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The impact of atmospheric storminess on the sensitivity of Southern Ocean circulation to wind stress changes

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ABSTRACT

The influence of changing the mean wind stress felt by the ocean through alteration of the variability of the atmospheric wind, as opposed to the mean atmospheric wind, on Southern Ocean circulation is investigated using an idealised channel model. Strongly varying atmospheric wind is found to increase the (parameterised) near-surface viscous and diffusive mixing. Analysis of the kinetic energy budget indicates a change in the main energy dissipation mechanism. For constant wind stress, dissipation of the power input by surface wind work is always dominated by bottom kinetic energy dissipation. However, with time-varying atmospheric wind, near surface viscous dissipation of kinetic energy becomes increasingly important as mean wind stress increases. This increased vertical diffusivity leads to thicker mixed layers and higher sensitivity of the residual circulation to increasing wind stress, when compared to equivalent experiments with the same wind stress held constant in time. This may have implications for Southern Ocean circulation in different climate change scenarios should the variability of the atmospheric wind change rather than the mean atmospheric wind.

1. Introduction

The Southern Ocean (SO) is believed to have a strong influence on global climate via its Residual Meridional Overturning Circulation (RMOC) and the Antarctic Circumpolar Current (ACC) (Meredith et al., 2011). These lead to the upwelling of deep water masses and a zonal connection between major ocean basins, respectively. 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 due to the combination of large zonal wind stress and strong zonal ocean currents (Wunsch, 1998).

Mesoscale eddies play a prominent role in the momentum budget of the Southern Ocean (Munk and Palmén, 1951; Johnson and Bryden, 1989). They flux a large amount of heat southwards (Bryden, 1979; Jayne and Marotzke, 2002; Meijers et al., 2007) and dominate the dissipation of kinetic energy at the bottom of the water column (Cessi et al., 2006; Cessi, 2008; Abernathey et al., 2011). The use of eddy-resolving, or at least eddy-permitting, numerical models allows the emergence of two dynamical phenomena that have been dubbed eddy saturation and eddy compensation.

Eddy saturation refers to the loss of sensitivity of the volume transport of a circumpolar current to changes in wind stress (Hallberg and Gnanadesikan, 2006; Tansley and Marshall, 2001). This loss of sensitivity can extend to the limit of no zonal wind stress (Munday et al., 2013) and changes in the sensitivity can be linked to the zonal momentum balance of the current (Munday et al., 2015). The degree of eddy saturation that a given model configuration achieves is subject to subtleties due, for example, to the inclusion of shallow coastal areas (Hogg and Munday, 2014) or the structure of the wind forcing (Nadeau and Straub, 2009; 2012).

Eddy compensation is the reduced sensitivity to changes in wind stress of the RMOC when eddies are resolved or permitted (Viebahn and Eden, 2010; Abernathey et al., 2011). Although complimentary to eddy saturation, eddy compensation is dynamically distinct (Meredith et al., 2012; Morrison and Hogg, 2013). Like eddy saturation, the degree to which a particular model's RMOC is compensated depends on several different aspects of the model including, but not limited to, whether the surface buoyancy forcing is fixed flux vs. restoring to a fixed buoyancy (Abernathey et al., 2011, henceforth AMF11) and even the particular timescale used in the restoring condition (Zhai and Munday, 2014, henceforth ZM14).

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Investigations into eddy saturation and eddy compensation using numerical models typically involve varying the magnitude of the mean wind stress in the Southern Ocean, without concern as to whether this variation is due to changes in the mean atmospheric wind or atmospheric variability. In practice, changes of the mean stress may be brought about by either, owing to the nonlinear dependence of the wind stress on the wind (Zhai, 2013). This is illustrated in Fig. 1a, which shows the mean zonal wind (blue line) from the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al., 1996) as well as the square root of the Eddy Kinetic Energy (EKE) of the atmospheric wind (red line). Clearly the variability of the wind is significant at every latitude, with particularly large values in the Southern Ocean. In Fig. 1b we show the time-mean wind stress (blue line), which includes data from every timestep of the reanalysis, and the wind stress calculated from the mean wind alone using the bulk formula of Large and Pond (1981) (red line). This highlights how variability of the atmospheric wind makes a large contribution to the mean wind stress felt by the ocean, particularly at mid and high latitudes (Zhai, 2013).

Variability of the atmospheric wind results in time-varying wind stress, which is capable of exciting near-inertial motions in the surface ocean. Recent studies (Furuichi et al., 2008; Zhai et al., 2009; Rath et al., 2014) show that the majority of the wind energy input to the near-inertial motions is dissipated and lost to turbulent mixing within the upper 200 m, contributing to deepening of the mixed layer and cooling of the sea surface temperature. Jouanno et al. (2016) demonstrate that the passage of storms over an idealised Southern Ocean leads to a slight enhancement of both mean and eddy kinetic energy. Energy dissipation at depth is also increased, in part due to the generation of more near-inertial waves. In their experiments with storms, there is a shift in the energy balance such that more energy is dissipated by vertical viscous processes with respect to a stormless control experiment. This enhanced dissipation is found to be sensitive to the strength of the wind stress and the propagation speed and strength of the storms, with increases in any of these leading to further enhancement of the viscous dissipation.

Turbulent mixing associated with energy dissipation is also likely to contribute to water mass transformation processes in the surface diabatic layer. Wind stress variability can play a direct role in mode water formation via the destruction or creation of potential vorticity at ocean fronts (Thomas, 2005) or by generating wave-induced vertical mixing (Shu et al., 2011). Changes in the mode of variability of atmospheric wind, i.e. ENSO or the Southern Annular Mode, has been observed to change the dominant creation mechanism for Subantarctic Mode Water (Naveira Garabato et al., 2009). In other words, there may be a role for wind-induced near-inertial energy and/or wind variability to play in the emergence of eddy saturation and compensation due to changes in the mode and intensity of near surface dissipation.

In this paper we aim to investigate how changing the wind stress felt by the ocean via an increase in the variability of the atmospheric wind, instead of the mean wind, impacts upon eddy saturation and eddy compensation. In Section 2 we give a brief description of the experimental design and model domain. Section 3 describes the circulation achieved at the control wind stress. Section 4 discusses the sensitivity to wind stress of the model's energy budget under conditions of varying wind. Section 5 discusses the sensitivity of the Southern Ocean circulation to wind stress changes. We close with a summary and discussion of our results in Section 6.

2. Experimental design

In order to investigate the impact of time-varying atmospheric wind on Southern Ocean dynamics we adopt the idealised MIT general circulation model (MITgcm, see Marshall et al., 1997a, b) configuration of AMF11, adapted to a coarser grid spacing by ZM14 and used by Munday and Zhai (2015, henceforth MZ15) to investigate the role of relative wind stress, in which the effect of ocean current speed on surface wind stress is taken into account, on Southern Ocean circulation. The model domain is a zonally reentrant 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 10m at the surface to 250 m for the bottom-most level.

Table 1
Model parameters.

Parameter	Symbol	Value	Units
Domain size	L_x, L_y	1000, 1990	km
Latitude of sponge edge	L_{sponge}	890	km
Domain depth	H	2985	m
Boussinesq reference density	ρ_0	1000	$kg\,m^{-3}$
Thermal expansion coefficient	α	2×10^{-4}	K^{-1}
Coriolis parameter	f_0	-1×10^{-4}	km
Gradient in Coriolis parameter	β	1×10^{-11}	$m^{-1}s^{-1}$
Surface heat flux magnitude	Q_0	10	$W\,m^{-2}$
Peak wind speed	U_0	7	$m\,s^{-1}$
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 (momentum)	A_v	10^{-3}	$m^2\,s^{-1}$
Horizontal hyperviscosity	A_4	10^{10}	$m^4\,s^{-1}$

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. Strictly speaking, this grid spacing makes the model eddy-permitting, rather than eddy-resolving, since it does not resolve the first baroclinic deformation radius throughout the model domain. In particular, it cannot resolve the eddy formation process. However, when mature, i.e. at their maximum size/strength, eddies are typically several deformation radius across. Furthermore, this grid spacing is fine enough that substantial eddy saturation of the zonal transport occurs in domains with bottom bathymetry (Munday et al., 2015). 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. The equation of state is linear and only temperature variations are considered. The model is set on a β-plane. Parameter values for bottom friction, viscosity, etc, are as given in Table 1. The schematic in Fig. 2 indicates the meridional cross-section of the model configuration and forcing, including the northern boundary sponge (see below for details).

The model's potential temperature, θ, is forced by a constant heat flux at the surface and restored to a prescribed stratification in a sponge layer within 100 km of the northern boundary. The surface heat flux is given by

$$Q(y) = \begin{cases} -Q_0 sin(3\pi y/L_y), & \text{for } y < L_y/3 \\ 0, & \text{for } y > L_y/3 \end{cases} \tag{1}$$

where Q_0 is the magnitude if the flux and L_y is the meridional extent of the domain, as per AMF11 and ZM14, with $y = 0$ km placed at the centre of the domain following MZ15. This broadly describes the observed distribution of surface buoyancy flux around the SO (see Fig. 1 of AMF11). Within 100 km of the northern boundary, potential temperature is restored to the stratification given by

$$\theta_N(z) = \Delta\theta\left(e^{z/h_e} - e^{-H/h_e}\right)/\left(1 - e^{-H/h_e}\right). \tag{2}$$