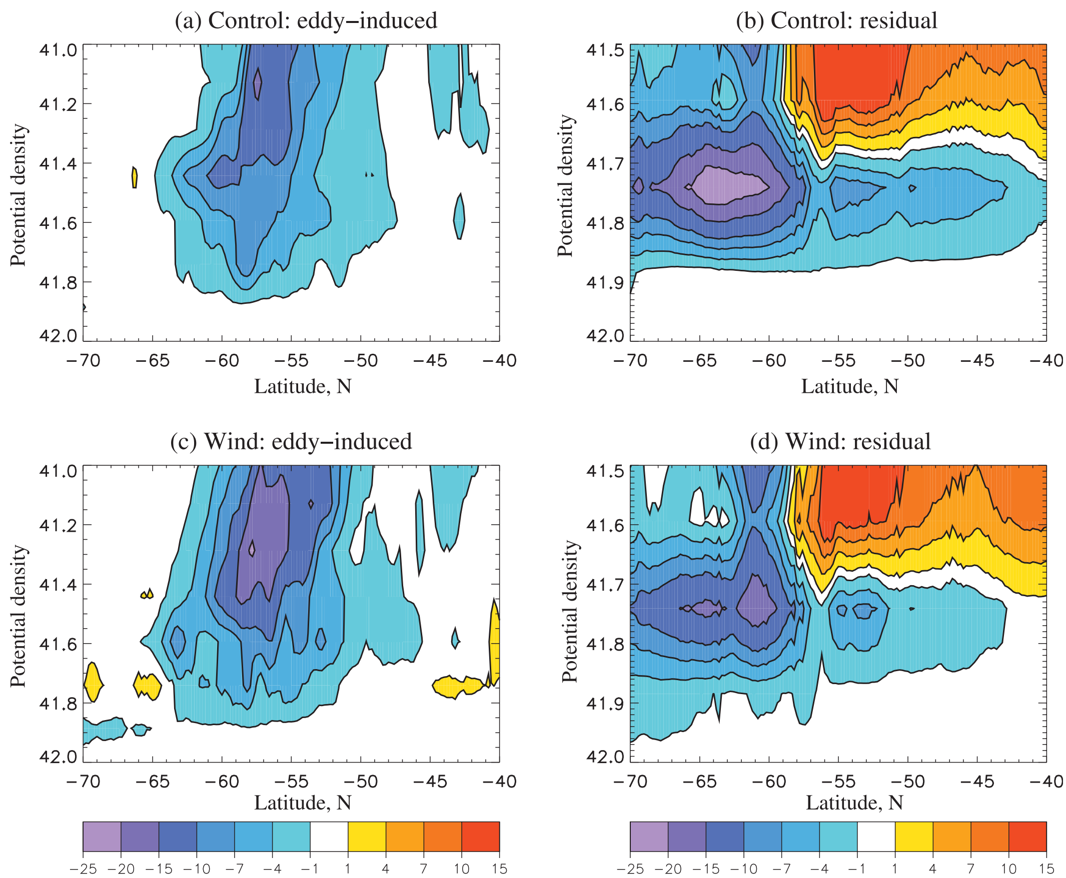

FIG. 4. Overturning circulation (Sv) in the Southern Ocean as a function of potential density (σ_3) in the (a),(b) Control and (c),(d) Wind experiments: (a),(c) eddy-induced and (b),(d) residual within the densest water classes. Negative values indicate counterclockwise circulation. The experiments were run for 40 yr. In each case, the corresponding overturning was determined from the last 5 yr of 5-day snapshots. Note that the vertical scale in (b),(d) is only half of that in (a),(c).

4. Discussion: Energetic arguments

Observational estimates have revealed vast areas of enhanced turbulence in the Southern Ocean, above complex bathymetry and along the major fronts (e.g., Naveira Garabato et al. 2004; Sloyan 2005). It has been argued that the associated values of diapycnal diffusivity could have a strong impact on the large-scale circulation (Naveira Garabato et al. 2007; Saenko 2008). While the energy sources that sustain the observed enhanced turbulence in the Southern Ocean are not clear, evidence has been presented suggesting that some of it might come from the deep and energetic flows of the Antarctic Circumpolar Current (Naveira Garabato et al. 2004; Nikurashin and Ferrari 2011). It has also been estimated that the bottom layer drag is not the primary sink for the wind energy input to the Southern Ocean, leaving breaking internal waves over rough topography as one of the key candidates (Sen et al. 2008). Taken together, it is possible that a fraction of the Southern Ocean eddy energy, which ultimately comes from the wind, is converted locally to diapycnal mixing (Tandon and Garrett 1996).

Neither processes nor parameterizations converting eddy kinetic energy to diapycnal mixing are present in the models considered in the previous sections. On the other hand, most of the coarse-resolution climate models rely on the Gent and McWilliams (1990) scheme to parameterize the eddy-induced transport. As typically implemented, however, the Gent and McWilliams scheme implicitly assumes pure viscous dissipation of the available potential energy release (Tandon and Garrett 1996; Gent et al. 1995). It can be shown that if the eddy energy were lost locally in the interior and were accompanied by small-scale turbulence (both are speculative assumptions) with a mixing efficiency of Γ [$\approx$0.2; Osborn (1980)], the Gent and McWilliams (1990) scheme would imply the following link between the diapycnal diffusivity k and the eddy transfer coefficient K (Tandon and Garrett 1996; Gent et al. 1995):

$$k = \Gamma s_\rho^2 K, \quad (2)$$

where s_ρ is the slope of the mean isopycnals. Assuming further that the zero-order dynamics of the along-isopycnal and across-isopycnal flows in the Southern Ocean interior are given, respectively, by

$$\Psi = \overline{\Psi} + K s_\rho \quad (3)$$

and

$$\Psi = \frac{k}{s_\rho}, \quad (4)$$

one can obtain a scaling for Ψ by combining (2)–(4). This gives

$$\Psi = -\overline{\Psi}\frac{\Gamma}{1-\Gamma}. \quad (5)$$

Equation (3) follows from the zonally averaged budget of momentum in the Southern Ocean (e.g., Ito and Marshall 2008), whereas Eq. (4) represents an approximate balance that can be derived from the zonally veraged buoyancy budget of the form $J_{y,z}(\Psi,\overline{b}) = (k\overline{b}_z)_z$ (where $J_{y,z}$ is the Jacobian in the y–z plane and $\overline{b}$ is the mean buoyancy field) under the assumption of weak stratification in the near-bottom layer (see also Ito and Marshall 2008; Nikurashin and Vallis 2011). If K is itself proportional to the slope of the isopycnals (i.e., $K = \gamma|s_\rho|$, with γ being a positive scaling constant), such as suggested by Visbeck et al. (1997), the final result [Eq. (5)] is the same. For $\Gamma = 0.2$, it gives $\Psi = -0.25\overline{\Psi}$. A similar relation can be obtained from Eq. (1) if $\phi \sim 1$, which is plausible. However, under the assumptions made in this section, it is not straightforward to separate the effects of mixing and wind on the overturning [by, for example, assuming that $|\overline{\Psi}| \ll |Ks_\rho|$, so that combining Eqs. (3) and (4) would lead to $\Psi = -\sqrt{kK}$, such as in Ito and Marshall (2008)]. Energetically, the whole system is driven by the wind: wind steepens isopycnals and generates eddies, and the eddies generate small-scale mixing. The latter creates potential energy, which drives large-scale overturning in the abyss. The scaling [Eq. (5)] implies a stronger AABW transport under stronger Southern Ocean winds.

5. Conclusions

Several recent studies have proposed that the export of AABW from the Southern Ocean is coregulated by wind stress, mesoscale eddies, and diapycnal mixing (Ito and Marshall 2008; Nikurashin and Vallis 2011; Kamenkovich and Goodman 2000). Here, we focused on the role of winds. The simulations based on coarse-resolution climate models suggest that the lower-limb overturning in the Southern Ocean would weaken if the projected warmer twenty-first-century climate were accompanied by stronger zonal winds in the Southern Ocean. The weakening is mainly due to the stronger southward flux associated with the Eulerian-mean transport within the densest water classes. This tends to be compensated by the eddy-induced northward transport in the abyss, although the compensation is not complete.

The employed high-resolution model suggests a stronger response of the eddy-induced circulation to a comparable increase in the Southern Ocean wind stress. As a result, the simulated residual overturning is only weakly affected. One possibility to account for this in coarse-resolution climate models would be to make the eddy transfer coefficient dependent on the slope of the isopycnals (e.g., Visbeck et al. 1997). However, the use of improved ocean physics, and hence the ability to better predict the abyssal overturning, could be attenuated by the large range in the projected changes of the Southern Ocean wind stress.

Assuming that a fraction of the eddy kinetic energy can be converted to diapycnal mixing, the effects of wind stress, mesoscale eddies, and diapycnal mixing on the overturning circulation are not easy to separate. In this case, the whole system is controlled by the wind: wind steepens isopycnals and generates mesoscale eddies, and the eddies generate small-scale mixing, which drives the overturning. As a result, a stronger Southern Ocean wind may lead to a stronger AABW outflow. This idea needs further research.

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