Influence of Wind Stress, Wind Stress Curl, and Bottom Friction on the Transport of a Model Antarctic Circumpolar Current

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ABSTRACT

Eddy-permitting simulations of a wind-driven quasigeostrophic model in an idealized Southern Ocean setting are used to attempt to describe what sets the wind-driven circumpolar transport of the Antarctic Circumpolar Current (ACC). For weak forcing, the transport is well described as a linear sum of channel and basin components. The authors’ main focus is on stronger forcing. In this regime, an eddy-driven recirculation appears in the abyssal layer, and all time-mean circumpolar streamlines are found to stem from a Sverdrup-like interior. The Sverdrup flux into Drake Passage latitudes can then be thought of as the sum of one part that feeds the circumpolar current and another that is associated with the recirculation. The relative fractions of this partitioning depend on the bottom drag, the midchannel wind stress, and the wind stress curl. Increasing the strength of the bottom drag reduces the recirculation and increases circumpolar transport. Increasing a zero-curl eastward wind stress reduces the upper-layer expression of the recirculation and increases the transport. Increasing the curl-containing portion of the forcing (while holding the midchannel stress constant) increases the recirculation and decreases the transport.

The weakly forced regime is also considered, as are the relative roles of large and small-scale eddies in transporting momentum vertically through the water column in the Drake Passage latitude band. It is found that the vertical momentum flux associated with transient structures can be used to distinguish between different regimes: these structures transmit momentum upward when the dynamics is dominated by the large-scale recirculation gyre and downward when it is not.

1. Introduction

What sets the eastward transport of the Antarctic Circumpolar Current (ACC) remains unclear. There has long been a consensus that topographic form drag is the main mechanism by which eastward momentum input by the winds is removed. Similarly, interfacial form drag is widely believed to be the dominant mechanism by which momentum is transferred vertically from the upper to the abyssal ocean. Reynolds stresses are thought to play a less dominant role (e.g., Rintoul et al. 2001). There also appears to be a consensus that the transport is linked to the strength of the stratification (e.g., Gnanadesikan and Hallberg 2000; Hallberg and Gnanadesikan 2001) and that mesoscale eddies are crucial (e.g., Doos and Webb 1994; Henning and Vallis 2005; Hallberg and Gnanadesikan 2006), except perhaps in a regime where forcing is weak and diapycnal fluxes are strong (Hallberg and Gnanadesikan 2001). It is accepted that, averaged over the path of the current, mesoscale transients transfer alongstream momentum downward. Whether these transients are directly responsible for transferring eastward momentum downward over the Drake Passage latitude band remains, however, less clear (e.g., Treguier and McWilliams 1990).

Also unclear, and of particular relevance to us, is the relative role of the wind stress curl compared to that of the wind stress itself. Hallberg and Gnanadesikan (2006) used a 1/6°-resolution primitive equation model in a realistic Southern Ocean geometry and showed an increase or decrease of 20% in the wind stress to result in changes of less than 3% to the zonal transport. This suggests that
the ACC transport is in a saturation regime. Saturation regimes are also found in more idealized quasigeostrophic models, which do not allow for feedbacks between the wind forcing and the stratification (e.g., Tansley and Marshall 2001; Hogg and Blundell 2006). Hogg and Blundell (2006) considered a zonally reconnecting channel with topography and attributed this behavior to eddy saturation. Eddy saturation theories (e.g., Straub 1993; Marshall et al. 1993) suggest a regime where the interface slope (thermocline) is approximately given by the condition of baroclinic instability. This, in turn, determines the baroclinic portion of the transport. When a basin region is present to the north of the reconnecting channel, a saturation regime is also observed (e.g., Tansley and Marshall 2001); however, in this case the link with eddy saturation theories is less clear. In a recent set of experiments, Nadeau and Straub (2009, hereafter NS09) argued that saturation in this setting might instead be explained by the Sverdrup flow observed in the basin-like portion of the domain. In their experiments, the vertical structure of the southward Sverdrup flux into Drake Passage latitudes was a key factor in explaining the transport behavior.

Sverdrup ideas have been used to explain transport in barotropic channel models for which topography blocks the geostrophic contours (e.g., Wang and Huang 1995; Krupitsky and Cane 1994). In a baroclinic ocean, one anticipates a band of zonally reconnecting potential vorticity contours in the upper ocean. That is, one anticipates a “channel like” region in the Southern Ocean in which Sverdrup dynamics break down. This has led to Sverdrup dynamics often being dismissed as altogether irrelevant to the ACC. Nonetheless, the possibility remains that basin-like dynamics in regions adjacent to Drake Passage latitudes may play a significant role in determining the transport. The idea dates back to Stommel (1957), Wyrtki (1960), and Gill (1968). They suggest scenarios whereby a southward Sverdrup flux turns eastward, joining—or in fact forming—the ACC. In Stommel’s and Wyrtki’s picture, there is no band of unblocked latitude circles, whereas, in Gill’s picture, a “channel region” (where Sverdrup dynamics do not apply) lies adjacent to a basin-like region (where Sverdrup dynamics do apply). We will refer to this general framework as Stommel’s idea or model.

Evidence supporting this picture comes from the clear east–west asymmetry of the ACC. For example, model simulations show the majority of the current to make a sharp northward turn just east of Drake Passage (e.g., Killworth 1992; Hallberg and Gnanadesikan 2006; Mazloff et al. 2010). This is consistent with Stommel’s picture, in which the ACC feeds into the Falkland–Malvinas Current. A similar northward turn feeding a western boundary region is also seen along the continental margin southeast of New Zealand. Additional evidence comes from numerical experiments. Allison et al. (2010) used a general circulation model to examine the effect of the Pacific and Indian basins on the ACC. When these basins were replaced with land north of Cape Horn, a considerable reduction in circumpolar transport was observed. They suggest that the transport is set by an integral of the wind stress over the path of the circumpolar current. Both this integral and the transport were smaller with the basins north of Cape Horn removed. One can also interpret this result in the context NS09; that is, filling in the basins reduces the basin contribution to the total transport.

The analytic model of NS09 explains well the saturation behavior seen in their simple two-layer quasigeostrophic model; however, their simulations did not test for the possible effect of an adding a zero-curl wind stress to the wind forcing profile. One anticipates that addition of a eastward zero-curl wind stress will lead to an increase in transport, as predicted by channel theories and seen in realistic geometry numerical simulations (e.g., Ivenhiko et al. 1999; Gent et al. 2001). Basin dynamics are mainly determined by the vorticity balance set by the wind stress curl, whereas channel dynamics also depend strongly on the wind stress itself. Thus, one might anticipate that increasing the eastward wind stress in channel (i.e., Drake Passage) latitudes will increase a “channel contribution” to the total transport. In this paper, we investigate the effect of the wind stress, the wind stress curl, and the strength of the bottom drag on circumpolar transport in a simple quasigeostrophic model with an idealized Southern Ocean geometry (a zonally reconnecting channel with walls to represent Patagonia and the Antarctic Peninsula).

A review of the analytic model of NS09 is presented below, and section 2 then briefly describes the numerical model used for this study. Section 3 presents results. First, a study of the influence of the bottom drag is considered. We find that bottom drag influences the strength of a large-scale abyssal recirculation in the southwest portion of the domain (east of Drake Passage), and that this influences the transport. The relative roles of wind stress and wind stress curl are then considered. For a fixed curl, adding an eastward wind stress increases the transport. Conversely, for a fixed wind stress magnitude in the middle of the Drake Passage latitude band, addition of a curl (in the form of a double-gyre forcing) decreases the transport. The relative roles of large-scale standing eddies, a topographically trapped eddy, and mesoscale transients in transferring eastward momentum downward in the channel latitude band are also considered.
Analytic model

It is useful to first give a brief review (following NS09) of why saturation occurs in the strongly forced regime. In their two-layer quasigeostrophic model, circumpolar free flow was blocked in the lower layer by a Scotia Ridge–like topography. Simulations comparing cases where Drake Passage was open with cases where it was not showed nearly identical circulation patterns over much of the region north of the model Drake Passage latitude band. Moreover, adjacent to the channel latitudes and away from the western boundary region, the time-averaged barotropic flow closely obeyed the Sverdrup relation. The Sverdrup flux into the Drake Passage latitude band was entirely top trapped (confined to the upper layer) for weak forcing. For stronger forcing, it was distributed over both layers in the western portion of the domain and remained top trapped in the east. NS09 assumed that only the top-trapped portion of the Sverdrup flux fed a basin contribution to the circumpolar transport (Fig. 1). The zonal position $x_0$ separating Sverdrup streamlines that feed the ACC from those that instead are associated with a recirculation was modeled as the point at which the separatrix geostrophic contour (or characteristic) intersected the Drake Passage latitude band. A basin contribution to the ACC $T_{basin}$ was then the zonal integral of the Sverdrup flux east of $x_0$. This is unlike previous Sverdrup theories, in which all the southward Sverdrup flux is assumed to feed the ACC.

The total circumpolar transport was then modeled as the sum of this basin contribution with a channel contribution, unrelated to the gyres,

$$T = T_{channel} + T_{basin}$$  

Here, $T_{channel}$ was found to roughly coincide with the analytic prediction of Straub (1993); $T_{basin}$ increased linearly with the forcing strength (in agreement with Stommel’s idea) for weak forcing and then reached a saturation value $T_{sat}$ for stronger forcing. The value of $T_{sat}$ was given by

$$T_{sat} \sim \beta L^2_{\mu} H L.$$  

Here, $\beta$ is the meridional derivative of the Coriolis parameter $f$, $H$ is the fluid depth, $L_{\mu}$ is the Rossby radius, and $L$ is the meridional distance from the north side of Drake Passage to the zero wind stress curl line lying farther north. For simplicity, the wind stress is assumed to be a function of the meridional coordinate $y$ only.

Numerical simulations showed the analytic model predictions to be reasonable. Even in situations where eddies drove a weak abyssal flow east of $x_0$, the net basin contribution to the transport was nonetheless well represented by the simple theory. Nothing in this simple theory, however, explicitly depends on bottom drag or on a zero-curl component to the wind forcing. Intuitively, however, one expects addition of a uniform eastward wind stress to increase the transport, and indeed channel theories of the ACC generally link the transport to the strength of the eastward wind stress in the Drake Passage latitude band (e.g., Johnson and Bryden 1989; Marshall and Radko 2003). It also seems intuitive that the transport should vary with the strength of the bottom drag, although in exactly what way is less obvious. The main focus of this work is on how the transport depends
on these factors in the strongly forced (or saturation) regime.

2. Numerical model and experimental design

a. Model details

Numerical experiments are performed using a model based on the quasigeostrophic equations, truncated to only two layers (or levels) in the vertical (Pedlosky 1996),

\[
D_t \left[ \nabla^2 \psi_1 - F_1(\psi_1 - \psi_2) + \beta y - F_0 \psi_1 \right] = -A_h \nabla^2 \psi_1 + \frac{\mathbf{k} \cdot \nabla \times \tau}{\rho_1 H_1} \quad \text{and}\quad (3)
\]

\[
D_t \left[ \nabla^2 \psi_2 + F_2(\psi_1 - \psi_2) + \beta y + \frac{f_0}{H_2} h_b \right] = -A_h \nabla^2 \psi_2 - r \nabla^2 \psi_2, \quad (4)
\]

where \( D_t (\cdot) = (\partial / \partial t) (\cdot) + J[\psi, (\cdot)] \) is the total time derivative, \( F_0 = f_0^2 / g H_1, F_1 = f_0^2 / g' H_1, F_2 = f_0^2 / g' H_2, f_0 \) is the mean Coriolis parameter, \( g \) is the gravitational acceleration, \( g' = g(\Delta \rho / \rho) \) is the reduced gravity, \( H_1 \) and \( H_2 \) are layer thicknesses, and \( h_b \) is bottom topography. Dissipation takes the form of a linear Rayleigh (Stommel) drag in the lower layer (with drag coefficient \( r \)) and an additional biharmonic dissipation. As in NS09, slip conditions appropriate for a biharmonic dissipation operator were applied at lateral boundaries. The coefficient \( A_h \) is taken to be \( A_h = \beta dx^3 \), where \( dx \) is the grid spacing; this ensures that the (biharmonic version of the) Munk layer is resolved. Other model details are as in NS09. A third-order Adams–Bashforth scheme is used for time stepping, the Arakawa (1966) kinetic energy and enstrophy-conserving scheme is used for the Jacobian operator, and an multigrid method is used to do the elliptic inversions.

Note that only the curl of the wind stress appears in (3) and (4). Moreover, in a channel setting, the elliptic inversion in nonunique and mass and momentum conservation laws are needed to close the problem. We follow a standard procedure similar to that described by McWilliams et al. (1978). Because this determines the transport and may not be familiar to all readers, a few details are given below. The momentum budget is closed using the zonal momentum equation integrated over a latitude circle in the channel

\[
\int_0^{L_x} \left[ D_t [u_{0k}] - f_0 v_{1k} - \beta y v_{0k} \right] dx = -g \partial_y \eta_1 - \delta_{N,k} r u_{0k} - A_H \nabla^2 u_{0k} + \delta_1 \left[ \frac{\tau}{\rho_1 H_1} \right] dx, \quad (5)
\]

where \([u_0, v_0] = [-\partial_x \psi, \partial_y \psi]\) is the horizontal velocity and \( \eta_k = (f_0 / g_k) \psi_k \) is the interface height. Note that \( \int_0^{L_x} [\partial_x \eta_1] dx \) is zero. The wind stress itself now appears explicitly in Eq. (5). The order Rossby number correction in a more complete model wind and buoyancy forcing can alter the stratification and in turn the transport (e.g., Hallberg and Gnanadesikan 2001; Henning and Vallis 2005). This indirect effect of wind forcing on ACC transport is absent in our model. Moreover, in the quasigeostrophic framework, the domain is bounded by two streamlines so that the transport is independent of the longitude. On the other hand, the two-layer quasigeostrophic framework is a simple comprehensible system for which it is easier to understand the basic behavior. It also allows us to cheaply perform long integrations of a large number of eddy-permitting simulations.

The reference wind forcing is also similar to that used in NS09; that is,

\[
\tau_{\text{ref}} = i r_0 \sin^2(\pi y / L_y). \quad (7)
\]

We will be interested in trying to isolate the respective effects of the stress and the curl. Specifically, we wish to vary separately the strength of the curl and that of the stress in the center of the channel, \( y = y_C \). To this end, it is useful to define \( \tau_{\text{curl}} \) as above, but with a constant subtracted so that a zero stress on \( y = y_C \) results,

\[
\tau_{\text{curl}} = \tau_{\text{ref}} - \tau_{\text{ref}}|_{y=y_C} = i r_0 [\sin^2(\pi y / L_y) - \sin^2(\pi y / L_y)] . \quad (8)
\]

Our most general forcing function is then

\[
\tau = \tau_{\text{curl}} + i r_1 . \quad (9)
\]

Note that, when \( r_1 \) corresponds to the value of \( \tau_{\text{ref}} \) on \( y = y_C \), our reference forcing profile is recovered. This forcing profile will be used in section 3a. In
section 3b, we will take $\tau_1 \propto \tau_0$, with proportionality constants such that the stress along $y = y_C$ varies between zero and twice that of our reference profile (for a given forcing amplitude). In section 3c, we will consider sets of experiments in which $\tau_0$ is varied independently of $\tau_1$: that is, in which the curl is varied in a way that leaves the stress along $y = y_C$ unaffected.

Model parameters can be found in Table 1. Simulations are carried out in a large domain (approximately 4000 km $\times$ 10 000 km), with a horizontal resolution of 15 km and a Rossby deformation radius $L_\rho$ of 32 km. This corresponds to just over two grid points per Rossby radius. Typically, one expects baroclinic instability to occur at wavenumbers smaller than $k_D \sim 1/L_\rho$. Note that this corresponds to a wavelength larger than $L_\rho$ by a factor of 2. In a flat-bottom channel setting, bottom friction removes momentum input to the upper ocean by the wind. Increasing $r$ then decreases the zonal transport. When topography is added, the effect of the bottom friction is less clear. For example, Hogg and Blundell (2006) observed a 20% increase in transport when $r$ was increased by a factor 2.5 and suggested that this phenomenon is due to eddy saturation (cf. discussion in the introduction). One also expects an increase in bottom drag to weaken the lower-layer flow, so that the southward Sverdrup flux into channel latitudes becomes more top trapped. In the context of NS09, increased bottom drag would then imply an increase in the saturation value of transport. Below, results from a series of simulations forced using $\tau_{\text{ref}}$ are presented for a range of coefficients $r$.

1) TRANSPORT

Figures 3a,b show the zonal transport in the top and bottom layers, plotted as a function of the forcing amplitude and for different values of bottom friction. We consider experiments with $r = 0.5, 1, 2,$ and $10 \times 10^{-7}$ s$^{-1}$ and will refer to $r_0 = 10^{-7}$ s$^{-1}$ as our reference value.

For the reference simulation, transport has a similar behavior to that described in NS09. In the weak forcing regime, a minimum background channel contribution to the total transport is evident, and the basin contribution increases more or less linearly with the forcing amplitude. In the strong forcing regime, the upper-layer transport reaches a saturation and lower-layer transport becomes weakly negative as the forcing strength in increased.

The weak forcing regime is insensitive to the value of $r$. This is expected because, in this regime, the Sverdrup flux into the channel region is top trapped at all longitudes. As such, an increased bottom drag does not alter the

### Table 1. Model parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rossby deformation radius</td>
<td>$L_\rho$</td>
<td>32 km</td>
</tr>
<tr>
<td>Horizontal resolution</td>
<td>$\Delta x = \Delta y$</td>
<td>15 km</td>
</tr>
<tr>
<td>Typical time step</td>
<td>$\Delta t$</td>
<td>1 h</td>
</tr>
<tr>
<td>Typical spinup time</td>
<td>—</td>
<td>150 yr</td>
</tr>
<tr>
<td>Typical averaging time</td>
<td>—</td>
<td>50 yr</td>
</tr>
<tr>
<td>Length of channel</td>
<td>$L_x$</td>
<td>9600 km</td>
</tr>
<tr>
<td>Width of channel</td>
<td>$L_y$</td>
<td>3840 km</td>
</tr>
<tr>
<td>Width of Drake Passage</td>
<td>$L_{\text{gap}}$</td>
<td>240 and 720 km</td>
</tr>
<tr>
<td>Position of $\mathbf{V} \times \mathbf{r} = 0$</td>
<td>$L_y/2$</td>
<td>1920 km</td>
</tr>
<tr>
<td>Top layer</td>
<td>$H_1$</td>
<td>1200 m</td>
</tr>
<tr>
<td>Bottom layer</td>
<td>$H_2$</td>
<td>2800 m</td>
</tr>
<tr>
<td>Coriolis parameter</td>
<td>$f_0$</td>
<td>$-1.3 \times 10^{-4}$ s$^{-1}$</td>
</tr>
<tr>
<td>Beta parameter</td>
<td>$\beta$</td>
<td>$1.5 \times 10^{-11}$ m$^{-1}$ s$^{-3}$</td>
</tr>
<tr>
<td>Gravitational acceleration</td>
<td>$g$</td>
<td>9.8 m s$^{-2}$</td>
</tr>
<tr>
<td>Reduced gravity</td>
<td>$g'$</td>
<td>0.02 m s$^{-2}$</td>
</tr>
<tr>
<td>Reference density</td>
<td>$\rho_1$</td>
<td>1035 kg m$^{-3}$</td>
</tr>
<tr>
<td>Bottom friction coef</td>
<td>$r$</td>
<td>$0.5 \times 10^{-7}$ to $10 \times 10^{-7}$ s$^{-1}$</td>
</tr>
<tr>
<td>Biharmonic dissipation coef</td>
<td>$A_{\beta}$</td>
<td>$1.14 \times 10^{10}$ m$^{-3}$ s$^{-1}$</td>
</tr>
</tbody>
</table>

A difference is that topography in NS09 included a continental rise along the eastern side of the Patagonia peninsula, and this is absent here. Note that topography is sufficiently high so as to block geostrophic contours in the lower layer. Because of this, circumpolar flow is expected to be weak in the bottom layer. Note, however, that an eddy-driven mean flow is expected in the lower layer. This can include both a closed circulation and a lower-layer contribution to the transport. In our simulations, the lower-layer transport remains small compared to the upper-layer transport. It is largest (and typically westward) in situations where the wind stress curl is large and the bottom drag is relatively small.

3. Results

a. Bottom friction

In a flat-bottom channel setting, bottom friction removes momentum input to the upper ocean by the wind. Increasing $r$ then decreases the zonal transport. When topography is added, the effect of the bottom friction is less clear. For example, Hogg and Blundell (2006) observed a 20% increase in transport when $r$ was increased by a factor 2.5 and suggested that this phenomenon is due to eddy saturation (cf. discussion in the introduction). One also expects an increase in bottom drag to weaken the lower-layer flow, so that the southward Sverdrup flux into channel latitudes becomes more top trapped. In the context of NS09, increased bottom drag would then imply an increase in the saturation value of transport. Below, results from a series of simulations forced using $\tau_{\text{ref}}$ are presented for a range of coefficients $r$. 

![Image](http://journals.ametsoc.org/doi/pdf/10.1175/JPO-D-11-058.1)
vertical structure of the Sverdrup flux. The value of $t_0$ at which a transition between the weakly and strongly forced regimes occurs, however, does depend on $r$. Specifically, the transition occurs at larger $t_0$ for increased drag. Also dependent on $r$ is the nature of the strongly forced regime. For intermediate values, a saturation occurs, whereas, for our weakest drag, transport decreases with $t_0$ and, for our strongest drag, it increases with $t_0$ (at least over the range considered). This behavior appears to be related to an influence of bottom drag on a large-scale recirculating gyre, as discussed below.

Time-averaged streamfunctions and related fields are shown in Fig. 4 for weakly and strongly damped simulations. The top panels show the barotropic streamfunction. In the interior and north of Drake Passage, the two are very similar. The bottom panels show the difference between the Sverdrup streamfunction and observed barotropic streamfunction, labeled $\psi_{\text{diff}}$. Away from boundaries and in the basin part of the domain, the barotropic flow is in approximate Sverdrup balance. The vertical structure of the Sverdrup flux is, however, different in the two cases. The points $x_{0.5}$ and $x_{10}$ mark the observed division separating recirculating streamlines to the west from circumpolar streamlines to the east. In these simulations, this also coincides with the eastward extent of the abyssal recirculation along the northern edge of the Drake Passage latitude band, $y = y_{DP}$. In NS09, this position (indicated as $x_Q$ in the figure) was predicted by the geostrophic contours, whose structure does not depend on $r$. We will refer to $x_{0.5}$ and $x_{10}$ as the analytic and observed division separating recirculating from circumpolar streamlines, respectively. The characteristics

![Fig. 2. Snapshots of the relative vorticity for the upper layer at different horizontal resolutions: (a) 7.5, (b) 15, (c) 30, and (d) 60 km. The wind stress amplitude is 0.2 N m$^{-2}$. In each case, values are capped at $\pm 10^{-6}$ s$^{-1}$.](image)

![Fig. 3. Average zonal transport across Drake Passage at equilibrium (a) in the top layer and (b) in the bottom layer for different values of bottom friction coefficient: $r = 5 \times 10^{-8}$, $10^{-7}$ ($t_0$), $2 \times 10^{-7}$, and $10^{-6}$ s$^{-1}$. All experiments are forced using $t_{\text{ref}}$; the horizontal axis corresponds to different values of $t_0$.](image)
calculated using the time-averaged streamfunction are shown in the third row panels. Because the interior is in near-Sverdrup balance, the separatrix contour corresponds reasonably well with its analytic analog.

The observed position $x_0$ separating recirculating from circumpolar streamlines is, however, poorly predicted by the characteristics. It lies to the west of the predicted longitude when $r$ is strong and to the east when $r$ is weak. Figure 4 shows the strength of the recirculating gyre to vary inversely with $r$. Large $r$ then implies a weak abyssal flow and a top-trapped Sverdrup flux, which is consistent with the linear (Stommel) regime extending to larger $r_0$ and with an incomplete saturation in the strong forcing regime. Conversely, weak damping implies a strengthened abyssal gyre, which can extend beyond the separatrix contour. As such, less of the Sverdrup flux is top trapped. $x_0$ is displaced eastward (to $x_{0.5}$ in Fig. 4), and $T_{\text{basin}}$ is reduced. At higher values of $r_0$, this effect is enhanced, so that rather than a saturation $T_{\text{basin}}$ decreases with $r_0$ in the weakly damped, strongly forced regime.

It is also interesting to note from Fig. 4 that, even in these cases where a saturation in transport is not observed, the strong forcing regime is nonetheless characterized by circumpolar streamlines that all stem from the Sverdrup interior lying to the north. It will also be of interest later that, in addition to the large eddy-driven abyssal recirculation, a tighter recirculation also appears over the topography. This topographically trapped

FIG. 4. Time-mean barotropic streamfunction $\psi_B$, lower-layer streamfunction $\psi_2$, geostrophic contours $\Theta$, and the difference between the analytic and modeled barotropic streamfunction $\psi_{\text{diff}}$ for two simulations. Forcing uses $r_{\text{ref}}$ with $r_0 = 0.4$ N m$^{-2}$. (left) $r = 5 \times 10^{-8}$ and (right) $r = 10^{-6}$. The scale of $\psi_2$ is changed slightly compared to that for $\psi_B$ to help visualize the circulation. Zonal locations $x_{0.5}$ and $x_{10}$ mark the division between recirculating and circumpolar streamlines on $y = y_{\text{DP}}$. For comparison, the analytic prediction $x_{0}$ is also shown.
recirculation is evident in the bottom-left corner of the left panels and contributes to the interfacial form drag, as discussed below.

2) MOMENTUM TRANSFER

North of the channel latitudes, meridional barriers support a western boundary current that closes the Sverdrup circulation and establishes a net westward pressure force to close the zonal momentum budget. In the channel latitudes, one expects Sverdrup dynamics to fail and the zonal momentum balance to involve topographic and interfacial form drags. Because it appears that the eddy-driven abyssal recirculation helps determine the baroclinic structure of the Sverdrup flux—and, by extension, helps determine $T_{\text{basin}}$—it is interesting to see if we can extract a signature of this gyre in the channel momentum balance. Using angle brackets $\langle \rangle$ to represent zonal and temporal averages, the dominant momentum balance at statistical equilibrium in each layer can be written as

$$
\langle \psi_{1x}\psi_{1yy} \rangle - \frac{f_0^2}{gH_1}\langle \psi_{2y}\psi_{1x} \rangle - \frac{1}{\rho_0H_1}(\tau) = 0 \quad \text{and} \quad (10)
$$

$$
\langle \psi_{2x}\psi_{2yy} \rangle - \frac{f_0^2}{gH_2}\langle \psi_{1y}\psi_{2x} \rangle + \frac{f_0}{H_2}\langle \psi_{2x}h_b \rangle = 0, \quad (11)
$$

where the first two terms in each equation are the Reynolds stress and interfacial form stress, respectively, and the third term represents either wind forcing or topographic form drag.

In most of our simulations, interfacial form drag dominates over the Reynolds stresses. An exception is the weakly damped case, for which Reynolds stresses were larger. In this case, a basin mode also appears and associated with this is increased variability in the transport. Excepting in this low drag simulation, the overall balance is essentially what one would expect: wind input of eastward momentum is transferred downward by interfacial form drag and is removed by topographic form drag. The direct effect of bottom drag and biharmonic dissipation on the momentum balance was small in all experiments. Figure 5 decomposes the interfacial form stress according to zonal wavenumber $k$. To do this, we first express snapshots of the streamfunction in each layer as a Fourier series, $\psi_i = \sum_n A_n(k) \cos(k) + B_n(k) \sin(k)$, for several latitudes in the channel. The interfacial form stress is then calculated as $T(k) = 0.5k[A_1(k)B_1(k) - B_1(k)A_1(k)] - B_1(k)A_2(k)$, and a time and latitudinal average of this is shown in the figure. Plotted is $kT(k)$ as a function of $\ln(k)$; normalized in this way, the area under the curve is roughly proportional to the stress.

For the strong drag case, the interfacial stress peaks around wavenumber 30 and is very weak at large spatial scales ($k < 10$). For weaker drag, $T(k)$ shows a more complex structure. Near $k = 15$, a peak of upward momentum transfer is evident and seen to strengthen with decreasing $r$. Additionally, at low $k$, a strong downward momentum flux develops as $r$ is decreased. A calculation of $T(k)$ using time-averaged streamfunctions showed similar results for $k < 20$, but not very much transfer at higher wavenumbers (not shown). In other words, the low $k$ structure in $T(k)$ is associated with standing eddies or structures, whereas the high $k$ peaks are associated with transients. Note that the high $k$ peaks correspond to wavenumbers on the order of a few hundred kilometers. Although this is considerably larger than the Rossby radius, it corresponds roughly to the meanders evident in quasi-zonal jets seen in our simulations (e.g., Figs. 2, 13). Dynamically, these are similar to meanders in the Gulf Stream extension, which also are large compared to the Rossby radius. We associate these transient features with the high $k$ peaks discussed above. As pointed out by an anonymous reviewer, Rossby waves might also play a role.

To recap, for strong drag the bulk of the downward momentum flux is accomplished by mesoscale transients, whereas, for weak drag it is accomplished by a large-scale form drag associated with the time-mean flow. As $r$ is decreased, both the eddy-driven abyssal recirculation and the large-scale time-mean form drag strengthen. Simultaneously, a change of sign is observed.
in the interfacial stress associated with mesoscale transient eddies.

3) Variability

Bottom friction also has a strong influence on the variability of the transport. Figure 6 (bottom) shows time series of the transport for different values of bottom drag. As discussed above, zonal transport increases with $r$ in the saturation regime. Variability of the transport, on the other hand, decreases with $r$ in this same regime. In particular, a large vacillation of short time scale appears in our $r = 0.5r_0$ simulation. To better understand this, we compare the streamfunction averaged over periods where transport is low with that averaged over periods when it is high (specifically, one standard deviation above or below the mean). The averaged $\psi_B$ fields for these periods when transport is maximal and minimal is shown in Fig. 6 (top left). Negative values correspond to dashed contours. From the figure, it appears that there is a strong basin mode (zonal wavenumber $\sim 6$) superimposed on the Sverdrup circulation. This is evident in both the maximum and minimum $\psi_B$ fields. Thin dotted lines in the figure show that features in the $\psi_B$ (min) field appear approximately aligned with similar features (having the opposite phase) in the $\psi_B$ (max) field. This is more evident in the right panels, in which the Sverdrup gyres have been removed. It is clear from the figure that minima in the transport occur when the basin mode is in one phase and maxima occur when it is in the other. A movie of this oscillation shows what appears to be large-scale barotropic Rossby waves propagating westward and breaking at the western boundary (not shown). Both the strength of this oscillation and of the variability in transport decrease as the bottom drag coefficient is increased. The strength of the recirculating gyre also decreases as $r$ is increased; however, it is not clear whether the oscillation identified above plays a role in maintaining this gyre.

b. Effect of a zero-curl wind stress

In a channel setting, transport is widely thought to be controlled by the wind stress, rather than by the wind stress curl. We are therefore interested to see how a constant zonal wind stress affects transport in our ACC geometry and whether any changes in transport might...
also be associated with changes in the recirculating gyre identified in the previous section.

Referring to Eqs. (8) and (9), we consider a range of \( \tau_0 \) (as before) and choose several values for \( \tau_1 \), ranging from 0 to 2\( \tau_{\text{ref}} \). Recall that the stress on the center latitude of the channel is equal to \( \tau_1 \). Thus, with \( \tau_1 = 0 \), there is no stress along \( y = y_C \), irrespective of the forcing amplitude \( \tau_0 \). With \( \tau_1 = \tau_{\text{ref}}(y_C) \), the reference profile used in the previous section is recovered. With \( \tau_1 = 2\tau_{\text{ref}} \), the stress along \( y = y_C \) is twice that of our reference profile, for a given forcing amplitude. To simplify notation, we define \( \tau C = \tau_{\text{ref}}(y_C) \). We will therefore refer to these wind profiles as the zero stress, 0.5 \( \times \tau_C \), 1 \( \times \tau_C \), 1.5 \( \times \tau_C \), and 2 \( \times \tau_C \) profiles: that is, based on the magnitude of the wind stress along \( y = y_C \) relative to that for our experiments forced using \( \tau_{\text{ref}} \). In all cases, we take \( r \) to be the reference value: \( r = \tau_0 = 10^{-7} \text{ s}^{-1} \). The reference wind stress experiment, 1 \( \times \tau_C \), is then identical to the experiment labeled \( r_0 \) in Fig. 3.

1) TRANSPORT

Upper-layer transport values obtained for this set of experiments are shown in Fig. 7. The various curves appear similar, except for an offset related to the addition of the zero-curl wind stress. Thus, clearly, the zero-curl portion of the wind stress plays a key role in determining the total transport, as expected from channel theories. For a fixed curl profile, more stress in the channel translates to more circumpolar transport. Conversely, for a fixed value of stress in the channel, more curl seems to imply less transport (e.g., compare \( \tau_0 = 0.1, 0.2, \) and 0.4 on the 2 \( \times \tau_C \), 1 \( \times \tau_C \), and 0.5 \( \times \tau_C \) curves, respectively).

Note also that the general form of a Stommel regime followed by a saturation regime is also roughly respected for both stratifications and irrespective of the channel wind stress.

Figure 8 shows sample time-averaged barotropic streamfunctions for our 0 \( \times \tau_C \), 1 \( \times \tau_C \), and 2 \( \times \tau_C \) simulations. For very small wind amplitude (\( \tau_0 = 0.02 \text{ N m}^{-2} \)), adding a zero-curl wind stress mainly affects the channel contribution to the transport. In the 0 \( \times \tau_C \) experiment, a westward countercurrent is observed along southern boundary. This disappears in the 1 \( \times \tau_C \) and 2 \( \times \tau_C \) experiments, for which \( T_{\text{channel}} \) is observed to increase with the stress, whereas the basin contribution is left unaffected. For stronger forcing amplitudes, the zero-curl stress affects the strength of the upper-layer recirculation. This is seen in Fig. 8 as a displacement of the point \( x_0 \) marking the observed division between recirculating and circumpolar streamlines. As the stress in the channel is increased, the recirculation shrinks and \( x_0 \) moves to the west.

On casual inspection, an increased constant stress at strong forcing amplitudes appears equivalent to an increased bottom drag in that both lead to a westward displacement of \( x_0 \) and to an increase in the transport. Also similar to the bottom drag experiments is that the barotropic flow is in approximate Sverdrup balance in the basin region, north of Drake Passage (see \( \psi_{\text{diff}} \) in Fig. 9). There is, however, a difference between the two experiments. Changing \( r \) changed the zonal extent of the abyssal recirculation. Here, this recirculation is similar between the three experiments and extends roughly to the position predicted by the separatrix characteristic (see \( \psi_{\text{diff}} \) in Fig. 9). What differs is that, in the 0 \( \times \tau_C \) experiment, some of the top-trapped Sverdrup flux entering the channel region recirculates, whereas, in the 2 \( \times \tau_C \) experiment, some of the southward Sverdrup flux that is not top trapped nonetheless feeds the circumpolar transport.

To summarize, it appears that addition of a uniform eastward wind stress does not significantly affect the lower-layer circulation. At low forcing amplitude, it tends to increase \( T_{\text{channel}} \). In the stronger forcing regime, addition of a uniform eastward wind stress decreases the upper-layer expression of the eddy-driven recirculation gyre. Associated with this is an increase in the saturation level for the circumpolar transport.

2) MOMENTUM TRANSFER

We next consider the eastward momentum balance in the Drake Passage latitude band. As before, the vertical transfer of eastward momentum is dominated by the interfacial form drag, and we focus on a spectral
decomposition of this term as a function of horizontal wavenumber (Fig. 10). Note that the averaged momentum transfer roughly cancels in the zero-stress case because there is little net zonal momentum input to channel latitudes in these simulations. Note also that the stress integrated over all wavenumbers differs between the experiments, in contrast to Fig. 5. Nonetheless, several familiar features are evident. As previously, a time decomposition of these spectra shows the low $k$ structure ($k < 20$) to be associated with standing eddies, whereas the high $k$ features are associated with transients. The low $k$ structure is similar to that seen for the bottom friction experiments, with a peak of upward momentum flux near $k = 15$ and strong downward momentum flux for $k < 10$.

The interfacial stress associated with transient eddies (high $k$) is weakly upward for the 0, 0.5, and $1 \times \tau_C$ experiments but becomes strongly downward for the $2 \times \tau_C$ experiment. Note also that the momentum transfer at large and intermediate scales changes significantly between the 0.5 and $1 \times \tau_C$ experiments. Thus, for zero to moderate values of the stress along $y_C$, most of the downward momentum transfer is performed by the large-scale standing eddies. For stronger stresses, however, the large-scale downward transfer increases significantly and an intermediate $k$ upward transfer appears. Together, these do not accomplish the required net downward transfer and in this limit the direct effect of transients becomes increasingly important in transferring the momentum downward. In this sense, the role of transients in the channel latitude band becomes similar to their role in the momentum budget calculated along time-mean streamlines in the strong stress limit.

c. Effect of the wind stress curl

In this section, we show that the transport is also affected by changing the wind stress curl, while holding the midchannel wind stress constant. We find that increasing $\tau_0$ while keeping $\tau_1$ fixed tends to decrease the transport. This trend was noted in passing in the previous section but will be considered in more detail here. Also, in the previous section, we chose a fairly narrow channel width (240 km). Because this might tend to lessen the role of channel-like dynamics on determining the transport, a larger value (720 km) is considered here. Because the channel is wider, we also changed the precise form of the topography. It is wider and elongated...
so as to extend across the wider gap at the model Drake Passage.

1) TRANSPORT

We begin by comparing simulations forced using $\tau_{ref}$ with experiments forced using a uniform eastward wind stress ($\tau_0 = 0$) [cf. Eqs. (7)–(9)]. The wind stress in the center of the channel is identical in the two sets of simulations. Figure 11 shows upper-layer transports for these experiments. The experiments forced using the uniform wind stress are labeled as zero curl and those forced using the reference wind stress profile are labeled as curl. For the curl experiments, the 240-km gap results are included for comparison. Transport is significantly higher in the zero-curl experiments. As expected, a clear saturation at higher levels of forcing is seen for the reference wind profile, whereas saturation is less evident in the zero-curl experiments. In the strongly forced regime, the curl experiments have transports smaller by a factor of 2 or more compared to the no-curl experiments. As we will see below, the structure of the flow is also quite different.

Note that the saturation seen in the curl experiments appears to be independent of the width of the gap. This is as expected in the analytic model of NS09. Also consistent with the analytic model is that the Stommel regime is clear at low values of the forcing in the 240-km case and is almost absent in the 720-km gap experiments. This is because the basin part of the domain is smaller in latitudinal extent for the 720-km gap simulations.

![Fig. 10](image-url) As in Fig. 5, but for different values of stress in the middle of the channel.
We wish to more systematically compare simulations having identical stresses in the center of the channel but different amplitudes of the curl. Starting with several of the zero-curl simulations from Fig. 11, we add a $\tau_{\text{curl}}$ profile by progressively increasing $\tau_0$. Each set of experiments is labeled according to the value of stress along $y = y_C$, which does not change as the curl is increased. Specifically, we label the experiments $\tau_1 = 0.02$, $\tau_1 = 0.1$, $\tau_1 = 0.2$, and $\tau_1 = 0.4$ (cf. the horizontal axis in Fig. 11). Transports obtained for this set of experiments are shown in Fig. 12. Except in the $\tau_1 = 0$ case, the upper-layer transport decreases as the amplitude of the curl is increased. Note that there is also a westward transport in the lower layer. This increases with the strength of the curl and is relatively independent of the stress applied in the middle of the channel.

Upper-layer streamfunction and velocity magnitude for the $\tau_1 = 0.1$ experiments with different amplitudes of curl forcing $\tau_0$ are shown in Fig. 13. In the zero-curl simulations (Fig. 13, top), motion in the basin portion of the domain is very weak. The circumpolar transport is composed of two zonal jets near the channel latitudes and a strong southern boundary current. This boundary current is particularly strong in the western half of the domain (i.e., immediately east of the gap), where it corresponds to almost 40% of the total upper-layer transport. It becomes smaller as the wind stress curl is progressively added and changes direction for $\tau_0 = 0.2$.

North of the gap, a Sverdrup circulation appears and increases in strength with the curl. With it, a recirculating gyre also appears. For lower values of $\tau_0$, the recirculation is seen only in the lower layer, whereas for higher values it is also visible in the upper layer. For example, it is clearly visible in Fig. 13 (bottom left). A significant decrease of zonal transport is correlated with the appearance of this recirculating gyre. The right-hand panels show instantaneous velocity magnitudes. Zonal jets become increasingly predominant in the basin region as the curl is increased, and associated with this is an increase in eddy activity (not shown).

2) MOMENTUM TRANSFER

As before, we consider the interfacial form drag in the channel region for this set of experiments. Also as previously, the main balance in both experiments is that the wind stress inputs eastward momentum, which is transferred downward via an interfacial form stress and removed by topographic form drag. The Reynolds stress term becomes significant only in the zero-curl simulations and always remains small compared to the interfacial form drag (not shown).

![Fig. 11. Upper-layer time-averaged zonal transport. The horizontal axis is $\tau_0$ for the curl experiments and $\tau_1$ for the zero-curl experiments. The stress along $y = y_C$ is thus equal in both sets of experiments for a given position along the horizontal axis.](http://journals.ametsoc.org/doi/pdf/10.1175/JPO-D-11-058.1)

![Fig. 12. Average zonal transport (a) in the top layer and (b) in the bottom layer for different stresses in the center of the channel in function of the wind curl amplitude.](http://journals.ametsoc.org/doi/pdf/10.1175/JPO-D-11-058.1)
Figure 14 shows a spectral decomposition of the interfacial stress for different amplitudes of the wind stress curl. As before, positive values correspond to downward transfers of eastward momentum and low $k$ structure ($k < 20$) is associated with standing eddies. When the wind stress curl is set to zero, all downward momentum transfer is accomplished by mesoscale transient eddies. However, for strong curl, the momentum transfer is performed by the large-scale standing eddies associated with the time-mean flow. This suggests that transient eddies transfer momentum downward (in channel latitudes) when no other mechanism is available, but they can transfer momentum upward when large-scale recirculation patterns appear in the time-mean flow. Note that, even in the latter case, transient eddies play an indirect role in that they help determine the large-scale recirculation patterns.

4. Discussion

We considered the robustness of circumpolar transport in a quasigeostrophic model with an idealized Southern Ocean geometry to the magnitude of the bottom drag and to the relative magnitudes of the wind stress and wind stress curl. Our focus was on the strong forcing, or saturation regime. In this regime and except for forcing profiles with small curls, essentially all of the upper-layer circumpolar flow was fed by a southward Sverdrup flux into Drake Passage latitudes. A fraction of this same Sverdrup flux also fed into a recirculation gyre lying above an eddy-driven gyre. What fraction fed the ACC and what fraction fed the gyre were strongly dependent on both the bottom drag coefficient and zero-curl portion.
of the wind stress. Associated with this, larger drag and stronger eastward wind stress were both found to yield larger circumpolar transports.

It may seem counterintuitive that zonal transport should increase with bottom friction because the opposite is true in the absence of topography. A similar dependence was found by Hogg and Blundell (2006), who explained it using eddy saturation ideas. Here, this dependence on \( r \) can be explained in the NS09 framework. In this framework, the transport is given by the top-trapped portion of the Sverdrup flux into Drake Passage latitudes. For large \( r \), the lower-layer circulation is weak, so that almost all of the Sverdrup flux is top trapped (and hence feeds the ACC). For smaller \( r \), a large-scale eddy-driven recirculation (or gyre) occurs. In this case, part of the upper-layer Sverdrup flux into the Drake Passage latitude band feeds the recirculation and as such does not add to the circumpolar transport. NS09 assumed the eddy-driven abyssal circulation to be contained inside the separatrix geostrophic contour following Rhines and Young (1982). For low drag, it extends farther east and for high drag it does not extend eastward to the separatrix. Because of this, the fraction of the southward Sverdrup flux into the Drake Passage latitude band that is top trapped did not correspond to the analytic prediction.\(^1\) Associated with this, a saturation was not observed when \( r \) was either relatively high or relatively low.

It is not counterintuitive that zonal transport should increase with an increased eastward wind stress; rather this is anticipated by analogy with channel flows. It is perhaps counterintuitive, however, that in the saturation regime this added transport does not represent an increased “channel contribution.” Instead it is linked to a change in the fraction of the Sverdrup flux feeding the circumpolar flow. NS09 assumed that, where the Sverdrup flux is top trapped, it feeds the circumpolar current and, where it is not, it feeds the recirculation. Instead, it appears that stress in the channel influences the upper-layer expression of the recirculation gyre, and, with it, the relative fractions of Sverdrup flux feeding the recirculation and circumpolar current. Stronger eastward winds allow for some of the upper-layer Sverdrup flux overlying the abyssal recirculation to feed the circumpolar current.

We also found that adding a wind stress curl to a uniform wind stress forcing profile (without altering the midchannel wind stress) tends to reduce the transport. Addition of a curl in the form of a double-gyre forcing creates an area inside the separatrix characteristic where eddy activity readily leads to the development of an abyssal recirculation. Once again, the strength of this abyssal gyre helps to control the partitioning of the Sverdrup flux into Drake Passage latitudes between circumpolar and recirculating flow.

The strength of the abyssal recirculating gyre was found to (i) decrease with increasing bottom drag, (ii) be relatively insensitive to the wind stress in the channel, and (iii) increase with increasing curl. The level of stress in the channel instead affects the upper-layer expression of the recirculation, with more stress implying less upper-layer recirculation (and hence more transport). We also found that, where the upper-layer recirculation was more prevalent (e.g., weak bottom drag, weak stress, or strong curl), the bulk of the interfacial form drag was associated with the large-scale, time-mean flow. In these regimes, the direct effect of mesoscale transients was to accomplish an upward transfer of eastward momentum in the channel latitude band. In limits where the recirculation was weak (e.g., strong drag, strong stress, or weak curl), standing eddies contributed little or insufficiently to the interfacial form drag and mesoscale transients transferred eastward momentum downward.

A limited number of simulations were also carried out at higher resolution. Although the precise value of the transport for a given choice of other parameters was found to be resolution dependent, the general behavior described above was not (not shown). For example, simulations similar to those in sections 3a and 3c were carried out at 7.5-km resolution. As with the 15-km-resolution simulations, these showed transport to increase with \( r \) and decrease with the curl (for a fixed midchannel wind stress). Moreover, this behavior was associated with changes in the recirculating gyre, as reported above. Simulations in a five-layer setting with different values of \( r \) were also carried out (at 15-km resolution). Like our two-layer simulations, these also showed transport to increase with decreasing \( r \). As above, this was linked to the partitioning of the Sverdrup flux into the Drake Passage latitude band between recirculating and circumpolar streamlines. The main difference found was that increased resolution (both horizontal and vertical) led to a strengthening of the tight recirculation immediately above our Scotia Ridge–like topography (similar to that in Fig. 4).

A simple theory for what sets the transport of the ACC, even in our two-layer quasigeostrophic setting, remains elusive. Nevertheless, several avenues offer promise. Gyre dynamics clearly play an important role; however, to translate this to an estimate of the transport would require a better understanding of what sets the level eddy activity in the lower layer and how this affects both the

\(^1\) NS09 showed that the introduction of a continental-rise-like topography on the western boundary could also allow for zonal jets to extend beyond the separatrix, resulting in a reduction of the transport similar to that seen for weak drag.
abyssal recirculation gyre and its upper-ocean expression. A focus on the dynamics of momentum transfer in the Drake Passage latitude band also offers insight. Here, that dynamics was made complex by a strong topographically trapped gyre sitting over our model Scotia Ridge. It could be that our choice of such a localized topography complicates the dynamics and that a more realistic topography will serve to distribute the bottom form drag over a broader range of longitudes. In this case, it is not clear that such a large (and arguably unrealistic) topographic recirculation needs to develop. Preliminary results show this feature to be much reduced when a more realistic topography is used (not shown). A third perspective is to focus on the momentum budget integrated along the path of the ACC rather than along latitude circles. This has the advantage of removing standing eddies from consideration, because the standing eddy contribution to form drag depends on $\psi$ varying along the integration path. A problem with this approach, though, is that this path is itself part of the solution. It thus seems likely that some combination of these approaches will be needed. For example, it may be that basin dynamics can be exploited to give an a priori estimate of the circumpolar path and that some measure of the wind stress integrated along that path determines the transport, at least over a range of drag coefficients and other model parameters.

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