The Connection between Southern Ocean Winds, the Atlantic Meridional Overturning Circulation, and Indo-Pacific Upwelling

M. JOCHUM

Niels Bohr Institute, University of Copenhagen, Copenhagen, Denmark

CARSTEN EDEN

Institut für Meereskunde, Universität Hamburg, Hamburg, Germany

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ABSTRACT

Coupled GCM simulations are analyzed to quantify the dynamic effect of Southern Ocean (SO) winds on transports in the ocean. It is found that the closure for skew diffusivity in the non-eddy-resolving ocean model does not allow for a realistic eddy saturation of the zonal transports in the SO in response to the wind changes and that eddy compensation of the meridional transports in the SO is underestimated too. Despite this underestimated eddy compensation in the SO, however, and in contrast to previous suggestions, the Atlantic meridional overturning circulation (AMOC) strength is almost insensitive to SO winds. In the limit of weak SO winds the AMOC waters upwell not in the SO but rather in the tropical Indo-Pacific. Through their effect on sea ice, weaker SO winds also lead to less production of Antarctic Bottom Water and therefore a deeper and stronger AMOC.

1. Introduction

The meridional heat transport of the Atlantic meridional overturning circulation (AMOC) is a key component of the Earth climate system, but observational and theoretical challenges keep us from fully understanding and predicting it. Earlier ideas that the AMOC strength is sustained by ocean turbulence (Munk and Wunsch 1998) have been challenged by speculations on the possible role of winds in the Southern Ocean (SO) based on the oceanic distribution of carbon isotopes (Toggweiler and Samuels 1995). Their heuristic arguments were also supported by numerical simulations, the interpretation of which were unfortunately hampered by the Veronis effect (Veronis 1975; this refers to spurious diapycnal mixing caused by the mixing tensor not being aligned with isopycnals). Furthermore, it quickly became clear that the wind-driven Eulerian transports in the SO are compensated to a large extent by mesoscale eddy-driven transport of opposite sign (Johnson and Bryden 1989). This pronounced eddy effect for the meridional overturning of the SO remains unaccounted for in the model by Toggweiler and Samuels (1995) but has since shown to be important in both realistic (Hallberg and Gnanadesikan 2006) and idealized numerical models (Viebahn and Eden 2010; Morrison and Hogg 2013; Munday et al. 2013; Hogg et al. 2008) as well as analytical studies (Marshall and Radko 2003; Olbers and Visbeck 2005).

In accordance with the nonacceleration theorem of Andrews and McIntyre (1976)—first discussed in the context of the SO by Johnson and Bryden (1989)—it was found that eddy compensation by eddy-driven meridional transports in the SO is correlated to the strength of the wind-driven Eulerian transports, such that an increase in SO winds only partly leads to a stronger residual overturning (defined as the sum of Eulerian and eddy-induced overturning). This effect is called eddy compensation. A similar effect can be seen in the analysis of zonal transports: increase in winds leads to increased eddy activity, but not necessarily to a stronger Antarctic Circumpolar Current (ACC; Hallberg and Gnanadesikan 2006; Meredith and Hogg 2006; Böning et al. 2008; Hogg et al. 2008). This effect is called eddy saturation.

To limit the computational costs in global GCMs, mesoscale eddies are often parameterized using the

Corresponding author address: M. Jochum, Niels Bohr Institute, University of Copenhagen, Juliane Maries Vej 30, 2100 Copenhagen, Denmark.
E-mail: mjochum@nbi.dk

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closure after Gent et al. (1995), which involves in turn the specification of a skew diffusivity. Viebahn and Eden (2010) and Munday et al. (2013) tested the ability of the parameterizations by Gent et al. (1995) using different choices of the skew diffusivity to account for eddy compensation. The conclusion from these tests is that in some closures for the skew diffusivity the eddy compensation effect on the meridional transports in the SO exists, but is too weak when compared to eddy-permitting model versions. The departing point of the present study is the result of Gent and Danabasoglu (2011) that their eddy parameterization provides the right amount of eddy compensation and eddy saturation. Here, we use the same parameterization, albeit in the coarse-resolution version of CCSM4, and focus on the connection between SO winds and the meridional transports in the Atlantic Ocean.

Several different theoretical models of the global circulation based on zonally averaged dynamics account for eddy compensation and show that meridional transports in the Atlantic are tied to SO winds (Gnanadesikan 1999; Shakespeare and Hogg 2012; Nikurashin and Vallis 2012). Their prediction of a strong linkage between the transports in the SO and the Atlantic is in fact supported by idealized high-resolution models: Munday et al. (2013) find that a doubling of the SO winds leads to an increase in AMOC strength of 30%–50%, depending on the resolution. Note, though, that while the agreement between zonally averaged models and idealized permitting models is remarkable, neither kind of model includes a separate Indo-Pacific basin. We revisit these results here using a realistic but non-eddy-resolving coupled GCM and find almost no sensitivity of AMOC strength to SO winds, despite an underestimated eddy compensation effect. In the limit of weak winds, the Atlantic water simply upwells in the Indo-Pacific region instead of the SO.

The next section provides a description of model and experiments, section 3 discusses the results, and the final section summarizes the results.

2. Model setup

The numerical experiments are performed using the Community Climate System Model, version 4 (CCSM4), in its T31 $\times$ 3 configuration. It consists of fully coupled atmosphere, ocean, land, and sea ice models. The ocean model has a zonal resolution that varies from 340 km at the equator to 40 km around Greenland, and a meridional resolution that varies from 70 km at the equator to 40 km around Greenland and 350 km in the North Pacific. This spatially varying resolution is achieved by placing the north pole of the numerical grid over Greenland, and allows for an acceptable resolution of both the equatorial current system and North Atlantic deep convection. In the vertical there are 60 depth levels; the uppermost layer has a thickness of 10 m and the deepest layer has a thickness of 250 m. The atmospheric model uses T31 spectral truncation in the horizontal (about 3.75° resolution) with 26 vertical levels. The sea ice model shares the same horizontal grid as the ocean model and the land model is on the same horizontal grid as the atmosphere. This setup has been developed specifically for long climate integrations, and a detailed description of this version and comparisons with higher-resolution versions can be found in Shields et al. (2012).

Two particular details that are of relevance for the present study are the representation of diapycnal mixing away from the mixed layer and the representation of mesoscale eddies. The former is described in Jochum (2009): a background value of 0.17 $\mathrm{cm\,s^{-1}}$ with slight enhancements at the critical latitudes of 29°N and 29°S, and reduced values along the equator. The latter is described in Danabasoglu and Marshall (2007) and represents an empirical fit to the vertical profiles of skew diffusivities relevant for the parameterization of Gent and McWilliams (1990) from inverse model estimates (Ferreira et al. 2005): The skew diffusivity in CCSM4 depends on the background stratification with maximum values of 4000 $\mathrm{m^2\,s^{-1}}$ near the surface and minimum values of 400 $\mathrm{m^2\,s^{-1}}$ in the abyss.

Several 500-yr-long experiments were performed; all were initialized using the Polar Science Center Hydrographic Climatology dataset (PHC2) of potential temperature and salinity data [representing a blending of the Levitus et al. (1998) and Steele et al. (2001) data for the Arctic Ocean] and a state of rest in the ocean model. The analyses were performed on the average of the last 50 years of these simulations. The control simulation (CONT) is identical to the one described in Shields et al. (2012), and the design of the three perturbation experiments is that of Gent and Danabasoglu (2011): the wind speed is computed by the atmospheric model, the coupler computes the wind stress, and south of 35°S this stress is then multiplied by 0 (experiment NULL), 0.5 (HALF), and 2 (TWO). Between 35° and 25°S these multipliers are reduced linearly to 1 (Fig. 1). It is only the wind stress seen by the ocean model that is manipulated in this way, while all other forcing fields of the ocean model are left untouched. Of course, a strict separation of momentum and buoyancy forcing is not possible (Saenko 2009), but we do think that the experimental setup will allow for a lowest-order assessment sensitivity of the ocean circulation on SO winds. It should also be noted that our experiments are nevertheless fully coupled simulations, so the actual wind stress cannot be controlled. For the

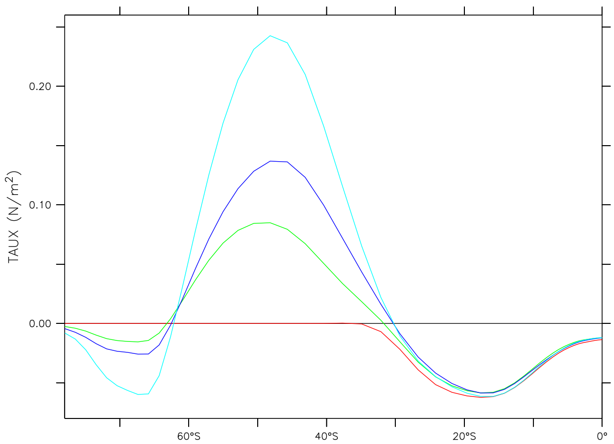

FIG. 1. Zonally averaged mean wind stress: NULL (red), HALF (green), CONT (blue), and CONT (light blue).

interpretation of the results it is fortuitous that the structure of the zonally averaged wind stress is the same in all simulations.

Throughout this article the term ''residual overturning'' will be used as defined by Marshall and Radko (2003): the sum of Eulerian overturning and eddy-induced overturning averaged at fixed heights in geometrical space, except that here the eddy-induced overturning is diagnosed from the parameterized eddy transports. It is also possible to diagnose overturning or tracer transport in density space (Döös and Webb 1994), which is equivalent to using the isopycnal thickness weighted averaging framework of McDougall and McIntosh (2001). The averaging frameworks in geometrical space and density space are not the only possible ones, but can both be used to consistently identify diapycnal or diffusive eddy effects (Eden et al. 2007). While the former is a standard model diagnosis in a z-level coordinate model like the present one, transforming the overturning transports from geometric space to density space and back is more elaborate and still leaves ambiguities in the interpretation with respect to standing eddy effects (Viebahn and Eden 2012). Thus, since the present focus is AMOC strength, which is similar in all averaging frameworks, we chose the much simpler form of residual overturning in geometric space to illustrate the model results.

3. Results and discussion

a. Insensitivity of the AMOC

The most surprising result of the present simulations is the insensitivity of the AMOC to changes in SO wind stress (Fig. 2, Table 1); in particular, there is a vigorous AMOC even without SO winds. Increasing the SO wind stress from NULL to HALF and CONT in fact reduces the AMOC instead of enhancing it, and only a doubling of the winds in TWO leads to a slight increase in AMOC strength (Table 1).

While the AMOC strength is roughly constant under changing wind stress in the SO, the location of its upwelling changes. Figure 2 shows the residual meridional overturning circulation (MOC) in the Indo-Pacific sectors of CONT and NULL. Upwelling in the Indo-Pacific increases drastically going from CONT to NULL, which demonstrates that without SO winds the North Atlantic Deep Water (NADW) upwells in the Indo-Pacific. The SO upwelling and thus also the residual MOC do not completely vanish in NULL, which means that thermohaline surface forcing still drives some overturning in the SO in NULL, but the contribution by the Indo-Pacific upwelling dominates. In the extreme case TWO the upwelling in the Indo-Pacific basin almost vanishes, while the AMOC is compensated almost exclusively by SO upwelling (Fig. 2).

The SO upwelling is wind driven, but in the Indo-Pacific it is commonly assumed that upwelling is balanced by vertical mixing (Munk and Wunsch 1998), for which the fixed diffusivity is not directly affected by winds in our model. On the other hand, it was shown with simple analytical models by Gnanadesikan (1999), Shakespeare and Hogg (2012), and Nikurashin and Vallis (2012) that the SO winds control the stratification in the zonally bounded parts of the idealized models. Those models do not include the Indo-Pacific, but we might speculate that the stratification in the Indo-Pacific is affected in the same way as suggested by the idealized models. Therefore, even for a fixed vertical diffusivity the idealized models suggest that stronger SO winds will lead to weaker Indo-Pacific stratification and thus weaker upwelling, which is in accordance with our model sensitivity (Table 1).

There are two previous studies that have bearing on the present results: Spence et al. (2009) and Farneti and Delworth (2010) both suggest that the connection between AMOC and SO is resolution dependent. However, they disagree on the sign of the dependence, and the eddy-permitting integrations are shorter than 100 years in both (i.e., less than the 200–300 years it typically takes to equilibrate the AMOC; Griffies et al. 2009), but we note that resolution and resolved eddy activity might affect our results with respect to the insensitivity of the AMOC to SO winds.

b. Eddy compensation

Table 1 lists the maximum values of the wind-driven Eulerian transport and the eddy-driven compensating

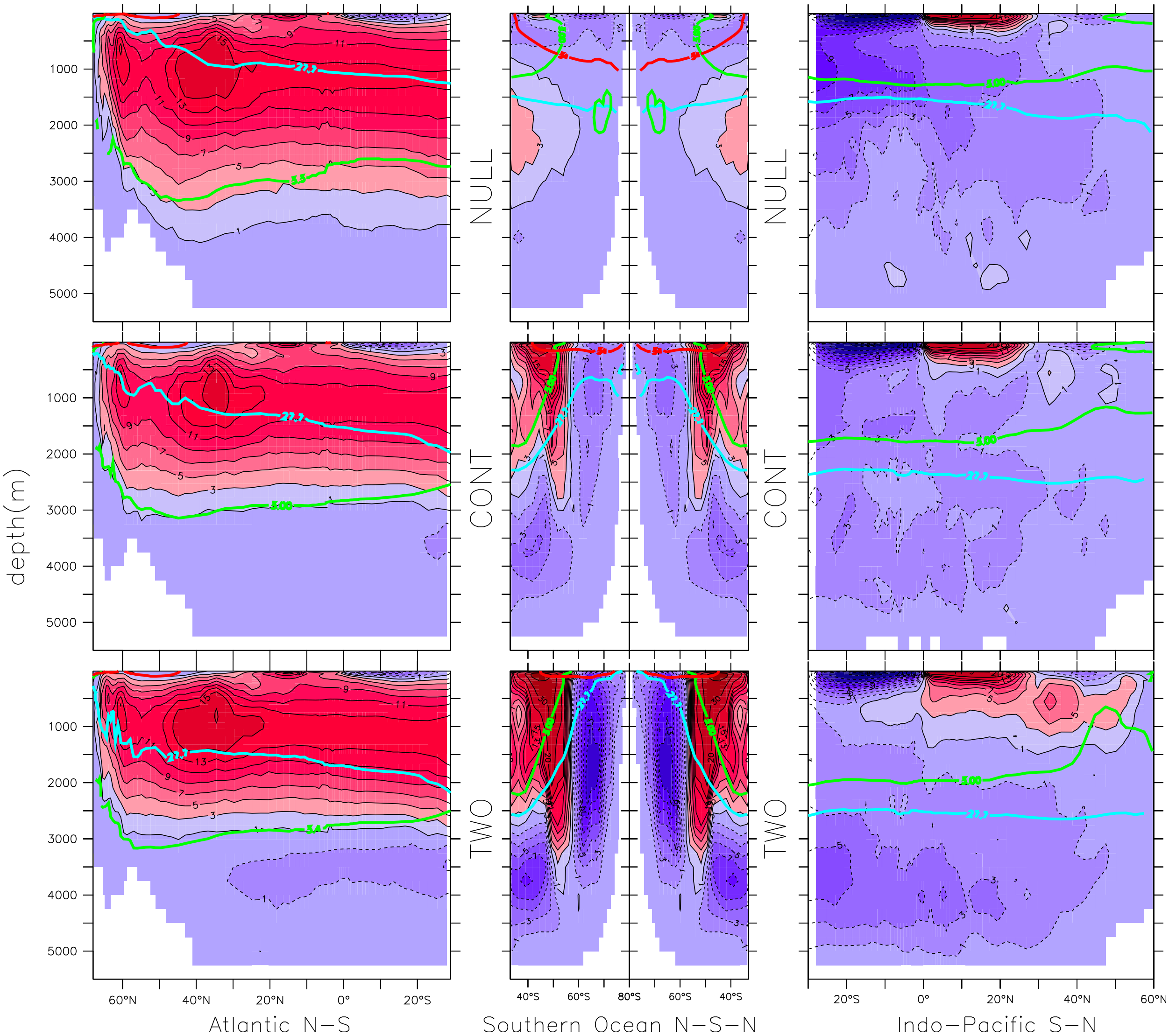

FIG. 2. Zonally averaged (in depth space) mean residual overturning in the (left) Atlantic, (center) Southern Ocean (north–south and south–north), and (right) Indo-Pacific, for (top) NULL, (middle) CONT, and (bottom) TWO. The superimposed lines indicate the basin averaged depths of the $\sigma_\theta = 27.7\,\mathrm{kg\,m^{-3}}$ isopycnal (blue), the 3°C isotherm (green), and the $34.3\,\mathrm{g\,kg^{-1}}$ isohaline (red; visible only in the Southern Ocean and the near-surface high-latitude North Atlantic). Note that to visualize the connection between the basins, the sense of circulation on the left and center six panels is opposite to the one on the right and center six panels (i.e., red shading indicates counterclockwise flow on the left panels and clockwise flow on the right panels). The exact strengths of the various cells are given in Table 1. The circulation patterns and strengths of HALF are roughly halfway between CONT and NULL; they are not shown here for sake of clarity.

transport in the SO. As expected, both transports are approximately linearly correlated with the maximum zonal mean wind stress, but the eddy component correlates at a lesser rate (i.e., a 70% increase in wind stress leads to a 70% increase in the Eulerian component but only a 40% increase of the eddy component; Table 1).

Clearly, the eddy parameterization is responding to the changed wind stress, but an increase in the eddy-driven transports can also be seen with a constant skew diffusivity and is thus no indication for the quality of a closure. It is revealing to analyze the residual overturning. In the present coarse-resolution version, there is no eddy compensation at all, and a 70% increase in wind stress is matched by a 70% increase in residual overturning (Table 1, Fig. 2). In the idealized model version of Viebahn and Eden (2010) the use of a constant skew diffusivity leads to a 20%

TABLE 1. List of experiments with key overturning metrics, with $taux_{max}$ being the maximum of the zonally averaged zonal wind stress ($N\,m^{-2}$) and $AMOC_{eq}$ the residual cross-equatorial zonally averaged transport in the Atlantic; $Up_{Pacific}$ quantifies the net upwelling across 1200 m in the Indo-Pacific basin, SO_{Euler} and SO_{eddy} denote the maximum Eulerian and minimum eddy component of the SO overturning, and SO_{resid} the maximum of their sum (all five in Sv). Also, κ_{GM} is the Gent and McWilliams mixing (GM) coefficient averaged zonally and between 200 and 1000 m, and 66° and 38°S.

Exp	$taux_{max}$	$AMOC_{eq}$	$Up_{Pacific}$	SO_{Euler}	SO_{eddy}	SO_{resid}	κ_{GM}
NULL	0.00	13	9	5	−7	4	500
HALF	0.09	12	7	20	−11	15	570
CONT	0.14	12	3	35	−15	26	610
TWO	0.25	14	1	60	−22	42	700

overestimation of the residual MOC when compared to the equivalent eddy-resolving model version. A similar overestimation of the residual MOC in an idealized model with the parameterization of Gent and McWilliams (1990) is also found by Munday et al. (2013). Viebahn and Eden (2010) find that to reproduce the compensation found in the eddy-resolving model, a doubling of wind stress has to be balanced by doubling of the thickness diffusion coefficient. Similarly, Abernathey et al. (2011) find that the effective magnitude of diffusivity is proportional to the square root of the wind stress, although the detailed level of compensation does depend on the surface buoyancy fluxes as well. In the closure by Danabasoglu and Marshall (2007) that we use in the present model the coefficient is prognostic, but it changes in fact only by approximately 10% when the wind stress is doubled (Table 1).

It is still not known what the right amount of eddy compensation should be. Nonidealized eddy-permitting OGCM studies suggest a doubling in wind stress should lead to a less than 50% increase in SO residual overturning (i.e., substantial eddy compensation): Spence et al. (2010) find that a 25% increase in SO wind stress leads to a similar increase in eddy heat fluxes, and Farneti et al. (2010) find an only 15% increase of the SO residual overturning in response to a 70% increase in SO wind stress. Thus, combined with the results of the cited idealized simulations it is clear that eddy compensation in the present simulations is too weak. The lack of eddy compensation, however, does not invalidate the results of the relation of the AMOC, the SO transports, and the Indo-Pacific upwelling. In fact, it suggests that the slight increase of AMOC strength in TWO is an overestimation, so that one would not expect much of a change at all when eddies are resolved.

TABLE 2. List of experiments with key ACC metrics, with $taux_{max}$ being the maximum of the zonally averaged zonal wind stress, $taux_{dp}$ the zonal wind stress at the southern tip of South America, $taux_{ave}$ the zonal wind stress averaged over the circumpolar barotropic streamlines (all in $N\,m^{-2}$), ACC_{Drake} the ACC transport (Sv) through Drake Passage, and κ_{GM} the GM coefficient averaged zonally and between 200 and 1000 m, and 66° and 38°S.

Exp	$taux_{max}$	$taux_{dp}$	$taux_{ave}$	ACC_{Drake}	κ_{GM}
NULL	0.00	0.00	0.00	0	500
HALF	0.09	0.06	0.02	110	570
CONT	0.14	0.06	0.03	100	610
TWO	0.25	0.14	0.05	180	700

c. *Eddy saturation*

In contrast to the results of Gent and Danabasoglu (2011), there is also no eddy saturation in CCSM4. Table 2 shows that the Drake Passage transport generally increases with increase in maximum zonal mean wind stress over the SO going from NULL to TWO. Doubling roughly the wind stress from CONT to TWO, the Drake Passage transport also roughly doubles. For vanishing wind stress the transport also vanishes, confirming numerical results of Saenko (2009) and theoretical results of Shakespeare and Hogg (2012): the ACC consists of a wind- and a buoyancy-driven component, where the wind-driven one vanishes by construction in NULL but there is still the possibility of a significant buoyancy-driven component. The surface buoyancy fluxes are, however, also affected by the wind changes since the absence of wind also prevents the equatorward export of freshwater from sea ice melt and precipitation in our coupled model. The resulting freshwater cap (Fig. 2, center top) not only inhibits formation of Antarctic Bottom Water (AABW), but also leads to a meridional salinity gradient that opposes the temperature gradient created by the surface heat fluxes. This leads to an almost perfect density compensation, with the effect of vanishing meridional density gradients and thus vanishing geostrophic flow. Note that this almost perfect density compensation might be particular to this model. In other models the balance might be different, and could even lead to a westward ACC. Clearly, this issue needs further investigation.

The comparison of HALF and CONT yields the curious results that in HALF the transport is larger than in CONT, although the maximum zonal mean wind stress is lower by 35%. We do not know the exact reason for this result; however, the ACC transport is sensitive to details of wind field, stratification, and topography, which makes it challenging to design a clean set of experiments in a realistic setting (Allison et al. 2010; Nadeau and Straub 2012). Most recently Langlais et al.

(2015) decomposed the ACC transport into a barotropic and a baroclinic component, and showed that only the baroclinic component is partially saturated; the barotropic component is freely accelerated by the southernmost part of the westerlies. Thus, the level of eddy saturation will depend on the latitude of the maximum wind stress changes—which in the present study is far from Antarctica and should lead to significant saturation. In any case, CCSM4 behavior differs from idealized eddy-resolving model experiments, where the zonal transport is almost insensitive to wind stress changes (Munday et al. 2013). We therefore have to conclude that, in our model setup, the closure by Danabasoglu and Marshall (2007) does not provide eddy saturation.

In all experiments the latitude of the maximum zonally averaged wind stress is the same at 48°S. However, this measure of wind stress strength—which is also used by Gent and Danabasoglu (2011)—might not be appropriate to be related to the zonal transports in the SO. It is often argued that the wind stress at the Drake Passage is a better measure to be related to the Drake Passage transport (Gent et al. 2001; Mazloff 2012), and Allison et al. (2010) show for a reduced gravity model that the baroclinic ACC transport is set by average value of the wind stress over the ACC path. Table 1 shows that the wind stress over the Drake Passage correlates indeed well with the transport, in fact much better than the other two metrics: while the stress over Drake Passage in HALF and CONT is $0.06\,\mathrm{N\,m^{-2}}$ and the transport is 100 and 110 Sv ($1\,\mathrm{Sv} \equiv 10^6\,\mathrm{m^3\,s^{-1}}$), respectively, both the stress and the transport increase by 43% and 56%, respectively in TWO, while both stress and transport vanish in NULL. Thus, with this measure of wind stress magnitude there is no eddy saturation either.

4. Summary and conclusions

The sensitivity of the AMOC to the strength of Southern Ocean winds is explored in a state-of-the-art general circulation model. The model is fully coupled and over the SO the wind stress that is passed on to the ocean model is modified in strength. The setup is reminiscent of the original study by Toggweiler and Samuels (1995), but it also contains a sea ice module and an adiabatic parameterization of eddy effects (Gent et al. 1995), including a flow-interactive skew diffusivity. In this particular model it is found that there is only a weak connection between SO winds and AMOC strength; in the limit of weak winds the AMOC even strengthens as the winds are reduced. The model produces an AMOC of 13 Sv without any SO winds at all. It is the key finding of this study that SO winds do not control the AMOC magnitude; instead they determine where the NADW returns to the surface: the stronger the winds the less diabatic upwelling happens in the Indo-Pacific basin. SO winds do have a secondary impact on AMOC strength, though. Doubling them leads to a 10% increase in AMOC strength through increased upwelling in the SO, and switching them off also leads to a 10% increase due to reduced AABW production and increased NADW penetration depth.

One shortcoming of the present study is that eddies are parameterized. Since eddy compensation is underestimated in our model one can therefore speculate that the contribution of SO winds to AMOC strength is even overestimated here, and therefore negligible. It would be worthwhile though to repeat our experiments with a high-resolution model version to verify our results, or alternatively with an idealized eddy-resolving model with and without a Pacific Ocean basin to verify that the insensitivity of the AMOC to SO winds is indeed due to basin geometry.

The insensitivity of the AMOC strength to SO winds in our coupled GCM is in contrast to theoretical expectations (Gnanadesikan 1999; Shakespeare and Hogg 2012; Nikurashin and Vallis 2012) and idealized single-basin numerical models (e.g., Munday et al. 2013), which tie AMOC strength to SO upwelling. The recent eddy-permitting numerical results by Munday et al. (2013) suggest that a doubling of the SO winds should lead an increase in AMOC strength of 30%–50%, depending on the resolution. Apart from permitting eddies, the simulations by Munday et al. (2013) are quite idealized, so there are several differences that could explain the difference in AMOC sensitivity: topography, basin geometry, air–sea buoyancy fluxes, or sea ice. In fact, closer inspection shows that in NULL AABW is absent, so that NADW can penetrate to greater depth (Fig. 2). Without SO winds, freshwater and sea ice accumulate at the surface of the polar SO and inhibit AABW production (Fig. 2). Increasing the winds allows for sea ice export, brine rejection, and the formation of cold, heavy AABW (e.g., Duffy et al. 2001), limiting the vertical extent of NADW. The most important difference, however, is the missing large Indo-Pacific basin in the idealized models because it opens a second, diabatic pathway for Atlantic water masses to return to the surface.

The crude representation of small-scale ocean turbulence is likely the main shortcoming of the present study. Ocean turbulence is limited by stratification and the amount of available mechanical energy (e.g., Osborn 1980; Huang 1999), whereas in CCSM4, away from bottom topography, it is parameterized as constant diffusivity. Thus, the sensitivities reported here may change when using an energetically consistent mixing scheme.

The Southern Ocean does not only provide a dynamic connection between the three main ocean basins, but it is also a net source of atmospheric carbon dioxide (Gruber et al. 2009). Therefore, variations in SO winds have been postulated to have a controlling influence on the atmospheric carbon dioxide concentration (e.g., Sarmiento and Toggweiler 1984), although observational validation is challenging (e.g., LeQuéré et al. 2007). At least for CCSM4, the present results suggest that because of lacking eddy compensation wind control of outgassing is overestimated (e.g., Lovenduski et al. 2013). Furthermore, for longer time scales the present results suggest that SO winds cannot control oceanic carbon storage. Carbon-rich NADW will always return to the surface, if not in the SO then in the tropical Indo-Pacific.

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