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Eddy-mediated transport of warm Circumpolar Deep Water across the Antarctic Shelf Break

Andrew L. Stewart,¹ and Andrew F. Thompson²

The Antarctic Slope Front (ASF) modulates ventilation of the abyssal ocean via the export of dense Antarctic Bottom Water (AABW), and constrains shoreward transport of warm Circumpolar Deep Water toward marine-terminating glaciers. Along certain stretches of the continental shelf, particularly where AABW is exported, density surfaces connect the shelf waters to the mid-depth CDW offshore, offering a pathway for mesoscale eddies to transport CDW directly onto the continental shelf. Using an eddy-resolving process model of the ASF, the authors show that mesoscale eddies can supply a dynamically significant transport of heat and mass across the continental shelf break. The shoreward transport of surface waters is purely wind-driven, whilst the shoreward CDW transport is entirely due to mesoscale eddy transfer. The CDW flux is sensitive to all aspects of the model’s surface forcing and geometry, suggesting that shoreward eddy heat transport may be localized to favorable sections of the continental slope.

1. Introduction

The Antarctic continental shelf is almost completely encircled by the Antarctic Slope Front [ASF, Jacobs, 1991]. The exchange of water masses across the ASF is critical to the global ocean circulation and the stability of the Antarctic ice sheet. Antarctic Bottom Water (AABW) exported from the continental shelves of the Weddell and Ross Seas supplies the deepest branch of the global overturning circulation [Orsi et al., 1999; Lumpkin and Speer, 2007]. AABW fills around 50% of the sub-surface ocean [Gebbie and Hughes, 2011], “ventilating” the abyssal ocean with relatively oxygen-rich waters [Orsi et al., 2001]. This abyssal circulation stores around 30 times as much carbon as the atmosphere, so small changes in AABW export could substantially alter the atmospheric/oceanic partitioning of CO₂ [Skinner et al., 2010]. Meanwhile the transport of relatively warm Circumpolar Deep Water (CDW) onto the continental shelf supplies the heat needed to maintain coastal polynyas and melt the bases of the Antarctic ice shelves [Nicholls et al., 2009; Hattermann et al., 2014]. In the Bellingshausen and Amundsen Seas there is no distinguishable ASF, and CDW floods the local ice shelf cavities. The observed retreat of the region’s marine-terminating glaciers has been attributed to changes in the thickness of this CDW layer [Jacobs et al., 2011; Schimidt et al., 2014].

Several recent studies have highlighted that mesoscale eddies could make an important contribution to exchanges across the Antarctic shelf break. For example, Nast et al. [2011] and Hattermann et al. [2014] showed that eddies carry CDW across the shelf break and beneath Fimbul ice shelf. St-Laurent et al. [2013] showed that the presence of a coastal trough in the Bellingshausen Sea continental shelf can enhance the heat transport to the adjacent ice shelves by accumulating warm anticyclonic eddies from offshore. Nakayama et al. [2014] showed that eddies carry newly-formed AABW across the continental slope from a coastal polynya off Cape Darnley, in East Antarctica. In a more general configuration Stewart and Thompson [2012, 2013] showed that the Antarctic continental slope suppresses mesoscale eddies, resulting in a strong sensitivity of the AABW export to the Antarctic easterly winds. However, the warm, mid-depth CDW layer is absent in these buoyancy-only simulations.

The present study is motivated by stretches of the Antarctic continental shelf over which AABW is formed and exported, particularly the western Weddell Sea [Thompson and Heywood, 2008]. In such regions the export of dense water onto the continental slope modifies the stratification in such a way that density surfaces, or isopycnals, directly connect the shelf waters with the CDW offshore [Nicholls et al., 2009]. This isopycnal connection offers a route for CDW to cross onto the continental shelf via the action of mesoscale eddies, as illustrated in Figure 1(a). Recent glider observations capture this process in the northwest Weddell Sea [Thompson et al., 2014]. In this article we test the hypothesis that these stretches of the ASF support a dynamically important shoreward eddy transport of CDW, and thus heat. We further investigate how the ASF’s structure and cross-slope exchange are shaped by local conditions along various stretches of the Antarctic continental slope. We approach this problem by constructing a high resolution, eddy-resolving process model of the ASF, described in §2. In §3 we explore the sensitivity of the ASF’s cross-slope transport and structure to various aspects of the surface forcing, continental slope geometry, and numerical discretization. In §4 we discuss the implications of our results and the outlook for further research.

2. An eddy-resolving model of the ASF

The configuration of our model study is strongly constrained by the requirement that mesoscale eddy motions be adequately resolved. In §3 we show that a horizontal grid spacing of 1 km or finer is needed to capture the shoreward transport of CDW. It is also important that each simulation reach statistically steady state, as otherwise the results could be strongly dependent on arbitrary choices made in setting the initial conditions [c.f. St-Laurent et al., 2013]. To adequately explore the physical controls on the cross-slope transport it is also necessary to explore a wide range of simulation parameters, which multiplies the computational burden. These constraints motivate the use of an idealized model configuration in a small domain. Yet a realistic representation of the water masses is desirable for the purpose of

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comparison with the real ocean, and because it allows characterization of water mass transformation between neutral density classes [Jackett and McDougall, 1997]. The latter is important because in the ASF there is often no potential density that is stably stratified, due to non-linearities in the equation of state.

Our compromise between these constraints is illustrated in Figure 1(c–d), which shows an isometric view of the instantaneous potential temperature and salinity in our reference simulation. The potential temperature most clearly distinguishes the surface freshwater layer, which we refer to for simplicity as Antarctic Surface Water (AASW), from the warm, salty CDW at mid-depth and the cold, fresh AABW outflow along the ocean bed. The AASW/CDW boundary forms a characteristic ‘V’-shape over the continental slope, which supports a surface-intensified slope current that flows with the shelf break to the left (see Supplementary Figure S1). Eddy stirring of the CDW onto the continental shelf is visible in the potential temperature snapshot. The salinity snapshot highlights the presence of large (50–100 km) mesoscale eddies in the deep ocean and generation of energetic submesoscale eddies over the continental shelf. The eddies are less in evidence over the continental slope, where they are suppressed by the strong topographic vorticity gradient [Isachsen, 2011; Stewart and Thompson, 2013].

We conducted our simulations using the MIT general circulation model [MITgcm, Marshall et al., 1997a, b], which we configured in a 450 km (across-slope) by 400 km (along-slope) by 3000 m (depth) re-entrant channel, as indicated in Figure 1(d). The bathymetry is symmetric in the along-shore direction, while in the across-shore direction it follows a tanh-like profile characterized by a minimum depth $H_{\text{shelf}} = 500$ m, a maximum depth $H_{\text{deep}} = 3000$ m, and a slope of width $W_{\text{slope}} = 150$ km. This profile is plotted in Figure 1(b). The slope is centered at $y = Y_{\text{slope}} = 200$ km, where $y$ is the cross-slope Cartesian coordinate with $y = 0$ at the shoreward boundary. The domain lies on an $f$-plane with Coriolis parameter $f_0 = -1.31 \times 10^{-4}$ rad s$^{-1}$.

Away from the shoreward boundary ($y > 50$ km) we impose a steady, along-shore surface wind stress, oriented such that the continental shelf lies to the left of the winds. The wind stress has a piecewise-sinusoidal profile, vanishing at $y = 50$ km and $y \geq 450$ km. The wind stress maximum of $\tau_{\text{max}} = 0.075$ N m$^{-2}$ is offset from the center of the continental slope by an offshore distance of $L_{\text{wind}} = 25$ km. This profile is plotted in Figure 1(b). To parametrize thermodynamic...
Table 1. Model parameters varied in our sensitivity study. Reference values are used in our reference simulation, shown in Figure 1(c)–(d). Range indicates the variation of each parameter in our sensitivity study, with all other parameters held constant and equal to their Reference values.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Reference</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_{\text{max}}$</td>
<td>Wind stress maximum</td>
<td>0.075 N m$^{-2}$</td>
<td>0–0.1 N m$^{-2}$</td>
</tr>
<tr>
<td>$\Sigma_{\text{polynya}}$</td>
<td>Polynya salt input</td>
<td>$2.5 \times 10^{-3}$ g m$^{-2}$ s$^{-1}$</td>
<td>$1.5-3.5 \times 10^{-3}$ g m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$H_{\text{shelf}}$</td>
<td>Continental shelf depth</td>
<td>500 m</td>
<td>300–700 m</td>
</tr>
<tr>
<td>$W_{\text{slope}}$</td>
<td>Continental slope width</td>
<td>150 km</td>
<td>50–250 km</td>
</tr>
<tr>
<td>$L_{\text{wind}}$</td>
<td>Wind stress max. offset</td>
<td>25 km</td>
<td>−75–125 km</td>
</tr>
<tr>
<td>$\Delta z$</td>
<td>Model grid spacing</td>
<td>1 km</td>
<td>0.5–10 km</td>
</tr>
</tbody>
</table>

3. Sensitivity to surface forcing and continental slope geometry

To investigate the dynamical controls on the cross-slope transports of AASW, CDW and AABW, we conduct a series of sensitivity experiments varying the wind stress maximum $\tau_{\text{max}}$, the polynya salt forcing $\Sigma_{\text{polynya}}$, the shelf depth $H_{\text{shelf}}$, the continental slope width $W_{\text{slope}}$, the wind stress offset $L_{\text{wind}}$, and the grid spacing $\Delta z$. These parameter variations are summarized in Table 1. In each parameter configuration we first integrate the model at 2 km horizontal resolution until the total mean energy and eddy kinetic energy (EKE) reach a statistically steady state, typically requiring around 40 years. We then interpolate the instantaneous model state onto a 1 km grid and continue the integration until the steady state is re-established, typically a further 5–20 years. We derive our results using averages over the final 5 years of model integration.

We find that the cross-slope transport can consistently be partitioned into a shoreward flux of AASW ($F_{\text{AASW}}$), a shoreward flux of CDW ($F_{\text{CDW}}$), and an offshore flux of AABW ($F_{\text{AABW}} = F_{\text{AASW}} + F_{\text{CDW}}$), as illustrated in Figure 1. To diagnose these fluxes we first compute the model’s overturning circulation streamfunction $\psi$, which is most accurately calculated using the fluxes within density surfaces [Doös and Webb, 1994; Stewart and Thompson, 2012]. In general there is no choice of potential density that increases monotonically with depth, whereas neutral density is stably stratified almost everywhere [Jackett and McDougall, 1997]. To avoid the heavy computational burden associated with computing neutral density throughout the domain at frequent time intervals, we use the Transformated Eulerian Mean (TEM) based on the locally-referenced potential density at each depth [McIntosh and McDougall, 1996], which closely approximates the mean isoneutral fluxes computed from the daily-averaged model output (see Supplementary Figure S2). Note that the along-shore periodicity of our model domain makes this calculation possible: it is not possible to derive a closed overturning streamfunction in a regional model due to the open boundaries.

In each case we compute $F_{\text{AABW}}$ following Stewart and Thompson [2012]: we first identify all offshore-flowing streamlines that connect the shelf break ($y = 100$ km) continuously to the edge of the sponge region ($y = 400$ km). Then we define $F_{\text{AABW}}$ as the maximum strength of overturning streamfunction $\psi$ over all such streamlines. This excludes any streamlines that form closed loops in the interior of the domain, and therefore do not contribute to the net export of AABW. We similarly define $F_{\text{CDW}}$ using shoreward-flowing streamlines for salinities greater than $S = 34.543 \text{ kg m}^{-3}$, and finally calculate $F_{\text{AASW}} = F_{\text{AABW}} - F_{\text{CDW}}$. This calculation procedure is illustrated in Supplementary Figure S2. In our reference simulation we obtain $F_{\text{AASW}} = 0.20 \text{ Sv}$, $F_{\text{CDW}} = 0.07 \text{ Sv}$, and $F_{\text{AABW}} = 0.27 \text{ Sv}$. Note that these transports appear small because our domain spans an along-shore distance of only 400 km.
Figure 2. Cross-slope transports of Antarctic Surface Water (\(F_{\text{AASW}}\)), Circumpolar Deep Water (\(F_{\text{CDW}}\)) and Antarctic Bottom Water (\(F_{\text{AABW}}\)), which are sketched in Figure 1 and defined in §3, to surface forcing and bathymetry. The panels show the sensitivity to (a) the wind stress maximum amplitude \(\tau_{\text{max}}\), (b) the brine rejection rate on the continental shelf \(\Sigma_{\text{polynya}}\), (c) the depth of the continental shelf \(H_{\text{shelf}}\), (d) the width of the continental slope \(W_{\text{slope}}\), (e) the offset of the wind stress maximum from the center of the continental slope \(L_{\text{wind}}\), and (f) the model’s horizontal grid spacing \(\Delta x\). Panel (a) also shows the theoretical wind-driven southward surface Ekman transport \(F_{\text{Ekman}} = \tau_{\text{max}} / \rho_0 |f|\), which agrees closely with the shoreward transport of AASW.

The shoreward transport of AASW varies almost linearly with the wind stress maximum \(\tau_{\text{max}}\), but is insensitive to all of the other model parameters varied in this study. In Figure 2(a) we demonstrate that \(F_{\text{AASW}}\) is closely predicted by the theoretical maximum shoreward wind-driven Ekman transport, \(F_{\text{Ekman}} = \tau_{\text{max}} / \rho_0 |f|\). This indicates that the shoreward AASW transport is wind-driven, with diabatic transformations in the surface layer supported by heat and salt exchanges with the sea ice. Our re-entrant domain cannot support a geostrophic shoreward transport, so the shoreward flux of CDW in the ocean interior is accomplished by mesoscale eddy transport [c.f. Marshall and Radko, 2003]. An interior shoreward mean flow can also be supported by eddy momentum flux divergence [e.g. Spall, 2013], but this mechanism makes a negligible contribution to the cross-slope transport in our model. Figure 2(a) shows that \(F_{\text{CDW}}\) decreases almost linearly with \(\tau_{\text{max}}\), in such a way that the AABW export \(F_{\text{AABW}}\) remains approximately unchanged, varying by around 20% over the range of wind stresses considered here. By contrast, in panels (b)–(f) the AABW export is sensitive to all other aspects of the model forcing, geometry and discretization, and that this sensitivity is almost entirely due to changes in the shoreward CDW transport \(F_{\text{CDW}}\): the shoreward AASW transport is essentially fixed by \(\tau_{\text{max}}\). Remarkably, varying any one of the six parameters selected for this sensitivity study can shift the ASF from a regime in which almost no CDW crosses the shelf break to one in which \(F_{\text{CDW}}\) is comparable to \(F_{\text{AASW}}\).

The sensitivity of \(F_{\text{CDW}}\) to each individual parameter may be understood as follows. Increasing \(\tau_{\text{max}}\) steepens the isopycnals in the ASF, deepening the pycnocline at the shelf break until CDW can no longer mix across it (see Figure 3). Increasing \(\Sigma_{\text{polynya}}\) increases the input of available potential energy on the continental shelf, which is partly released as EKE on the continental shelf and slope, enhancing the rate of eddy CDW transport. Note that the isopycnals close to the ocean bed over the continental slope are shoaled to
Figure 3. Comparison between simulated and observed potential temperature (colors) and neutral density (black contours). (a–d) Time- and along-shore-mean properties in our process model for a range of surface wind stress maxima. (e–h) Cross-slope hydrographic sections from, the eastern Somov Sea [Gordon et al., 2004], the western Weddell Sea [Thompson and Heywood, 2008], the Bellingshausen Sea and the eastern Weddell Sea [Orsi and Whitworth, 2005]. Vertical white dashed lines indicate cast locations. (i) Map of the 1000 m depth contour around Antarctica (black) [Amante and Eakins, 2009] with selected hydrographic sections (red). For each section we estimated the local wind stress $\tau$ directed along the 1000 m depth contour from the CORE.v2 dataset [Large and Yeager, 2009].

the south (see Figure. 3), so both the surface wind stress and mesoscale eddies serve to extract available potential energy (APE), and thus the APE must be sourced from $\Sigma_{\text{polynya}}$. Increasing $H_{\text{shelf}}$ deepens the continental shelf, allowing more CDW isopycnals to connect with the shelf waters. Increasing $W_{\text{slope}}$ reduces the topographic potential vorticity gradient, allowing the slope to support larger eddies that transport CDW shoreward more rapidly. Shifting the surface wind stress maximum simply changes the effective wind stress over the continental slope, which in turn modifies $F_{\text{CDW}}$ by changing the pycnocline depth at the shelf break. Finally, reducing the numerical grid spacing $\Delta x$ resolves baroclinic instabilities at shorter wavelengths, which increases the conversion of potential energy to EKE on the continental slope, and thus enhances the eddy transport of CDW. We find that there is no further increase in $F_{\text{CDW}}$ for $\Delta x \leq 1$ km because at this point the model resolves the first
Rossby radius of deformation, which is around 4 km on the continental slope.

Of all of the parameters examined in this study, the ASF’s structure and transport are particularly sensitive to the along-slope wind stress. Our simulations suggest that much of the regional variation in the structure of the ASF can be explained by the local alongshore winds. To illustrate this idea, in Figure 3(a)–(d) we plot the simulated time- and alongshore-mean potential temperature and neutral density obtained for increasing wind strengths $\tau_{max}$, with all other aspects of the model configuration held fixed. Panels (e)–(h) show hydrographic sections from the Bellingshausen, Southern and Weddell Seas that are subject to comparable mean wind stress along the 1000 m depth contour [Amante and Eakins, 2009]. In each case we computed the average alongslope wind stress along a ~300 km stretch of the 1000 m depth contour, using the CORE v2 surface wind stress data [Large and Yeager, 2009] averaged between 1996 and 2006. Note that this dataset does not account for sea ice, which can enhance the ocean surface wind stress by up to 50% [e.g. Fujisaki et al., 2010]. Our idealized model has not been tailored to simulate these sections of the ocean, and consequently there are several clear differences between the simulations and the observations, most obviously the presence of AABW in panels (a) and (d). The agreement is arguably worst for the Bellingshausen (panels (a) and (e)), where the overturning circulation is fundamentally different, with a warm inflow along the ocean bed on the continental shelf and surface outflow of fresh water from the ice shelves [e.g. Potter et al., 1988]. However, the striking qualitative agreement in the isopycnal structure over the continental slope and the penetration of CDW onto the continental shelf suggests that the local alongshore wind stress controls the pycnocline depth at the shelf break, and thereby limits the eddy-driven shoreward CDW transport.

4. Discussion

This study elucidates the dependence of the ASF’s structure and cross-slope transport on various aspects of the local surface forcing and continental shelf geometry. If we regard our model as a conceptual representation of the mean surface forcing and continental shelf geometry. If we re-eddy-driven shoreward heat transport. This corresponds to a longitudinal resolution of around 1/48°, which would incur an excessive computational cost in any global ocean model; even the Southern Ocean State Estimate (SOSE) uses only 1/6° resolution [Mazloff et al., 2010]. In models coarser than this 1/48° cutoff, the cross-slope eddy transfer coefficient must be parametrized carefully to ensure that the shelf water masses are accurately represented [Stewart and Thompson, 2013].

Our model’s sensitivity to the along-shore wind stress, shown in Figure 2(a), and the qualitative agreement between the observed and modeled ASF structure at various wind strengths, shown in Figure 3, suggest that the Antarctic easterlies exert substantial influence over transport and mixing across the continental shelf break [c.f. Stewart and Thompson, 2012]. This implies that past and future changes in the along-shore winds could substantially alter the exchanges of mass, heat, salt, and biogeochemical tracers across the Antarctic shelf break, with potentially dramatic implications for the ocean’s abyssal circulation, ventilation and carbon storage [Jacobs, 2004]. Consistent with our findings, Spence et al. [2014] have shown that a southward shift of the mid-latitude westerlies, weakening the Antarctic easterlies, warms the Antarctic subsurface coastal waters by up to 2°C due to weakening of the shoreward surface Ekman transport. Further work is required to quantify the contribution of the along-shore winds to the variety of ASF structures observed around the Antarctic margins.
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