The Southern Ocean Limb of the Global Deep Overturning Circulation*

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ABSTRACT

Nine hydrographic sections are combined in an inverse box model of the Southern Ocean south of \( \sim 12^\circ \)S. The inclusion of independent diapycnal flux unknowns for each property and air–sea (heat, freshwater, and momentum) fluxes make it possible to estimate the three-dimensional deep water circulation. The authors find a vigorous \( 50 \times 10^6 \) m\(^3\) s\(^{-1}\) deep overturning circulation that is dominated by an equatorward flow of Lower Circumpolar Deep Water and Antarctic Bottom Water and poleward flow of upper deep water including Indian and Pacific Deep Water below 1500 dbar. In the subtropical Indian and Pacific Oceans the deep overturning cell is essentially isolated from the thermocline and intermediate waters of the subtropical gyre. The southward flowing upper deep water shoals south of the Antarctic Circumpolar Current, where air–sea fluxes convert outcropping upper deep water to Antarctic surface water and drive a net air–sea transformation of \( 34 \times 10^6 \) m\(^3\) s\(^{-1}\) to lighter intermediate water. It is the outcropping of upper deep water and transformation by air–sea fluxes that connects the deep and intermediate circulation cells. The significant poleward transport of relatively light (i.e., above all topography at the latitude of Drake Passage) upper deep water, as required here to balance lateral and diapycnal divergence and air–sea exchange, provides observational evidence that advection by standing and transient eddies carries significant meridional transport in the Southern Ocean.

1. Introduction

The overturning circulation refers to the flow of dense deep and bottom waters away from their sources and the compensating return flow of less dense upper-ocean water. The sources of deep and bottom water are limited to a few high latitude sites in the present-day ocean. In the North Atlantic, dense overflow water spills across numerous sills in the Greenland–Scotland Ridge. Subsequent mixing of this dense water in the northern Atlantic results in a prominent high salinity, relatively cold water mass, the North Atlantic Deep Water (NADW), that penetrates southward into the South Atlantic (Dickson and Brown 1994). Bottom waters are also produced at specific sites around the Antarctic continent, including the Weddell Sea, Ross Sea, and Adélie Land (Gill 1973; Carmack 1990; Fahrbach et al. 1994; Rintoul 1998; Wong et al. 1998). Newly formed Antarctic Bottom Water (AABW) mixes with overlying deep water as it spreads northward and fills the abyssal basins of the Atlantic, Indian, and Pacific Oceans (Orsi et al. 1993; Mantyla and Reid 1995).

Much of the focus to date has been on the size and path of the NADW component of the deep circulation (Gordon 1986; Rintoul 1991; Saunders and King 1995; Macdonald and Wunsch 1996). On the other hand, Gordon (1975) estimated a total production rate of AABW of \( 35 \times 10^6 \) m\(^3\) s\(^{-1}\) to \( 40 \times 10^6 \) m\(^3\) s\(^{-1}\), roughly twice that of NADW. Despite the fact that this early estimate, based on a simple heat budget calculation for the abyssal ocean, suggested a significant Southern Ocean role in the deep overturning circulation, attention has remained fixed on NADW. More recently, a number of studies have renewed interest in the Southern Ocean’s contribution to the global overturning circulation, including its influence on NADW production (Toggweiler and Samuels 1993a; Rahmstorf and England 1997; Goodman 1998; Toggweiler and Samuels 1998), and its role in closing the NADW cell by converting deep water to intermediate water (Di"øs and Coward 1997).

The sinking of NADW and AABW, and subsequent
spreading into the abyssal basins, forms the lower limb of the overturning circulation. The upper-ocean return limb of the overturning flow is also important, particularly for estimates of the meridional transport of heat, salt, and other properties. Estimates of property transports to date have usually been confined to a single section or single basin (Rintoul 1991; Toole and Warren 1993; Saunders and King 1995; Robbins and Toole 1997; Tsimplis et al. 1998). Notable exceptions include the global analyses of Schmitz (1995), Macdonald and Wunsch (1996), Schmitz (1996a,b), and MacDonald (1998). Schmitz combined a large number of published transport estimates from regional studies in an effort to map out the large-scale circulation. Perhaps surprisingly, he found that the individual transport estimates were largely consistent with each other, allowing the construction of a global circulation scheme. Macdonald and Wunsch (1996) and Macdonald (1998) derived a global circulation scheme using an inverse model similar in spirit, but different in several important respects, to the model described here.

Our picture of the overturning circulation in the Southern Ocean has changed little from that developed by Sverdrup and Deacon in the 1930s. They traced property extremum along meridional sections across the Southern Ocean and inferred a poleward transport of deep water, balanced by an equatorward transport split between denser bottom water and less dense intermediate water. To date, this schematic view of the Southern Ocean overturning circulation has yet to be quantified using observations, although a number of recent numerical modeling studies have focused on this issue.

An important element in this circulation scheme is the diapycnal fluxes, driven by air–sea buoyancy forcing and interior mixing, which are needed to transfer water between density layers and close the overturning cell. These diapycnal transports must be included and quantified in any complete description of the overturning circulation scheme. Macdonald and Wunsch (1996) and Macdonald (1998) derived a global circulation scheme using an inverse model similar in spirit, but different in several important respects, to the model described here.

The purpose of this study is to describe the circulation of deep and bottom waters in each sector of the Southern Ocean. From these an estimate of the overall Southern Ocean overturning circulation is derived. Companion papers present other results from the inverse model: a discussion of the interior diapycnal fluxes is given in Sloyan and Rintoul (2000) and the formation and circulation of mode and intermediate waters is given in Sloyan and Rintoul (2001).

The paper is set out in the following way. In section 2 we provide details of the hydrographic data used, a priori assumptions, and give a brief description of the design of the inverse model, including the treatment of air–sea forcing. The net mass, heat, and salt fluxes across each section are described in section 3. Section 4 discusses the deep circulation in each basin. The results from individual basins are drawn together in section 5 to provide an integrated description of the overall overturning circulation in the Southern Ocean. The conclusions are summarized in section 6.

2. Inverse model description

a. Hydrographic data, domain, and layers

Nine hydrographic sections are used to define six “boxes” in the Southern Hemisphere oceans, as shown in Fig. 1. The sections used are recent (1984–94) high quality hydrographic sections, apart from the Ind18 section, which was occupied in 1976 (Table 1). The sections across the Pacific, in Drake Passage, and south of Australia are part of the World Ocean Circulation Experiment (WOCE) dataset.

Twenty-three layers (Table 2) are chosen to span the water masses in the model domain. The layers are defined by neutral density surfaces calculated using the Jackett and McDougall (1997) algorithm. Table 2 also gives the average potential temperature and salinity of each layer at each section. [Note that cold, fresh Antarctic Surface Water (AASW) lowers the average temperature and salinity in layers 13 to 15 at the Southern Ocean choke point sections, relative to the lower latitude sections.]

The temperature and salinity data are used to determine the baroclinic or relative velocity between adjacent stations along the hydrographic sections. Numerous methods have been used to estimate the property transports in the bottom triangles below the deepest common depth of adjacent stations (Wunsch 1996). In this study the horizontal density gradient at the deepest common depth is held constant through the bottom triangle area of each station pair. The associated property transport is then included in the deepest layer at each station pair.

The Ekman property fluxes are calculated normal to the station pairs along the hydrographic section. The depth of the Ekman layer is assumed to be 60 dbar for
all sections and the Ekman flux is calculated using Hellerman and Rosenstein (1983) annual mean wind stress. The Ekman property fluxes are calculated as a weighted mean from the sea surface to 60 dbar assuming an exponential decay of the Ekman velocity. The Ekman property fluxes are then added to the top layer of each station pair.

Across the meridional sections (SAVE2, SAVE4, Ind18, Ind32, and Pac32) the initial reference level is the boundary between the northward flowing AABW and southward flowing North Atlantic Deep Water (NADW), Indian Deep Water (IDW), and Pacific Deep Water (PDW). In the Argentine basin of SAVE4 and the three Southern Ocean choke point sections (DrakeP, SAfrica, and SAust) the initial reference level is taken to be the deepest common depth at each station pair.

b. Model constraints, weighting, and error estimates

A box inverse model, following that of Wunsch (1978), is designed. Mass, heat, and salt are conserved in all layers while the silica constraint is imposed as a total box constraint. Two novel aspects of the inverse model are the use of independent diapycnal flux unknowns for each property [mass ($w^m$), heat ($w^T$), and salt ($w^S$)], and the explicit inclusion of air–sea fluxes of heat, freshwater, and momentum (wind) and the water mass transformation they drive.

The inclusion of independent diapycnal fluxes for each property represents the net diapycnal flux that results from all mixing processes that act to transfer mass, heat, or salt between water masses in the ocean interior. (Here “interior” is taken to mean the entire ocean beneath the sea surface. Diapycnal fluxes across isopycnals outcropping in the surface mixed layer, for example, are included in the interior diapycnal fluxes.) These mixing processes include advection, diffusion, cabbeling, and eddy fluxes. McIntosh and Rintoul (1997) and Sloyan and Rintoul (2000) show that inverse models using this representation of diapycnal fluxes can reproduce the “true” diapycnal fluxes when tested using output from a numerical model. In contrast, inverse models that use a single unknown $w^*$ (so that the diapycnal mass, heat, and salt fluxes are given by $w^*$, $w^*T$, $w^*S$, where the
Table 2. The 22 neutral density surfaces (γ') that bound the 23 layers and the water masses they define at the hydrographic sections. Also shown are the layer-average potential temperature (upper, °C) and salinity (lower, psu) at each station. Asterisks indicate no layer present in section.

<table>
<thead>
<tr>
<th>Layer</th>
<th>γ'</th>
<th>SAVE2</th>
<th>SAVE4</th>
<th>DrakeP</th>
<th>SAfrica</th>
<th>Wedsea</th>
<th>Ind18</th>
<th>Ind32</th>
<th>SAust</th>
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<td>*</td>
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<td>*</td>
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<td>26.10</td>
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<td>34.720</td>
<td>34.712</td>
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</tbody>
</table>

Overbar represents the average on an isopycnal surface (e.g., Wunsch 1984; Macdonald 1998), or which integrate the residuals in the mass conservation equations to infer the diapycnal flux (e.g., Wunsch et al. 1983; Roemmich and McCallister 1989), are unlikely to accurately represent the diapycnal fluxes (McDougall 1987; McIntosh and Rintoul 1997; Sloyan and Rintoul 2000).

Exchange of heat and freshwater with the atmosphere results in net buoyancy forces that can transfer fluid from one density layer to another (Walin 1982). To date, inverse methods have not explicitly included these effects on the mass, heat, or salt conservation equations. Inverse models usually either downweight the conservation requirement in layers that outcrop and interact with the overlying atmosphere or only consider con-
servation below the thermocline (Metzl et al. 1990; Rintoul 1991; Macdonald 1998). Rather than throw away the information contained in the conservation constraints for outcropping layers, it is desirable to include the effects of air–sea interaction explicitly. This is particularly so at high latitudes, where many layers outcrop. More importantly, the diapycnal fluxes driven by air–sea interaction are a fundamental part of the three-dimensional circulation of the ocean. An inverse model that seeks to determine this circulation should include these processes in a physically consistent manner.

The effect of air–sea fluxes of heat, freshwater, and momentum can be included explicitly in the model following the strategy of Walin (1982), Tziperman (1988), Schmitt et al. (1989), and Speer and Tziperman (1992). The mass flux \( f_m \) due to air–sea fluxes of heat and freshwater, and Ekman transport is

\[
f_m = -\frac{\alpha H}{C_p} + \rho \beta Q S - \rho Q + \frac{\partial P E}{\partial \rho}.
\]

In (1) \( \rho \) is the sea surface density, \( \alpha \) thermal expansion coefficient, \( \beta \) saline contraction coefficient, \( H \) heat flux, \( Q \) freshwater flux (evaporation \( - \) precipitation \( + \) run-off), \( S \) sea surface salinity, \( E \) the Ekman transport, and \( C_p \) specific heat. Each of these terms is a function of time and position. The first two terms of (1) represent the effect of heat and freshwater fluxes on density, while the last two terms are direct input/loss of freshwater and the horizontal Ekman transport, respectively. The total mass flux across an isopycnal with density \( \rho \), for one year, as a function of sea surface density is

\[
F_m(\rho) = \int_{\text{year}} dt \int \Delta f_m \delta(\rho - \rho').
\]

where the \( \delta \) function samples the density flux at surface \( \rho \). The mass convergence in a layer bounded by isopycnals \( \rho \) and \( \rho + d\rho \) is

\[
C_m(\rho)d\rho = -[F_m(\rho + d\rho) - F_m(\rho)].
\]

In a similar manner, the convergence of heat and salt in density layers due to air–sea fluxes of heat, freshwater, and momentum can be found.

The heat flux \( f_h \) across outcropping density surfaces is

\[
f_h = -\frac{H}{C_p} + \frac{\partial P E \theta_{E}}{\partial \rho};
\]

while the salt flux \( f_s \) is

\[
f_s = \frac{\partial P E S_{E}}{\partial \rho},
\]

where \( \theta_{E} \) and \( S_{E} \) are the weighted mean temperature and salinity over the Ekman layer \( (D_E) \).

The mass, heat, and salt fluxes across each outcropping layer are calculated from climatological datasets. As the data are discontinuous, the property transformations across outcropping neutral density surfaces are summed over a discrete interval surrounding the outcropping density (Speer and Tziperman 1992). The delta function is now a boxcar (\( \Pi \)) of width \( \Delta \rho \):

\[
F_m(\rho) = \sum_{n=1}^{12} \Delta t \sum_{i,j} \Delta A_{ij} \left[ \frac{-\alpha H_{ij}}{C_p} + \rho \beta Q S_{ij} \right]
\]

\[
-\int \Delta t \sum_{i,j} \Delta A_{ij} \left[ \frac{H_{ij}}{C_p} + E_{ij} \theta_{Eij} \right] \Pi(\rho - \rho').
\]

The climatological data used for the air–sea fluxes are: heat flux \( (H) \) COADS by da Silva et al. (1994); wind stress \( (E) \) Hellerman and Rosenstein (1983) and; freshwater \( (Q) \) from GASP by Budd et al. (1995). We use Levitus and Boyer (1994) to find the outcropping position of the neutral density surfaces for each month, the weighted-mean Ekman layer temperature and salinity, and the annual mean depth and property concentration of the neutral density surfaces found within each box.

The datasets used to calculate the air–sea fluxes contain errors, including errors in the bulk formulae used and a lack of oceanographic and meteorological observations, especially at high latitudes and over subtropical regions away from commercial shipping routes (Speer and Tziperman 1992; Barnier et al. 1995). As a result, we treat these climatologies as an initial guess in the inverse model and use the inverse solution to determine corrections to the climatological datasets, which make them consistent with the hydrographic data and the model physics. A complete description of the method, the derived corrections to the air–sea fluxes, and the water mass formation driven by the air–sea fluxes can be found in Sloyan (1997).

The conservation equations, including the diapycnal flux unknowns for each property and the air–sea flux–driven transformations, take the form

\[
\sum_{j=1}^{N} \Delta x_j \int_{b_j}^{t_{n+1}} (\rho C_{j}(v_{j} + b_{j})dP) + E_{j}\rho C_{j} + (w^{*}_{j}A\rho C)_{m} - (w^{*}_{j}A\rho C)_{m+1} + (F_{e} + F^{*}_{e}) - (F_{e} + F^{*}_{e})_{m+1} = 0.
\]

In (9) \( \Delta x_j \) is the station spacing at pair \( j \), \( C_{j} \) is the property value/unit mass at pair \( j \). The baroclinic or relative velocity \( (v_{j}) \) is determined from the hydrographic data, \( E_{j}C_{j} \) is the Ekman property flux at pair \( j \), and \( F_{e} \) is the total flux across an outcropping isopycnal driven by buoyancy forcing. The subsequent system of
simultaneous equations is solved for the unknown reference level velocities \((\mathbf{b})\), the diapycnal fluxes for mass, heat, and salt \((w^{m}, w^{h}, w^{s})\) and the corrections to the air–sea climatologies \((\mathbf{F}^{*})\) (9).

The importance of appropriate column and row weighting has been discussed by McIntosh and Rintoul (1997). The rows are weighted by the property norm and the columns are weighted by the area norm. A scaling factor is also applied to the reference velocity unknowns, the diapycnal property flux unknowns, and air–sea climatology corrections such that the condition number (the ratio of largest eigenvalue to smallest eigenvalue) is minimized. The salt conservation equations are expressed as salt anomaly, \(S = 35.00\). Errors due to the combination of nonsynoptic hydrographic sections lead us to choose a rank where the data residual norms are of \(O[1\text{ to } 2\times10^{6} \text{ m}^{3} \text{ s}^{-1}]\).

Table 3 contains information on the a priori constraints applied to the inverse model. Peterson and Whitworth (1989) and Peterson (1992) find large bottom velocities beneath the Malvinas Current that increase the transport substantially above that estimated from geostrophy relative to a deep reference level. This information is included by setting the initial choice of the reference velocity at the first six station pairs adjacent to the South American coast to 0.132 m s\(^{-1}\) (13.2 cm s\(^{-1}\)), comparable to observations from near-bottom current meters in this region (Peterson and Whitworth 1989; Peterson 1992). Mooring array PCM-9 was deployed at 32\(^{\circ}\)S in the Pacific Ocean along a 1000-km line east at the Tonga–Kermadec Ridge (Whitworth et al. 1999). The mooring transport constraint is applied to deep and bottom water below neutral density surface \(\gamma^* = 28.0,\) across the hydrographic stations that span the longitude range of the mooring array (181\(^{E}\)–192.5\(^{E}\))E. Therefore, only 1000 km of the 6000-km wide Southwest Pacific basin is constrained. No prior constraint is placed on the transport of the Antarctic Circumpolar Current (ACC). Total silica conservation in each box is a weak constraint (large box imbalance is assumed), reflecting the uncertainty in the silica flux. The error bars presented are the sum of the noise and null space errors, which are calculated following Wunsch et al. (1983) and Rintoul (1991). These errors represent the formal errors of the inverse method—the error associated with determining the barotropic velocity, and diapycnal fluxes. These formal errors do not take into account errors due to the asynoptic data, solution sensitivity to the first guess, and ocean variability. As the formal errors do not include these other error sources they are unlikely to reflect the true uncertainty in the estimates. In recognition of the inadequacy of the formal error bars, Macdonald and Wunsch (1996) and Macdonald (1998), for example, chose to increase the uncertainty of all their heat flux estimates by a somewhat arbitrary \(\pm 0.25\text{ PW}\) [based on Holfort (1994) error analysis in the South Atlantic]. They believed this value more accurately accounted for the effect of (the unmeasured) oceanic variability. The value used is much larger than the formal error bar provided by the inverse method. Here we have chosen to present the estimates with the formal error bars so that it is clear what the error bars represent. If a good estimate exists of the uncertainty introduced by error sources not included in the formal error bars (e.g., Holfort’s value of 0.25 PW for the South Atlantic), this value should be added to the error bars shown here.

Ultimately better assessments of the uncertainty in ocean flux estimates—however made—require measurements of ocean variability (and forcing), which do not yet exist for most of the ocean. One exception is the WOCE SR3 section south of Australia, which was occupied six times between 1991 and 1996, including each season. Sloyan (1997) tested the sensitivity of the inverse solution by rerunning the model using a realization of the SR3 section from a different year and different season (winter 1995). Overall, the changes were small (layer transport changes of \(<1 \times 10^{6} \text{ m}^{3} \text{ s}^{-1}\) in almost all layers, and usually \(<0.5 \times 10^{6} \text{ m}^{3} \text{ s}^{-1}\), except at the SR3 section itself). The main difference from the solution described below is an increase of \(5 \times 10^{6} \text{ m}^{3} \text{ s}^{-1}\) in the eastward flow south of Australia, which is balanced by increased northward flow in the Pacific, increased Indonesian Throughflow, and increased southward flow in the Indian Ocean. The net heat flux across roughly 30\(^{\circ}\)–40\(^{\circ}\)S (SAVE4, Ind32, and Pac32) is almost unchanged (difference of 0.02 PW), despite the increase in strength of the circum-Australia
flow. The pattern and magnitude of the lateral and diapycnal fluxes—in particular, the strength and structure of the deep overturning circulation, the subject of this paper—are similar in both models. While using different realizations of a single section clearly does not provide a complete assessment of the impact of oceanic variability on our circulation estimates, this sensitivity test and those described in the appendix supports the assertion that robust estimates of the large-scale circulation and fluxes can be made from models such as the one used here.

3. Net meridional and zonal fluxes

In the following discussion we give a brief description of the net mass, heat, and salt fluxes across each section. The overall circulation generally agrees with our current understanding of the circulation of the Southern Ocean and adjacent basins. We also describe the corrections to the climatological air–sea heat and freshwater flux fields required to ensure consistency between the forcing fields and the hydrography.

a. Mass transport

The integrated mass, heat, and salt fluxes across the hydrographic sections are shown in Fig. 2 and Table 4. The individual layer mass fluxes for each section are given in Table 5.

In the Southern Ocean the Antarctic Circumpolar Current dominates the three choke point sections with an eastward mass flux of $134.9 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ at Drake Passage increasing to $145.6 \pm 1.2 \times 10^6$ m$^3$ s$^{-1}$ south of Australia. Across the Atlantic SAVE2 and SAVE4 sections there is a small southward mass flux of $-0.77 \pm 2.9 \times 10^6$ m$^3$ s$^{-1}$ and $-0.80 \pm 2.7 \times 10^6$ m$^3$ s$^{-1}$. This corresponds to the leakage of Pacific water through Bering Strait and into the North Atlantic Ocean. In the Indian Ocean there is a southward mass flux at 18°S and 32°S of $-10.4 \pm 3.1 \times 10^6$ m$^3$ s$^{-1}$ and $-10.3 \pm 2.3 \times 10^6$ m$^3$ s$^{-1}$, respectively. This southward mass flux through the Indian Ocean is within reasonable agreement of recent estimates of the size of the Indonesian Throughflow (Cresswell et al. 1993; Meyers et al. 1995). The net southward flux from the Indian Ocean results in a net increase in flow south of Australia, accounting for the mass flux increase between south of Africa and Australia. At 32°S in the Pacific Ocean there is a northward mass flux of $7.4 \pm 2 \times 10^6$ m$^3$ s$^{-1}$. [The small mass imbalances in each box result from choosing a solution (rank) where the layer residual norms are of $O(1$ to $2 \times 10^6$ m$^3$ s$^{-1}$). The imbalance does not represent a convergence of freshwater. The freshwater convergence for each box is given by the air–sea freshwater flux and is discussed below.] The mass fluxes across each of the sections (Fig. 2) are in reasonable agreement with previous studies (Rintoul 1991; Toole and Warren 1993; Sloyan and Rintoul 1996).

Table 4. Total section property fluxes (+ve–northward/eastward). Heat fluxes are relative to $0^\circ$C and converted to PW (1 PW = $244 \times 10^{15}$ W). Units: mass $\times 10^6$ m$^3$ s$^{-1}$; heat PW $\times 10^9$; salt $\times 10^8$ kg s$^{-1}$; freshwater $\times 10^9$ kg s$^{-1}$; silica kmol s$^{-1}$.

<table>
<thead>
<tr>
<th>Section</th>
<th>Mass</th>
<th>Heat</th>
<th>Salt</th>
<th>Freshwater</th>
<th>Silica</th>
</tr>
</thead>
<tbody>
<tr>
<td>SAVE2</td>
<td>$-0.77 \pm 2.9$</td>
<td>$0.42 \pm 0.05$</td>
<td>$-28.3 \pm 29.5$</td>
<td>$-0.85 \pm 3.0$</td>
<td>$124 \pm 194$</td>
</tr>
<tr>
<td>SAVE4</td>
<td>$-0.80 \pm 2.7$</td>
<td>$0.28 \pm 0.04$</td>
<td>$-27.6 \pm 55.5$</td>
<td>$-0.83 \pm 2.9$</td>
<td>$295 \pm 339$</td>
</tr>
<tr>
<td>Drake P</td>
<td>$134.9 \pm 0.75$</td>
<td>$1.41 \pm 0.01$</td>
<td>$4785.9 \pm 14.1$</td>
<td>$134.2 \pm 0.83$</td>
<td>$9071 \pm 73$</td>
</tr>
<tr>
<td>S. Africa</td>
<td>$135.4 \pm 2.7$</td>
<td>$1.19 \pm 0.03$</td>
<td>$4795.1 \pm 51.7$</td>
<td>$134.5 \pm 2.8$</td>
<td>$8450 \pm 407$</td>
</tr>
<tr>
<td>Weddell Sea</td>
<td>$-0.35 \pm 2.5$</td>
<td>$-0.05 \pm 0.01$</td>
<td>$-13.7 \pm 57.3$</td>
<td>$-0.35 \pm 2.6$</td>
<td>$-26 \pm 339$</td>
</tr>
<tr>
<td>Ind18</td>
<td>$-10.4 \pm 3.1$</td>
<td>$-1.27 \pm 0.05$</td>
<td>$-371.7 \pm 56.3$</td>
<td>$-10.3 \pm 3.0$</td>
<td>$261 \pm 679$</td>
</tr>
<tr>
<td>Ind32</td>
<td>$-10.3 \pm 2.3$</td>
<td>$-0.87 \pm 0.06$</td>
<td>$-369.1 \pm 43.9$</td>
<td>$-10.2 \pm 2.4$</td>
<td>$425 \pm 316$</td>
</tr>
<tr>
<td>S. Aust.</td>
<td>$145.6 \pm 1.0$</td>
<td>$1.83 \pm 0.01$</td>
<td>$5164.4 \pm 18.8$</td>
<td>$144.7 \pm 1.2$</td>
<td>$8017 \pm 99$</td>
</tr>
<tr>
<td>Pac32</td>
<td>$7.4 \pm 1.8$</td>
<td>$0.13 \pm 0.04$</td>
<td>$261.1 \pm 43.7$</td>
<td>$7.32 \pm 2.0$</td>
<td>$821 \pm 283$</td>
</tr>
</tbody>
</table>
### Table 5. Layer mass (×10⁶ m³ s⁻¹) fluxes for each section (+ve=northward/eastward).

<table>
<thead>
<tr>
<th>Layers</th>
<th>SAVE2</th>
<th>SAVE4</th>
<th>DrakeP</th>
<th>SAfrica</th>
<th>Wedsea</th>
<th>Ind18</th>
<th>Ind32</th>
<th>SAust</th>
<th>Pac32</th>
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<td>*</td>
<td>*</td>
<td>*</td>
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<td>*</td>
<td>*</td>
</tr>
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<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>1.66 ± 0.03</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>4</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>1.17 ± 0.08</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>5</td>
<td>0.24 ± 0.01</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>0.30 ± 0.05</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>6</td>
<td>1.48 ± 0.04</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>0.03 ± 0.06</td>
<td>-0.20 ± 0.00</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>7</td>
<td>0.79 ± 0.06</td>
<td>0.12 ± 0.01</td>
<td>*</td>
<td>0.05 ± 0.00</td>
<td>*</td>
<td>0.71 ± 0.06</td>
<td>-0.11 ± 0.00</td>
<td>*</td>
<td>1.24 ± 0.05</td>
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<tr>
<td>8</td>
<td>2.55 ± 0.08</td>
<td>0.94 ± 0.03</td>
<td>*</td>
<td>-1.01 ± 0.01</td>
<td>*</td>
<td>0.08 ± 0.06</td>
<td>-3.23 ± 0.03</td>
<td>*</td>
<td>0.28 ± 0.07</td>
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<tr>
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<td>3.25 ± 0.07</td>
<td>0.59 ± 0.04</td>
<td>*</td>
<td>-0.35 ± 0.01</td>
<td>*</td>
<td>0.12 ± 0.08</td>
<td>-2.03 ± 0.12</td>
<td>*</td>
<td>9.07 ± 0.12</td>
</tr>
<tr>
<td>10</td>
<td>1.97 ± 0.10</td>
<td>0.88 ± 0.14</td>
<td>1.13 ± 0.01</td>
<td>-0.36 ± 0.06</td>
<td>*</td>
<td>0.33 ± 0.11</td>
<td>-1.53 ± 0.41</td>
<td>0.02 ± 0.01</td>
<td>0.43 ± 0.14</td>
</tr>
<tr>
<td>11</td>
<td>2.32 ± 0.13</td>
<td>3.36 ± 0.16</td>
<td>1.40 ± 0.01</td>
<td>1.85 ± 0.15</td>
<td>*</td>
<td>-0.04 ± 0.13</td>
<td>2.33 ± 0.46</td>
<td>3.75 ± 0.04</td>
<td>-0.73 ± 0.11</td>
</tr>
<tr>
<td>12</td>
<td>1.95 ± 0.14</td>
<td>2.48 ± 0.24</td>
<td>2.30 ± 0.01</td>
<td>4.61 ± 0.13</td>
<td>*</td>
<td>0.80 ± 0.22</td>
<td>-1.11 ± 0.37</td>
<td>14.74 ± 0.17</td>
<td>-1.20 ± 0.13</td>
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<tr>
<td>13</td>
<td>1.79 ± 0.18</td>
<td>1.28 ± 0.20</td>
<td>9.34 ± 0.07</td>
<td>6.62 ± 0.15</td>
<td>0.41 ± 0.01</td>
<td>2.04 ± 0.37</td>
<td>-3.67 ± 0.36</td>
<td>12.38 ± 0.09</td>
<td>1.57 ± 0.33</td>
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<tr>
<td>14</td>
<td>1.73 ± 0.30</td>
<td>-0.39 ± 0.47</td>
<td>18.51 ± 0.11</td>
<td>15.31 ± 0.32</td>
<td>-0.15 ± 0.03</td>
<td>-0.84 ± 0.21</td>
<td>-4.35 ± 0.26</td>
<td>15.61 ± 0.09</td>
<td>-1.61 ± 0.27</td>
</tr>
<tr>
<td>15</td>
<td>0.94 ± 0.40</td>
<td>-1.53 ± 0.48</td>
<td>19.29 ± 0.11</td>
<td>18.89 ± 0.35</td>
<td>-0.21 ± 0.04</td>
<td>-1.70 ± 0.41</td>
<td>-8.52 ± 0.44</td>
<td>15.88 ± 0.13</td>
<td>-3.55 ± 0.25</td>
</tr>
<tr>
<td>16</td>
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<td>-1.96 ± 0.56</td>
<td>19.48 ± 0.13</td>
<td>19.32 ± 0.45</td>
<td>0.66 ± 0.06</td>
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<td>-5.67 ± 0.54</td>
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<td>-8.42 ± 0.41</td>
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<td>17</td>
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<td>30.45 ± 0.32</td>
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<td>-1.48 ± 0.20</td>
<td>-1.24 ± 1.02</td>
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<td>-9.69 ± 2.03</td>
<td>-10.70 ± 1.40</td>
<td>15.98 ± 0.32</td>
<td>22.54 ± 1.25</td>
<td>-1.24 ± 0.26</td>
<td>-3.21 ± 1.36</td>
<td>8.71 ± 1.11</td>
<td>21.72 ± 0.44</td>
<td>-0.37 ± 0.88</td>
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<td>19</td>
<td>1.79 ± 1.48</td>
<td>0.74 ± 1.14</td>
<td>12.90 ± 0.38</td>
<td>18.25 ± 1.31</td>
<td>-2.34 ± 0.53</td>
<td>5.64 ± 2.48</td>
<td>4.23 ± 1.99</td>
<td>13.12 ± 0.51</td>
<td>18.56 ± 0.77</td>
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<tr>
<td>20</td>
<td>3.40 ± 0.62</td>
<td>5.49 ± 1.22</td>
<td>4.18 ± 0.42</td>
<td>1.47 ± 1.17</td>
<td>7.47 ± 1.62</td>
<td>1.11 ± 0.04</td>
<td>10.40 ± 0.85</td>
<td>5.40 ± 0.46</td>
<td>8.02 ± 0.28</td>
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<td>21</td>
<td>*</td>
<td>1.11 ± 0.51</td>
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<td>4.31 ± 1.16</td>
<td>1.72 ± 1.82</td>
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<td>-0.67 ± 0.45</td>
<td>9.74 ± 0.55</td>
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<td>23</td>
<td>*</td>
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<td>*</td>
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</tr>
</tbody>
</table>
Saunders and King 1995; Macdonald and Wunsch 1996; Robbins and Toole 1997; Rintoul and Bullister 1999).

b. Western boundary current transports

The western boundary currents are reproduced in the inverse solution. In the Atlantic, at 12°S (SAVE2) the weak, shallow beginnings of the Brazil Current (−2.5±0.4 × 10⁶ m³ s⁻¹) are found. In the Argentine Basin the strong northward Malvinas Current (57.7±3.2 × 10⁶ m³ s⁻¹), resulting from the bottom velocities imposed there in the initial guess, is maintained. The Malvinas constraint concentrates the southward NADW transport to the Brazilian Basin at SAVE2, and increases the southeastward flow of intermediate and deep water over the mid-Atlantic Ridge at SAVE4 (Sloyan 1997).

Adjustments to the circulation in the Atlantic sector in response to the Malvinas constraint also affect the Indian and Pacific Oceans, resulting in a 2 × 10⁶ m³ s⁻¹ decrease in the net southward, eastward, and northward transport across the Indian Ocean, south of Australia and at 32°S in the Pacific, respectively.

In the Indian Ocean at 18°S the East Madagascar Current transport is −28.1±5 × 10⁶ m³ s⁻¹; previous estimates vary from −41 × 10⁶ m³ s⁻¹ to −20 × 10⁶ m³ s⁻¹ (Lutjeharms et al. 1981; Schott et al. 1988; Swallow et al. 1988). At 32°S in the Indian Ocean the Agulhas Current transport is −48.3±6 × 10⁶ m³ s⁻¹ within 400 km of the African coast. This is composed of −52.3 × 10⁶ m³ s⁻¹ southward flow above 2000 dbar and 4 × 10⁶ m³ s⁻¹ northward below 2000 dbar. Using the same section Toole and Warren (1993) estimate a topo-bottom transport of −79.5 × 10⁶ m³ s⁻¹ (−80.7 × 10⁶ m³ s⁻¹ southward above 2000 dbar and 1.2 × 10⁶ m³ s⁻¹ northward below 2000 dbar) and a net southward transport of −85 × 10⁶ m³ s⁻¹ was found within 400 km of the African coast. At the same 32°S section, Robbins and Toole (1997) and Macdonald (1998) estimate an Agulhas Current of −90 × 10⁶ m³ s⁻¹ and −93±2 × 10⁶ m³ s⁻¹, respectively. Estimates from a Lowered Acoustic Doppler Current Profiler (ADCP) within 230 km of the African coast show a volume transport of −71 × 10⁶ m³ s⁻¹ (Beal and Bryden 1997).

The LADCP data show a V-shaped zero velocity surface (ZVS) with a maximum depth of 2500 m some 80 km offshore (Beal and Bryden 1997). Inshore there is evidence of a northward undercurrent close to the continental slope. In the Natal Basin Toole and Warren (1993) use a zero velocity surface of 2000 m, which increases to 3000 m in the deepest part of the Natal Basin. [Robbins and Toole (1997) and Macdonald (1998) employ the Toole and Warren (1993) zero velocity surface.] Beal and Bryden (1997) find that the difference between their Agulhas transport and that of Toole and Warren (1993) is due to their use of the V-shaped ZVS. In this study all ZVSs are neutral density surfaces that may deepen or shoal across an ocean basin. In the Natal Basin the ZVS is neutral density surface γ'' = 28.0; at station pairs where this layer is not found a deepest common ZVS is used. Here γ'' = 28.0 shows a similar, although not as sharp, V-shaped ZVS to that of Beal and Bryden (1997)—it is above 2000 dbars adjacent to the African coast; in the middle of the Natal Basin it reaches a maximum depth of 2500, and then shoals slightly before “grounding” at the Mozambique Ridge. The different ZVS employed in this study may partly explain the smaller Agulhas Current compared to those of Toole and Warren (1993), Robbins and Toole (1997), and Macdonald (1998). A more significant factor that affects the strength of the Agulhas Current is the difference in the vertical extent of the deep overturning circulation between our studies and those previously quoted. In Toole and Warren (1993), Robbins and Toole (1997), and Macdonald (1998) northward flowing deep and bottom water is balanced by southward flow of intermediate and thermocline water in the Agulhas Current. Our deep overturning cell is confined to below 1500 dbars and the balancing southward flow occurs east of 90°E, as discussed below.

In the Pacific Ocean at 32°S we find a relatively weak East Australian Current of −21.3±2 × 10⁶ m³ s⁻¹, which agrees with recent estimates (Ridgway and Godfrey 1994; Macdonald 1998).

c. Heat transport

All heat fluxes are temperature fluxes, relative to 0°C, and scaled to PW, where 1 PW (≈10¹⁵ W) = 244 × 10⁶ C m³ s⁻¹. Across SAVE4 (nominally 37°S) there is a northward heat flux of 0.28 ± 0.04 PW. This estimate agrees with that of Rintoul (1991) but is slightly smaller than more recent estimates of 0.49 ± 0.25 PW by Macdonald (1998) and 0.5 ± 0.1 PW by Saunders and King (1995). The northward heat flux at SAVE2 is 0.42 ± 0.05 PW. A heat input of 0.14 PW (positive heat flux represents heat gain by the ocean) in the subtropical Atlantic results from the COADS heat fluxes and the inverse corrections (~10%) (Fig. 3). This heat input is larger than Talley (1984) but similar to estimates of Trenberth and Solomon (1994). The inverse model of Macdonald (1998) implies an even larger heat input in the subtropical Atlantic, the result of weak heat loss (~0.16 ± 0.25 PW) between 27° and 23°S and strong heat gain between 23° and 11°S (0.56 ± 0.25 PW).

In the subtropical Indian Ocean the heat flux is −1.27 ± 0.05 PW at 18°S and −0.87 ± 0.06 PW at 32°S. The large southward heat flux at 18°S is the result of both heat gained by the ocean in the northern Indian Ocean and heat entering from the Pacific Ocean via the Indonesian Throughflow. At the same sections Macdonald (1998) estimates a southward flux of −1.45 ± 0.30 PW and −1.30 ± 0.28 PW, respectively, while at 32°S Toole and Warren (1993) estimate a southward flux of −1.67 PW. These previous heat flux estimates are significantly larger than found in this study, although all studies have a similar sized Indonesian Throughflow of −10 × 10⁶
Our smaller southward heat flux results from the reduced vertical extent of the deep and bottom overturning cell when compared to the previous studies, as discussed in detail in the following section.

In the subtropical Indian Ocean between 18° and 32°S the air–sea fluxes (COADS plus inverse correction) result in an oceanic heat loss of −0.40 PW. This is similar to estimates of Trenberth and Solomon (1994) but larger than estimates of Talley (1984) and the −0.15 ± 0.30 PW found by Macdonald (1998).

At 32°S in the Pacific Ocean there is a northward heat flux of 0.13 ± 0.04 PW. Macdonald (1998) estimates a northward heat flux of 0.26 ± 0.28 PW at 43°S and a small southward heat flux of −0.04 ± 0.30 PW at 28°S. Wunsch et al. (1983), using the same 43°S and 28°S section as that of Macdonald (1998), estimate a south-
ward flux of $-0.03 \pm 0.35$ PW and $-0.18 \pm 0.22$ PW, respectively. [Note, however, that Wunsch et al. (1983) set the throughflow to zero.]

As the circulation of the Indian and Pacific Oceans are linked via the Indonesian Throughflow it is more appropriate to consider the combined heat flux of these oceans. At $32^\circ$S the southward Indian–Pacific heat flux is $-0.73 \pm 0.10$ PW. Including the heat flux at SAVE4 provides us with an estimate of the net meridional heat flux at nominally $35^\circ$S (SAVE4, Ind32, and Pac32) of $-0.46 \pm 0.14$ PW. A similar calculation by Macdonald (1998) at $30^\circ$S gives an estimated southward heat flux of $-0.9 \pm 0.3$ PW.

Across the Southern Ocean choke point sections the eastward heat fluxes are $1.41 \pm 0.01$ PW at Drake Passage, $1.19 \pm 0.03$ PW south of Africa, and $1.83 \pm 0.01$ PW south of Australia. [The reader is reminded that the errors quoted represent only the formal error and therefore may not reflect the true uncertainty in the estimates (see end of section 2b).] In the Southern Ocean Atlantic sector there is also a southward flux of $-0.05 \pm 0.01$ PW into the Weddell Sea. The heat fluxes at Drake Passage and south of Africa are similar to estimates of Macdonald (1998) while south of Australia there is a $0.10$ PW difference in the estimated heat flux. The inverse corrected air–sea heat fluxes show that the Atlantic sector (Box II) gains $+0.11$ PW from the atmosphere. This heat gain and the lateral heat flux through Drake Passage balance the ocean heat export south of Africa, into the Weddell Sea, and northward across SAVE4. In the Southern Ocean Indian sector the corrected air–sea fluxes indicate a heat loss from the ocean to the atmosphere of $-0.22$ PW. This heat loss is more than compensated by the southward heat flux across $32^\circ$S in the Indian Ocean. In the Pacific sector the ocean loses $-0.31$ PW to the atmosphere. The ocean–atmosphere heat loss and the northward ocean heat flux at $32^\circ$S balance the difference between the eastward heat flux south of Australia and Drake Passage.

The air–sea heat fluxes in the Southern Ocean Indian and Pacific sectors in this study are similar to those found by Georgi and Toole (1982). The similarities in the estimated air–sea heat flux result in part from the use of the same observation data between the two studies; that is, COADS is derived from the heat flux estimates used by Georgi and Toole (1982). In the Atlantic sector Georgi and Toole (1982) estimate an ocean to atmosphere heat loss of $-0.16$ PW. This ocean heat loss is estimated from the integration of air–sea heat fluxes of Taylor et al. (1978) and Gordon (1981) south of $40^\circ$S. The difference in the air–sea heat flux in the Southern Ocean Atlantic sector between our study and Georgi and Toole (1982) results in part from our use of more recent estimates suggesting smaller heat loss [$-16$ W m$^{-2}$ (Gordon and Huber 1990; Fahrbach et al. 1994)] over the Weddell Sea.

In the Southern Ocean Macdonald (1998) estimates a larger air–sea ocean heat gain in the Atlantic sector of $+0.3 \pm 0.1$ PW, a larger heat loss in the Indian sector of $-0.7 \pm 0.4$ PW, and a smaller heat loss in the Pacific sector of $-0.1 \pm 0.3$ PW. The most significant difference in the estimated air–sea heat flux between Macdonald (1998) and this study occurs in the Southern Ocean Indian sector. This difference may be linked to the difference in southward heat flux at $32^\circ$S in the Indian Ocean. Macdonald (1998) has a large southward ocean heat flux across $32^\circ$S in the Indian Ocean. To reconcile this large heat input into the Southern Ocean Indian sector with the heat divergence in the ACC between Australia and Africa, a large air–sea ocean heat loss is required. In our model the smaller southward heat flux across $32^\circ$S is compatible with the ACC heat divergence between Australia and Africa and the COADS heat flux (Fig. 3).

The most significant difference between this study and the earlier study of Georgi and Toole (1982) is the calculation of ocean heat flux across the three choke point sections. Georgi and Toole (1982) assume that the ACC mass transport is constant at $127 \times 10^6$ m$^3$ s$^{-1}$ at all choke points, and effectively solve a heat balance assuming no Indonesian Throughflow. In our model the ACC transport at Drake Passage and south of Africa is approximately $135\pm2 \times 10^6$ m$^3$ s$^{-1}$ while south of Australia the ACC transport increases to $145\pm1 \times 10^6$ m$^3$ s$^{-1}$ as required to offset the $10\pm3 \times 10^6$ m$^3$ s$^{-1}$ Indonesian Throughflow. The differences in mass transport result in significantly different ocean heat fluxes at the choke point sections, and hence in the meridional heat fluxes across $40^\circ$S inferred from the choke point heat transport divergences.

d. Salt, freshwater, and silica transport

The net salt flux in the Atlantic at SAVE2 and SAVE4 is $-28.3\pm29.5 \times 10^6$ kg s$^{-1}$ and $-27.6\pm55.5 \times 10^6$ kg s$^{-1}$, respectively. This salt flux is in reasonable agreement with the salt flux of $26.7 \times 10^6$ kg s$^{-1}$ from the Pacific Ocean to the Atlantic through Bering Strait (Wijffels et al. 1992). (The net salt fluxes include the salt flux associated with a nonzero mass flux at a section and the freshwater divergence in the box.) The import of salt from the subtropical Atlantic increases the salt flux south of Africa relative to Drake Passage. In the subtropical Indian Ocean there is a net southward salt flux associated with the input of Pacific water into the Indian Ocean via the Indonesian Throughflow (Table 4). The input of salt from the Indian Ocean results in a larger eastward salt flux south of Australia than south of Africa. The import of salt into the Southern Ocean in the Atlantic and Indian sector is balanced by a northward flux of salt in the Pacific Ocean across $32^\circ$S.

The air–sea freshwater fluxes from inverse-corrected GASP data (Fig. 4) show that the subtropical Atlantic (Box I) and subtropical Indian (Box IV) Oceans lose $-0.34 \times 10^9$ kg s$^{-1}$ and $-0.33 \times 10^9$ kg s$^{-1}$ of freshwater via evaporation while the Southern Ocean Atlantic
The silica flux is dominated by large eastward fluxes at the Southern Ocean choke point sections (Table 4). The net northward silica flux at 32°S in the Indian Ocean of $424.7 \pm 316.4$ kmol s$^{-1}$ is significantly smaller than the northward flux of 1532 kmol s$^{-1}$ estimated by Toole and Warren (1993), but is larger than the estimate of 21
kmol s$^{-1}$ by Robbins and Toole (1997). Independent estimates of the global silica budget suggest that the Southern Ocean is a net sink for silica (DeMaster 1981; Tréguer et al. 1995). Therefore, the size of the northward silica flux at 32°S in the Pacific of 821.4 ± 283 kmol s$^{-1}$ and the large net meridional flux of silica across 30°–40°S are of concern and led to additional experiments as described in the Appendix. The transports of mass and other properties in the standard model and the silica experiments are the same within the given errors. A close examination of the fluxes in individual layers reveals how the solution responds to the requirement that silica be more exactly conserved. The relatively tight silica constraints in experiments 2 and 3 (appendix) are met by converting intermediate water (IW) in the Indian and Pacific regions to upper deep water (UDW), and producing an eastward flow of extreme AABW south of Africa and Australia. The conversion of low silica, high oxygen IW to high silica, low oxygen intermediate deep water (IDW) within the Indian Ocean is hard to reconcile with our understanding of the circulation and tracer patterns in this basin. Recent studies also support westward flow of extreme AABW south of Australia and Africa (Frew et al. 1995; Rintoul 1998), as found in the standard model. Therefore we consider the solutions with the tight silica constraints (experiments 2 and 3) to be unlikely. The changes required to meet the weaker constraints of experiment 1 [fluxes constrained to be 0 ± 200 kmol s$^{-1}$ and total silica conserved to within $O(500$ kmol s$^{-1}$)] are more modest, and the resulting circulation is plausible given our understanding. In any case, the inclusion of tighter constraints on silica does not change the main conclusions of the study: the deep overturning circulation is increased when silica is more tightly constrained, and the role of air–sea fluxes in converting UDW to IW is unchanged.


In the following sections we present a detailed analysis of the deep overturning circulation in the individual ocean basins, including both lateral and diaxial fluxes. The flux estimates are compared to previous studies, in particular the global analyses of Schmitz (1995, 1996b) and Macdonald (1998). The multiplicity of water mass definitions in the literature, many of them regionally specific, can create confusion. In the following, we specify the layers to which each water mass corresponds; the neutral densities defining each layer and their average temperature and salinity are listed in Table 2. Note, for example, that bottom water here refers to water denser than $\gamma^* = 28.2$, a slightly broader definition than the common criterion of water with potential temperature less than 0°C.

a. Subtropical Atlantic

About $20 \times 10^6$ m$^3$ s$^{-1}$ of North Atlantic Deep Water (layer 17 to 18) enters the subtropical (South) Atlantic across SAVE2 at 12°S (Fig. 5). The southward flow of NADW occurs principally in the western Brazil Basin ($-17\pm 5 \times 10^6$ m$^3$ s$^{-1}$) with only a small southward flux in the eastern basin over the Guinea Rise ($-3\pm 3 \times 10^6$ m$^3$ s$^{-1}$). Below NADW there is a northward flux of $5\pm 2 \times 10^6$ m$^3$ s$^{-1}$ of Lower Circumpolar Deep Water (LCDW, layer 19) and Antarctic Bottom Water (AABW, layer 20). The magnitude of NADW flowing south and its principal confinement to the western basin agrees with previous studies (McCartney and Curry 1993; Dickson and Brown 1994). At 23°S, Macdonald (1998) estimates a southward flux of $-20 \times 10^6$ m$^3$ s$^{-1}$ NADW and a northward flux of $6 \times 10^6$ m$^3$ s$^{-1}$ AABW, although the southward flow of NADW is evenly split between the western and eastern basins.

The southern boundary of the subtropical Atlantic region (Box I) is SAVE4. Because SAVE4 is not strictly zonal (Fig. 1), it is useful to divide discussion of the deep fluxes into the western Argentina Basin, the Mid-Atlantic Ridge and Angola Basin, and the eastern Cape Basin. In the Argentine Basin $36\pm 2 \times 10^6$ m$^3$ s$^{-1}$ of Upper Circumpolar Deep water (UCDW layers 15 to 17) and $17\pm 2 \times 10^6$ m$^3$ s$^{-1}$ LCDW (layer 18 and 19) move northward into the subtropical Atlantic (Fig. 5). Below these layers there is a $6\pm 2 \times 10^6$ m$^3$ s$^{-1}$ northward flux of AABW (layer 20) and colder Weddell Sea Deep Water (WSDW, layer 21). Southeastward flow of deep water (layers 15–19) occurs over the mid-Atlantic ridge and in the Angola Basin, with $43\pm 3 \times 10^6$ m$^3$ s$^{-1}$ of UCDW/NADW and $26\pm 3 \times 10^6$ m$^3$ s$^{-1}$ of NADW/LCDW. A small southward flux of NADW ($2\pm 1 \times 10^6$ m$^3$ s$^{-1}$) is also found in the Cape Basin, while northward penetration of dense AABW into the Cape Basin is restricted by the Walvis ridge.

This flow structure is clearly evident when looking at the cumulative layer fluxes (west to east) across SAVE4 (Fig. 6). The recirculating southward–northward flow of UCDW/NADW (15–17) and NADW/LCDW (18–19) between 55° and 40°W in the Argentine Basin is associated with a salinity maximum ($\approx 34.8$ psu). Northward fluxes of UCDW, LCDW, and AABW occur in the eastern part of the Argentine Basin. The northward flow of UCDW and LCDW is more than compensated by returning southward flow of these waters and a net input of NADW over the mid-Atlantic Ridge and in the Angola Basin. Saunders and King (1995) show a similar cumulative southward flux of deep water between 14°W and 6°E, although they find more NADW moving southward in the Cape Basin than suggested in this study. [Differences in the fluxes in the deep layers result in part from the inclusion of our lightest deep water (15) in intermediate water by Saunders and King (1995).] The northward flow in the western basin, southeastward flow over the ridge, and small
southward flow in the eastern basin agrees with that shown by Reid (1989, his Figs. 22–25).

The saline NADW flowing to the south at SAVE2 is sandwiched between fresher UCDW and Antarctic Intermediate Water above and fresher LCDW and AABW below. Sloyan and Rintoul (2000) show that the diapycnal fluxes between these water masses along with isopycnic mixing between deep water varieties can explain the changes in temperature and salinity in these layers in the subtropical Atlantic. The mixing of AABW with overlying NADW/LCDW results in the conversion of $3 \times 10^6 \text{m}^3\text{s}^{-1}$ of AABW to NADW/LCDW. This mixing results in the warmer, saltier NADW/LCDW observed to move northward across SAVE2 into the North Atlantic, and the cooler, fresher NADW at SAVE4 relative to SAVE2. Across the upper boundary of NADW (layer 17) there is a salt flux into less saline UCDW and intermediate water [Antarctic Intermediate Water (AAIW) and Subantarctic mode water (SAMW)]. This salt flux drives an interior $8 \times 10^6 \text{m}^3\text{s}^{-1}$ mass flux of NADW into overlying UCDW, of which $4 \times 10^6 \text{m}^3\text{s}^{-1}$ is further modified into AAIW and SAMW. The mass and salt flux from NADW into overlying UCDW, AAIW, and SAMW results in a salinity increase of these water masses in the subtropical Atlantic and an erosion of the NADW salinity maximum.

In the subtropical South Atlantic Macdonald and Wunsch (1996) find a net upwelling of $1 \times 10^6 \text{m}^3\text{s}^{-1}$ across the 3.5°C isotherm. This isotherm is slightly above the salinity maximum core (~2000 dbar) at 12°S but in the eastern (Angola and Cape) basins at 30°S it is in the salinity minimum AAIW water. The interpretation of this upward flux in terms of mixing of specific water masses is not clear due to the isotherm (in the South Atlantic) intersecting two water masses.

Schmitz (1996b) estimates $3 \times 10^6 \text{m}^3\text{s}^{-1}$ of AABW is converted to less dense NADW/LCDW, similar to our estimate (see his Fig. II-155, between “Agulhas Leakage” and the equator). He finds $8 \times 10^6 \text{m}^3\text{s}^{-1}$ of denser NADW upwells into lighter UCDW, again similar to our study (8±3 $10^6 \text{m}^3\text{s}^{-1}$ flux of NADW/LCDW to UCDW). The main difference between our study and Schmitz (1996b) is that we estimate approximately half of the diapycnal flux between NADW and UCDW is eventually converted to intermediate water by interior mixing, while Schmitz (1996b) has no conversion of NADW to intermediate water in the subtropical South Atlantic.

b. Atlantic sector of the Southern Ocean

The ACC carries bottom and deep water masses eastward across the Atlantic sector (Box II) of the Southern Ocean (Fig. 5). At Drake Passage $69 \pm 1 \times 10^6 \text{m}^3\text{s}^{-1}$ of UCDW (layers 15 to 17), $29 \pm 2 \times 10^6 \text{m}^3\text{s}^{-1}$ of LCDW (layers 18 and 19), and $4 \pm 1 \times 10^6 \text{m}^3\text{s}^{-1}$ of

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**Fig. 5.** Lateral and diapycnal volume fluxes ($\times 10^6 \text{m}^3\text{s}^{-1}$) in bottom and deep layers for the Atlantic region (Box I and II). (top) The volume flux driven by air–sea buoyancy forcing acting on the lightest outcropping layer (15). Diapycnal fluxes due to interior mixing are indicated by curly arrows, solid arrow shows flux across lower interface, and dashed arrow flux across upper interface. Upward diapycnal arrows represent upwelling and downward arrows downwelling across the interface. Error estimates for the lateral and diapycnal fluxes are given in the text.
AABW (layer 20) enter the region. Across SAVE4 there is a net import of $17 \pm 3 \times 10^6$ m$^3$ s$^{-1}$ of modified NADW (layer 15–19) into the Southern Ocean and an export of $6 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ of AABW into the Argentine Basin. Across the Weddell Sea section an overturning circulation carries a southward flux of $-11 \pm 3 \times 10^6$ m$^3$ s$^{-1}$ of LCDW/AABW/UCDW balanced by a northward flux of $11 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ of denser WSDW/WSBW (Weddell Sea Bottom Water). The net exchange of LCDW/AABW with WSDW/WSBW in the Weddell Sea agrees with a recent estimate of Fahrbach et al. (1994). South of Africa there is a net import of $3 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ of extreme AABW (layer 21) from the Indian Ocean. Import of extreme bottom water into the Atlantic sector from the Indian Ocean is supported by evidence for AABW formation off Enderby Land (Jacobs and Georgi 1977; Mantisi et al. 1991), and recent studies showing bottom water formed along the Adélie coast flows westward through the Princess Elizabeth Trough to enter the Atlantic (Frew et al. 1995; Rintoul 1998).

The exchange of AABW, CDW, and NADW between the Atlantic sector of the Southern Ocean and neighboring basins results in changes to the volume flux and temperature and salinity characteristics of these deep water masses between Drake Passage and south of Africa (Sloyan and Rintoul 2000). The $14 \pm 5 \times 10^6$ m$^3$ s$^{-1}$ of extreme bottom water (layers 21 to 23) entering the Atlantic from the Weddell Sea and Indian sector mixes with LCDW entering through Drake Passage to produce warmer, saltier AABW (layer 20). The modified AABW spreads northward into the Argentine Basin and replaces water in this layer lost to the Weddell Sea. One-third, or $5 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ of AABW (layers 20 to 23), is eventually converted to LCDW (Fig. 5), resulting in cooler, fresher NADW/LCDW exiting the basin south of Africa. There is only a small interior diapycnal flux of $1 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ of UCDW/NADW to denser NADW/LCDW.

In the Southern Ocean Atlantic sector the deep water isopycnals shoal and the UCDW/NADW (layers 15–17) outcrops. The outcropping of upper deep waters exposes them to air–sea fluxes (wind, heat, and freshwater), which modify their temperature and salinity properties resulting in formation of cold, fresh AASW. The buoyancy forcing drives a conversion of $8.4 \times 10^6$ m$^3$ s$^{-1}$ of AASW to lighter water across outcropping surface

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**Fig. 6.** Cumulative flux in AABW (open circle), NADW/LCDW (open square), and UCDW/NADW (open diamond) across SAVE4 from the western Argentine Basin to the eastern Cape Basin. Positive is northward. Lower plot is the salinity (solid: psu) distribution along the section, the bounding neutral density (dot-dash: kg m$^{-3}$) of the deep water masses: UCDW/NADW $\gamma^\prime < 28.0$, NADW/LCDW $28.0 < \gamma^\prime < 28.2$ and, AABW $\gamma^\prime > 28.2$. Southward mass flux is shaded.
15 (Fig. 5). This diapycnal transport driven by air–sea fluxes is partly compensated by an interior diapycnal flux \( (4 \pm 2 \times 10^6 \text{ m}^3 \text{ s}^{-1}) \) of intermediate water (layer 14) into the lightest deep water layer (15).

The mixing of NADW with LCDW and conversion of UCDW to intermediate water presented here differs from the circulation scheme proposed by Schmitz (1996b) (see his Fig. II-155, south of the Agulhas Leakage). He indicates a conversion of NADW/LCDW \( (10 \times 10^6 \text{ m}^3 \text{ s}^{-1}) \) to intermediate water, with the rest of the southward flowing LNADW \( (4 \times 10^6 \text{ m}^3 \text{ s}^{-1}) \) being converted to AABW. He also shows \( 8 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) of UNADW flowing southward into the Southern Ocean of which \( 4 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) is converted to intermediate water and the remainder is converted to AABW. In contrast we find that NADW spreads south along density surfaces to mix with LCDW entering the basin through Drake Passage, thereby reinforcing the salinity maximum of LCDW leaving the basin south of Africa. The injection of NADW into CDW balances the LCDW lost in the production of WSDW/WSBW and subsequent entrainment to form AABW, as well as the transformation of UCDW to intermediate water driven by air–sea fluxes. If NADW were to be directly converted to intermediate water in the Atlantic sector of the Southern Ocean, it must be accomplished by interior mixing processes, as these layers do not outcrop in this sector. However, the interior diapycnal fluxes found here show a net conversion of AABW to LCDW, rather than NADW/LCDW to intermediate water. The main impact of this study on the diagram of Schmitz (1996b) is to remove the \( 10 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) (light blue in his figure) conversion of NADW to intermediate water. Instead, the NADW exported from the Atlantic increases the volume flux of LCDW leaving the basin south of Africa over that entering through Drake Passage. [The divergence of LCDW in the Atlantic sector of the ACC is balanced by a convergence of SAMW, as in Rintoul (1991) and Sloyan and Rintoul (2001).]

Across the 3.5°C isotherm Macdonald and Wunsch (1996) estimate a \( 9 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) downwelling. In the Southern Ocean Atlantic sector this isotherm is essentially the boundary of UCDW/NADW and intermediate water. This large downwelling from intermediate water to deep water disagrees with the net conversion of AASW (air–sea modified upper deep water) to intermediate water found in this study.

c. Subtropical Indian Ocean

Across 18°S, in the subtropical Indian Ocean (Box IV) a small amount of AABW moves northward in the western Mascarene Basin \( (1 \pm 1 \times 10^6 \text{ m}^3 \text{ s}^{-1}) \) while a net transport of \( 6 \pm 3 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) of LCDW (layer 19) is achieved by northward flow in the eastern Western Australian Basin and overlying the bottom water in the Mascarene Basin (Fig. 7). The northward flow of Antarctic origin water is balanced by \( 7 \pm 3 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) of Indian Deep Water returning southward above CDW/AABW in the Madagascar and Central Indian Basin. (The \( 3 \pm 1 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) southward flow in layer 18 occurs in the central Indian Basin). The inflow of water of Antarctic origin (layers 20 and 19) in the Mascarene and West Australian Basin agrees with water mass distributions on potential density surfaces shown by Mantyla and Reid (1995).

At 18°S, 105°E there is an increase in the northward transport of LCDW (Fig. 7), associated with a decrease in the temperature and increase in density of LCDW (below 3000 db) in comparison with water directly to the west [see Plates 1 and 3 of Warren (1981)]. The deep water below 3000 db at 105°E has similar properties to deep water found at the western boundary of the West Antarctic Basin. The cold LCDW at 105°E must be fed by a southern source. At 105°E the eastern Indian Ocean is directly open to the South Australian Basin through the Perth Basin (Fig. 1). The northward flow of LCDW near 105° could be influenced by topographic features at 20°S (see Fig. 1, unlabeled topographic features), which rise to above 3000 dbar and a string of topographic features (not shown) farther south that rise above 4000 dbar. These topographic features may enhance/maintain direct northward flow of AABW/LCDW near 105°E in the Perth and West Australian Basins. This suggests that there are two deep western boundary currents in the eastern Indian Ocean: adjacent the Ninetyeast Ridge, and along 105°E.

Farther south at 32°S the Indian Ocean is divided into numerous basins by ridges and plateaus. The effect of topography is to guide and restrict the \( 23 \pm 4 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) northward flow of AABW (layer 20, \( 10 \pm 2 \times 10^6 \text{ m}^3 \text{ s}^{-1} \)) and LCDW (layer 18 and 19, \( 13 \pm 2 \times 10^6 \text{ m}^3 \text{ s}^{-1} \)) into the western Natal, Mozambique, Madagascar, Crozet, and Perth Basins (Fig. 8). The northward transport of Antarctic bottom and lower deep water is partially balanced by \( 20 \pm 2 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) of southward IDW/UCDW (layers 17–15). The southward flow of IDW/UCDW occurs east of 80°E in the Central Indian Basin, over the Broken Plateau, and in the Perth Basin, although there is some southward flow in the western (Natal and Madagascar) basins.

The northward flow of Antarctic origin bottom and deep water in the numerous western boundary currents agrees with previous analysis of water properties (Gordon and Tchernia 1972; Toole and Warren 1993; Mantyla and Reid 1995). The southward flow of IDW between 80°E to Australia is associated with an oxygen minimum and nutrient (phosphate and nitrate) maximum. These property characteristics are derived from the north. Much of the southward flow of IDW occurs over and adjacent to the Broken Ridge and in the eastern Perth Basin, Toole and Warren (1993), using the same 32°S section, find a net northward transport of water below 2000 dbar in the Perth Basin, but note that strong southward transport of deep water is found in the eastern part of the basin. They also find weak transport over
the Broken Plateau below 1000 dbar. Robbins and Toole (1997) find a small net southward transport of IDW (their layer 5) across the entire subtropical basin. Talley and Baringer (1997) find southward transport above 2000 dbar between 100° and 103.5°E and east of 109°E. South of Australia at 115°E Hufford et al. (1997) finds an eastward flowing deep boundary current of $1 \times 10^6$ m$^3$ s$^{-1}$ to $19 \times 10^6$ m$^3$ s$^{-1}$, which has water properties similar to the southward flowing IDW in the Perth Basin.

The southward flow of IDW/UCDW (layers 15–17) across 32°S is supplied by an upward diapycnal flux of $20 \pm 3 \times 10^6$ m$^3$ s$^{-1}$ of LCDW to IDW/UCDW between 32° and 18°S and southward flow ($4 \pm 1 \times 10^6$ m$^3$ s$^{-1}$) of IDW (layers 17–15) across 18°S (Fig. 9). The net upward diapycnal flux of LCDW to IDW/UCDW results from modification of LCDW, AABW ($9 \pm 2 \times 10^6$ m$^3$ s$^{-1}$), and IDW (layer 18) to less dense IDW/UCDW. Most of the upwelled bottom and lower deep water remains in the UCDW and IDW returning to the Southern Ocean sector and only a small portion ($4 \pm 2 \times 10^6$ m$^3$ s$^{-1}$) is converted further into intermediate and thermocline water masses (Fig. 9). This means that the large ($23 \pm 3 \times 10^6$ m$^3$ s$^{-1}$) overturning Indian Ocean circulation found in this study is effectively balanced below 1500 dbar.

Estimates of the strength of the deep overturning circulation in the Indian Ocean vary considerably (Table 6). Schmitz (1996b) provides two scenarios of the overturning circulation (Fig. II-130 and Fig. II-156), the former based on $13 \times 10^6$ m$^3$ s$^{-1}$ overturning from Robbins and Toole (1997) while the latter is similar to the circulation scheme in Schmitz (1995). The difference of $11 \times 10^6$ m$^3$ s$^{-1}$ between the southward flow of IDW/UCDW is balanced by a difference in the northward flux of bottom and lower deep water across 32°S in the two scenarios. Opposed to the hydrographic studies is the estimate of a small ($4 \times 10^6$ m$^3$ s$^{-1}$) deep inflow from a primitive equation model that assimilates the Levitus and Boyer (1994) climatological temperature and salinity fields (Lee and Marotzke 1997). This climatology provides a smoothed temperature and salinity field that may not adequately resolve the deep western boundary currents. As shown in Fig. 8 it is at the western boundaries that strong northward flow of LCDW and AABW is found—not resolving these features will likely lead to smaller northward fluxes of these...
FIG. 8. Cumulative flux in AABW (open circle), LCDW (open square), and IDW/UCDW (open diamond) across Indian 32°S from the western Natal Basin to the eastern Perth Basin. Positive is northward. Lower plot is the oxygen (solid: μmol kg⁻¹) distribution along the section, the bounding neutral density (dot-dash: kg m⁻³) of the deep water masses: IDW/UCDW γ° < 28.0, LCDW 28.0 < γ° < 28.2, and AABW γ° > 28.2. Southward mass flux is shaded.

water masses and a small overturning circulation, as also noted by Robbins and Toole (1997).

MERIDIONAL FLUXES

The Toole and Warren (1993) overturning estimate results in a large net northward silica flux (Table 6). This large silica flux from the Southern Ocean is difficult to reconcile with global silica budgets, which suggest that the Southern Ocean is a net sink for silica (DeMaster 1981; Tréguer et al. 1995). In Robbins and Toole (1997) the small northward silica flux is achieved by reducing the northward flux of Antarctic origin bottom and deep water. The present study still maintains a large overturning circulation but has a much reduced silica flux compared to Toole and Warren (1993). The reduction of the northward silica flux, while still maintaining a large overturning circulation, is achieved by a significant deepening of the returning southward flow. Figure 10 shows a comparison of the overturning scheme of Toole and Warren (1993), Robbins and Toole (1997), and this study. (Note that the discrepancy in the deepest layer between our results and the other studies is largely due to different layer definitions; the sum over their deepest two layers covers a similar density range to our layers 19 and 20, and the sum of the fluxes in these layers is similar in each of the studies.) In this study the inflowing AABW and LCDW is balanced below 1500 dbar by southward flow of UCDW and IDW, while in the previous studies the northward flow of AABW and LCDW is largely balanced by intermediate and thermocline water. Indian Deep Water and UCDW have a significantly higher silica concentration than intermediate and thermocline water, resulting in the smaller northward silica flux found here compared to Toole and Warren (1993).

The difference in the vertical structure of the overturning circulation affects not only the silica balance but also the heat flux. In this study the heat flux in the Indian Ocean at 18°S is −1.27 ± 0.05 PW. Macdonald (1998) estimates a southward heat flux of −1.45 ± 0.3 PW, with a similar-sized Indonesian Throughflow (10.4 ± 10 × 10⁶ m³ s⁻¹). The large southward heat flux at 18°S results from both the input of warm Pacific water via the Indonesian Throughflow and water warmed in the northern Indian Ocean where heat is gained from the atmosphere. We can partition the southward heat flux at 18°S into its two sources, assuming that there is little
Fig. 9. Lateral and diapycnal volume fluxes ($10^6$ m$^3$ s$^{-1}$) in bottom and deep layers for the Indian region (Box IV and V). (top) The volume flux driven by air–sea buoyancy forcing acting on the lightest outcropping layer (15). Diapycnal fluxes due to interior mixing are indicated by curly arrows, solid arrow shows flux across lower interface, and dashed arrow flux across upper interface. Upward diapycnal arrows represent upwelling and downward arrows downwelling across the interface. At 18°S total northward flow of AABW and LCDW and smaller southward flow of layer 18 are shown separately. Error estimates for the lateral and diapycnal fluxes are given in the text.

d. Indian sector of the Southern Ocean

In the Southern Ocean Indian sector the ACC dominates with strong eastward transports of CDW south of Africa ($112\pm4 \times 10^6$ m$^3$ s$^{-1}$) and south of Australia ($96\pm3 \times 10^6$ m$^3$ s$^{-1}$) (Fig. 9). In the bottom water class $5\pm1 \times 10^6$ m$^3$ s$^{-1}$ of generic warmer, saltier AABW (layer 20) moves eastward south of Australia, while there is a westward flux of denser "new" AABW (Ross Sea and Adélie Land Bottom Water, layers 21 and 23) south of Australia and Africa. The westward flow of dense AABW suggests production of AABW along the continental margin between 140°E and 0°. Recent analysis suggests AABW formed along the Adélie coast contributes up to 25% of the global inventory of AABW (Rintoul 1998). Bottom and deep water flow northward
in the Indian sector to fill the abyssal basins of the Indian Ocean. As AABW moves eastward and northward, its property characteristics are modified due to mixing with overlying LCDW. The involvement of LCDW in the production and modification of AABW results in a net diapycnal flux of $17 \pm 3 \times 10^6$ m$^3$ s$^{-1}$ of LCDW to AABW. The LCDW lost in both the formation of and subsequent entrainment into AABW is replaced by a downward diapycnal flux of $25 \pm 4 \times 10^6$ m$^3$ s$^{-1}$ of UCDW. The downwelling of slightly fresher, less oxygenated UCDW (IDW) into LCDW results in the dilution of the salinity maximum and reduced oxygen concentration of the LCDW leaving the Indian Basin south of Australia relative to that entering south of Africa (Sloyan and Rintoul 2000).

In addition to the diapycnal fluxes resulting from mixing in the ocean interior, the shoaling of the upper deep layers near the Antarctic continent exposes them to air–sea fluxes, which change the temperature and salinity characteristics to AASW and result in air–sea flux driven transformation of $18 \times 10^6$ m$^3$ s$^{-1}$ of AASW to lighter water masses (Fig. 9). Some of this water is balanced by interior diapycnal mixing ($12 \pm 3 \times 10^6$ m$^3$ s$^{-1}$) of lower intermediate water to UCDW (layer 15). Sloyan and Rintoul (2001) discuss the circulation and interactions of interior and air–sea diapycnal fluxes between AASW, intermediate, and thermocline waters.

For the Southern Ocean Indian sector, Schmitz (1996b) first scenario (overturning strength $13 \times 10^6$ m$^3$ s$^{-1}$) lacks details of the circulation south of the Polar Front zone and there is no conversion of UCDW to intermediate water. The second scenario ($24 \times 10^6$ m$^3$ s$^{-1}$ overturning) has a conversion of $5 \times 10^6$ m$^3$ s$^{-1}$ of deep water into intermediate water, which is smaller than the $18 \times 10^6$ m$^3$ s$^{-1}$ found in this study.

e. Subtropical Pacific

Across 32°S there is northward flow of $8 \pm 0.5 \times 10^6$ m$^3$ s$^{-1}$ of AABW (layer 20) and $18 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ of LCDW (layer 18 and 19), which results in a $26 \pm 3 \times 10^6$ m$^3$ s$^{-1}$ net northward flux of Antarctic origin water into the Pacific Ocean (Fig. 11). The northward a priori transport constraint of $11 \times 10^6$ m$^3$ s$^{-1}$ of lower deep water (layer 18 and 19) and bottom water (layers 21–23) between 181°E and 192.5°E is maintained (Fig. 12). [From the same current mooring array (PCM-9)
Moore and Wilkin (1998) estimate a northward flux of AABW and LCDW of $20.4 \times 10^6$ m$^3$ s$^{-1}$ below 2000 dbar and $17.5 \times 10^6$ m$^3$ s$^{-1}$ below 3000 dbar. The transport estimate of Moore and Wilkin (1998) is larger than that of Whitworth et al. (1999), as their analysis takes into account the vertical shear of adjacent current moorings to fill missing current records. Further northward transport of AABW is shown immediately east of 192.5°E, while northward flow of LCDW continues to accumulate across much of the southwest Pacific Basin (Fig. 12). Reid (1986) also shows a broad northward flow in his adjusted steric height maps at 2000 and 3000 dbar between 190° and 230°E. Broad northward flow is also shown in a recent analysis of the Pacific 32°S section (Wijffels et al. 2000, manuscript submitted to J. Geophys. Res.).

The northward flow of AABW and LCDW in this model is significantly larger than previous estimates, which vary from between $20 \times 10^6$ m$^3$ s$^{-1}$ and $7 \times 10^6$ m$^3$ s$^{-1}$ (Warren 1981; Wunsch et al. 1983; Macdonald 1998; Tsimplis et al. 1998). The northward transport of AABW and LCDW is balanced by return flow below 1500 dbar of $-25\pm3 \times 10^6$ m$^3$ s$^{-1}$ of Pacific Deep Water (PDW) and UCDW. The balance between northward flowing CDW/AABW and middepth southward flowing PDW/UCDW agrees with previous studies (Reid 1973; Wunsch et al. 1983; Toggweiler and Samuels 1993b; Wijffels 1993).

Macdonald (1998) estimates northward transport of deep and bottom water of $8\pm5 \times 10^6$ m$^3$ s$^{-1}$ at 18°S and $9\pm6 \times 10^6$ m$^3$ s$^{-1}$ at 28°S. These estimates are clearly much smaller than our estimate of $26\pm3 \times 10^6$ m$^3$ s$^{-1}$ at 32°S. Macdonald (1998) discusses difficulties with reconciling her deep Pacific circulation with that of previous studies. At similar latitudes, her northward transport of deep water at 10° and 24°N is only half that estimated by Wijffels (1993) and Roemmich and McCallister (1989), respectively. More importantly, she finds at 10°N the northward transport of abyssal water is balanced by southward thermocline transport, rather than by PDW. This deep to thermocline Pacific overturning between 10° and 24°N disagrees with other studies that show a deep to middepth overturning circulation in the Pacific Ocean (Toggweiler and Samuels 1993b; Wijffels 1993). The deep water transport discrepancy between this study and that of Macdonald (1998) is apparently not due to the fact that we impose an a priori constraint on the northward transport of deep and bottom water adjacent to the Tonga–Kermadec Ridge, as Sloyan (1997) shows that models with no deep water transport constraint have a similar or slightly larger northward transport of deep and bottom water. A more likely explanation for the discrepancy in Pacific deep water transports is the different representations of diapycnal fluxes used in the two studies, as discussed in section 5.

At 11°S the northward flow of LCDW/AABW is restricted to deep passages in the western basin, namely the Samoan Passage, Robbie Ridge, and on the eastern flank of the Manihiki Plateau. Current moorings in these restrictions give a mean (17 month) northward transport of $10.6\pm1.7 \times 10^6$ m$^3$ s$^{-1}$ for water colder than 1.1°C (Roemmich et al. 1996). Below ~1°C we estimate a northward flow of $8\pm0.5 \times 10^6$ m$^3$ s$^{-1}$ at 32°S. This
suggests that there is little conversion of AABW to less extreme LCDW between 32°S and 11°S.

**f. Pacific sector of the Southern Ocean**

The ACC transports $60 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ of UCDW (layers 15-17) and $36 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ of LCDW (layers 18 and 19) to the east south of Australia (Fig. 11). During its transit of the Pacific, the transport of UCDW increases to $69 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ and LCDW decreases to $29 \pm 2 \times 10^6$ m$^3$ s$^{-1}$ on leaving the basin at Drake Passage. The Pacific sector also exports $2 \pm 0.5 \times 10^6$ m$^3$ s$^{-1}$ of extreme AABW (layer 21) into the Indian sector, while $5 \pm 1 \times 10^6$ m$^3$ s$^{-1}$ of warmer AABW (20) enters the Pacific from the Indian sector. At Drake Passage there is also a small ($4 \pm 1 \times 10^6$ m$^3$ s$^{-1}$) eastward flux of warmer, saltier AABW (layer 20).

The southward flow of PDW/UCDW and northward flow of LCDW/AABW across 32°S result in the observed changes of transport in these layers between Australia and Drake Passage. Sloyan and Rintoul (2000) show that the temperature and salinity characteristics of these layers are also substantially modified by diapycnal fluxes in the Pacific sector of the Southern Ocean. A downward diapycnal flux of $24 \pm 5 \times 10^6$ m$^3$ s$^{-1}$ of UCDW into LCDW replaces both the net export of LCDW to the north and east, and the $11 \pm 5 \times 10^6$ m$^3$ s$^{-1}$ of LCDW involved in the formation of AABW in this sector. The $11 \pm 5 \times 10^6$ m$^3$ s$^{-1}$ includes LCDW consumed in the process of forming dense water over the continental shelf, as well as subsequent entrainment of LCDW into newly formed AABW to form the warmer, saltier, less dense AABW that moves northward in the Pacific Ocean and through Drake Passage.

The interior diapycnal flux of UCDW to LCDW is not the only diapycnal process acting on UCDW. UCDW shoals and outcrops south of the Polar Front. Air-sea fluxes convert outcropping UCDW to AASW by cooling and freshening and drive a net conversion of $8 \times 10^6$ m$^3$ s$^{-1}$ of water in this density class to lighter intermediate waters. This air-sea flux driven transformation is more than compensated by interior mixing converting $15 \pm 4 \times 10^6$ m$^3$ s$^{-1}$ of lower intermediate water into the upper deep water (layer 15).

Schmitz (1996b) details the large-scale meridional circulation for the Pacific Ocean [see his Fig. II-69, the portion from just north of the Subantarctic Frontal Zone (SFZ) to Antarctica]. He estimates that the $17 \times 10^6$ m$^3$...
s$^{-1}$ of LCDW/AABW moving northward into the Pacific Ocean is balanced by a return flow in upper deep layers. In the Southern Ocean the UCDW is principally converted into denser LCDW and AABW. This circulation path is similar to that found in this study, although our overturning circulation is larger (26±3 × 10$^6$ m$^3$ s$^{-1}$).

Schmitz (1996b) also indicates 2 × 10$^6$ m$^3$ s$^{-1}$ of UCDW is converted to intermediate water and 10 × 10$^6$ m$^3$ s$^{-1}$ of Antarctic surface water is converted into intermediate water between the SFZ and Polar Frontal Zone (PFZ). Schmitz inferred such a transformation on the basis of published estimates of layer fluxes at lower latitudes, but could not determine the mechanism responsible for the diapycnal fluxes. Here we are able to explicitly identify the combination of air–sea buoyancy and momentum forcing and interior diapycnal mixing responsible for the water mass conversions observed. We estimate a net 8 × 10$^6$ m$^3$ s$^{-1}$ transformation of AASW to lighter intermediate water by air–sea fluxes, which is effectively balanced by interior diapycnal fluxes from lower intermediate water to the upper layer (15) of UCDW.

Macdonald and Wunsch (1996) indicate a downwelling of 9 × 10$^6$ m$^3$ s$^{-1}$ across the 3.5°C isotherm in the Southern Ocean Pacific sector. In this study we estimate interior mixing of 15±4 × 10$^6$ m$^3$ s$^{-1}$ of intermediate water to upper deep water (layer 15). As stated above, this interior flux of intermediate water to deep water is partially offset by the 8 × 10$^6$ m$^3$ s$^{-1}$ air–sea flux of AASW to intermediate water.

5. The Southern Ocean contribution to the deep overturning circulation

The overturning circulations in each basin, and their zonal integral, are illustrated in Fig. 13 by summing over layers corresponding to five broadly defined water masses. The export of 17 × 10$^6$ m$^3$ s$^{-1}$ of NADW from the Atlantic is balanced by a northward flow of thermocline water (TW) and intermediate water (IW) (10 × 10$^6$ m$^3$ s$^{-1}$) and bottom water (BW) (6 × 10$^6$ m$^3$ s$^{-1}$) across SAVE4 (30°–50°S). As a result of upwelling in the subtropical Atlantic, at SAVE2 (12°S) the upper layers carry a larger fraction of the northward flow required to balance the export of NADW (14 × 10$^6$ m$^3$ s$^{-1}$ out of 18 × 10$^6$ m$^3$ s$^{-1}$). In the Atlantic sector of the Southern Ocean, 11 × 10$^6$ m$^3$ s$^{-1}$ of BW is exported from the Weddell Sea. The net northward flow of 4 × 10$^6$ m$^3$ s$^{-1}$ of BW from the Weddell Sea is a balance between southward flowing CDW and warmer AABW (layer 20, indicated by thin arrow in Fig. 13) and newly formed Weddell Sea Deep and Bottom Water (layers 21–23). Comparing fluxes in density layers shows that more IW enters the Atlantic through Drake Passage, and more LDW leaves the basin south of Africa (boxed numbers in Fig. 13). The small net transports of intermediate water at 32°S in the Indian and Pacific Oceans are the difference between much larger gross fluxes to the north and south (31 × 10$^6$ m$^3$ s$^{-1}$ and 25 × 10$^6$ m$^3$ s$^{-1}$, respectively). The intermediate circulation of the model is discussed in Sloyan and Rintoul (2001).

The subtropical Indian Ocean is dominated by a strong, deep overturning circulation. Inflow of 23 × 10$^6$ m$^3$ s$^{-1}$ of LDW and BW is largely balanced by outflow of 20 × 10$^6$ m$^3$ s$^{-1}$ of UDW (IDW in this case). Most of the upwelling required to close this circulation occurs between 32° and 18°S. A net southward flow of 14 × 10$^6$ m$^3$ s$^{-1}$ in the upper layers balances the 10 × 10$^6$ m$^3$ s$^{-1}$ Indonesian Throughflow and 4 × 10$^6$ m$^3$ s$^{-1}$ of UDW converted to IW in the subtropical Indian Ocean. The ACC carries more IW (18 × 10$^6$ m$^3$ s$^{-1}$) out of the Indian basin south of Australia than enters south of Africa; the reverse is true for UDW (−10 × 10$^6$ m$^3$ s$^{-1}$) and LDW (−6 × 10$^6$ m$^3$ s$^{-1}$). The Indian Ocean export of BW (7 × 10$^6$ m$^3$ s$^{-1}$) is achieved by westward transport of “new” dense bottom water south of Africa and an eastward transport of warmer bottom water south of Australia. That is, the Indian basin as a whole imports UDW, LDW, and throughflow water, and exports IW and BW.

The strong, deep overturning (26 × 10$^6$ m$^3$ s$^{-1}$) in the Pacific Ocean at 32°S is similar to that of the Indian Ocean. The inflow of IW in the eastern subtropical Pacific is nearly compensated by outflow in this density class in the western Pacific, so the net transport of IW across 32°S is small. The throughflow is largely fed by 8 × 10$^6$ m$^3$ s$^{-1}$ of TW flowing northward across 32°S. The Pacific sector of the Southern Ocean imports IW and LDW from the Indian sector, and exports UDW to the Atlantic sector.

The zonal sum of the fluxes in density layers is completely dominated by the strong deep overturning cells in the Indian and Pacific Oceans. Roughly 50 × 10$^6$ m$^3$ s$^{-1}$ of LDW and BW are exported to lower latitudes, balanced by an equal poleward flux of slightly less dense UDW. The deep overturning cell is closed below about 27.4, or a depth of about 1500 m. The total transport of deep water entering and leaving the Southern Ocean is even larger due to partial compensation in the zonal sum by flows of opposite sign in different basins in each layer. The total southward flow across roughly 30°–40°S is 62±12 × 10$^6$ m$^3$ s$^{-1}$ (17±5, Atlantic; 20±2, Indian; 26±3, Pacific). The total northward flow is 56±9 × 10$^6$ m$^3$ s$^{-1}$ (7±2, Atlantic; 23±4, Indian; 26±3, Pacific).

The buoyancy (heat) input needed to drive the deep upwelling of this model is estimated from the temperature change required to convert AABW/LCDW to UCDW (1.5°C) and to convert UDW/NADW/IDW to intermediate water (2°C–3°C) in the Atlantic and Indian Oceans. In the subtropical Atlantic 3 × 10$^6$ m$^3$ s$^{-1}$ of AABW is converted to LCDW (AT 1.5°C), which requires a buoyancy input of 0.02 PW. The conversion of 4 × 10$^6$ m$^3$ s$^{-1}$ of UCDW/NADW to intermediate water results in a buoyancy input of 0.04 PW. In the subtropical Indian Ocean the 20 × 10$^6$ m$^3$ s$^{-1}$ conversion of
Fig. 13. A schematic five-layer view of the overturning circulation \((\times 10^6 \text{ m}^3 \text{ s}^{-1})\) in each Southern Hemisphere basin, and the zonal sum for SW: surface water, TW: thermocline water, IW: intermediate water, UDW: upper deep water, LDW: lower deep water, and BW: bottom water. The neutral density surfaces used to define each layer are shown. Air–sea flux-driven diapycnal fluxes are shown by bold, dashed arrows at the sea surface. Diapycnal fluxes due to interior mixing are indicated by thin dashed arrows. Fluxes in boxes represent the net convergence (+ve) or divergence (−ve) of a particular water mass in that sector of the Southern Ocean due to meridional and diapycnal fluxes; mass is conserved by a compensating divergence in zonal transport of the ACC. Two-headed arrows in intermediate water highlight that the net flux is the difference of nearly balancing northward and southward fluxes. Small [2–3 \((\times 10^6 \text{ m}^3 \text{ s}^{-1})\)] imbalances result from choosing a solution where layer residual norms are of \([1–2 \times 10^6 \text{ m}^3 \text{ s}^{-1}]\) and rounding to nearest whole number.

AABW/LCDW to UCDW requires an input of 0.12 PW, while the conversion of UCDW to intermediate water requires a further 0.05 PW input. North of 32°S in the Pacific the \(26 \times 10^6 \text{ m}^3 \text{ s}^{-1}\) conversion of AABW/ LCDW to UCDW/PDW requires an input of 0.16 PW. The \(50 \times 10^6 \text{ m}^3 \text{ s}^{-1}\) overturning circulation requires a net input of 0.39 PW.

The buoyancy input that drives the deep upwelling
in the subtropical oceans is supplied by lateral advection and diapycnal mixing. The heat required to convert AABW/LCDW to UCDW is supplied by warm deep water flowing southward across the northern boundary of the subtropical gyre. Diapycnal mixing between this warm upper deep water and the underlying cold water provides the buoyancy input required to convert LCDW to UCDW/IDW/PDW. Munk and Wunsch (1998) suggest that tidal and wind provide the energy source for the mechanical mixing, with surface buoyancy fluxes and hydrothermal heating playing a secondary role. Enhanced mixing over rough topography (Polzin et al. 1997) is also likely to play a part in driving the deep mixing that the large-scale budgets require. Further conversion of UCDW to intermediate water in the subtropical Atlantic and Indian Ocean is driven by downward diapycnal heat fluxes from overlying thermocline and intermediate water (Sloyan and Rintoul 2000).

The vigorous deep overturning circulation requires similarly vigorous diapycnal fluxes. The transformation of water from one density class to another is accomplished both by air–sea fluxes of heat and freshwater (hence buoyancy) and momentum (Ekman), and by interior diapycnal mixing. Air–sea fluxes convert $3.4 \times 10^8$ m$^3$ s$^{-1}$ of UDW to IW in the Southern Ocean, with more than half of this occurring in the Indian sector. In a steady-state ocean the conversion of UDW to IW by air–sea fluxes must be compensated by interior mixing when the global extent of the isopycnal is considered (Tziperman 1988). In this model we find that the compensating flux occurs in the Southern Hemisphere oceans south of $30^\circ$–$40^\circ$S where interior mixing converts $3.1 \times 10^8$ m$^3$ s$^{-1}$ of IW to denser UDW, although it is not balanced locally within each Southern Ocean sector. [Recall that “interior” in this context refers to the entire ocean beneath the sea surface.] As shown in Fig. 13, within each sector the transformation and interior mixing contributions to the flux of IW and UDW are not balanced: more IW is produced by air–sea fluxes than is consumed by mixing in the Atlantic and Indian sectors, while the opposite is true in the Pacific. In addition, there is large $O(80 \times 10^8$ m$^3$ s$^{-1})$ export of new (cold, fresh) IW to the subtropics, which is nearly balanced by a return flow of modified (warm, salty) IW from the subtropics (Sloyan and Rintoul 2001). Intermediate water gains heat and salt along its circumpolar and subtropical circulation path. The interior compensation of slightly warmer, saltier IW for cold, fresh AASW results in a transfer of heat and salt south of the Polar Front (Sloyan and Rintoul 2001).

In the Southern Ocean, $50 \times 10^6$ m$^3$ s$^{-1}$ of UDW is converted to denser LDW; $27 \times 10^6$ m$^3$ s$^{-1}$ is converted to even denser BW. Because our air–sea flux climatologies do not include the intense buoyancy loss taking place near the Antarctic margin, our model cannot distinguish the relative contributions of air–sea fluxes and interior mixing to this transformation. (It should also be kept in mind that in Fig. 13 the layer called “BW” includes all water denser than $\gamma = 28.2$; that is, this layer includes newly formed AABW and LDW entrained as new AABW sinks from the Antarctic shelf and spreads northward. Our results should not be interpreted to mean that $27 \times 10^6$ m$^3$ s$^{-1}$ of dense water sinks from the Antarctic shelf.) The $50 \times 10^6$ m$^3$ s$^{-1}$ of LDW and BW exported to the lower latitudes is converted to UDW by interior diapycnal mixing, largely in the Indian and Pacific Oceans. The substantial diapycnal fluxes in both the Southern Ocean and subtropical latitudes can explain the changes in water mass properties observed on isopycnals (Sloyan and Rintoul 2000), as well as the changes in transport described here.

Our results are similar to those of Schmitz (1995, 1996b), and the numerous observational studies on which his synthesis was based. The net overturning circulation across about $40^\circ$S inferred from Schmitz (1995, plate 7) consists of a southward deep-water transport of $52 \times 10^6$ m$^3$ s$^{-1}$, balanced by a northward transport of $48 \times 10^6$ m$^3$ s$^{-1}$ of bottom water and $4 \times 10^6$ m$^3$ s$^{-1}$ of intermediate water. The updated version of this figure in Schmitz (1996b, Fig. II-156) shows a deep overturning circulation of similar strength: $48 \times 10^6$ m$^3$ s$^{-1}$ of bottom water flowing north across $40^\circ$S, roughly balanced by $53 \times 10^6$ m$^3$ s$^{-1}$ of deep water flowing south.

The results discussed here differ in several important respects from the global inversion of Macdonald (1998). The zonally integrated overturning circulation in deep water classes is $32 \times 10^6$ m$^3$ s$^{-1}$ in her model. The Indian Ocean overturning is weaker and extends much higher in the water column, and the overturning in the Pacific is much weaker in her model than found here. Part of the reason for the differences in the two models may lie in the representation of the diapycnal fluxes, McIntosh and Rintoul (1997) and Sloyan and Rintoul (2000) find that use of a single unknown for the diapycnal flux across each interface, as in Macdonald (1998), fails to reproduce the known diapycnal fluxes when such a model is applied to output from a numerical model. The choice of weights may also play a role; Macdonald (1998) found that the deep transports in the Pacific increased in a model that was weighted to better resolve the diapycnal fluxes.

In this study the deep overturning circulation is completely dominated by a CDW–BW cell. The volume transports involved in the NADW cell are relatively small in comparison. This suggests that the common practice of equating the global overturning (or thermohaline) circulation with the NADW cell is misleading. Goodman (1998) reached a similar conclusion on the basis of a coarse resolution ocean general circulation model (OGCM).

Our results also provide some insight into the pathways involved in the NADW cell. In order to compensate the sinking of dense water in the North Atlantic, somewhere deep water must be converted to less dense intermediate or thermocline water to close the circulation. We do not find any evidence for upwelling from
deep water through the thermocline in low latitudes. Rather, inflowing NADW (modified by mixing in the ACC) enters the Indian and Pacific Oceans, upwells, and is converted to slightly less dense deep water before being returned to the Southern Ocean. These results are consistent with the radiocarbon and OGCM results discussed by Toggweiler and Samuels (1993b) and Toggweiler and Samuels (1998). The TW/IW and DW circulations are largely isolated from each other in the subtropics, in contrast to the traditional assumption that uniform upwelling balances the sources of deep water, an assumption that underlies a number of conceptual models of the deep circulation (e.g., Stommel and Arons 1960; Munk 1966; Gordon 1986). The deep water does interact with intermediate water in the Southern Ocean, where the deep layers outcrop. More specifically, the UDW exported from the Indian and Pacific Oceans is sufficiently light to outcrop at high latitudes. The NADW itself is mostly too dense to outcrop. At the outcrop, air–sea fluxes of heat and freshwater convert UCDW to AASW and ultimately to IW. The NADW cell is closed by first being converted to UDW in the Indian and Pacific Oceans, followed by conversion of UDW to IW by air–sea exchange, and a return flow of IW to the Atlantic.

That it is UCDW, rather than denser LCDW/NADW, which upwells at high latitude and is converted to lighter water by air–sea buoyancy fluxes has important implications for the Southern Ocean overturning, as well as the NADW cell. The continuity constraints and air–sea flux climatology require a significant net southward flux of UCDW at 30°–40°S. UCDW occupies a depth range of 1500–2500 dbar between 30° and 40°S. As this layer shoals to the south across the latitude of the Drake Passage gap, it lies at depths both blocked and unblocked by topography. The poleward transport of UCDW above topography required to balance the equatorward Ekman transport and air–sea transformation to lighter layers must therefore occur in an eddy-driven transport dynamically balanced by divergence of interfacial form stress. The eddy-driven transport can occur at all scales, from the large-scale standing eddies to small-scale mesoscale eddies. The large-scale standing eddies and mesoscale eddies result in northward or southward movement of the ACC and the vertical displacement of particular water masses along a latitude band. The possibility that part of the UCDW southward flow is accomplished by eddies is supported by observations that significant meridional gradients of potential vorticity exist in the UDW layer (Speer et al. 2000). OGCMs that include a parameterization of eddy advection (Gent and McWilliams 1990) show a maximum in the eddy-induced transport in the ACC (Gent et al. 1995), where eddies carry UCDW poleward. Similarly, Gilyard et al. (1999) use a coupled atmosphere–ocean model to show that poleward advection of UCDW by eddies supplies the upwelling required to feed the water mass transformation driven by air–sea buoyancy fluxes. That

the three-dimensional circulation derived here includes a large poleward transport of relatively light (i.e., above topography) UCDW provides observational evidence that advection by standing and transient eddies is an essential physical process in the Southern Ocean.

6. Conclusions

A box inverse model is used to determine the circulation of the Southern Hemisphere oceans, including diapycnal fluxes due to interior mixing and water mass transformation driven by air–sea buoyancy fluxes. The fluxes due to interior mixing and water mass transformation are large terms in the layer balances. A consistent description of the three-dimensional ocean circulation must include an adequate representation of both contributors to the diapycnal flux.

The section-integrated fluxes and transports of major currents are generally consistent with earlier estimates, but some significant discrepancies with earlier work are found. In the Atlantic, the export of $17 \times 10^6 \text{m}^3\text{s}^{-1}$ of NADW is balanced primarily by inflow of intermediate and bottom water. The deep circulation of both the Indian and Pacific Oceans is dominated by strong, deep overturning cells. In the Indian, $23 \times 10^6 \text{m}^3\text{s}^{-1}$ of northward flowing lower deep and bottom water are converted to upper deep water, which leaves the basin to the south below 1500 dbar. The restricted vertical extent of the deep overturning is in contrast to the overturning cells found by Toole and Warren (1993), Robbins and Toole (1997), and Macdonald (1998), where the northward flow of deep water was balanced by southward flow of thermocline and intermediate water. As a result of the small difference in temperature between northward and southward flowing limbs in the deep Indian overturning, the meridional heat flux at 32°S found here is smaller than that found by the earlier authors. The circulation of the deep Pacific consists of a deep overturning cell similar in strength and vertical extent to that of the Indian Ocean. The deep circulation is essentially isolated from the thermocline circulation in the subtropics.

The zonal integral of the meridional flow across roughly 30°–40°S reveals a vigorous 50 × 10^6 m^3 s^-1 deep overturning circulation consisting of poleward flow of UCDW (including IDW and PDW) and equatorward flow of denser abyssal water. The cell is closed at high latitude by conversion of UCDW to LCDW/BW by interior mixing and/or buoyancy loss to the atmosphere, and at low latitudes by interior mixing converting abyssal water to lighter deep water. NADW exported from the Atlantic Ocean makes a relatively small contribution (in terms of mass transport) to the global deep overturning circulation.

The UCDW also participates in a counterrotating overturning cell south of 30°–40°S. The explicit inclusion, in this model, of air–sea fluxes and the transformation they drive shows that UCDW upwells at high
latitude and is driven north in the Ekman layer. Air–sea fluxes convert upwelling UCDW to cooler, fresher Antarctic Surface Water (AASW), and transform $34 \times 10^6$ m$^3$ s$^{-1}$ of AASW to lighter intermediate water. Some of the intermediate water formed in this way flows north in the Atlantic to close the NADW overturning cell. The Ekman transport is largely compensated by a return flow at slightly greater density, rather than by poleward flow in density layers below topography. The significant meridional transports in density layers above topography suggest that advection by standing and transient eddies plays a central role in the Southern Ocean overturning circulation.

This study has shown that diapycnal transports driven by interior mixing and air–sea buoyancy fluxes are a key part of the large-scale circulation of the Southern Ocean, and that they can be determined with a suitably designed inverse model. The bulk, area-averaged estimates of diapycnal transport due to mixing in the ocean interior provided by such an inverse model complement the small, but growing, number of direct observations of ocean mixing now available. The dataset used here is limited, and improved estimates of large-scale lateral and diapycnal fluxes will undoubtedly result from applying similar inverse modeling techniques to the WOCE dataset, with its much more complete spatial and temporal coverage, additional constraints (e.g., tracers, floats), and repeat measurements with which to assess the impact of oceanic variability.

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APPENDIX

Sensitivity Tests and the Silica Constraint

In our "standard model" a weak silica constraint [box conservation $\mathcal{O}(1000$ kmol s$^{-1}$)] and no flux constraint at 30°–40°S is applied to the inverse model. Although this reduces the silica flux across 32°S in the Indian Ocean compared to Toole and Warren (1993), the meridional silica flux across 30°–40°S is still large. In order to investigate the sensitivity of the circulation and the silica constraint a series of additional experiments were performed. The experiments differ from the standard model by the tightness of the net box silica conservation and the addition of a silica flux constraint at SAVE4, Ind32, and Pac32. In experiment 1 the meridional silica fluxes across the Atlantic, Indian, and Pacific are $0 \pm 200$ kmol s$^{-1}$ and total box conservation is $\mathcal{O}(500–500$ kmol s$^{-1}$). In experiment 2 the meridional silica flux constraints across the Atlantic, Indian, and Pacific are tightened to $0 \pm 150$ kmol s$^{-1}$ and the box conservation is $\mathcal{O}(100–200$ kmol s$^{-1}$). Experiment 3 constrains the meridional silica fluxes across the Atlantic, Indian, and Pacific to $0 \pm 25$ kmol s$^{-1}$ and box conservation of $\mathcal{O}(30–50$ kmol s$^{-1}$). In all experiments a mass imbalance of $\mathcal{O}(1–2 \times 10^6$ m$^3$ s$^{-1}$) is maintained.

Table A1 shows that the net mass flux in the additional silica experiments and the standard model are the same within the given error. (As stated, the errors are only the formal errors of the inverse method. As such they are likely to underestimate the true uncertainty in the estimates.) The property fluxes (not shown) show a similar agreement between the standard model and silica experiments. The silica experiments maintain the deep northward transport ($43\pm16 \times 10^6$ m$^3$ s$^{-1}$, expt 1) across 30°–40°S. As in the standard model the silica experiments also show that the northward transport of deep and bottom water is essentially balanced by southward transport of UDW ($53\pm10 \times 10^6$ m$^3$ s$^{-1}$, expt 1). The air–sea conversion of $34 \times 10^6$ m$^3$ s$^{-1}$ of UDW to IW is present in all experiments, although there are differences in the interior diapycnal flux between AABW, LCDW, and UCDW in the Southern Ocean sector between the standard and silica experiments. These differences result from changes in the transport and direction of AABW south of Australia and Africa (discussed below).

Although the mass and property fluxes are similar between the models, the silica flux at SAVE4, Ind32, and Pac32 are significantly reduced (Table A2). The differences in the meridional silica flux result from small changes in the overturning circulation in the subtropical

<table>
<thead>
<tr>
<th>Table A1. Comparison of the total mass flux ($\times 10^6$ m$^3$ s$^{-1}$) between the standard model and the three silica experiments.</th>
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<tbody>
<tr>
<td></td>
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<tr>
<td>SAVE4</td>
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<tr>
<td>Drake P</td>
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<tr>
<td>S. Africa</td>
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<tr>
<td>Weddell Sea</td>
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<td>S. Aust</td>
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<td>Pac32</td>
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basins. Figures A1 and A2 show the overturning circulation for the Indian Ocean north of 18°S and the Pacific Ocean north of 32°S for the standard model and the additional silica experiments. At 18°S in the Indian Ocean the standard model and experiment 1 show that northward transport of lower deep and bottom water is balanced by southward transport of upper deep water. The small northward transport of intermediate water is converted to surface water. In experiments 2 and 3 the southward transport of upper deep water is balanced by northward transport of lower deep and bottom water and intermediate water. In these models the silica flux at 18°S in the Indian Ocean is reduced by the conversion of low silica concentration intermediate water to higher silica concentration upper deep water.

The standard model and experiment 1 show the greatest similarity with respect to individual water mass fluxes. Difference in the layer fluxes between these models occur in the Indian and Pacific Oceans and in the transport of extreme AABW (layer 21 and 22) and generic AABW (layer 20) south of Africa and Australia. Across 32°S in the Indian Ocean, experiment 1 has a smaller net northward flux of bottom (4 ± 3 × 10^6 m^3 s^-1) and larger northward transport of deep water (17 ± 5 × 10^6 m^3 s^-1), which results in a small decrease in northward transport of bottom and deep water when compared to the standard model. The redistribution and slight reduction of the northward deep and bottom transport results in a decrease of the silica flux across 32°S over that of the standard model. Across 32°S in the Pacific Ocean, experiment 1 has a small northward flux of IW (3.0 ± 2.8 × 10^6 m^3 s^-1), which is balanced by an increase in the southward return flow of PDW (Fig. A2). The large deep overturning circulation in the Pacific Ocean is maintained in experiment 1. However, the conversion of low silica concentration intermediate water to higher silica concentration PDW increases the southward transport of PDW and reduced the net silica flux across 32°S. South of Australia and Africa experiment 1 reverses (now eastward) and reduces the transport of extreme AABW (layer 21–23) and increase the eastward transport of generic AABW (layer 20) over that of the standard model. The increased eastward transport of AABW increases the eastward silica flux.

Given 1) the sensitivity of the silica flux to relatively small changes in the overturning and horizontal circulations; 2) the uncertainty as to how tightly the silica constraints should apply, given present understanding of the silica cycle; and 3) that the conversion of high oxygen, low silica IW to low oxygen, high silica UDW in the subtropical gyres is inconsistent with tracer patterns, we choose the standard model as our preferred

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**Table A2. Comparison of the total silica flux (kmol s^-1) between the standard model and the three silica experiments.**

<table>
<thead>
<tr>
<th></th>
<th>Standard</th>
<th>Expt 1</th>
<th>Expt 2</th>
<th>Expt 3</th>
</tr>
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<tr>
<td>SAVE2</td>
<td>124 ± 194</td>
<td>-58 ± 133</td>
<td>-47 ± 113</td>
<td>-7 ± 37</td>
</tr>
<tr>
<td>SAVE4</td>
<td>295 ± 339</td>
<td>-25 ± 149</td>
<td>-26 ± 109</td>
<td>-2 ± 25</td>
</tr>
<tr>
<td>Drake P</td>
<td>9071 ± 73</td>
<td>9070 ± 658</td>
<td>8818 ± 644</td>
<td>8726 ± 622</td>
</tr>
<tr>
<td>S. Africa</td>
<td>8450 ± 407</td>
<td>8968 ± 662</td>
<td>8791 ± 636</td>
<td>8717 ± 623</td>
</tr>
<tr>
<td>Weddell Sea</td>
<td>-26 ± 339</td>
<td>2 ± 54</td>
<td>-1 ± 53</td>
<td>-1 ± 37</td>
</tr>
<tr>
<td>Ind18</td>
<td>261 ± 679</td>
<td>395 ± 373</td>
<td>106 ± 216</td>
<td>8 ± 51</td>
</tr>
<tr>
<td>Ind32</td>
<td>425 ± 316</td>
<td>51 ± 192</td>
<td>38 ± 140</td>
<td>7 ± 25</td>
</tr>
<tr>
<td>S. Aust</td>
<td>8017 ± 99</td>
<td>8581 ± 633</td>
<td>8679 ± 627</td>
<td>8716 ± 622</td>
</tr>
<tr>
<td>Pac32</td>
<td>821 ± 283</td>
<td>25 ± 190</td>
<td>4 ± 138</td>
<td>3 ± 25</td>
</tr>
</tbody>
</table>

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**Fig. A1.** Comparison of the overturning cell (×10^6 m^3 s^-1) north of 18°S in the Indian Ocean between the standard model, exp 1, exp 2, and exp 3 for SW: surface water, TW: thermocline water, IW: intermediate water, IDW: Indian Deep Water, LDW: lower deep water, and BW: bottom water.
solution. In any case, the most significant conclusions of this analysis regarding the role of the Southern Ocean in the global overturning circulation are not sensitive to the weight imposed on the silica constraint.

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