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Sensitivity of Southern Ocean overturning to wind stress changes: Role of surface restoring time scales

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ABSTRACT

The influence of different surface restoring time scales on the response of the Southern Ocean overturning circulation to wind stress changes is investigated using an idealised channel model. Regardless of the restoring time scales chosen, the eddy-induced meridional overturning circulation (MOC) is found to compensate for changes of the direct wind-driven Eulerian-mean MOC, rendering the residual MOC less sensitive to wind stress changes. However, the extent of this compensation depends strongly on the restoring time scale: residual MOC sensitivity increases with decreasing restoring time scale. Strong surface restoring is shown to limit the ability of the eddy-induced MOC to change in response to wind stress changes and as such suppresses the eddy compensation effect. These model results are consistent with qualitative arguments derived from residual-mean theory and may have important implications for interpreting past and future observations.

1. Introduction

Upwelling in the Southern Ocean, driven by the prevailing westerly winds, plays a key role in closing the Meridional Overturning Circulation (MOC) of the global ocean (e.g. Marshall and Speer, 2012). Changes of the strength of this upwelling branch of the MOC associated with changes of the Southern Ocean winds have been proposed as an important mechanism for regulating global climate, in particular, through enhancing or reducing the communication between the carbon-rich deep ocean and the surface (e.g. Toggweiler and Russell, 2008; Anderson et al., 2009). Projections from state-of-the-art climate models suggest that the Southern Ocean westerlies are likely to strengthen as well as become stormier over the next few decades (e.g. Solomon et al., 2007; Chang et al., 2012), both of which act to enhance the Southern Ocean surface wind stress (e.g. Zhai et al., 2012; Zhai, 2013). However, the robust response of the Southern Ocean overturning circulation to changes of the wind field is yet to be determined.

The problem of how the Southern Ocean responds to changes in surface wind stress has been investigated previously in both ocean-only and coupled general circulation models (e.g. Fyfe and Saenko, 2006; Hallberg and Gnanadesikan, 2006; Meredith and Hogg, 2006; Farneti et al., 2010; Viebahn and Eden, 2010; Abernathey et al., 2011; Meredith et al., 2012; Munday et al., 2013). Models that resolve mesoscale ocean eddies are generally found to be less sensitive to wind stress changes than those with parameterised eddies in terms of both circumpolar volume transport/global pycnocline depth and MOC. This insensitivity comes from the subtle balance between the wind-driven Eulerian-mean MOC that acts to steepen isopycnals and the eddy-induced MOC that acts to flatten them out; this balance largely determines the net residual MOC in the Southern Ocean (e.g. Marshall, 1997). Note that it is the residual circulation that advects temperature, salinity, CO_2 and other climatically-important tracers in the eddying ocean.

In eddy-resolving ocean models, an increase in the Southern Ocean wind stress results in enhanced Ekman divergence and convergence that acts to tilt the isopycnals further and increase the mean available potential energy (APE) of the system. This leads to the generation of a more vigorous eddy field that releases the newly-increased APE and at least partially compensates for changes of the wind-driven overturning. As a result, the residual MOC is rendered less sensitive to changes of wind stress, that is, changes of the residual MOC are much smaller than those of the direct wind-driven Eulerian-mean MOC (the so-called *eddy compensation* effect; Viebahn and Eden, 2010). It is, however, unlikely to have perfect eddy compensation due to the different depth dependence of the Ekman and eddy-induced transports; changes of the Ekman transport are strongly surface-intensified

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whereas changes of the eddy-induced transport spread over the whole water depth (e.g. Morrison and Hogg, 2013).

The extent to which changes in the eddy-induced MOC compensate for changes in the wind-driven Eulerian-mean MOC varies among different eddy-resolving models. For example, relatively weak sensitivity of the residual MOC to altered wind forcing is found in an eddying model of Hallberg and Gnanadesikan (2006), while greater sensitivity is found in the models of Viebahn and Eden (2010) and Munday et al. (2013). Recently, Abernathey et al. (2011) showed that the sensitivity of the Southern Ocean residual MOC to changes of the wind forcing depends on the surface boundary condition for buoyancy: a fixed surface buoyancy flux boundary condition severely limits the ability of the residual MOC to change, whereas the use of a Haney-type restoring boundary condition for buoyancy (Haney, 1971) leads to greater sensitivity. Since in thermodynamic equilibrium the residual MOC matches the buoyancy forcing (e.g. Walin, 1982; Watson and Naveira, 2006; Badin and Williams, 2010), the higher degree of freedom at which surface buoyancy flux can vary under the restoring boundary condition implies a higher sensitivity of the residual MOC.

In Abernathey et al. (2011), a surface restoring time scale of 30 days was used for model experiments under the restoring boundary condition. In the ocean, due to the lack of observations, it remains unclear on what time scales the surface turbulent heat fluxes damp the sea surface temperature anomalies, although the spatial scales of these anomalies are believed to be important (e.g. Bretherton, 1982; Frankignoul, 1985).[1] For example, studies based on heat flux data derived from ship and satellite observations suggest that the restoring time scales can vary from less than one month to almost one year in the Southern Ocean, depending on season and location (e.g. Park et al., 2005). Recently, Shuckburgh et al. (2011) studied the mixed layer lateral eddy fluxes mediated by air–sea interaction and found a large sensitivity of surface eddy diffusivity to prescribed surface restoring time scale. However, the question of whether and how the sensitivity of the Southern Ocean MOC to changes in wind stress depends on the surface restoring time scale is, to our knowledge, yet to be explored.

The aim of this study is to investigate the effect of different surface restoring time scales on the response of the Southern Ocean overturning to wind stress changes, extending the recent work by Abernathey et al. (2011). We begin in Section 2 by presenting some qualitative arguments based on the residual-mean framework of Marshall and Radko (2003) to illustrate the influence of different surface boundary conditions. After describing the numerical model setup and experiment design in Section 3, we present and discuss changes of the eddy-induced and residual MOCs in response to wind stress changes in experiments with various restoring time scales in Section 4. We close with a summary in Section 5.

2. Role of surface restoring on Southern Ocean response

Here we adopt the residual-mean framework of Marshall and Radko (2003) to illustrate the influence of different surface restoring time scales on the response of the Southern Ocean to wind stress changes. The time and zonally-averaged buoyancy equation is given by

$$J(\Psi_{res}, \bar{b}) = \frac{\partial \bar{B}}{\partial z}, \tag{1}$$

where $b = -g(\rho - \rho_0)/\rho_0$ is buoyancy, B is the buoyancy forcing, Ψ_{res} is the streamfunction of the residual circulation in the meridional plane (MOC), and overbars denote time and zonal averaging. Following Marshall and Radko (2003), the residual MOC can be written as a combination of the Eulerian-mean MOC ($\bar{\Psi}$) and the eddy-induced MOC (Ψ^*), i.e.

$$\Psi_{res} = \bar{\Psi} + \Psi^* = -\frac{\tau}{\rho_0 f} + Ks, \tag{2}$$

where τ is zonal wind stress, ρ_0 is reference density, f is the Coriolis parameter, $s = -\bar{b}_y/\bar{b}_z$ is the mean isopycnal slope and K is the eddy thickness diffusivity.

Using mixing length theory, the eddy diffusivity can be expressed as

$$K \simeq V_e L_e, \tag{3}$$

where V_e denotes a characteristic eddy velocity and L_e denotes a characteristic eddy length scale. Following Visbeck et al. (1997) and Marshall et al. (2012), we assume that $V_e \simeq \sigma L_e$, where σ is the Eady growth rate, given by

$$\sigma = \frac{f}{\sqrt{Ri}} = \frac{f}{N/|\bar{u}_z|} = N|s|. \tag{4}$$

Here N is the buoyancy frequency with $N^2 = \bar{b}_z$. Eq. (4) shows that the eddy growth rate depends linearly on the mean isopycnal slope. Combining Eqs. (2)–(4), while noting that s is always negative in our model (see Fig. 1), the eddy diffusivity is then given by

$$K \simeq -L_e^2 N s \tag{5}$$

and the eddy-induced MOC is given by

$$\Psi^* \simeq -L_e^2 N s^2. \tag{6}$$

The eddy-induced MOC is therefore anticlockwise and depends quadratically on the mean isopycnal slope (e.g. Visbeck et al., 1997).

Following Marshall and Radko (2003), we assume zero stratification within the surface mixed layer and neglect the entrainment fluxes at its base. Integrating Eq. (1) over the depth of the surface mixed layer h_m while noting $\Psi_{res} = 0$ at the surface gives

$$\Psi_{res|_{z=-h_m}} \frac{\partial \bar{b}_s}{\partial y} = \bar{B}, \tag{7}$$

where $\bar{B}$ is interpreted as the effective buoyancy forcing that includes both air–sea buoyancy fluxes and lateral diabatic eddy fluxes in the mixed layer.

In the ocean interior, we assume the buoyancy forcing is weak, i.e., $B = 0$, and Eq. (1) reduces to

$$J(\Psi_{res}, \bar{b}) = 0, \tag{8}$$

meaning that the residual circulation remains constant along the mean isopycnals, i.e., $\Psi_{res} = \Psi_{res}(\bar{b})$.

At the northern boundary of our model, the buoyancy distribution throughout the water column is prescribed through a restoring boundary condition at a short time scale, i.e.,

$$\bar{b} = \bar{b}_N(z). \tag{9}$$

Physically, $\bar{b}_N$ is set by ocean adjustment to global diabatic processes further to the north of our model domain (Munday et al., 2011). Fig. 1 shows a schematic of the conceptual model used by this study. We now consider surface restoring boundary conditions at two limits.

2.1. Strong surface restoring

In the limit of strong surface restoring ($\lambda \gg \sigma$, where λ^{-1} is the surface restoring time scale), buoyancy at the surface, b_s, is

[1] The situation for the sea surface salinity (SSS) is very different because it does not rain preferentially over regions of positive SSS anomalies nor evaporate preferentially over regions of negative SSS anomalies (e.g. Zhai and Greatbatch, 2006a,b).