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Sensitivity of Southern Ocean circulation to wind stress changes: Role of relative wind stress

D.R. Munday[a,b,*], X. Zhai[c,d]

[a] British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK
[b] Atmospheric, Oceanic and Planetary Physics, Department of Physics, University of Oxford, Oxford OX1 3PU, UK
[c] Centre for Ocean and Atmospheric Sciences, School of Environmental Sciences, University of East Anglia, Norwich, UK
[d] School of Marine Science, Nanjing University of Information Science and Technology, Nanjing, China

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ABSTRACT

The influence of different wind stress bulk formulae on the response of the Southern Ocean circulation to wind stress changes is investigated using an idealised channel model. Surface/mixed layer properties are found to be sensitive to the use of the relative wind stress formulation, where the wind stress depends on the difference between the ocean and atmosphere velocities. Previous work has highlighted the surface eddy damping effect of this formulation, which we find leads to increased circumpolar transport. Nevertheless the transport due to thermal wind shear does lose sensitivity to wind stress changes at sufficiently high wind stress. In contrast, the sensitivity of the meridional overturning circulation is broadly the same regardless of the bulk formula used due to the adiabatic nature of the relative wind stress damping. This is a consequence of the steepening of isopycnals offsetting the reduction in eddy diffusivity in their contribution to the eddy bolus overturning, as predicted using a residual mean framework.

1. Introduction

The transfer of momentum between the atmosphere and ocean is usually parameterised as a stress applied at the surface. Arguments originating from the theory of vertical turbulent transfers give rise to the following expression for the applied stress

$$\boldsymbol{\tau}_{relative} = \rho_a c_d |\mathbf{U}_{10} - \mathbf{u}_s| (\mathbf{U}_{10} - \mathbf{u}_s), \quad (1)$$

where $\mathbf{U}_{10} = (U_{10}, V_{10})$ is the 10 m (atmospheric) wind velocity, $\mathbf{u}_s = (u_s, v_s)$ is the surface ocean velocity, ρ_a is air density, and c_d is a drag coefficient, which itself may be a weak function of $\mathbf{U}_{10} - \mathbf{u}_s$. We will refer to the use of Eq. (1) to calculate wind stress as using "relative wind stress." In the limit that $\mathbf{u}_s \ll \mathbf{U}_{10}$, known as the resting ocean approximation, Eq. (1) can be simplified to

$$\boldsymbol{\tau}_{resting} = \rho_a c_d |\mathbf{U}_{10}| \mathbf{U}_{10}. \quad (2)$$

The use of relative wind stress leads to a slight decrease in the stress felt by the ocean, relative to the resting ocean approximation. This contributes to a reduction of the power input to the ocean circulation by ~20–35% (Duhaut and Straub, 2006; Hughes and Wilson, 2008; Zhai and Greatbatch, 2007; Zhai et al., 2012). Since the power input from the wind is a major source of energy to the ocean (Ferrari and Wunsch, 2009; Wunsch and Ferrari, 2004) this could have significant consequences for the large-scale ocean circulation, its variability, and its sensitivity to changes in surface wind stress.

Relative wind stress exerts a torque on individual eddies that opposes their circulation and so directly damps them. This is due to the increase in the velocity *difference* between ocean and atmosphere from one side of the eddy to the other (see Fig. 1 of Zhai et al., 2012). This acts as a drag at the surface of the ocean and significantly increases the rate of spindown of waves and eddies via the introduction of "top friction" (Dewar and Flierl, 1987). In regions in which mesoscale eddies play an important role in ocean circulation/dynamics, such as the Southern Ocean, this could indicate an important role for relative wind stress.

The Southern Ocean is subject to strong atmospheric winds and makes a large regional contribution to the global integral of mechanical power input to the ocean (Wunsch, 1998). It has a strong influence on global climate, via its Residual Meridional Overturning Circulation (RMOC) and the Antarctic Circumpolar Current (ACC) (Meredith et al., 2011). Mesoscale eddies play prominent roles in the momentum (Johnson and Bryden, 1989; Munk and Palmén, 1951), heat (Bryden, 1979; Jayne and Marotzke, 2002; Meijers et al., 2007), and kinetic energy (Abernathey et al., 2011; Cessi, 2008; Cessi et al., 2006) budgets of the Southern Ocean. The role that relative wind

* Corresponding author at: British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK. Tel.: +44 1223 221 365.
E-mail addresses: danday@bas.ac.uk (D.R. Munday), xiaoming.zhai@uea.ac.uk (X. Zhai).

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stress might play in the dynamics and circulation of the Southern Ocean can be usefully framed in terms of a residual mean treatment of the RMOC.

In residual mean theory, the streamfunction of the RMOC is written as the combination of the Eulerian mean MOC ($\overline{\Psi}$) and the eddy-induced bolus overturning (Ψ^*) (see, e.g., Marshall and Radko, 2003), i.e.

$$\Psi_{\rm res} = \overline{\Psi} + \Psi^* = -\frac{\overline{\tau}_x}{\rho_0 f} + Ks. \tag{3}$$

In Eq. (3), $\overline{\tau}_x$ is the time-mean zonal wind stress, ρ_0 is the Boussinesq reference density, f is the Coriolis parameter, K is the quasi-Stokes/eddy diffusivity for the buoyancy field ($b = -g(\rho - \rho_0)/\rho_0$) and $s = -\overline{b}_y/\overline{b}_z$ is the isopycnal slope. There are a considerable number of ways to formulate the dependence of K on external parameters. For the current purpose, the most informative is to use mixing length theory (Prandtl, 1925) to relate K to the product of an eddy length and eddy velocity scale, i.e. $L_{\rm eddy}$ and $U_{\rm eddy}$, such that $K = L_{\rm eddy}U_{\rm eddy}$ (see, e.g., Green, 1970; Stone, 1972; Eden and Greatbatch, 2008).

In Eq. (3), it is the mean wind stress that plays a role in setting the residual overturning. Relative wind stress can therefore directly impact the residual overturning by reducing $\overline{\tau}_x$. Furthermore, the direct damping of the eddy field can be reasonably expected to alter both $L_{\rm eddy}$ and $U_{\rm eddy}$, i.e. K, and, hence, the eddy-induced bolus overturning and net RMOC. Intuition suggests that damping the eddy field will reduce $U_{\rm eddy}$ and K, and hence Ψ^*.

A further indirect effect can also occur through the isopycnal slope, s, which can be related to the zonal volume transport of the ACC via thermal wind. Eddies play a large role in setting the stratification of the ocean (e.g. Karsten et al., 2002) as part of a dynamic balance with other processes. Damping eddies at the surface may alter the balance between processes that set the stratification and so change s. This would then have a knock-on effect on the bolus overturning and zonal transport of the ACC. As an example, in the quasi-geostrophic Southern Ocean simulations of Hutchinson et al. (2010) the use of relative wind stress results in a 38 Sv *increase* in circumpolar transport. This comes about due to steepening of isopycnals and an increase in the geostrophic velocity field via thermal wind shear.

The above discussion is framed in terms of a particular wind stress and the ocean circulation/stratification that results. However, when the wind stress over the Southern Ocean changes, the mesoscale eddy field also responds. This leads to a decrease in the sensitivity of the circumpolar transport of the ACC (Hallberg and Gnanadesikan, 2001; Tansley and Marshall, 2001) and of the RMOC (Farneti et al., 2010; Hallberg and Gnanadesikan, 2006) to changes in wind stress when the eddy field is resolved instead of parameterised. These phenomena are known as eddy saturation (Straub, 1993) and eddy compensation (Viebahn and Eden, 2010). Although there are subtleties to the degree of eddy saturation/compensation that a particular model may exhibit, e.g. the presence of shallow coastal shelves (Hogg and Munday, 2014) or surface breaking continents (Munday et al., 2015) and the use of fixed heat/buoyancy fluxes vs. restoring to a fixed temperature/buoyancy profile (Abernathey et al., 2011; Zhai and Munday, 2014, henceforth AMF11 and ZM14, respectively), their emergence upon resolution of an eddy field is robust in many respects.

Many of the above cited papers use idealised model configurations to investigate the effect changing wind stress on circumpolar transport and/or the RMOC. In doing so, they usually use a specified wind stress (e.g. AMF11; ZM14; Morrison and Hogg, 2013; Munday et al., 2013). Applying a constant wind stress is certainly within the idealised spirit and design of such experiments. However, it rules out the direct damping of the mesoscale eddy field that takes place under relative wind stress and the role that this might play in setting the sensitivity of the RMOC and/or stratification to changing winds.

In this paper we seek to answer the following questions: (1) Can the impact of relative wind stress be modelled simply by accounting for the reduced mean wind stress? (2) Does the direct damping of the mesoscale eddy field have implications for Southern Ocean dynamics? (3) Does relative wind stress significantly alter the sensitivity of the circumpolar transport and the RMOC to wind stress changes?

We begin in Section 2 with a brief description of the experimental design and model domain. The control simulations of three suites of experiments are discussed in Section 3. Section 4 briefly derives a simplified mechanical energy budget for the ocean including the effects of relative wind stress. The sensitivity to wind stress changes across the full suite of experiments is discussed in Section 5. We close with a summary and discussion of our results in Section 6.

2. Experimental design

In order to investigate the impact of relative wind stress, and its associated eddy damping effects, on Southern Ocean dynamics we adopt the idealised MIT general circulation model (MITgcm, see Marshall et al., 1997a; 1997b) configuration of AMF11, adapted to a coarser grid spacing by ZM14. This model domain is a zonally re-entrant channel that is 1000 km in zonal extent, nearly 2000 km in meridional extent, and 2985 m deep with a flat bottom. There are 33 geopotential levels whose thickness increase with depth, ranging from 10 m at the surface to 250 m for the bottom-most level.

The horizontal grid spacing is chosen to be 10 km, which is sufficiently fine so as to permit a vigorous eddy field without incurring undue computational cost. This grid spacing makes the model eddy-permitting, rather than eddy-resolving, with the control wind stress (see below for forcing details) giving a first baroclinic Rossby radius in the range of ~5 km near the southern boundary and ~25 km near the northern. It is important to note that the eddies are generally several multiples of the deformation radius in size and that use of a 10 km grid spacing does not preclude the emergence of a high degree of eddy saturation (Munday et al., 2015) and as such we deem it sufficient for our purposes.

We employ the K-profile parameterisation (KPP) vertical mixing scheme (Large et al., 1994) and a linear bottom friction in addition to the much weaker drag from a noslip bottom boundary condition. The equation of state is linear and only temperature variations are considered. The model is set on a β-plane and lateral boundaries are noslip. Parameters values for bottom friction, viscosity, etc., are as given in Table 1.

The model's potential temperature, θ, is forced by a heat flux at the surface given by

$$Q(y) = \begin{cases} -Q_0 \sin(3\pi y/L_y), & \text{for } y < L_y/3 \\ 0, & \text{for } y > L_y/3 \end{cases} \tag{4}$$

Table 1
Model parameters.

Parameter	Symbol	Value	Units
Domain size	L_x, L_y	1000, 1990	km
Latitude of sponge edge	L_{sponge}	1890	km
Domain depth	H	2985	m
Reference density	ρ_0	1000	kg m^{-3}
Thermal expansion coefficient	α	2×10^{-4}	K^{-1}
Coriolis parameter	f_0	-1×10^{-4}	s^{-1}
Gradient in Coriolis parameter	β	1×10^{-11}	m^{-1}s^{-1}
Surface heat flux magnitude	Q_0	10	W m^{-2}
Control wind speed	U_0	12	m s^{-1}
Bottom drag coefficient	r_b	1.1×10^{-3}	m s^{-1}
Sponge restoring timescale	t_{sponge}	7	days
Sponge vertical scale	h_e	1000	m
Horizontal grid spacing	Δx, Δy	10	km
Vertical grid spacing	Δz	10–250	m
Vertical diffusivity (θ)	κ_v	10^{-5}	m^2 s^{-1}
Horizontal diffusivity (θ)	κ_h	0	m^4 s^{-1}
Vertical viscosity ($\mathbf{u}$)	A_v	10^{-3}	m^2 s^{-1}
Horizontal hyperviscosity ($\mathbf{u}$)	A_4	10^{10}	m^4 s^{-1}