Reshaping the Antarctic Circumpolar Current via Antarctic Bottom Water export

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ABSTRACT

Zonal momentum input into the Antarctic Circumpolar Current (ACC) by westerly winds is ultimately removed via topographic form stress induced by large bathymetric features that obstruct the path of the current. These bathymetric features also support the export of Antarctic Bottom Water (AABW) across the ACC via deep geostrophically-balanced northward flows. These deep geostrophic currents modify the topographic form stress, implying that changes in AABW export will alter the ocean bottom pressure and require a rearrangement of the ACC in order to preserve its zonal momentum balance. A conceptual model of the ACC momentum balance is used to derive a relationship between the volume export of AABW and the shape of the sea surface across the ACC’s standing meanders. This prediction is tested using an idealized eddy-resolving ACC/Antarctic shelf channel model that includes both the upper and lower cells of the Southern Ocean meridional overturning circulation, using two different topographic configurations to obstruct the flow of the ACC. Eliminating AABW production leads to a shallowing of the sea surface elevation within the standing meander. To quantify this response, the authors introduce the “surface-induced topographic form stress”, the topographic form stress that would result from the shape of the sea surface if the ocean were barotropic. Eliminating AABW production also reduces the magnitude of the eddy kinetic energy generated downstream of the meander, and the surface speed of the ACC within the meander. These findings raise the possibility that ongoing changes in AABW export may be detectable via satellite altimetry.

1. Introduction

The Antarctic Circumpolar Current (ACC) in the Southern Ocean has a number of unique features. Most importantly, the lack of continental boundaries in the east-west direction means that circumpolar flow is uninterrupted, leading to continuous eastward flow. The first to recognize that this continental configuration would lead to special dynamics were Munk and Palmén (1951), who speculated that momentum input from surface winds would be balanced by form stress at the bottom of the ocean. Munk and Palmén’s view differed from the Sverdrup balance arguments that provided a first-order description of gyre circulation in other ocean basins (Stommel 1948; Munk 1950), and ultimately lead to a protracted debate over the true natures of the mechanisms driving the ACC (Warren et al. 1996; Olbers 1998; Hughes 1997). These disparate views were (mostly) reconciled by Hughes and de Cuevas (2001), who demonstrated the equivalence of vorticity-based (Sverdrup balance) arguments and zonal momentum (form stress) balances. The result is that, whilst ACC dynamics can be most clearly understood from the perspective of the zonal momentum balance, entirely analogous arguments can be constructed from a barotropic vorticity perspective.

Viewed from the momentum balance perspective, modern ocean models have confirmed Munk and Palmén’s prediction that momentum input from surface winds is balanced by topographic form stress (Stevens and Ivchenko 1997). Topographic form stress is strongly dependent on standing meanders (Thompson and Naveira Garabato 2014), which arise as standing barotropic Rossby waves (Marshall 2016). Transient eddies act to distribute horizontal momentum in the vertical direction within the water column (Tréguier and McWilliams 1990). During a transient spin-up, topographic form stress reacts quickly to the surface stress forcing (Ward and Hogg 2011) implying that the bulk of topographic form stress can be considered to occur via a rapid adjustment of the barotropic Rossby waves. Subsequent internal readjustment of the vertical profile of zonal momentum can then occur through interfacial form stress associated with transient eddies.

The distribution and magnitude of topographic form
stress varies around the Southern Ocean, with maxima near major obstacles such as Kerguelen Plateau, Campbell Plateau and Drake Passage (Gille 1997). Eddy-permitting models indicate that most of the form stress occurs at mid-depth, with the deep ocean contributing little to net form stress (Masich et al. 2015). This result is consistent with the notion that the deep ocean form stress may be dominated by large gyres, recirculating around bathymetric contours (Nadeau and Ferrari 2015).

Understanding the ACC momentum budget is required to predict the future structure and transport of the ACC which, in turn, alters meridional and vertical transport in the Southern Ocean. Despite the role of westerly winds in contributing momentum to the ACC, the net transport is relatively insensitive to the magnitude of wind stress (Hogg et al. 2015). The barotropic response of the ACC to wind stress variations can occur rapidly (Meredith et al. 2004), but over longer timescale the adjustment of interior momentum is modified by a transient eddy field that is roughly proportional to wind stress (Meredith et al. 2012). The result is that the sensitivity of ACC transport to wind stress changes is reduced (Hogg et al. 2015), and that this sensitivity may be mostly derived from the action of wind near the southern boundary (Langlais et al. 2015).

On the other hand, ACC transport in ocean models is sensitive to surface buoyancy fluxes (Gent et al. 2001; Borowski et al. 2002; Hogg 2010). The mechanisms at play here can be described as follows. Enhancing the loss of buoyancy in the southern part of the Southern Ocean can increase the strength of Antarctic Bottom Water (AABW) formation (e.g. Behrens et al. 2016; Lynch-Stieglitz et al. 2016). This enhanced AABW formation drives a stronger meridional overturning circulation, leading to a baroclinic torque via the Coriolis force, which produces westward flow in the lower layers and eastward flow higher in the water column. Westward bottom flow can then interact with topography, thereby reducing the loss of eastward momentum by topographic form stress (Howard et al. 2015). Thus, enhanced AABW formation may accelerate the ACC by drawing eastward momentum from the solid earth.

The consequence of AABW modifying the ACC transport is that future ACC transport may be reduced by changing surface fluxes owing to near-surface warming or freshening of the Southern Ocean (e.g. Downes and Hogg 2013). However, these predictions are difficult to verify from observations, as the rate of formation of AABW is poorly constrained. The solution to this problem may lie in the dynamics that lead to rapid barotropic adjustment of the topographic form stress in response to changes in wind stress. That is, if AABW-driven changes in topographic form stress occur, then the barotropic adjustment may work in reverse to produce a surface signature. If true, then the surface response to AABW changes may be more readily observable than direct measurements of the rate of AABW formation.

In this paper, we test the hypothesis that a change in the rate of production of AABW modifies the topographic form stress, and thereby creates a surface response. If this hypothesis holds, then there is the possibility that the surface signature of AABW changes may be detectable using satellite altimetry. We begin by elaborating, on theoretical grounds, the possible modes of adjustments of the ACC to AABW changes (§2), then test these predictions in an idealized, eddy-resolving model of the ACC region (§3). In §§4 and 5 we diagnose the responses of this model’s mean and transient surface flows, respectively, to changes in AABW export. Finally, in §6 we summarize our results and discuss the applicability of our findings to the ocean.

2. AABW and the zonal momentum balance

In this section we present a conceptual model to explain why changes in the force balance of the ACC due to AABW export are expected to alter the ACC’s mean circulation and sea surface height (SSH). We present this argument from an Eulerian perspective because it concerns only the net topographic form stress, rather than the mechanism via which momentum is transferred vertically (Warren et al. 1996). However, as discussed further below, the result is intrinsically linked to the transfer of momentum via form stresses between isopycnal layers (Hughes 1997; Olbers 1998), and our discussion could be equally well posed in density coordinates.

Fig. 1(a) shows a conceptual barotropic model of the force balance of the ACC along a latitude band. The total force associated with the eastward surface wind stress is balanced by a westward pressure gradient force supported by zonal variations in the bathymetry (e.g. Olbers et al. 2004), here sketched as a single Gaussian-like ridge. More formally,

$$F_{\text{wind}}(x) + F_{\text{topog}}(x) = 0 \quad (1)$$

where

$$F_{\text{wind}}(x) = \int \tau(x) \, dx, \quad \text{and} \quad F_{\text{topog}}(x) = - \int p_b \frac{\partial \eta}{\partial x} \, dx \quad (2)$$

are the total eastward wind force per unit latitude and total topographic form stress (defined positive eastward) per unit latitude, respectively. Here we denote the eastward surface wind stress as $\tau(x)$, the pressure at the ocean bed as $p_b$, and the height of the ocean bed as $z = \eta(x)$. In the case of a barotropic, hydrostatic ACC, the bottom pressure and topographic form stress can be written explicitly as

$$p_b = \rho_0gh_0 \implies F_{\text{topog}}(x) = - \int \rho_0gh_0 \frac{\partial \eta}{\partial x} \, dx \quad (3)$$

where $h_0 = \eta_0 - \eta_b$ is the thickness of the fluid layer. For the symmetric ridge illustrated in Fig. 1(a), the east-to-west increase in bottom pressure across the ridge requires that
Fig. 1. Schematic of the Antarctic Circumpolar Current (ACC) force balance and its adjustment to accommodate AABW export. (a) The canonical ACC force balance illustrated for a barotropic eastward flow, in which eastward surface wind stress is opposed by a topographically-supported westward pressure gradient force. (b) Introducing a geostrophically-balanced northward export of Antarctic Bottom Water (AABW) introduces an eastward anomaly in the pressure gradient force, as dense water piles up on the eastern side of the topography. (c,d) Hypothetical situations in which the force balance is re-established by adjustment of (c) the sea surface and (d) interior isopycnal surfaces overlying the AABW layer.

In general, SSH must be higher on the western side of the ridge than the eastern side. This asymmetry is achieved via the formation of a standing meander (Viebahn and Eden 2012; Thompson and Naveira Garabato 2014), in which the ACC is deflected northward around the ridge.

In Fig. 1(b) we illustrate the impact of a northward-flowing layer of AABW on the ACC’s force balance. Denoting the upper surface of this layer as $z = \eta_A(x)$ and its density as $\rho_A$, the new bottom pressure can be written as

$$p_b^* = \rho_0 g (\eta_0 - \eta_b) + \rho_0 g_A' (\eta_A - \eta_b),$$

where $g_A' = g(\rho_A - \rho_0)/\rho_0$ is the reduced gravity. Thus the bottom pressure anomaly associated with the introduction of the AABW layer is

$$\Delta p_b = p_b^* - p_b = \rho_0 g_A' h_A,$$

where $h_A = \eta_A - \eta_b$ is the thickness of the AABW layer. Here we have assumed that the AABW is exported via a geostrophically-balanced mean flow, as suggested by export pathways derived from observations (Orsi et al. 1999) and the Southern Ocean State Estimate (Mazloff et al. 2010; van Sebille et al. 2013). To achieve a northward geostrophic transport, the AABW layer must pile up against the eastern side of the ridge, i.e. $h_A^{\text{east}} > h_A^{\text{west}}$. Equation (5) demonstrates that this preferentially increases the bottom pressure on the eastern side of the ridge relative to the western side, and thereby introduces an eastward pressure gradient force anomaly that serves to reduce the topographic form stress,

$$\Delta F_{\text{topog}}^{(x)} = - \oint dx \, \rho_0 g_A' \frac{\partial \eta_b}{\partial x}. \quad (6)$$

Note that (6) and the results derived below apply whether
the AABW layer outcrops from the topography, as shown in Fig. 1(b), or not. We therefore retain the closed contour notation for integrals, with the understanding that \( h_A = 0 \) and \( \eta_A = \eta_h \) where the AABW layer is absent.

The preceding paragraph and Fig. 1(b) argue qualitatively that inserting a geostrophic export of AABW should serve to oppose the topographic form stress. However, it remains unclear whether this effect is dynamically significant, so we now attempt to quantify the topographic form stress anomaly \( \Delta F_{\text{topog}}^{(x)} \). First, to obtain an order-of-magnitude estimate we consider the case of a motionless upper layer \( v_0 = 0 \) and substitute \( \eta_h = \eta_A - h_A \) into (6) to obtain

\[
\Delta F_{\text{topog}}^{(x)} = -\oint dx \rho_0 g A^2 h_A \frac{\partial \eta_A}{\partial x} = -\rho_0 f T_A^* \quad \text{(7)}
\]

Here the first equality follows from an integration by parts, and \( v^* = (g/f) \partial \eta_A/\partial x \) and \( T_A^* > 0 \) are estimates of the northward AABW geostrophic velocity and volume flux respectively. These variables are decorated with stars * to indicate that (7) is only valid in the limit of a quiescent and infinitely deep overlying layer of ocean.

Using (7) we can estimate the magnitude of the change in topographic form stress due to the introduction of the AABW layer. For \( T_A = 30 \text{Sv} \) (Talley 2013), we obtain \( -\rho_0 f T_A \sim 3 \times 10^8 \text{N m}^{-1} \). For comparison, we scale the total wind force per unit latitude using representative values for the eastward wind stress, \( \tau(x) = 0.15 \text{ N m}^{-2} \), and zonal length of the ACC, \( L_x = 20,000 \text{ km} \), which yields \( F_{\text{wind}}^{(x)} \sim 3 \times 10^8 \text{ N m}^{-1} \). Thus the anomaly in the topographic form stress induced by the addition of an AABW layer is comparable in magnitude to the circumpolar-averaged surface wind forcing.

The above estimate neglects meridional flow in the upper layer: we now reinstate \( v_0 \) using the two-layer case shown in Fig. 1(b) to derive an exact relation between the form stress and the layer transports. Following a similar procedure as in (7),

\[
F_{\text{topog}}^{(x)} = -\oint dx \rho_b \frac{\partial \eta_b}{\partial x} = -\rho_0 f (T_0 + T_A) \quad \text{(8)}
\]

Here

\[
T_0 = \oint dx v_0 h_0 = \oint dx h_0 \frac{g \partial \eta_0}{f} \quad \text{(9)}
\]

is the northward geostrophic transport in the upper layer, and

\[
T_A = \oint dx v_A h_A = \oint dx h_A \left( \frac{g \partial \eta_0}{f} + \frac{g_A \partial \eta_A}{f} \right) \quad \text{(10)}
\]

is the geostrophic transport in the AABW layer, under the Boussinesq approximation. Equation (8) simply states that the topographic form stress is proportional to the total southward geostrophic transport. The requirement that (1) holds is equivalent to requiring that the total geostrophic meridional transport opposes the northward wind-driven Ekman transport, as can be seen by dividing both sides of (1) by \( (f \rho_0) \),

\[
\oint dx \frac{\tau(x)}{\rho_0 f} = T_0 + T_A \quad \text{(11)}
\]

In the two-layer configuration depicted in Fig. 1(b), the AABW-induced imbalance between the surface wind stress and topographic form stress must be resolved by increasing the geostrophic southward transport in the upper layer. This scenario corresponds to a reshaping of the SSH in the standing meander, as illustrated in Fig. 1(c), in order to re-establish the hydrostatic pressure gradient across the ridge.

However, the ACC is continuously stratified, and it is conceivable that the bottom pressure anomaly induced by AABW export may be accommodated via adjustment of lighter density classes, rather than an adjustment of the SSH. As illustrated by Fig. 1(d), this corresponds to southward flow in an intermediate density layer that we might identify as Lower Circumpolar Deep Water (LCDW) with density \( \rho_L \) and thickness \( h_L \). In this case it may be shown that (8) extends to

\[
F_{\text{topog}}^{(x)} = -\oint dx \rho_b \frac{\partial \eta_b}{\partial x} = -\rho_0 f_0 (T_0 + T_L + T_A) \quad \text{(12)}
\]

where the superscript † indicates that this expression refers to the three-layer topographic form stress and \( T_L \) is the geostrophic northward transport in the LCDW layer. However, even if the northward AABW transport is balanced exactly by southward LCDW transport, i.e. \( T_L + T_A = 0 \), a rearrangement of the surface circulation is still required because now the upper layer thickness \( h_0 \) is much thinner than it was in the barotropic case, and the surface geostrophic velocity \( v_0 \) must change accordingly.

In reality, both of the configurations presented in Fig. 1(c–d) are oversimplifications of the ACC’s response to AABW formation, which is likely to include adjustment of both the SSH and the isopycnal surfaces associated with lighter density classes. For the purpose of quantifying the ACC’s surface response to AABW changes, we define the “surface-induced” contribution to the topographic form stress as

\[
F_{\text{surf}} = -\oint dx \rho_0 g \frac{\partial \eta_b}{\partial x}. \quad \text{(13)}
\]

This is equivalent to equation (3), and quantifies the topographic form stress that would result if the ocean were purely barotropic, i.e. if the pressure at the ocean bed at each horizontal location was simply set by the sea surface height above that location. In the case of purely barotropic flow it may be shown that \( F_{\text{surf}} = -\rho_0 f T_0 \), but this holds
only approximately once other density layers are added. The practical advantage of (13) is that it may be computed entirely from the sea surface elevation and the shape of the sea floor.

In summary, a mean geostrophic export of AABW demands an adjustment of the ACC’s SSH and/or isopycnal surfaces in order to preserve the topographic form stress, or equivalently to provide a southward return flow. The central hypothesis of this paper is that adjustments of SSH in response to AABW changes offer the possibility of detecting changes in AABW export from altimetry. However, the conceptual model outlined above does not constrain the amplitude, nor spatial pattern of the SSH response. In the following sections we explore this response in more detail using an idealized eddy-resolving numerical model of the ACC.

3. An ACC channel model with AABW outflow

In this section we describe the configuration of our idealized ACC model. Our goal is to achieve a model of minimum complexity that can be used to test the hypothesis set out in §2. Our modeling approach is constrained by the following requirements:

i. The model must simulate an AABW outflow that crosses the ACC across the ocean bed rather than sinking along isopycnals (e.g. Abernathey et al. 2011).

ii. We require a model with a relatively fine vertical grid in order to resolve the overturning circulation and the vertical distribution of form stress across bathymetry.

iii. Changes in the mean circulation of the ACC may be expected to be accompanied by changes in the eddy field (e.g. Hogg et al. 2008; Langlais et al. 2015). We therefore require that the model resolve mesoscale eddies.

iv. We require that the experiments be integrated for several decades to accurately diagnose the force balance of the ACC in a statistically steady state.

These requirements jointly motivate the use of an idealized model that can be affordably integrated with fine grid resolution for long periods, and in which the export of AABW can be controlled and its impact on the ACC unambiguously diagnosed.

a. Model configuration

Our idealized model is comprised of a re-entrant channel of the ACC and Antarctic continental shelf/slope, and closely resembles that of Stewart and Thompson (2013). The model geometry and forcing are summarized in Figs. 2 and 3. In this subsection we provide a brief overview of the model configuration, highlighting differences from that of Stewart and Thompson (2013) in the interest of reproducibility.

We solve the Boussinesq hydrostatic primitive equations using the MIT general circulation model (MITgcm Marshall et al. 1997a,b) in ocean-only configuration. The model parameters are listed in Table 1 for ease of reference. The ocean thermodynamics are simplified to a single-component equation of state that depends linearly on temperature with thermal expansion coefficient \( \alpha = 1 \times 10^{-4} {K^{-1}} \). The equations of motion therefore effectively simplify to the Boussinesq equations with a single buoyancy variable representing the thermodynamics (e.g. Vallis 2006), which is sufficient to produce a qualitatively realistic overturning circulation (e.g. Abernathey et al. 2011; Stewart and Thompson 2013) and topographic form stress (e.g. Munday et al. 2015). The equations are solved in a

| Table 1. List of parameters used in our Reference simulations. |
|------------------------|------------------------|------------------------|
| Param. | Value | Description |
| \( L_x \) | 2000 km | Zonal domain size |
| \( L_y \) | 2000 km | Meridional domain size |
| \( H \) | 4000 m | Maximum ocean depth |
| \( H_s \) | 500 m | Continental shelf depth |
| \( Y_p \) | 300 km | Continental slope latitude |
| \( W_s \) | 75 km | Continental slope half-width |
| \( X_b \) | -500 km | Bump/ridge zonal position |
| \( Y_b \) | 1200 km | Bump meridional position |
| \( W_b \) | 200 km | Bump/ridge zonal half-width |
| \( L_b \) | 400 km | Bump meridional half-width |
| \( H_b \) | 2000 m | Bump/ridge height |
| \( Y_0 \) | 500 km | ACC/ASF boundary |
| \( L_c \) | 100 km | Radius of shelf cooling |
| \( T_c \) | 7 days | Shelf cooling timescale |
| \( L_r \) | 100 km | Width of northern sponge |
| \( T_r \) | 7 days | Sponge timescale |
| \( \tau_e \) | 0.15 N m\(^{-2}\) | Eastward wind stress max. |
| \( \tau_w \) | 0.05 N m\(^{-2}\) | Westward wind stress max. |
| \( Q_n \) | 10 W m\(^{-2}\) | Northern heat flux |
| \( Q_s \) | 5 W m\(^{-2}\) | Southern heat flux |
| \( \alpha \) | \( 1 \times 10^{-4} {K^{-1}} \) | Thermal expansion coeff. |
| \( f_0 \) | \(-1.3 \times 10^{-4} {s^{-1}} \) | Reference Coriolis parameter |
| \( \beta \) | \( 9.6 \times 10^{-12} {(ms)^{-1}} \) | Meridional gradient of \( f \) |
| \( A_h \) | 12 m\(^2\) s\(^{-1}\) | Horizontal viscosity |
| \( A_v \) | \( 3 \times 10^{-4} {m^2 s^{-1}} \) | Vertical viscosity |
| \( A_{\text{grid}} \) | 0.1 | Grid biharmonic viscosity |
| \( C_{4 \text{leith}} \) | 1.0 | Leith vertical viscosity |
| \( C_{4 \text{leithD}} \) | 1.0 | Leith solenoidal viscosity |
| \( \kappa_v \) | \( 5 \times 10^{-6} {m^2 s^{-1}} \) | Vertical diffusivity |
| \( \tau_b \) | \( 1.1 \times 10^{-3} {m s^{-1}} \) | Bottom friction |
| \( \Delta_x, \Delta_y \) | 4.2, 4.0 km | Horizontal grid spacing |
| \( \Delta_z \) | 10.5 m–103.8 m | Vertical grid spacing |
| \( \Delta t \) | 287 s | Time step size |
Fig. 2. Model domain and applied surface forcing. (a) Downward surface heat flux \( Q(y) \) and (b) applied surface wind stress \( \tau^x(x) \). In panels (a–b) the solid lines correspond to our Reference simulations, the dotted lines correspond to our simulations with no Antarctic Bottom Water (AABW) export, and the dashed lines correspond to our simulations with halved eastward wind stress. (c) Zonal- and time-mean zonal velocity (colors) and temperature (contours) for our RIDGE reference simulation. The arrow indicates the volume over which the shelf waters are cooled close to \( x = 0 \) in order to produce AABW.

Cartesian channel with zonal and meridional dimensions \( L_x = L_y = 2000 \text{ km} \) and maximum depth \( H = 4000 \text{ m} \). The channel is presumed to exist on a tangent plane to the Earth’s surface, with Coriolis parameter \( f_0 = -1.3 \times 10^{-4} \text{ s}^{-1} \) and \( \beta = 1 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1} \) appropriate for the Southern Ocean.

We impose steady fluxes of energy and momentum at the ocean surface that qualitatively capture the circum-polar patterns of surface forcing in the Southern Ocean (Large and Yeager 2009), with additional shelf cooling to aid bottom water formation. Surface fluxes are imposed as zonally-uniform functions \( Q(y) \) and \( \tau^x_w(y) \) that are split into “ACC” and “ASF” (Antarctic Slope Front) bands separated at \( y = Y_0 = 500 \text{ km} \): north of \( y = Y_0 \) we impose an eastward surface stress and surface heating, and south of \( y = Y_0 \) we impose westward surface stress and surface cooling\(^1\), as illustrated in Fig. 2. Specifically, we prescribe the

\(^1\)In our model the “ASF” latitude band is required to produce AABW that sinks to the deep ocean, but will otherwise not be discussed in this paper. The interested reader is referred to Stewart and Thompson (2015a,b, 2016) for further exploration of the shelf/slope dynamics.
stress as “ASF” latitude band. Similarly, we prescribe the surface “ACC” latitude band and \( L \) below. We define \( y < L \) nor in the shelf cooling region \( (\theta = 0) \) rest- ed to an exponential stratification that ranges from boundary of the domain the temperature is continuously where \( T \) and not.

\[ Q(y) = \begin{cases} Q_e \sin \left( \frac{\pi (y - Y_0)}{L_e} \right), & Y_0 \leq y \leq L_y - L_r, \\ -Q_w \sin \left( \frac{\pi (y - Y_s)}{L_w} \right), & L_c \leq y \leq Y_0, \end{cases} \]  

(14)

where \( Q_e = 10 \text{ W m}^{-2} \) is the maximum surface heating over the ACC and \( Q_w = 5 \text{ W m}^{-2} \) is the maximum surface cooling over the ASF. Note that no surface energy flux is over the ACC and \( Q \) below. We define \( L_y = L_y - L_c - Y_0 \) as the width of the “ACC” latitude band and \( L_w = Y_0 - L_c \) as the width of the “ASF” latitude band. Similarly, we prescribe the surface stress as

\[ \tau(x)(y) = \begin{cases} \tau_e \sin^2 \left( \frac{\pi (y - Y_0)}{L_e} \right), & Y_0 \leq y \leq L_y - L_r, \\ -\tau_w \sin^2 \left( \frac{\pi (y - Y_s)}{L_w} \right), & L_c \leq y \leq Y_0, \end{cases} \]  

(15)

where \( \tau_e = 0.15 \text{ N m}^{-2} \) is the maximum eastward wind stress over the ACC and \( \tau_w = 0.05 \text{ N m}^{-2} \) is the maximum westward wind stress over the ASF. The form of (15) ensures that the wind stress is both continuous and differentiable everywhere, thus avoiding discontinuities in the wind stress curl.

In order to produce a supply of AABW from our idealized Antarctic continental shelf, we impose a zonally and meridionally localized cooling of the water column on the continental shelf. Specifically, waters south of \( y = L_c \) are restored toward \( \theta = -2^\circ\text{C} \) with a time scale \( T \) that var- ies spatially following

\[ T(x,y,z) = T_c \frac{H_s}{|z|} \exp \left[ \left( \frac{x}{2L_c} \right)^2 + \left( \frac{y}{2L_c} \right)^2 \right], \quad y \leq L_c, \]  

(16)

where \( T_c = 7\text{ days} \) and \( L_c = 100\text{ km} \). At the northern boundary of the domain the temperature is continuously restored to an exponential stratification that ranges from \( \theta = 0^\circ\text{C} \) at the ocean bed to \( \theta = 12^\circ\text{C} \) at the ocean surface. The inverse relaxation time scale varies linearly from 1 week\(^{-1} \) at the \( y = L_y \) to zero at \( y = L_y - L_r \).

The model domain is discretized with a horizontal grid spacing of approximately 4 km. This provides several gridpoints per first Rossby radius of deformation in the core of the model ACC, where typically \( R_d \approx 15 \text{ km} \). The vertical discretization consists of 70 vertical levels ranging from approximately 100 m at the ocean bed to approximately 10 m at the surface. The near-bed grid spacing was chosen to resolve the vertical structure of the model’s overturning circulation and form stress, discussed further in §§3c and 4. The surface grid spacing was chosen to place several vertical grid points within the surface mixed layer, whose minimum depth of 50 m is imposed via the K-profile parameterization (KPP Large et al. 1994). KPP also supplies enhanced vertical mixing in the ocean interior that enhances the background vertical diffusivity of \( 5 \times 10^{-6} \text{ m}^2 \text{ s}^{-1} \) depending on the bulk Richardson number. We employ the Prather (1986) advection scheme for temperature to mini- mize spurious diabatic mixing due to numerical truncation errors (Hill et al. 2012). Kinetic energy is extracted from the simulated flow via a linear bottom drag with coefficient \( r_b = 1.1 \times 10^{-3} \text{ m s}^{-1} \), and grid-scale energy and enstrophy are controlled primarily using constant, Leith, and Leith-Plus biharmonic viscosities (Fox-Kemper and Menemenlis 2008).

b. Experiments

In this subsection we describe a series of experiments, based on the model configuration presented in §3a, to test our central hypothesis that AABW export has a substan- tial surface signature in the ACC (see §2). A complete list of our experiments and their defining parameter choices is given in Table 2. The constraints outlined at the beginning of this section make our model experiments computationally expensive, even in an idealized configuration, so we target our experiments to address the following question: how would the ACC respond to a complete shutdown of AABW export?

We define three configurations of the model’s momentum and buoyancy fluxes: We define our “Reference” con- figuration to match the parameter choices described in §3a. We then define a “No AABW” configuration in which AABW production is eliminated by removing the surface cooling in

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### Table 2. Simulation parameters varied between our model experiments. For parameter definitions, refer to Table 1.

<table>
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<th>Experiment</th>
<th>( \gamma )</th>
<th>( \tau_e ) (N/m(^2))</th>
<th>( \tau_w ) (N/m(^2))</th>
<th>( Q_s ) (W/m(^2))</th>
<th>( Q_r ) (W/m(^2))</th>
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<td>RIDGE_weakSAM</td>
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<td>0.075</td>
<td>0.05</td>
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<td>7</td>
</tr>
<tr>
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<td>0.05</td>
<td>10</td>
<td>5</td>
<td>7</td>
</tr>
<tr>
<td>BUMP_noAABW</td>
<td>0</td>
<td>0.15</td>
<td>0</td>
<td>10</td>
<td>0</td>
<td>( \infty )</td>
</tr>
<tr>
<td>BUMP_weakSAM</td>
<td>0</td>
<td>0.075</td>
<td>0.05</td>
<td>10</td>
<td>5</td>
<td>7</td>
</tr>
</tbody>
</table>
the “ASF” latitude band ($Q_w = 0$) and the cooling of the water column on the continental shelf ($T_c = \infty$). In concert with these changes in the buoyancy forcing, we found it necessary to additionally reduce the westward wind stress in the “ASF” latitude band to zero ($\tau_w = 0$). Retaining a non-zero westward wind stress led to intermittent, highly energetic baroclinic instabilities in the westward slope current that prevented the model from reaching a statistically steady state.

We contrast the impact of shutting off AABW formation with the impact of halving the surface wind stress ($\tau_e = 0.075 \text{ N m}^{-2}$), which we refer to as the “Weak SAM” case, referring to the close connection between the Southern Annular Mode (SAM) and the strength of the westerly winds (Langlais et al. 2015). Our conceptual model in §2 suggests that either reducing the eastward wind stress or eliminating AABW transport should lead to a shoaling of the SSH in the ACC’s standing meanders. The surface energy and momentum fluxes corresponding to our “Reference”, “No AABW” and “Weak SAM” configurations are shown in Fig. 2(a–b).

In §4 and §5 we investigate the impact of AABW production on the ACC circulation using two different topographic configurations, illustrated in Fig. 3(a–b), that qualitatively resemble those used in previous idealized eddy-resolving channel models of the ACC (e.g. Tréguier and McWilliams 1990; Bischoff and Thompson 2014; Abernathey and Cessi 2014; Chapman et al. 2015; Howard et al. 2015). The RIDGE topography consists of a 2000m-deep Gaussian ridge that spans the entire latitudinal extent of the model domain, while the BUMP topography consists of an isolated Gaussian bump centered beneath the eastward wind stress maximum. A key distinction between these topographies is that RIDGE does not admit any zonally re-entrant $f/|\eta_b|$ contours in the ACC, whereas in the BUMP case the $f/|\eta_b|$ contours are simply deflected northward around the topography (Nadeau and Ferrari 2015). The RIDGE topography might therefore be compared to Drake Passage in the real ACC, and the BUMP topography to Kerguelen Plateau.

Though the BUMP case yields arguably a less realistic representation of the ACC’s $f/|\eta_b|$ contours, which are blocked in a circumpolar sense (Nadeau and Ferrari 2015), we will show in §5 that its eddy field is much more sensitive to AABW formation because the eddy-driven component of the overturning circulation is as important as the mean in this case (see §3c). Diagnostics from the Southern Ocean State Estimate (Mazloff et al. 2013) and high-resolution coupled climate model runs (Bishop et al. 2016) indicate that this partitioning of the mean and eddy contributions to AABW export is more realistic. Furthermore, the BUMP case allows us to test the hypothesis set out in §2 in a qualitatively different model geometry; in §4 we show that the surface structure of the ACC exhibits a similar response to changes in AABW export despite the presence of zonally re-entrant $f/|\eta_b|$ contours.

Fig. 3. A visualization of the Antarctic Circumpolar Current (ACC) bathymetries used in our model experiments. (a) RIDGE, consisting of a ridge spanning the entire latitudinal extent of the domain. (b) BUMP, consisting of a single Gaussian bump in the center of the ACC. Each panel also visualizes an instantaneous snapshot from the model output, showing surfaces where the relative vorticity (normalized by the absolute Coriolis parameter) is equal to 0.1 (red) and –0.1 (blue). The black contours correspond to 20 year-averaged sea surface heights (SSH) of (a) {−0.1 m,−0.2 m,−0.3 m} and (b) {−0.2 m,−0.4 m,−0.6 m}, relative to the zonal-mean SSH at the northern boundary.
More precisely, we prescribe the model bathymetry $\eta_b(x,y)$ as a sum of “shelf” and “ridge” components,

$$\eta_b(x,y) = -H + \max(\eta_{\text{shelf}}(y), \eta_{\text{ridge}}(x,y)) .$$  \hfill (17)

Here the shelf component $\eta_{\text{shelf}}$ describes the continental shelf and slope,

$$\eta_{\text{shelf}}(y) = \frac{1}{2}(H - H_s) \left[ 1 - \tanh \left( \frac{y - Y_s}{W_s} \right) \right] ,$$  \hfill (18)

where $H_s = 500$ m is the depth of the continental shelf, $Y_s = 300$ km is the latitudinal position of the center of the continental slope, and $W_s = 75$ km is the half-width of the continental slope. This produces a continental slope with a maximum steepness of approximately 0.02, typical for the Antarctic margins, and ensures that the ocean bed is essentially flat in the ACC portion of the model domain ($y > 500$ km), as is visible in Fig. 2. The ridge component $\eta_{\text{ridge}}$ describes a topographic obstacle placed in the path of the ACC,

$$\eta_{\text{ridge}}(x,y) = H_b \exp \left[ - \left( \frac{x - X_b}{W_b} \right)^2 \right] \times \exp \left[ - (1 - \gamma) \left( \frac{y - Y_b}{L_b} \right)^2 \right] .$$  \hfill (19)

Here $\gamma \in \{0, 1\}$ is a switch: setting $\gamma = 1$ creates a continuous meridional ridge across the ACC, corresponding to our RIDGE topography in Fig. 3(a), while setting $\gamma = 0$ creates an isolated Gaussian seamount, corresponding to our BUMP topography in Fig. 3(b). All of the experiments described here are conducted as follows: the model is initialized from rest using a horizontal grid spacing of 8 km, and then integrated for $\sim 100$–150 years until a statistically steady state is reached. The solution is then interpolated onto a 4 km grid and the integration continued until the statistically steady state is re-established. In each case the model is judged to have reached a statistically steady state based on time series of the total kinetic energy, eddy kinetic energy and domain-averaged temperature. All results reported below are diagnosed from the final 20 years of model integration at 4 km horizontal resolution, during which the drift in the domain-averaged temperature was always smaller than 0.1°C/century.

c. Overturning circulation

As the export of AABW is a control parameter in our model experiments, we now verify that our experimental design allows us to independently activate/deactivate AABW production and vary the strength of the eastward wind stress over the ACC. In Fig. 4 we compare the overturning streamfunctions diagnosed from our Reference, No AABW and Weak SAM experiments. For each experiment we diagnose the overturning streamfunction $\psi$ by computing meridional volume fluxes between a series of discrete temperature surfaces, and then mapping those fluxes back to latitude/depth space via

$$\psi(y,z) = \int_{z^t = m}^{z^t = 0} \mathcal{H} \left( \theta(x, y, z^t, t) - \mathbf{\overline{\theta}}^{x,t}(y, z) \right) v dz^t .$$  \hfill (20)

Here an overline $\overline{\mathbf{x}}^{x,t}$ denotes a zonal and time average, and $\mathbf{\overline{\theta}}$ and $\mathbf{\overline{\theta}}$ are variables of integration. The remapping of the streamfunction from $(y, \theta)$ space to $(y, z)$ space approximately indicates the depths at which isopycnal volume fluxes occur, with the drawback that the streamfunction appears not to vanish at the top and bottom boundaries.

Fig. 4(a–b) shows that both the RIDGE_reference and BUMP_reference experiments exhibit similar two-cell overturning streamfunctions. Each has a relatively shallow clockwise upper cell with northward surface transport supported by our imposed heating, overlying a much more voluminous counter-clockwise cell associated with the export of AABW. In Table 3 we quantify the strengths of the upper and overturning cells in all of our model experiments.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Upper MOC strength (Sv)</th>
<th>Lower MOC strength (Sv)</th>
<th>ACC transport (Sv)</th>
<th>Northern gyre strength (Sv)</th>
<th>Surface-induced form stress $F_{\text{surf}}$ (TN)</th>
<th>Total EKE (PJ)</th>
<th>Baroclinic EKE production (MJ/s)</th>
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<td>0.68</td>
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<td>4.93</td>
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<td>31.6</td>
<td>9.74</td>
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<tr>
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<td>0.01</td>
<td>101.5</td>
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<tr>
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<td>0.63</td>
<td>98.3</td>
<td>1.3</td>
<td>0.17</td>
<td>23.9</td>
<td>6.85</td>
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</table>

Table 3. Scalar measures of the mean and transient circulation diagnosed from our model experiments. Upper MOC strength and Lower MOC strength are calculated as outlined in §3c. ACC transport and Northern gyre strength are calculated as outlined in §4a. Total EKE, Baroclinic EKE production and Barotropic EKE production are calculated as outlined in §5.
Fig. 4. A comparison of the meridional overturning circulation, calculated from fluxes between temperature surfaces and mapped back to latitude/depth space, between our model experiments. (a,b) Overturning streamfunction (colors) and zonal-/time-mean temperature surfaces (contours) for our RIDGE_reference and BUMP_reference experiments respectively. (c–f) Difference in the overturning streamfunction between the Reference experiments and the RIDGE_noAABW, BUMP_noAABW, RIDGE_weakSAM and BUMP_weakSAM experiments respectively (colors), and the absolute zonal-/time-mean potential temperature surfaces for each experiment (contours).

following Stewart and Thompson (2012). Figure 4(c–d) shows the difference between the Reference and No AABW cases, and can therefore be interpreted as the component of the Reference case associated with AABW export. Here, AABW export is entirely responsible for the lower cell, as expected, and the presence of all waters below 0°C. Perhaps more surprisingly, the addition of AABW leads to a slight weakening (5–10%) of the upper overturning cell (see Table 3), along with a substantial shoaling of around 500 m at the northern edge of the domain. In contrast, the eastward wind stress has a negligible impact on the strength of the deep overturning cell. Comparison of the Reference
cases with WeakSAM cases indicates a substantial (∼25%) decrease in the strength of the upper cell when the westerly wind stress is halved (Fig. 4e–f), consistent with previous analyses (e.g. Abernathey et al. 2011; Munday et al. 2013). This weakening is accompanied by a shoaling of the cell by around 100 m at the northern boundary.

A notable difference between our topographic configurations is that AABW is exported at shallower depths and warmer temperatures in RIDGE_reference than in BUMP_reference, with maxima of ψ around θ = 0.4°C and θ = 0°C respectively. This difference may be understood by decomposing the overturning streamfunction ψ into its “mean” and “eddy” components. We define the mean component $\psi_{\text{mean}}$ based on the volume fluxes carried by the time-mean meridional velocity $v_t$ between time-mean temperature surfaces $\theta_t$, and the eddy component $\psi_{\text{eddy}}$ as the residual between $\psi$ and $\psi_{\text{mean}}$ (see e.g. Wolfe and Cessi 2009; Stewart and Thompson 2015b),

$$
\psi_{\text{mean}}(y, z) = \int_{z'=0}^{z'} \mathcal{H} \left( \overline{\vec{v}}(x, y, z') - \overline{\vec{v}}^t(y, z') \right) \sigma^t \, dz',
$$

(21a)

$$
\psi_{\text{eddy}} = \psi - \psi_{\text{mean}}.
$$

(21b)

Here the superscripts $x$ and $t$ denote zonal and temporal averages, respectively. Note that this definition of $\psi_{\text{mean}}$ only approximates the time-mean velocities within density surfaces, and is exact only in the limit of vanishingly small deviations of the velocity and temperature from their respective means (McDougall and McIntosh 2001). For example, this formulation may misdiagnose intermittent outflows of water along the ocean bed, which are denser than the densest waters present in the time-mean model state, as a purely “eddy” component of the circulation. However, this does not substantially alter our results because AABW is exported in a layer several hundred meters thick (see Fig. 4).

Fig. 5 shows that our model’s upper cell is largely a
mean circulation with relatively modest opposition from transient eddies. This is consistent with the overturning circulation in the Southern Ocean State Estimate (SOSE, Mazloff et al. 2013): the upper cell is largely supported by a combination of standing meanders and wind-driven Ekman transport (not shown). The deep cell is also largely supported by mean flows in both the RIDGE and BUMP cases, but in this case eddies play a key role in transporting AABW northward across the core of the ACC (around $y = 1200$ km). In the BUMP case the AABW transport across this latitude is effected entirely by eddies, whereas in the RIDGE case they account for around 50% of the transport. In the BUMP case the eddies also necessarily become important close to the northern and southern boundaries of the ACC ($y = 1900$ km and $y = 500$ km respectively), where the height of the topographic bump vanishes. The latter requires dense water to accumulate around $y = 500$ km, visible in the mean temperature surfaces in Fig. 5(b) and (d), leading to a northward eddy thickness flux of AABW (e.g. Marshall and Speer 2012). In contrast, the extension of the RIDGE topography to the southern boundary permits a more efficient exchange of waters across the lower half of the continental slope, resulting in warmer temperatures on the continental shelf and warmer AABW.

4. Response of the ACC to AABW export

a. Barotropic circulation

In this section we examine the response of the mean circulation of our idealized ACC, and the surface signature of this response, to a shutdown of AABW export. In Fig. 6 we visualize the ACC’s zonal/meridional transport pathways via the barotropic streamfunction, defined for our purposes as

$$\Psi_{BT}(x, y) = \int_y^{L_y} \int_{\eta_n}^{0} \int_{\eta_b}^{0} dz \pi^t, \hspace{1cm} (22)$$

where $y'$ is a variable of integration. Fig. 6(a–b) shows that the ACC undergoes a northward excursion in order to pass the RIDGE or BUMP topography, consistent with previous studies (e.g. Bischoff and Thompson 2014; Abernathey and Cessi 2014; Thompson and Naveira Garabato 2014). In the lee of the topography two gyre-like recirculatory features abut the central jet of the ACC. Nadeau and Ferrari (2015) studied the development of these features in quasi-geostrophic and primitive equation simulations, and proposed a conceptualization of the ACC circulation as a superposition of “circumpolar mode” and a “gyre mode”. The key difference between the barotropic circulations in our two Reference simulations is that the circumpolar mode is weaker and the gyre mode stronger in the RIDGE case than in the BUMP case.

In Table 3 we quantify the ACC transport and the strength of the northern gyre for all of our model experiments. The ACC transport is calculated by computing contours of $\Psi_{BT}$ at 0.1 Sv intervals, and finding the contour with the largest value of $\Psi_{BT}$ that (i) is circumpolar and (ii) lies north of the center of the continental slope, $y = Y_s$, at $x = -1000$ km, where the ACC is typically at its widest. Condition (ii) excludes westward transport along the continental slope and eastward transport on the continental shelf at the southern edge of the domain. The northern gyre strength is simply computed by finding the domain-minimum value of $\Psi_{BT}$. In Table 3 we do not quantify the strength of the southern gyre because in several of the experiments (particularly those with halved eastward wind stress) it is very weak, and difficult to distinguish objectively from other local recirculations around the ACC topography and the continental shelf/slope.

The ACC response to halving of the eastward wind stress in this model is largely consistent with the findings of Nadeau and Ferrari (2015): in the RIDGE case the absence of zonally re-entrant $f/|\eta|$ contours leads to an almost perfectly eddy-saturated ACC (c.f. Munday et al. 2013). Instead, there is a substantial reduction in the strength of the northern gyre circulation from 24.8 Sv to 8.2 Sv. In contrast, halving the eastward wind stress in the BUMP case leads to a 20% reduction in the ACC transport. In this case there is also a large fractional reduction in the strength of the northern gyre circulation (5.6 Sv to 1.3 Sv), but this change is much smaller in absolute terms than in the RIDGE case. These responses are visualized in Fig. 6(e–f), which shows the difference in $\Psi_{BT}$ between our Reference and Weak SAM simulations.

Whereas the sensitivity of the ACC’s barotropic circulation to eastward wind stress depends strongly on topography, its response to AABW export is qualitatively similar in the RIDGE and BUMP cases. Eliminating AABW export reduces the ACC transport by 16% and 17% in our RIDGE and BUMP configurations respectively, which is comparable to the effect of halving the eastward wind stress in BUMP case. Eliminating AABW formation also strengthens the northern gyre, particularly in the BUMP case where the gyre strength more than doubles from 5.6 Sv in BUMP reference to 12.5 Sv in BUMP_noAABW. Fig. 6(c–d) shows that the change in $\Psi_{BT}$ due to the removal of AABW is concentrated over the topography in the ACC, consistent with the increase in circumpolar transport and the tendency of the ACC to be deflected northward around the topography (Fig. 6(a–b)).

From a detectability perspective, changes in the barotropic circulation of the ACC are less relevant than changes in the surface circulation. In Fig. 7 we reproduce Fig. 6 with $\Psi_{BT}$ replaced by SSH, which is proportional to the surface geostrophic streamfunction. The surface circulations in our Reference experiments, and the differences between those and our No AABW and Weak SAM experiments, are qualitatively similar to the barotropic circulations. This is to be
Fig. 6. A comparison of the barotropic streamfunction $\Psi_{BT}$ between our model experiments. (a,b) Barotropic streamfunction (colors and black contours) for our RIDGE_reference and BUMP_reference experiments respectively. White contours indicate isobaths between depths of 500 m and 3500 m at 500 m intervals. (c–f) Difference in the barotropic streamfunction between the Reference experiments and the RIDGE_noAABW, BUMP_noAABW, RIDGE_weakSAM and BUMP_weakSAM experiments respectively (colors and black contours).
Fig. 7. As Fig. 6, but showing SSH rather than barotropic streamfunction.
Fig. 8. Time-mean surface speed $|\bar{u}|$ (colors) in (a) RIDGE_reference, (b) BUMP_reference, (c) RIDGE_noAABW, (d) BUMP_noAABW, (e) RIDGE_weakSAM and (f) BUMP_weakSAM. White contours indicate isobaths between depths of 500 m and 3500 m at 500 m intervals.
expected based on the equivalent-barotropic vertical structure of the ACC (Killworth 1992). Fig. 7(c–d) shows that the ACC’s response to the introduction of a northward-flowing AABW layer is to develop a depression in the SSH of the standing meander, as postulated in Fig. 1 and discussed in our conceptual model in §2. This surface signature of AABW export in the ACC is clearly distinguished from the signature of stronger eastward wind stress in the RIDGE case; the latter produces an SSH signal primarily in the gyre-like circulations downstream of the ridge. In contrast, the surface response to changes in AABW export and changes in westerly wind stress bear qualitative resemblance to one another the BUMP case.

Another view of the surface circulation is presented in Fig. 8, which shows the model surface speed. This is approximately proportional to the absolute gradient of the SSH, and so is sensitive to slight re-arrangements of SSH contours. In the RIDGE case, weakening the westerlies produces little change in the surface speed around the meander, consistent with the insensitivity of the circumpolar transport to wind stress in this case (see Table 3). Removing AABW reduces the surface speed by almost a factor of two, associated with a weakening of the standing eddies over the topography, visible in Figs. 6 and 7. In the BUMP case, weakening the westerlies reduces the barotropic component of the circumpolar transport (Nadeau and Ferrari 2015) and reduces the surface speed in the meander by a factor of around 1/3. In contrast to the RIDGE case, removing AABW in the BUMP case produces almost no discernible change in the surface speed around the standing meander; the most prominent change is an increase in the surface speed downstream of the meander, associated with the strengthening of the northern gyre circulation (see Table 3).

b. Role of topographic form stress

In §2 we advanced a conceptual model for the ACC’s response to AABW export based on consideration of the topographic form stress. We now examine the topographic form stress in our idealized model experiments to explain the responses diagnosed in §4a.

We first verify that the surface wind stress is balanced locally by topographic form stress, as assumed in our conceptual model equation (1). In Fig. 9, we plot the contributions of terms in the zonal momentum budget

\[
\begin{align*}
\oint dx \tau^{(z)} & - \int_0^z \int_{\eta_b} d\eta \int_0^x dx' \frac{\partial \tau}{\partial x'} - \int_0^x dx' \tau_{bt}^{(z)} \\
\frac{\partial}{\partial y} \int_0^x dx' \int_{\eta_b} d\eta' \rho \nabla^t \tau^{(z)} & - \frac{\partial}{\partial y} \int_0^x dx' \int_{\eta_b} d\eta' \rho \nabla^t \tau^{(z)} \approx 0,
\end{align*}
\]

for our RIDGEreference and BUMPreference experiments. Here the primes on \(u'\) and \(v'\) denote deviations of the velocities from their respective time averages. Omitted from (23) are terms associated with horizontal viscosity and discretization errors due to the momentum equations being solved in vector-invariant form (rather than flux form), both of which are negligibly small. Fig. 9(b) shows that with BUMP topography the wind stress is almost perfectly balanced by topographic form stress, with a minor role for bottom drag (Munk and Palmén 1951; Olbers et al. 2004). With RIDGE topography the bottom drag is further diminished, but there is a non-negligible redistribution of the momentum due to mean advection by standing eddies, visible in Fig. 6(a) as closed contours of \(\Psi_{bt}\) around \(x = -250 \text{ km}, y = 1400 \text{ km}\). This redistribution does not invalidate the assumptions made in §2 on the scale of the entire ACC, but may expected to interfere locally with the distribution of the form stress over the topography.

Our conceptual model hinges on the contributions of water masses at different depths to the total topographic form stress \(F_{\text{topog}}^{(z)}\). In order to relate our conceptual model to the diagnosed circulation changes described in §4a, we define an effective pressure difference across topography \(\Delta p\), where

\[
\Delta p(y, z) = \frac{\partial}{\partial z} \int_0^x dx \int_{\eta_b}^z dz' \frac{\partial \tau}{\partial x}. \tag{24}
\]

This measures the effective zonal pressure gradient force exerted on the ocean per unit area in the latitude/depth plane, and has the property that its vertical integral is equal to the topographic form stress,

\[
F_{\text{topog}}^{(z)} = \int_0^x dz \Delta p. \tag{25}
\]

In Fig. 10 we plot \(\Delta p\) as a function of depth, integrated latitudinally over the ACC latitude band of our model \((y > 500 \text{ km})\). There is a very close correspondence between \(\Delta p\) and the meridional velocity averaged at constant depth \(\overline{\nabla^t}\) (not shown, see Warren et al. 1996), but not between \(\Delta p\) and our mean streamfunction \(\psi_{\text{mean}}\) because we perform zonal averaging at constant \(\overline{\nabla^t}\).
Fig. 9. Zonal momentum balance in experiments with (a) RIDGE and (b) BUMP topography, with Reference wind and buoyancy forcing. Quantities identified in the legend correspond to those identified in equation (23).

Fig. 10. Vertical distribution of the topographic form stress in our experiments with (a) RIDGE and (b) BUMP topography, with Reference wind and buoyancy forcing. In each case the topographic form stress has been integrated meridionally from $y = 500$ km to $y = 2000$ km.

In the BUMP case the vertical variation of $\Delta p$ is concentrated close to the ocean bed, and this latitudinally-integrated view yields little insight into the impact of the AABW outflow. We therefore focus our discussion on the RIDGE case shown in Fig. 10(a), which is characterized by two local maxima of $\Delta p$ close to the top and bottom of the topography ($z = 2000$ m and $z = 4000$ m respectively). Halving the eastward wind stress results in a decrease in
the magnitude of $\Delta p$ that is biased toward the deepest parts of the ocean. In contrast, removing the AABW outflow diminishes the maxima of $\Delta p$ close to the top and bottom of the topography, but increases the magnitude of $\Delta p$ at mid-depth (around $z = 3000$ m) such that the overall topographic form stress is unchanged. This response is suggestive of a northward AABW flow around $z = 3000$ m that is compensated by adjustments of lighter density classes and/or the ocean surface overlying isobaths around $z = 2000$ m and $z = 4000$ m, as hypothesized in Fig. 1.

It is not possible to attribute changes in $\Delta p$ to specific changes in the ACC circulation in this latitudinally-integrated view, so in Fig. 11 we map $\Delta p$ as a function of...
latitude and depth. Fig. 11(b) confirms that in the BUMP case, $\Delta p$ is concentrated close to the ocean bed, a result of the zonal asymmetry in the northward- and southward-flowing branches of the standing meander (see Figs. 1(a) and 6(b)). In the RIDGE case the ACC is forced to cross $f/|\eta_b|$ contours, resulting in $\Delta p$ being concentrated close to the top of the topography in the northern half of the ACC (Fig. 11(a)). Fig. 11(e) shows that weakening the eastward wind stress only impacts $\Delta p$ close to the ocean bed, associated with spin-down of the gyre circulations to preserve the zonal force balance (see Fig. 6(e) and Nadeau and Ferrari 2015). In contrast, removing AABW impacts both the deep and shallow contributions to $\Delta p$: Fig. 11(c) reveals that AABW acts to accelerate the ACC as it flows northward at mid-depth ($z \approx 3000$ m). This acceleration is compensated by enhanced shallow and deep negative contributions to $\Delta p$ that are created by preferentially depressing the SSH on the eastern side of the topography (see Fig. 7(c)). The same compensation is likely taking place in the BUMP experiments, but is not visible in Fig. 11(d) because the AABW-induced eastward acceleration and its compensation by the ACC both take place over isobaths close to $z = 4000$ m.

The diagnosed changes in the sea surface height, circulation and topographic form stress distribution support the connection between topographic form stress and the surface structure of the ACC posited in §2. However, the explicit prediction of our theory, which directly relates changes in SSH to changes in topographic form stress, is that the “surface-induced” form stress $F_{\text{surf}}$ should vary with changes in AABW export. Given that $F_{\text{surf}} \approx -\rho_0 f T_0$, equation (11) suggests that $F_{\text{surf}}$ should scale as the sum of the northward AABW and Ekman transports,

$$F_{\text{surf}} \approx \rho_0 f (T_A + T_{\text{Ek}}). \quad (26)$$

In Table 3 we list the computed $F_{\text{surf}}$ for each of our RIDGE and BUMP experiments. In the RIDGE case shutting off AABW export leads to an almost threefold reduction in $F_{\text{surf}}$, consistent with the changes in SSH over topography shown in Fig. 7(c), whereas halving the westerly wind stress has almost no impact on $F_{\text{surf}}$ because the changes in SSH in Fig. 7(e) occur predominantly in the gyre-like circulations downstream of the topography. In the BUMP case both shutting off AABW and halving the westerly wind stress lead to substantial reductions in $F_{\text{surf}}$, consistent with the qualitatively similar changes in SSH shown in Fig. 7(d,f).

As a further test of this connection, we performed a series of additional sensitivity experiments in which the AABW volume transport was varied over a larger range, compared to the transports summarized in Table 3. In these sensitivity we used a lower horizontal resolution ($\Delta x = 8.3$ km and $\Delta y = 8.0$ km) for computational efficiency. We removed the easterly winds ($\tau_w = 0$ N m$^{-2}$), the surface heat flux over the continental slope ($Q_s = 0$ W m$^{-2}$) and the temperature relaxation on the continental shelf ($T_{1r}^{-1} = 0$ days$^{-1}$), and instead applied a uniform heat loss of $Q_0 = \{0, 50, 100, 150\}$ W m$^{-2}$ over the southernmost 100 km of
the ocean surface. In Fig. 12(a) we plot $F_{\text{surf}}$ as a function of the mean overturning circulation in the abyss, computed as described in §3c. In each case $F_{\text{surf}}$ is calculated in the same way as $F_{\text{topog}}$, but using the surface pressure mapped down to the ocean bed, and is integrated latitudinally over the ACC sector of the domain (500 km $< y < 2000$ km). We have also included points corresponding to the simulations listed in Table 3, but using the same 8 km horizontal grid spacing to allow a fair comparison with our sensitivity runs. Fig. 12(a) demonstrates that systematically varying $Q_0$ leads to an almost-linear change in $T_A$, but the response of $F_{\text{surf}}$ is not as linear as (26) suggests.

5. Impact of AABW on ACC eddy generation

Given the close relationship between the mean circulation of the ACC and its mesoscale eddy field (e.g., Tréguier and McWilliams 1990; Ward and Hogg 2011), one might expect the AABW-induced response discussed in §4 to be accompanied by a substantial change in the EKE. This expectation offers an additional avenue for detecting changes in EKE downstream of the topography that resembles the shape of the topography. With BUMP topography, upstream of the EKE maximum. In the BUMP case, both weakening the westerlies and removing AABW have qualitatively similar impacts on the EKE generation, suppressing baroclinic instability in the lee of the topography. In our experiments the impact of removing AABW is actually more pronounced than weakening the westerlies, reducing the total baroclinic EKE production by 37% and 30% respectively (see Table 3). With RIDGE topography, there is a similar response in the Weak SAM case, but not in the noAABW case. Instead, removing AABW leads to large positive and negative anomalies over the topography, associated with changes in the strength and variability of the standing eddies there (see Fig. 6). These anomalies largely cancel one another, yielding only a 16% reduction in EKE production overall.
Fig. 13. A comparison of the depth-averaged eddy kinetic energy (EKE) between our model experiments. (a,b) EKE (colors) in our RIDGE_reference and BUMP_reference experiments respectively. (c–f) Difference in the EKE between the RIDGE_noAABW, BUMP_noAABW, RIDGE_weakSAM and BUMP_weakSAM experiments and their respective Reference experiments (colors and black contours). Contours indicate isobaths between depths of 500 m and 3500 m at 500 m intervals.
Fig. 14. As Fig. 13, but showing depth-integrated baroclinic production of eddy kinetic energy $\text{PE} \rightarrow \text{EKE}$. Note that we have used a nonlinear colorbar to visualize both localized peaks and domain-wide changes in EKE production/destruction.
In our model the EKE is largely sourced from the surface wind forcing, which supplies available potential energy (APE) by shoaling isopycnals to the south, and from the imposed buoyancy forcing. To elucidate the differing responses of our experiments to a shutdown of AABW export, we now jointly consider the production of PE from MKE and its subsequent baroclinic conversion to EKE. This energy pathway can be more accurately isolated by constructing a budget for APE (see Hogg et al. 2017), but for our purposes a PE budget is sufficient, and amounts to a decomposition of the vertical buoyancy flux. We define the conversion from PE to MKE as

$$\text{PE} \rightarrow \text{MKE} = \mathbf{w}^{\prime} \mathbf{b}^{\prime}$$

$$= \mathbf{w}^{\prime} \mathbf{b}^{\prime} \mathbf{t} + \mathbf{w}^{\prime} \mathbf{b}^{\prime} \mathbf{t} + \mathbf{w}^{\prime} \mathbf{b}^{\prime} \mathbf{t} \mathbf{t} \mathbf{t},$$

where daggers $\dagger$ indicate deviations from the zonal mean, e.g. $w^{\dagger} = w^{\prime} - \mathbf{w}^{\prime} \mathbf{t}$. Here we have further decomposed the mean buoyancy flux into a “zonal mean” component associated with the time-/zonal-mean vertical velocity and temperature ($\mathbf{w}^{\prime} \mathbf{b}^{\prime} \mathbf{t}$). A “standing eddy” component associated with zonal variations of the time-mean vertical velocity and buoyancy, and “cross terms” that vanish in a zonal integral. The standing eddy component quantifies how much of the mean buoyancy flux is achieved by the Eulerian-mean overturning circulation, i.e. that set by the surface Ekman transport and the geostrophic flows associated with the topographic form stress (see Fig. 11).

In Fig. 15 we plot the depth-dependencies of the zonal mean, standing eddy and transient eddy components of the vertical buoyancy flux in our RIDGE and BUMP simulations.
tions, integrated over the ACC latitude band ($y \gtrsim 500$ km$^2$). Despite the substantial deviation of the mean overturning streamfunction from the wind-driven Ekman transport (see Fig. 5), the mean vertical buoyancy flux is almost entirely achieved by the Eulerian-mean circulation. Standing eddies play a negligible role in the vertical buoyancy transport, while the upward buoyancy flux by transient eddies approximately balances the downward buoyancy flux associated with the Eulerian-mean flow. The total buoyancy flux is a relatively small residual of these terms, and is directed downward (upward) over the depths spanned by the upper (lower) overturning cell, respectively.

Fig. 15(a,e) shows that halving the eastward wind stress leads to a $\sim 50\%$ reduction in the mean buoyancy flux in both the RIDGE and BUMP cases. This result is consistent with a halving of the strength of the wind-driven Eulerian-mean circulation. To emphasize this point, we plot additional zonal mean buoyancy flux curves (dashed lines) constructed using $\vec{b}^{x,t}$ from our Reference simulations and $\vec{b}^{x,t}$ from the Weak SAM simulations. These curves closely follow the zonal mean buoyancy flux curves diagnosed solely from the Weak SAM simulations. In contrast, the change in the mean buoyancy flux between our Reference and No AABW simulations is almost entirely due to changes in the time-/zonal-mean buoyancy. We show additional curves in Fig. 15(a,e) that were constructed by taking $\vec{b}^{x,t}$ from our Reference simulations and $\vec{b}^{x,t}$ from our No AABW simulations. The resulting zonal mean buoyancy flux matches almost perfectly with that diagnosed solely from the No AABW simulations.

Fig. 15 allows us to interpret the change in EKE between our simulations as follows: The decrease in EKE production (i.e. $\vec{w}\vec{b}/T^{1/2}$) between our Reference and Weak SAM simulations occurs due to the halving of the strength of the wind-driven Ekman vertical velocity, which approximately halves the PE production, and therefore must be compensated by an approximate halving of the EKE production to preserve the total vertical buoyancy flux. Between our No AABW and Reference simulations the total upward buoyancy flux increases due to the addition of the lower cell of the MOC (see Fig. 4). In both the RIDGE and BUMP cases, this upward buoyancy flux is primarily supported by an increase EKE production, especially below $\sim 1000$ m. The much larger total vertical buoyancy flux, and thus change in EKE production, in BUMP-reference is commensurate with the more steeply sloped isopycnals in this simulation (see Fig. 4(a–b)). This is equivalent to the statement that the northward flux of AABW is more dependent on eddies in the BUMP case, whereas in the RIDGE case the topographically-supported mean flows can achieve an end-to-end latitudinal export of AABW, and thus the EKE is more sensitive to changes in AABW export with BUMP topography.

In support of this interpretation, in Fig. 12 we plot the domain-integrated EKE against the strength of the eddy streamfunction for all of our model experiments, including the additional sensitivity experiments discussed in §4b. The eddy MOC is quantified using the largest absolute value of $\psi_{\text{edd}}$ in the ACC sector of the model domain ($500$ km $y < 2000$ km). These quantities almost collapse to a line, having a Pearson’s correlation coefficient of $r^2 = 0.56$ and being statistically significant at the 5% level. This demonstrates that the EKE simply responds to changes in the eddy-driven component of the overturning circulation, which may change to compensate a reduced wind-driven mean overturning circulation (in the weak SAM cases) or to compensate a reduction in eddy-driven AABW export (in the noAABW cases). The latter is much less pronounced with RIDGE topography because AABW is almost entirely exported by the mean flow.

6. Discussion

This work was motivated by the prospect that changes in AABW export might produce changes in the surface circulation of the ACC via its impact on the topographic form stress, which is the largest sink of the ACC’s zonal momentum (Munk and Palmén 1951; Olbers et al. 2004). Our experiments using an idealized channel model of the ACC (see §3) support this hypothesis. In this section we review our key results, discuss their relevance to observing the real ACC, and provide an outlook for future work in this area.

Turning off AABW production reduces our model’s ACC transport by 15–20% and raises the SSH above the ACC’s topography by 10–20 cm. This result is consistent with the two-layer conceptual model developed in §2, in which the change in topographic form stress induced by the northward-flowing AABW layer is compensated by deepening of the SSH in the meander. More generally the surface response need not take the form of a depression/elevation of the SSH above the topography; the formal prediction of the theory is that changes in AABW export $T_A$ and wind-driven Ekman transport $T_{\text{Ek}}$ should each lead to a linear change in the “surface-induced” contribution to the topographic form stress $F_{\text{surf}}$, which we defined in equation (13). In §4b we

\footnote{For this calculation we extended the latitudinal extent of the integration southward by up to 100 km at each depth, such that the southern boundary coincided with a zero of the Eulerian-mean overturning streamfunction,}

$$\psi_{\text{EM}}(y,z) = \int_z^0 dz' \oint dz \vec{b}^{x,t}.$$  

This is necessary because our model is invariant under the addition of a constant offset to the buoyancy $b$ (e.g. by measuring temperature $\theta$ in Kelvin instead of Celsius). Increasing the latitudinal extent of our horizontal integration ensures that the zonal mean component of the vertical buoyancy flux, $\vec{b}^{x,t}/T^{1/2}$, is also invariant under the addition of a constant offset to the buoyancy.
showed that systematically varying the AABW export flux alone leads to a monotonic but non-linear increase in $F_{surf}$.

These results indicate that the two-layer model depicted in Fig. 1(c) approximately captures our idealized model’s ACC response to changes in AABW export. In hindsight the alternative adjustment of the ACC depicted in Fig. 1(d) appears less plausible because our idealized model’s MOC is largely accomplished by mean flows that have a strong surface expression (see Figs. 5 and 6). This is particularly true of the southern half of our model ACC, where the upper MOC is absent and the lower MOC occupies the entire depth of the water column, and a substantial fraction of the mean southward transport occurs at the surface.

In §5 we showed that the ACC’s eddy field may also undergo substantial modifications in response to changes in AABW export. However, whereas the response of the mean circulation and SSH to a shutdown of AABW export is consistent between our RIDGE and BUMP configurations, the response of the EKE varies widely with the ACC topography. In §5 we explored this difference via consideration of the model’s energy balance, and showed that the EKE response ultimately depends on the extent to which the northward AABW export is effected by eddies vs. mean flows.

Part of the motivation for this study stems from the prospect that changes in AABW outflow (e.g. Purkey and Johnson 2012; Desbruyères et al. 2016) may be detectable by monitoring changes in the surface circulation of the ACC. Specifically, we have shown that changes in AABW outflow may produce:

(i) Large-scale changes in SSH over topographic features in the ACC.

(ii) Changes in SSH variance, corresponding to changes in EKE, downstream of topography in the ACC.

(iii) Changes in mean SSH gradients, associated with changes in geostrophic surface speed, in the jets that circumnavigate major topographic obstacles in the ACC.

Of these responses, (i) appears to be the least sensitive to the shape of the sea floor, and so may offer the strongest prospect for a monitoring strategy based on satellite altimetry. Though all of the results presented in §4–5 were calculated using 20-year time averages, we found that our results pertaining to (i) were almost identical when recalculated using averaging windows as short as 2.5 years. This suggests that monitoring AABW export via SSH changes over sub-decadal time scales could be a realistic prospect.

The agreement between our scaling arguments in §2 and our model experiments is encouraging. However, it remains to be determined how (i)–(iii) might respond to changes in AABW export across the real ACC, and whether these responses can be distinguished from changes in surface wind stress or the stratification on the northern flank of the ACC. Our model experiments are too heavily idealized to translate our results directly to the real ocean; eddy-resolving simulations of the whole Southern Ocean will likely be required to develop any kind of observational strategy based on our results. A potentially crucial difference between our experiments and the real ACC is that topographic form stress can occur across lateral features like South America (Munday et al. 2015; Masich et al. 2015). It is unclear how this fundamental difference in the geometry will complicate the surface signature of the ACC’s response to changes in AABW export. The complicated geometry of the sea floor may also necessitate an extension of our proposed “surface-induced” topographic form stress $F_{surf}$ to a more general diagnostic analogous to bottom pressure torque, i.e. $\rho g \eta J(\eta_b, \eta_h)$. Our findings should therefore be regarded as a positive first step toward constraining the surface signature of AABW export in the real ACC.

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