An Antarctic Circumpolar Current driven by surface buoyancy forcing

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[1] Simulations of an idealised, but eddy-resolving, channel model of the Antarctic Circumpolar Current (ACC) are used to investigate the sensitivity of ACC transport to wind and surface buoyancy forcing. The results are consistent with theoretical predictions of the eddy-saturated limit, where transport is independent of wind stress. In this parameter regime, buoyancy forcing provides the primary control over ACC transport. **Citation:** Hogg, A. McC. (2010), An Antarctic Circumpolar Current driven by surface buoyancy forcing, *Geophys. Res. Lett.*, *37*, L23601, doi:10.1029/2010GL044777.

1. Introduction

[2] The Antarctic Circumpolar Current (ACC) is generally considered to be forced by strong westerly winds over the Southern Ocean [see, e.g., *Allison et al.*, 2010, and references therein]. However, a comprehensive theoretical prediction of the volume of water transported by the ACC as a function of wind stress remains elusive; recent and current theories include a linear dependence of transport upon wind stress [*Marshall and Radko*, 2003] (consistent with most coarse-resolution model simulations), a square root dependence [*Johnson and Bryden*, 1989] and the eddy-saturated limit [*Straub*, 1993; *Hallberg and Gnanadesikan*, 2006] in which transport is effectively independent of wind stress. There remains the possibility that surface buoyancy forcing modulates the ACC transport, either through local production of Antarctic Bottom Water (AABW) [*Gent et al.*, 2001] or interhemispheric influences [*Gnanadesikan and Hallberg*, 2000; *Fučkar and Vallis*, 2007]. Even laboratory analogues of the ACC [*Cenedese et al.*, 2004; *Hogg and Griffiths*, 2010] produce different interpretations of the sensitivity to mechanical (wind stress) forcing. Thus, much remains to be understood regarding the forces that drive the ACC.

[3] The eddy saturation hypothesis describes a flow state in which additional power input from the winds is absorbed entirely into eddy kinetic energy without changing the mean circumpolar flow. While the hypothesis appears counter-intuitive, it has gained currency with the advent of numerical models that can partially or fully resolve mesoscale eddies. It is now clear that quasigeostrophic channel models of the ACC result in transport that is completely independent of wind stress [e.g., *Hogg and Blundell*, 2006]; primitive equation models also appear to approach an eddy-saturated limit at eddy permitting resolution [*Hallberg and Gnanadesikan*, 2006; *Farneti et al.*, 2010]. Furthermore, models in the eddy-saturated parameter regime predict enhanced eddy kinetic energy following strong wind events, with a lag of several years; these predictions are consistent with observations [*Meredith and Hogg*, 2006], and can only be explained by the eddy-saturated model. Finally, long-term observations using float data indicate minimal change in the isopycnal slope in the Southern Ocean [*Böning et al.*, 2008], supporting a weak response of ACC transport to observed changes in the wind field. The debate over the extent to which the ocean is eddy-saturated depends upon the applicability of models using quasigeostrophic dynamics. For example, diabatic interaction between eddies and the stratification may be critical to control of the system, which would imply that eddy saturation may have limited applicability in a regular primitive equation ocean model, regardless of the extent to which eddies are resolved.

[4] In this paper we embark on eddy-resolving simulations in a primitive equation, diabatic channel model. The inclusion of diabatic terms (rather than specified stratification) requires that the model be forced with surface buoyancy forcing to generate a realistic model stratification, implying that long, computationally expensive, spinup simulations are essential. To reduce the computational demand we restrict our simulation to an idealised domain: a simple re-entrant channel only 4000 km long, driven by a combination of specified wind and buoyancy forcing. We examine whether flow in this model is eddy saturated, and also the dependence of transport upon buoyancy forcing.

2. Model Setup

[5] We use MITgcm [*Marshall et al.*, 1997] in hydrostatic mode with 24 vertical levels. The idealised model domain is 4000 km × 2500 km × 4000 m, with periodic boundary conditions in the zonal direction and vertical walls on the north and south. Horizontal grid spacing is 10 km and is spatially uniform. The topography (see Figure 1d) is chosen arbitrarily, with a pair of ridges which act to steer the flow while providing a source of bottom form stress, and a continental slope on the southern edge of the domain to facilitate sinking of deep water. (Sensitivity tests indicate that the results described below are insensitive to the form of topography.)

[6] The model forcing is shown in Figures 1a and 1b. Wind stress is specified to be a simple sinusoid, with non-zero stresses only in the central portion of the domain. Three wind stresses are defined, with peak values of 0.8, 0.12 and 0.16 N/m^2. We use a specified heat flux for computational convenience–the mean temperature in the channel does not evolve, so that equilibration time is substantially reduced. The model uses a linear equation of state, so that temperature is a proxy representing buoyancy variations due to both heat and salinity. Heat fluxes are applied in the northern and

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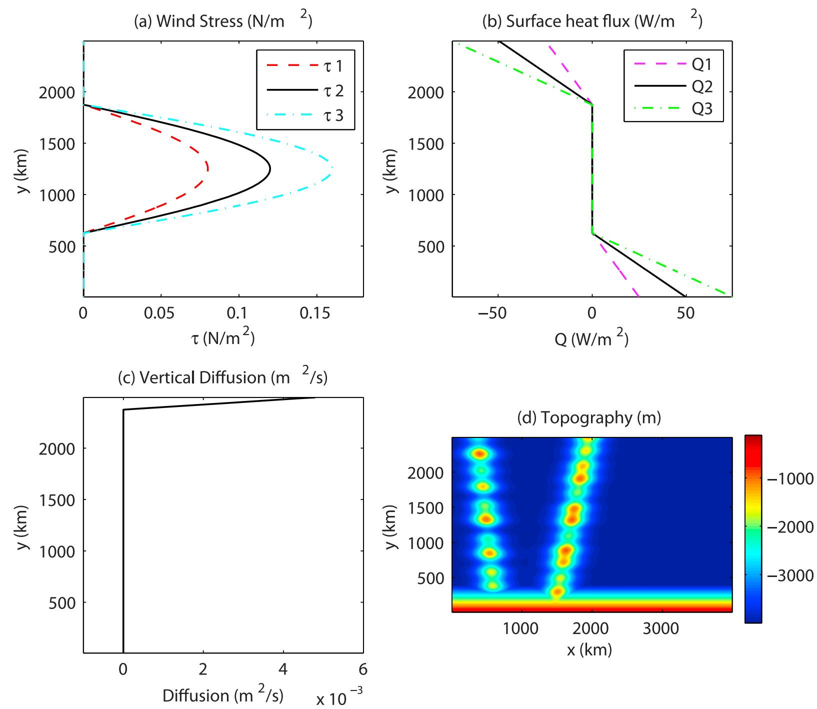

Figure 1. MITgcm forcing and topography. (a) Wind stress as a function of latitude for 3 different cases; (b) surface heat flux, for 3 different cases; (c) enhancement of vertical diffusivity (added to the background value of $K_\rho = 10^{-4}$ m^2/s); and (d) topography.

southern quarter of the domain (heating and cooling respectively) and three different cases are used with heat flux maxima of ±25, ±50 and ±75 W/m^2. The standard simulation uses buoyancy forcing of ±50 W/m^2 and peak wind stress 0.12 N/m^2. Importantly, we choose to decouple wind stress and buoyancy forcing by spatial separation, allowing us to independently vary each term; such a distinct separation is not possible in coupled models.

[7] A particular shortcoming of this model setup is that, with no heat flux through the northern boundary, a shallow thermocline is generated. To rectify this, we apply excess vertical diffusion near the northern edge of the domain (see Figure 1c). This additional diffusion acts to increase the thermocline depth (at a location 300 km north of the northernmost front), and hence amplify the ACC transport. Physically, it represents the overturning circulation, wind-driven circulation and vertical mixing acting throughout the global ocean; the caveat on these calculations is that changes in the stratification due to variations in surface buoyancy forcing are felt relatively quickly, compared with the likely millenial equilibration time in the real ocean [*Allison*, 2009]. The qualitative response to forcing perturbations in these experiments does not depend upon the value of additional diffusion.

3. Results

3.1. Mean Circulation

[8] Spinup simulations use a surface buoyancy forcing of ±50 W/m^2 for the first 50 years; after this time we impose wind stress forcing (0.12 N/m^2) and continue spinup simulations until an equilibrium state is reached (~ 250 years). In this state, circumpolar transport asymptotes to approximately 200 Sv, larger than accepted values for the ACC (~ 130 Sv [*Cunningham et al.*, 2003]) but consistent with the wide channel used in this case. The mean and instantaneous temperature and flow structure is shown in Figure 2. The impact of the diffusively deepened thermocline (at the northern edge of the domain) and dense water sinking down the continental slope (on the south, representing AABW formation) are shown in the temperature transect, while perturbations about the mean field indicate the presence of transient eddies. Multiple fronts are shown by sudden increases in thermocline depth. These fronts are also visible in the surface temperature and streamfunction field, while the eddy field is particularly enhanced in the lee of topography. The pattern of flow, as well as bulk diagnostics, are consistent with observations of the ACC, and indicate that this idealised model configuration is suitable for testing our understanding of the response of circumpolar currents to changes in forcing.

3.2. Sensitivity to Wind Stress Forcing

[9] We take the equilibrium model state at year 273 and run three different integrations for 80 years: the three cases span the 0.08, 0.12 and 0.16 N/m^2 peak wind stress cases illustrated in Figure 1a. Buoyancy forcing is unchanged (±50 W/m^2). The model diagnostics for the three experiments are shown in Figures 3a, 3c, and 3e), with the black line denoting the 0.12 N/m^2 case (i.e., no change in forcing). Kinetic energy (Figure 3a) is a strong function of wind stress, with a near-linear response to forcing changes and a relatively short response time. The response of Available Potential Energy (APE) (Figure 3c) is more complicated; the initial response (first 5 years) is proportional to wind stress, consistent with expected tilting of isopycnals by Ekman pumping. But this initial response decays away, presumably due to the enhanced eddy field relaxing isopycnals, and the

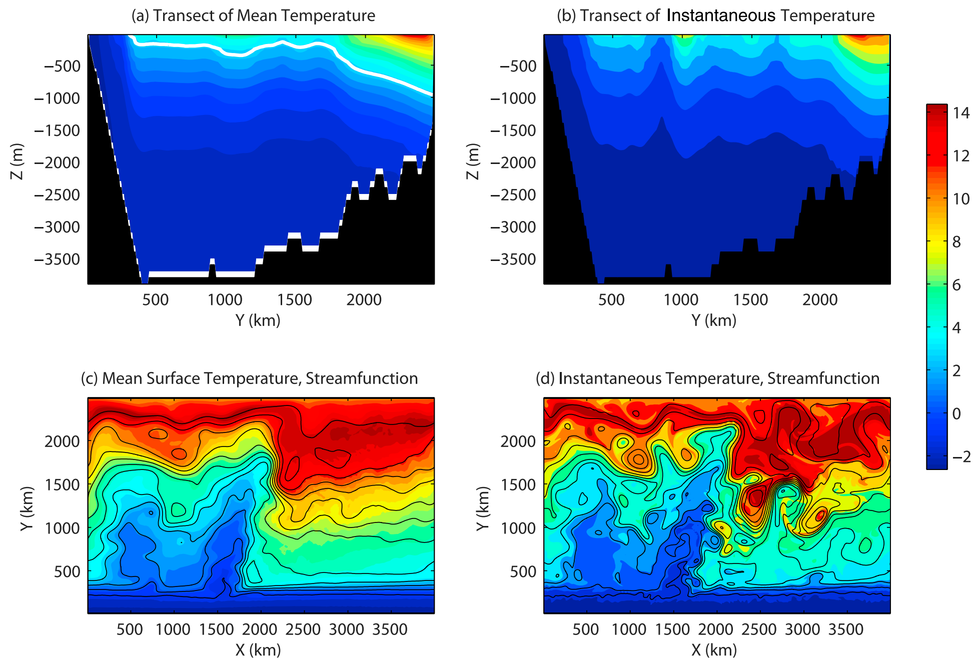

Figure 2. Equilibrium state of model with ±50 W/m^2 surface buoyancy forcing and 0.12 N/m^2 peak wind stress. Transects of (a) mean and (b) instantaneous temperature at X = 1800 km (the 3°C isotherm defining the thermocline depth is shown by the white contour). Surface temperature and streamfunction for (c) mean and (d) instantaneous flow.

final state shows no significant difference in APE between the three cases.

[10] There is a weak but systematic decrease in circumpolar transport with increased wind stress forcing (Figure 3e)–the opposite response to coarse-resolution models. This result is due to the eddy mixing of temperature (particularly in the upper thermocline) which acts to reduce the thermocline depth in the north, and thereby counteracts the buoyancy forcing of the mean flow. The zonal momentum input due to

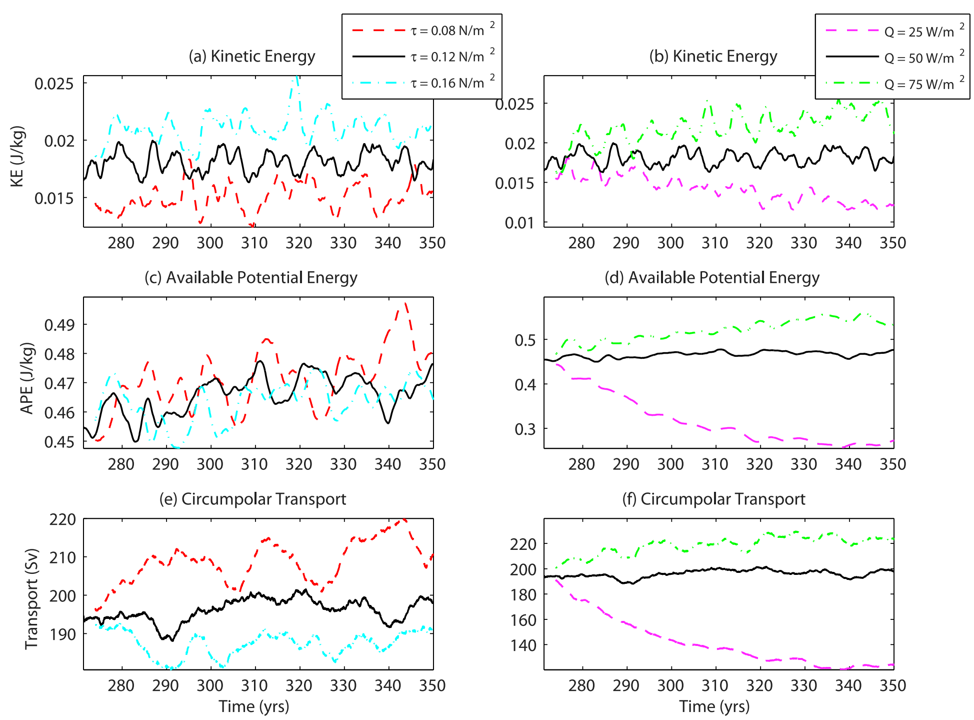

Figure 3. Response of (a) kinetic energy; (c) available potential energy; and (e) circumpolar transport to three different wind stress cases, and response of (b) kinetic energy; (d) available potential energy; and (f) circumpolar transport to three different surface buoyancy forcing cases.

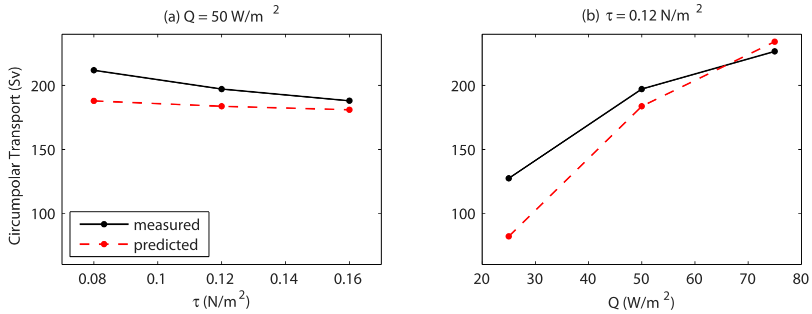

Figure 4. Comparison of circumpolar transport and that estimated by equation (1) and the observed temperature field as a function of (a) wind stress and (b) surface heat flux.

wind stress is balanced by eddy-enhanced bottom form stress, consistent with the eddy saturated parameter regime.

3.3. Sensitivity to Surface Buoyancy Forcing

[11] The above experiments are now repeated with the three buoyancy forcing cases shown in Figure 1b, with peak wind stress fixed at 0.12 N/m^2. The results (see Figures 3b, 3d, and 3f) indicate that, in this model, buoyancy forcing strongly controls the flow, with proportional responses in each of KE, APE and circumpolar transport. The response timescale of the system to buoyancy changes is > 50 years-substantially slower than the response to wind stress. In addition, an asymmetry in the response of APE and circumpolar transport to buoyancy forcing suggests an asymptote towards saturation with strong buoyancy forcing.

4. Discussion

[12] These idealised simulations of a circumpolar current provide a simple test of our understanding of the dynamics of the ACC. We have demonstrated that a diabatic, z-level model retains the intrinsic behaviour of the eddy-saturated parameter regime, where circumpolar transport is essentially independent of wind stress. In fact, there is a weak inverse relationship between wind forcing and transport, due to interaction between lateral eddy mixing and the buoyancy forcing. It is likely that eddy-resolving global circulation models will possess elements of the eddy-saturated state, in sharp contrast to coarse-resolution models.

[13] The second primary result from the present experiments is that, contrary to traditional conceptual models of ACC transport, buoyancy forcing exerted a much stronger control on circumpolar transport than wind stress. In fact, additional simulations (not shown) in which wind stress forcing is switched off entirely showed no significant decrease in transport; thus the usual assumption that eddy saturation only applies above a certain critical value of wind stress does not apply here.

[14] These results can be most easily understood in terms of a simple thermocline model, where the thermocline depth, D, is estimated from the 3°C isotherm (the white contour in Figure 2a). Then, the thermal wind relation can be used to predict circumpolar transport from the density field alone,

$$T_{\mathrm{ACC}} = \frac{g\alpha \Delta T D^2}{f}, \qquad (1)$$

where g is the gravitational acceleration, α the thermal expansion coefficient, f the Coriolis parameter and ΔT the vertical temperature difference. Figure 4 shows a comparison of measured circumpolar transport with that predicted from equation (1) and the observed temperature field. The accuracy of this simple relationship illuminates how enhanced surface buoyancy forcing directly affects transport, both by deepening the thermocline and warming surface waters. Wind forcing also acts to tilt the thermocline, but increased tilt from stronger wind stress is almost exactly offset by enhanced eddy fluxes.

[15] The are several caveats to place on this result: the enhanced diffusion near the northern boundary of this model is a poor proxy for the action of the global overturning circulation, the time-dependence of wind and buoyancy forcing are not considered and the complete separation (both spatial and temporal) of the buoyancy and wind forcing is unrealistic. (Additional simulations, not shown, indicate that wind stress forcing added over regions of buoyancy forcing acts to enhance transport, because meridional Ekman transport near the northern boundary generates subduction of warm water along that boundary and subsequent deepening of the thermocline.) However, the model does indicate that an eddying circumpolar current with some topography can reproduce the physics of the eddy-saturated limit previously observed in models with more idealised dynamics. This suggests that the impact of fully-resolved eddies upon the oceanic circulation in climate models is an urgent high priority.

[16] A possible interpretation of the present results would be to hypothesise that the ACC may be driven entirely by buoyancy. We do not make this claim; largely because the combination of buoyancy forcing and excess diffusion on the northern boundary of this model is a weak proxy for the (wind, buoyancy and diffusively forced) global ocean circulation. In addition, the interaction between wind and buoyancy forcing is difficult to disentangle in the real ocean, or in climate models, making evaluation of the buoyancy and wind-forced contributions difficult in practise. However, we suggest that hypotheses regarding the modulation of ACC transport by density gradients due to AABW formation [*Gent et al.*, 2001] or North Atlantic Deep Water formation [*Fučkar and Vallis*, 2007] need to be evaluated in eddy-resolving global climate models.

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