6.2 TESTING THEORIES OF THE ANTARCTIC CIRCUMPOLAR CURRENT AGAINST OBSERVATIONS

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1. INTRODUCTION

The Antarctic Circumpolar Current (ACC) dominates the dynamics of the Southern Ocean and plays a critical role in the global ocean circulation. While the ACC connects the ocean basins, it isolates the polar region since, by simple definition, the mean circumpolar flow cannot transport quantities poleward. (see streamlines on Fig. 1). Yet the ocean remains in a quasi-steady state, indicating that other processes transport quantities across the ACC to balance differential surface forcing.

The dominant forcing of the ACC is the strong westerly winds, the roaring forties, that blow over it. At the same time as the wind drives the ACC's eastward flow, it also drives an overturning cell—the Deacon Cell—associated with Ekman processes: fluid upwells in the south, travels northward at the surface, and is pumped to depth in the subtropics. The northward Ekman transport carries heat northward, requiring a balancing southward cross stream transfer of heat. This compensating flux is supplied by baroclinic eddies that feed off the potential energy in the stratification of the ACC. Mooring data have established that the poleward eddy heat flux is sufficient to balance the Ekman flux (Phillips and Rintoul, 2000).

The atmosphere also forces the ACC through surface buoyancy fluxes. The Southern Ocean is a region of significant heat loss and gain at ocean surface and strong surface salinity fluxes associated with evaporation and precipitation in the north and ice formation and melt in the south. These buoyancy fluxes are balanced by the residual circulation—the small imbalance between the eddies and the winds.

In previous work we examined how the balance between eddy transport and surface wind and buoyancy forcing set the stratification and transport of an idealized circumpolar current; see Marshall et al. (2001), Karsten et al. (2001). We discussed the implications of this balance on the dynamics of the ACC in Karsten and Marshall (2001). Here, guided by our previous work, we describe our attempts to diagnose from observations the meridional overturning of the ACC. We show that the transport associated with eddies is sufficient to establish a leading-order balance with the northward Ekman flux. We also calculate the residual circulation and connect it to surface buoyancy fluxes.

In doing so we provide a dynamical explanation of the Antarctic convergence and a means of calculating the position of the Polar Front. We also calculate the rate of formation and distribution of Antarctic Intermediate Water (AAIW). Not only is this water transformation a vital component of the thermohaline, it is also essential in the distribution of tracers in the Southern Ocean. We illustrate this by showing how the residual circulation explains the subduction of high-oxygen water formed south of the ACC.

2. DYNAMICAL BACKGROUND

We begin with the along-stream and temporally averaged equations in a steady state. For incompressible flow, the conservation of buoyancy can be written as

$$\overline{v}\frac{\partial \overline{b}}{\partial y}+\overline{w}\frac{\partial \overline{b}}{\partial z}+\frac{\partial}{\partial y}(\overline{v'b'})+\frac{\partial}{\partial z}(\overline{w'b'})=\frac{\partial B}{\partial z}, \quad (1)$$

where y (v) and z (w) are the cross-stream and vertical coordinates and corresponding velocities, respectively. In (1) the variables have been separated into mean (along stream and time) quantities, $\overline{b}$, and the perturbations about this mean, b'. The buoyancy forcing has been written as a vertical gradient of a buoyancy flux, B, with the flux taken as positive if it is directed into the ocean. For simplicity, we have ignored the small-scale diffusion.

The mean cross-stream advection is determined by the Ekman transport and can be written as

$$\overline{v}\frac{\partial \overline{b}}{\partial y}+\overline{w}\frac{\partial \overline{b}}{\partial z}=J(\overline{\Psi},\overline{b}), \quad (2)$$

with

$$\overline{\Psi}=-\frac{\overline{\tau}}{\rho_0 f}, \quad (3)$$

where $\overline{\tau}$ is the time-averaged along-stream wind stress, ρ_0 is the mean density, and f is the Coriolis parameter. The Jacobian operator J is defined by $J(A,B)=A_yB_z-A_zB_y$. The detailed structure of the flow in the Ekman layer and the return flow at depth are not required.

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We also interpret the divergence of the eddy fluxes as an advective flux thus:

$$\frac{\partial}{\partial y}(\overline{v'b'}) + \frac{\partial}{\partial z}(\overline{w'b'}) = J(\Psi^*, \overline{b}), \quad (4)$$

where Ψ^* is the eddy-induced streamfunction given by

$$\Psi^* = \frac{\overline{v'b'}}{\overline{b}_z}, \quad (5)$$

assuming that the eddy fluxes are adiabatic below the mixed layer. Again, we do not determine the form of Ψ^* within the mixed layer, but we do assume that $\Psi^* = 0$ along the surface.

Using (2) and (4) we can rewrite the Eulerian balance (1) in an alternative form, the transformed Eulerian mean equation:

$$J(\Psi_{res}, \overline{b}) = \frac{\partial B}{\partial z}, \quad (6)$$

where the residual streamfunction Ψ_{res} is given by

$$\Psi_{res} = \overline{\Psi} + \Psi^*. \quad (7)$$

If we integrate (6) in the mixed layer, where vertical gradients of buoyancy vanish, then we have

$$\Psi_{res_m} \frac{\partial \overline{b}_m}{\partial y} = B, \quad (8)$$

where Ψ_{res_m} is the residual streamfunction at the base of the mixed layer and $\overline{b}_m$ is the mixed-layer buoyancy (see also Marshall, 1997). Equation (8) states simply that the residual circulation in the mixed layer is forced locally by the surface buoyancy flux. Since the cross-stream surface buoyancy gradient is positive, buoyancy gain by the ocean, $B > 0$, is associated with a northward transport, $\Psi_{res} > 0$. The Antarctic convergence, where the residual circulation vanishes, coincides with the point where the buoyancy forcing changes sign.

If we consider (6) in the interior, away from the effect of surface buoyancy fluxes, we have

$$J(\Psi_{res}, \overline{b}) = 0. \quad (9)$$

Equation (9) implies that below the mixed layer the residual circulation is constant along isopycnals.

3. THE RESIDUAL CIRCULATION

In order to calculate the residual circulation, we use altimetry data to estimate eddy fluxes following Keffer and Holloway (1988). The eddy flux is assumed to be a down-gradient transfer, i.e.,

$$\overline{v'b'} = -K\frac{\partial b}{\partial y}, \quad (10)$$

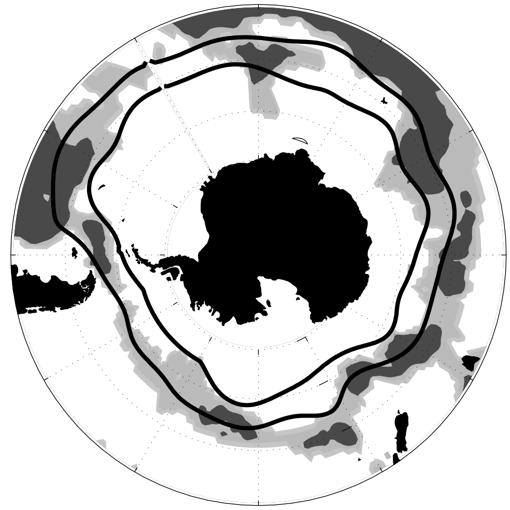

Figure 1: The eddy diffusivity K calculated using the formula (11). Dark areas correspond to very strong eddies with a diffusivity exceeding 2000 m s^{-2}, grey areas correspond to strong eddies with a diffusivity of 1500 to 2000 m s^{-2}. The solid lines are the streamlines that mark the boundaries of the ACC.

where the eddy diffusivity is assumed proportional to the root mean square of the geostrophic stream function variability, that is,

$$K = \alpha\frac{g}{f}\left(\overline{h'^2}\right)^{1/2}, \quad (11)$$

where g is the acceleration due to gravity, h' is the sea-surface height variability, and $\alpha = 0.5$ is a constant of proportionality.

In Fig. 1 we plot the estimate of K using (11) with TOPEX/POSIEDON sea surface height variability taken from Stammer (1998). The ACC, especially its northern boundary, is a relatively active region. In the path of the ACC, we see several hot spots of high eddy activity located at or downstream of locations where the ACC crosses large topographic features. The diffusivity increases as we move northward across the ACC with a mean value of roughly 1800 m s^{-2}.

Given the eddy diffusivity, we can calculate the eddy buoyancy flux and hence the eddy-induced transport given by (5). The net eddy-induced transport, plotted in Fig. 2, is a poleward transport increasing from 6 Sv to 27 Sv as we cross the ACC northward. Next, we use the wind stress from Josey et al. (1998) to calculate the Ekman transport. The net Ekman transport, shown in Fig. 2, is an equatorward transport of over 20 Sv throughout the ACC region.

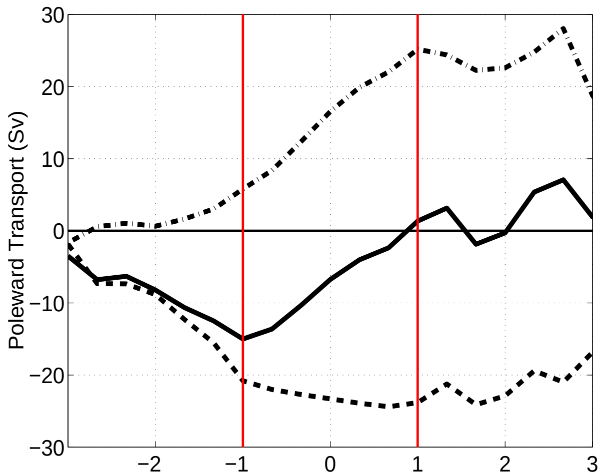

Figure 2: The net poleward transport due to the eddy-induced advection (dash-dot), the Ekman advection (dashed), and their sum—the residual circulation (solid). The vertical bars mark the boundaries of the ACC.

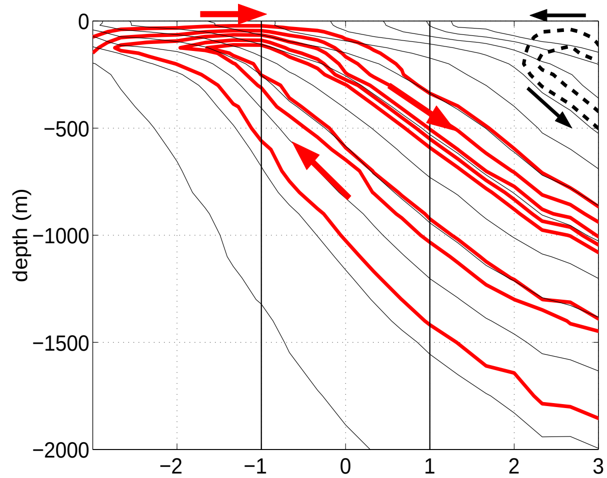

Figure 3: The residual circulation streamfunction versus depth and cross-stream coordinate. The solid (dashed) contours are positive (negative). The arrows mark the direction of flow, the faint lines are contours of buoyancy, and the vertical bars mark the boundaries of the ACC.

The net transport of the residual circulation is calculated from (7) and is shown in Fig. 2. There is a net transport of 15 Sv flowing northward into the ACC and 1 Sv flowing southward into the ACC—the Antarctic Convergence. Thus, within the ACC there must be the subduction of 16 Sv of water, a quantity similar to that discussed in Marshall (1997) but somewhat smaller than that discussed in Speer et al. (2000). At the northern boundary, the residual transport is especially weak in comparison to the Ekman transport and the eddy-induced transport and, thus, to leading order, the eddies balance the winds. However, at the southern boundary, the residual circulation actually exceeds the eddy-induced flow indicating that buoyancy forcing must enter the leading-order dynamics here.

We can map the residual circulation down to depth along the isopycnals using (9). In Fig. 3, we plot the mean residual circulation versus depth and a cross-stream coordinate. The circulation around the ACC is dominated by two meridional overturning cells. In the lower cell, Upper Circumpolar Deep water (UCD) is transported southward across the ACC, upwells to the surface south of the ACC, moves northward along the surface—gaining buoyancy to become Antarctic Surface Water (AASW)—and subducts to form AAIW. The upper, weaker cell has Subantarctic Mode Water (SAMW) moving southward—losing buoyancy—and subducting to form AAIW. These Diabatic Deacon cells have been discussed before (Marshall, 1997; Speer et al., 2000) but we believe we are the first to diagnose them from observations and estimates of eddy fluxes.

4. BUOYANCY FLUXES AND TRACERS

Using (8) and Fig. 2, we can infer a gain of buoyancy by the ocean at the southern boundary and a loss of buoyancy at the northern boundary agreeing with the flux patterns from COADS data (Speer et al., 2000). The magnitude of the flux is relatively weak—for a thermal expansion coefficient of 1×10^{-4} its maximum corresponds to a heat flux of roughly 5 W m^{-2}.

This pattern of buoyancy forcing is reasonable for the Southern Ocean, where the air and sea temperatures are in rough equilibrium with variations in the atmosphere occurring on shorter time scales than in the ocean. If a cold atmospheric anomaly moves north of the Polar Front, which separates the cold AASW from the warmer SAMW, the ocean will lose heat to this anomaly—a buoyancy loss. On the other hand, if a warm atmospheric anomaly moves south of the Polar Front, the ocean will gain heat from this anomaly—a buoyancy gain. Hence, the Polar Front—the point where the residual circulation vanishes and southward flowing SAMW meets northward flowing AASW—is also the boundary between regions of buoyancy gain and regions of buoyancy loss by the ocean—as it must be given (8).

In Fig. 4, we compare the distribution of dissolved oxygen to the residual circulation. The low-temperature surface waters south of the ACC have a very high concentration of oxygen. This region of well-oxygenated water also extends as a tongue under the ACC, marking the AAIW as well-ventilated water. As

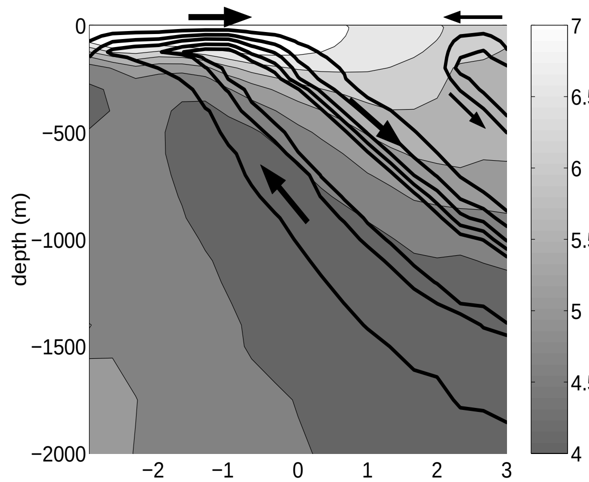

Figure 4: The shading marks contours of the mean dissolved oxygen. The dark solid lines are contours of the residual circulation with the arrows showing the direction of flow.

well, there is a tongue of low-oxygen water extending southward at greater depths, and then nearly rising to the surface. This marks unventilated UCD, which has its origin in the North Atlantic.

This pattern agrees with the flow determined by the residual circulation. Deep, low-oxygen waters (UCD) are transported southward across the ACC, rising upward until they are brought to the surface south of the ACC. At the surface these waters are ventilated, advected northward along the surface, and then subducted near the center of the ACC, forming AAIW.

5. CONCLUSIONS

We have established that the leading-order balance between eddy transport and Ekman transport exists, seen in the small magnitude of the residual circulation—the vanishing of the Deacon Cell. The balance is equivalent to a balance between the input of momentum at the ocean surface by wind stress and its transfer to bottom topography by interfacial form stress induced by eddies (Munk and Palmén, 1951). As well, we have shown that the residual circulation of the leading-order balance implies a surface buoyancy flux whose pattern agrees with the observations,

When the residual circulation is mapped to depth, the corresponding pattern of subduction near the center of the ACC also agrees with the tongue of high-oxygen water found in the hydrographic record. Thus, we have a method of establishing the position of the Polar Front—the front between the cold AASW and the warm SAMW where subduction occurs.

It has long been postulated that meridional overturning cells must exist in order to explain the observed water masses and the transformation of UCD to AAIW (Speer et al., 2000). However, we believe that we are the first to calculate the residual circulation responsible for these patterns from a dynamical analysis of observations. The fact that our calculated circulation agrees well with the surface-forcing patterns and oxygen distribution clearly indicates that the assumptions made in our calculations are reasonable.

More details and further discussion of the calculations presented here will be published in a paper currently in preparation. Preprints of this paper, and other works cited here, are available by request or at http://puddle.mit.edu/∼richard.

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