Does the sensitivity of Southern Ocean circulation depend upon bathymetric details?

Andrew McC. Hogg and David R. Munday

1 Research School of Earth Sciences and ARC Centre of Excellence for Climate System Science, The Australian National University, Canberra, Australian Capital Territory 0200, Australia
2 Atmospheric, Oceanic and Planetary Physics, Department of Physics, University of Oxford, Parks Road, Oxford OX1 3PU, UK

The response of the major ocean currents to changes in wind stress forcing is investigated with a series of idealized, but eddy-permitting, model simulations. Previously, ostensibly similar models have shown considerable variation in the oceanic response to changing wind stress forcing. Here, it is shown that a major reason for these differences in model sensitivity is subtle modification of the idealized bathymetry. The key bathymetric parameter is the extent to which the strong eddy field generated in the circumpolar current can interact with the bottom water formation process. The addition of an embayment, which insulates bottom water formation from meridional eddy fluxes, acts to stabilize the deep ocean density and enhances the sensitivity of the circumpolar current. The degree of interaction between Southern Ocean eddies and Antarctic shelf processes may thereby control the sensitivity of the Southern Ocean to change.

1. Introduction

The sensitivity of the Southern Ocean circulation to changes in forcing is one of the major uncertainties, and potential risk factors, of climate change. Most predictions of future climate indicate that we can expect increases in the strength of the westerly winds that influence flow in the Southern Ocean [1,2], as well as changes in surface buoyancy forcing through warming, the hydrological cycle and Antarctic runoff [3]. The effects of such changes on the ocean circulation may include enhancement of the Southern Ocean eddy field [4], a muted increase in
the Antarctic Circumpolar Current (ACC) transport [5] and changes in the rate of Southern Ocean upwelling which may accelerate outgassing of ancient carbon dioxide [6]. The response of the system depends upon the details of eddy dynamics in the Southern Ocean; parametrized eddies may not faithfully reproduce all aspects of the circulation response. In particular, proposed dynamical regimes incorporating the concepts of eddy saturation (i.e. loosely defined as the insensitivity of ACC transport, or global pycnocline depth, to wind stress) and eddy compensation (i.e. insensitivity of the overturning circulation to wind stress) imply the potential for dynamical eddy responses beyond the realm of most standard eddy parametrizations [7], although more sophisticated variable coefficient schemes do improve model fidelity of the overturning circulation response [8,9].

The need to understand eddying responses of the Southern Ocean to changing forcing must be balanced against the computational cost of running global ocean (or coupled ocean–atmosphere) simulations at sufficient resolution to include eddies. While some global eddy-permitting simulations have been conducted [10] and have led to an improved mean climate [11], very little in the way of parameter exploration can be achieved, and there remain doubts about whether or not such models can develop an equilibrium stratification over the necessarily short simulations [12]. The difficulty in offsetting computational cost with the necessity for ocean eddies has spawned a cottage industry of idealized eddy-permitting ‘sector’ models [7,13–16]. These models are usually configured within a reduced domain spanning 40° or less in longitude, while including polar oceans, simple geometry and idealized surface buoyancy and wind stress forcing. Thus, they include the majority of processes that are incorporated into conceptual models of the equilibrium circulation, while being sufficiently simple and efficient that equilibrium solutions can be obtained and a wide range in parameter space explored.

It is reasonable to expect that, with the ocean dynamics stripped to its bare essentials, these sector models would be similarly sensitive to surface forcing. However, this is not the case. For example, Jones et al. [13] pointed out that ocean stratification does adjust to a step change in the strength of Southern Ocean winds; eddy-permitting models indicate a deepening pycnocline after a century of adjustment and an associated acceleration of the circumpolar transport. Thus, on long timescales, the physics governing eddy saturation may break down, and the equilibrium circumpolar transport is sensitive to wind forcing ([15], hereafter SH12). On the other hand, Munday et al. ([16], hereafter MJM13), using the same numerical model in an almost identical configuration finds near complete insensitivity to wind stress forcing—an equilibrium eddy saturation. A direct comparison of the experiments of SH12 and MJM13 can be seen in figure 1a; note that, while the variations in wind stress shown here are within the bounds of that observed in the SOFS1 mooring [17], they far exceed the range of expected mean wind stress changes predicted by coupled climate model studies [1,2].

To compare different models, we define the fractional sensitivity of the circulation to wind stress, given by

\[
S_i = \frac{(\psi' - \psi_0)/\psi_0}{(\tau' - \tau_0)/\tau_0},
\]

where the \(\psi_0\) represents the circulation in the reference state about which the metric is linearized, driven by winds \(\tau_0\) and primed quantities \((\psi', \tau')\) represent the equilibrium state and wind stress forcing of the perturbed model. By this measure, the sensitivity of the ACC transport to wind stress forcing, \(S_{ACC}\), is essentially 0 in MJM13’s highest resolution, while SH12 yield approximately 0.25. Table 1 includes results from other experiments, showing the range of values that have emerged from different experiments.

Similar disparities can be seen when examining the overturning circulation. We follow MJM13’s definitions of overturning cells using the zonally averaged residual circulation, with the upper (or NADW) cell defined to be the maximum streamfunction in the Northern Hemisphere, and the lower (or AABW) cell to be the minimum streamfunction between 30° S and the equator. Figure 1b shows the overturning circulation comparison between MJM13 and SH12. Here, we find greater sensitivity in the upper cell circulation for SH12, and, most surprisingly, opposing
sensitivity in the lower cell. Most models in table 1 show partial eddy compensation (i.e. $S_{\text{NADW}} \sim 0.5$) of the upper cell, but there is little consensus on the sensitivity of the lower cell.

We reiterate that these idealized models contain the essential elements upon which conceptual models of the zonally averaged circulation are founded. It is in no way surprising that the magnitude of major current systems differs between models; but the factors controlling the sensitivity to wind stress forcing are not immediately apparent. The point of this paper is to outline a range of hypotheses that might explain these differences in model sensitivity (§2).

We then proceed in §3 to isolate variations in the model geometry which account for a large proportion of these differences, and in §4 explore the ramifications of these findings for idealized models and the Earth system.

2. Hypotheses

There are numerous minor differences in the configuration of the model between MJM13 and SH12. In the following, we list specific differences which may plausibly account for variations in model sensitivity and evaluate their impact:

![Figure 1.](image)

**Figure 1.** (a) Circumpolar transport and (b) overturning metrics for the upper cell (solid) and lower cell (dashed) as reported by MJM13 for three resolutions spanning $2^\circ - 1^\circ$ and SH12 ($1^\circ$ only). Note that the overturning metrics for SH12 have been reduced by a factor of 2 to account for differences in domain width. (Online version in colour.)

**Table 1.** Comparison of the sensitivity of major current systems from a series of idealized models.

<table>
<thead>
<tr>
<th>experiment</th>
<th>$S_{\text{ACC}}$</th>
<th>$S_{\text{NADW}}$</th>
<th>$S_{\text{AABW}}$</th>
<th>notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>MJM13 $1^\circ$</td>
<td>$-0.02$</td>
<td>$0.18$</td>
<td>$0.31$</td>
<td></td>
</tr>
<tr>
<td>MJM13 $1/6^\circ$</td>
<td>$0.13$</td>
<td>$0.32$</td>
<td>$0.19$</td>
<td></td>
</tr>
<tr>
<td>MJM13 $1/2^\circ$</td>
<td>$0.72$</td>
<td>$0.42$</td>
<td>$0.21$</td>
<td></td>
</tr>
<tr>
<td>SH12 $1/4^\circ$</td>
<td>$0.36$</td>
<td>$0.40$</td>
<td>$-0.28$</td>
<td></td>
</tr>
<tr>
<td>Morrison &amp; Hogg [7]</td>
<td>$0.25$</td>
<td>$0.5$</td>
<td>$0.0$</td>
<td>$1/12$, Southern Hemisphere only</td>
</tr>
<tr>
<td>Viebahn &amp; Eden [18]</td>
<td>—</td>
<td>$0.4$</td>
<td>—</td>
<td>reduced domain</td>
</tr>
<tr>
<td>Abernathy et al. [19]</td>
<td>—</td>
<td>$0.4$</td>
<td>$0.0$</td>
<td>zonal channel, prescribed Flux BCs</td>
</tr>
<tr>
<td>Abernathy et al. [19]</td>
<td>—</td>
<td>$1.2$</td>
<td>$-0.5$</td>
<td>zonal channel, restoring BCs</td>
</tr>
</tbody>
</table>
(a) Resolution

SH12 uses $\frac{1}{4}^{\circ}$ resolution on a Mercator grid, while MJM13 extends up to $\frac{1}{6}^{\circ}$ resolution. It is well established that improved resolution of eddies moderates the sensitivity of modelled circulation to wind stress [20]; hence, we first consider whether resolution dependence may explain some inter-model differences. However, SH12 gives a higher circumpolar transport and overturning sensitivity than even the $\frac{1}{2}^{\circ}$ model of MJM13, and resolution seems unable to explain the oppositely signed lower cell overturning sensitivity. Thus, we discard the hypothesis that increasing model resolution from $\frac{1}{4}^{\circ}$ to $\frac{1}{6}^{\circ}$ can account for the differing sensitivity between SH12 and MJM13. In what follows, all simulations are performed at $\frac{1}{4}^{\circ}$ resolution.

(b) Surface boundary conditions

SH12 employ fixed surface buoyancy fluxes, while MJM13 restore surface values of both temperature and salinity to prescribed meridional profiles. Restoring conditions act to stabilize the flow and to damp the eddy field; this factor has previously been shown to account for major differences in modelled overturning sensitivity [19] (table 1). However, we have run tests with both types of boundary condition, and find that the circumpolar transport sensitivity to wind is entirely independent of the type of surface boundary condition, or the restoring timescale used (over a wide range from 5 to 60 days; results not shown here). This result implies that the feedback between wind-driven changes in overturning and buoyancy forcing under restoring conditions—which is greatest near the core of the circumpolar current—has a minimal effect on the current, while buoyancy forcing in regions of bottom water formation which are important for circumpolar transport [15] is insensitive to wind. Thus, we proceed to other hypotheses.

(c) Nonlinearity of the equation of state

We have also tested the sensitivity of the flow state to the equation of state used. These tests indicate that changing from a linear to a nonlinear equation of state does indeed alter the value of the ACC transport (data not shown). This change in ACC transport occurs because the equation of state directly controls the density field for given values of temperature and salinity, thereby altering the current strength via thermal wind. Given a sufficiently strong buoyancy contrast, there is no reason to think that this effect could control the sensitivity of a model’s circumpolar transport to wind stress, and thus we do not pursue it further. However, note that we have not been able to positively exclude the impact of the nonlinear equation of state, and there may be subtle impacts of these nonlinearities on, for example, deep water formation rates owing to nonlinear processes.

(d) Topography

Finally, we address the question of the fidelity of the topography. SH12 use sloping sidewalls, as well as a region to the south of the ACC where topography blocks the zonal flow—akin to the Weddell Sea being constrained by the Antarctic Peninsula in the ocean. Having eliminated most other viable hypotheses, we now proceed to investigate the topographic control of the flow.

Do the details of model bathymetry alter the sensitivity of the circumpolar transport to wind stress? While it seems a remote possibility, having eliminated other options, we elect to test the sensitivity of the Southern Ocean circulation to two domains with differing bathymetry, shown in figure 2. Our ‘standard’ domain runs from 60° S to 60° N with a vertical wall on the southern side of the reconnecting channel (representing Drake Passage). This is the same domain as that used by MJM13. The second domain is ‘extended’ to 70° N/S. It includes a continental shelf and slope in both polar regions, as well as a peninsula, forming a Weddell Sea-like basin in the southern part of the domain. We note that there are other topographic variations in idealized models and that we have been unable to exhaustively test all variations.
Figure 2. Schematic showing meridional transects of (a) the standard domain used by MJM13 and (b) the new extended domain tested here. The position of a reconnecting channel (akin to Drake Passage) is shown by the dotted box, indicating the presence of a $1^\circ$ (four grid-point) wide peninsula-like feature on the southern side of the channel. (Online version in colour.)

We use MITgcm [21,22] with similar parameters to that in MJM13. The model resolution is $\frac{1}{4}^\circ$ in the horizontal on a spherical grid, with 42 vertical levels. The surface forcing conditions include a prescribed wind stress field with restoring on salinity (30 days) and temperature (10 days); both the timescales and the profile are identical to that used by MJM13. (The extended domain uses the $60^\circ$ N/S value of the temperature/salinity profiles for both polar regions.) We use a lateral biharmonic viscosity with a grid Reynolds number set at 0.15 (see MJM13 for details), a vertical viscosity of $10^{-3}$ m$^2$ s$^{-1}$ and a small subgridscale eddy parametrization coefficient of 2.5 m$^2$ s$^{-1}$. Each case is run for over 1000 years to reach equilibrium, and the peak value of wind stress forcing is varied over a wide range, from 0 to 0.6 N m$^{-2}$.

3. Results

(a) Stratification and overturning

We first examine the difference in mean state and circulation between the two simulations. Figure 3 shows the zonally averaged equilibrium position of four selected isopycnal layers for the reference wind forcing (which peaks at 0.2 N m$^{-2}$). The surface stratification shows no significant dependence upon the domain shape, due to the surface restoring conditions, but the thickness and average depth of isopycnal laters is altered beneath the surface. The extended domain shows substantial thickening of the 1036.0–1036.5 density layer, implying thinner surface and abyssal layers. The result is a deeper pycnocline on the northern side of the ACC channel which generates a greater ACC transport in this case.
A primary driver of the thickening mid-depth layers is the more efficient formation of deep water in the northern part of the domain. The presence of a shelf in the northern regions assists to minimize direct and eddy heat flux to the deep convection region, resulting in a stronger, denser NADW cell. This can be seen in a comparison of the overturning circulation (on density layers) for the two cases (figure 4). Here, the extended domain has a 20–25% stronger upper overturning cell, which is both denser (by 0.3–0.55 kg m$^{-3}$) and intrudes more efficiently into the Southern Hemisphere, with 25% more upwelling to the surface in the Southern Ocean. The difference in
abyssal circulation north of Drake Passage is negligible for these two examples, but there are changes of the order of 15–20% within Drake Passage itself.

These results indicate that the inclusion of an isolated polar sea and a sloping continental shelf south of the circumpolar band has only a weak effect on the mean circulation. Adjustments to the stratification are controlled by the favourable conditions for formation of deep water, enhancing the upper overturning cell in this system, while the abyssal cell is only weakly affected. We now examine the sensitivity of this system to variations in wind stress forcing.

(b) Sensitivity

Figure 5 shows the difference in sensitivity to wind stress forcing for the two domains. The circumpolar transport (figure 5a) is considerably more sensitive to wind in the extended domain. At zero wind stress, both simulations give a similar circumpolar transport of approximately 120 Sv. As wind increases, the extended domain shows increased circumpolar transport up to a peak wind stress value of 0.25 N m$^{-2}$. At a sufficiently high wind stress, both cases asymptote towards a constant circumpolar transport, but there remains a difference of almost 50 Sv between the two domains. This result implies that the details of domain bathymetry is at least partially responsible for the difference in wind stress sensitivity between the SH12 and MJM13 studies.
Figure 5b shows the difference in the magnitude of the upper overturning cell (defined as the maximum value of the residual overturning streamfunction at 30° S, following [7]) for the two simulations. At low wind stress, only a small fraction of the upper cell crosses the equator, but increased divergence of the Ekman transport with higher Southern Ocean wind stress enhances upwelling of deep water at Drake Passage latitudes, and further extends the upper cell into the Southern Hemisphere (as well as a moderate increase in the peak overturning at approx. 55° N, not shown). The amplification of the upper overturning cell with southern wind stress is clear in both domains but is considerably greater (by 30–50%) in the extended domain.

The sensitivity of the lower cell (figure 5c; defined as the minimum of the residual overturning streamfunction at 30° S) shows a more nuanced behaviour. Both domains have a relatively weak abyssal cell (even when accounting for the narrow domain). At low wind stress, the two domains have similar abyssal transport, but the extended domain has weaker sensitivity to wind. At the highest wind stress tested here, the standard domain has an abyssal cell 30% higher than that of the extended domain. The sensitivity of the abyssal cell to wind stress was a key difference between MJM13 and SH12; these experiments showed oppositely sloping trends (figure 1b). Here, while the slope is not reversed, the response of the abyssal transport remains markedly different.

To delineate the role of the abyssal cell and its potential feedback to the zonal circumpolar transport, we plot the differences in stratification in response to tripled winds for both domains in figure 6. Increased wind produces a similar upper ocean response for each of the domains: density decreases (i.e. isopycnals are depressed) throughout the upper 1000 m consistent with a deepening of the global pycnocline. However (and despite the minimal abyssal ocean differences
in the reference state; figures 3 and 4), below 3000 m the standard domain undergoes general warming (decrease in density), whereas the extended domain has roughly invariant stratification at this depth. Both domains show a local maxima in density difference at approximately 2000 – 3000 m depth (owing to enhanced Southern Ocean wind-induced upwelling acting to thin isopycnal layers at these depths), but this effect is partially offset by the decrease in abyssal density in the standard case. Therefore, only the extended domain has a positive density difference (blue colours in figure 6b) at these depths. Thus, while enhanced winds strengthen the upper overturning cell in both domains causing a deepening of the global pycnocline and thinning of mid-depth layers, the abyssal density is insulated from wind stress changes in the extended domain.

4. Discussion and conclusion

Predicting the future (and past) behaviour of the Southern Ocean is complicated by the strong role of eddies in governing the response of the circulation. Recent modelling experiments have shown how both the equilibrium circumpolar transport and overturning circulation may be modulated by the eddy response. But explicit simulation of eddies in global climate models is computationally intensive, and in many cases it is not possible to reach an equilibrium state which complicates the delineation of cause and effect. Much progress in recent years has been made by appealing to models with idealized and/or restricted domains.

The results outlined here constitute an attempt to reconcile the differing results reported from these idealized models. In short, we find that a substantial amount of the model sensitivity difference may be due to seemingly minor differences in the idealization of bathymetry. For example, inclusion of a continental shelf and slope in both the northern and southern parts of the domain enhances the upper overturning cell in the reference case, while also altering the sensitivity of the circulation to Southern Ocean winds.

The sensitivity of circumpolar transport to wind in the present simulations is largely controlled by the response of the abyssal cell to changing winds. Previous studies postulate a connection between the abyssal water formation and circumpolar transport [15,23–25]. This connection can be explained on dynamical grounds either as a consequence of the threshold for baroclinic instability being governed by vertical stratification [26] or alternatively by a simple restatement of thermal wind in governing the zonal transport [25]. In the extended domain shown here, there is almost no sensitivity of the abyssal cell strength or the abyssal density to wind stress forcing; however, the standard domain produces greater abyssal transport, with reduced bottom density.

We propose that these differences can be explained by the sensitivity of eddy fluxes of density across the circumpolar channel. In the standard domain, eddy fluxes in the circumpolar channel directly interact with the sinking plume of abyssal water. Increased wind strengthens the eddy kinetic energy [4] and corresponding eddy fluxes, thereby decreasing the density of newly formed abyssal water, while also increasing the rate of convection (leading to greater abyssal volume flux). This reduction in density balances the wind-induced increase in circumpolar transport owing to the depression of isopycnals on the northern side of the circumpolar channel (which MJM13 showed was a near exact balance at 1° resolution). Thus, the southward eddy fluxes contribute to an effective diabatic eddy saturation, albeit it differently from the adiabatic process which is usually envisaged [26].

In the extended domain, the dynamics is altered by the existence of a polar embayment to the south of the circumpolar channel, where abyssal water is formed. The embayment and continental slope create a circulation in which processes leading to bottom water formation are more distant from the baroclinically unstable eddy formation regions, and thus are not substantially influenced by eddy processes, especially at lower values of wind stress. For this reason, neither the abyssal density nor the volume transport into the abyss depends upon the strength of the wind or the eddy field. The deepest extent of wind stress forcing is felt at mid-depths, where enhanced upwelling acts to extract mid-depth water masses more efficiently.
The circulation and stratification in the extended domain is also altered by inclusion of a continental slope at the northern end of the basin. Eddy and direct fluxes of heat to this region are again reduced by topography, leading to a deeper and stronger NADW circulation cell. The northern processes, however, are not sensitive to changes in Southern Hemisphere winds.

It is not the purpose of this paper to determine which of these two domains is the closer analogue to the dynamics of the global ocean. The key hypothesis which arises from these results is that the equilibrium sensitivity of circumpolar transport to wind depends intrinsically upon the interaction between circumpolar eddy fluxes and abyssal water formation. While the extended domain appears to have a more realistic coastline, interaction between AABW overflow currents and circumpolar flow may occur in some regions due to local flow and topography variations. Thus, this paper should not be interpreted as predictive of future ACC behaviour, but instead to serve to illuminate a key aspect ACC sensitivity. In addition, these results also serve as a warning against the over-interpretation of idealized model results (or in fact results from any ocean/climate model) in application to the Earth system.

It is disconcerting that minor changes in the abyssal water formation regions in this type of model, and in fact in global climate models as well, may play a disproportionate role on governing the ocean response to changes in wind stress forcing. Recent indications that easterly wind stress near the Antarctic coastline may modulate dense water formation [27] support such a sensitivity. They point to a major challenge in Earth system modelling: that poorly constrained far southern processes (including sea ice and shelf processes) and bottom water transport place important controls on the sensitivity of ocean circulation to changes in surface forcing.

Acknowledgement. Our thanks to David Marshall for his comments on a draft of this manuscript, and to two anonymous reviewers helped to substantially improve the paper. This work was undertaken while A.M.H. was on sabbatical leave, hosted by Atmospheric, Oceanic and Planetary Physics, Department of Physics, University of Oxford.

Funding statement. A.M.H. was supported by Australian Research Council Future Fellowship FT120100842. D.R.M. was supported by the UK Natural Environment Research Council. This work made use of the facilities of HECToR, the UK’s national high-performance computing service, which is provided by UoE HPCx Ltd at the University of Edinburgh, Cray Inc and NAG Ltd, and funded by the Office of Science and Technology through EPSRC’s High End Computing Programme. Supporting simulations were undertaken on the NCI National Facility in Canberra, Australia, which is supported by the Australian Commonwealth Government.

References


