Supporting Information for “How does Antarctic Bottom Water Cross the Southern Ocean?”
A. Solodoch¹, A. L. Stewart¹, A. McC. Hogg²,³, A. K. Morrison²,⁴, A. E. Kiss²,³, A. F. Thompson⁵, S. G. Purkey⁶, and L. Cimoli⁶

¹Department of Atmospheric and Oceanic Sciences, University of California, Los Angeles, CA, USA
²Research School of Earth Sciences, Australian National University, Canberra, Australia
³ARC Centre of Excellence for Climate Extremes, Australia
⁴Australian Centre for Excellence in Antarctic Science, Australia
⁵California Institute of Technology, Environmental Science and Engineering, Pasadena, CA, USA
⁶Scripps Institution of Oceanography, UCSD, San Diego, CA, USA

Contents of this file
1. Text S1 to S7.
2. Figures S1 to S10.
3. Tables S1 to S4.

Corresponding author: A. Solodoch, Department of Atmospheric and Oceanic Sciences, University of California, Math Science Building, 520 Portola Plaza, Los Angeles, CA 90095, USA. (asolodoch@atmos.ucla.edu)
4. Captions for animations S1 to S4.

Additional Supporting Information (Files uploaded separately)

1. Animations S1-S4.

Introduction The supplemental material provides additional details about the ACCESS-OM2-01 numerical model (text S1). The model’s Meridional Overturning Circulation (MOC) is diagnosed in text S2, and model validation is provided through a comparison with observational estimates. Based on the diagnosed MOC, motivation for the AABW water mass definition (density threshold) used in this study is provided in text S3. In section S4, the methodology for calculating tracer transport is presented. In section S5, definitions of ocean basins used in main text figure 3 are provided. Model validation continues in text S6. The distribution of water mass transformations, in particular in those regions that correspond to formation sites of AABW source-waters, are examined in section S7.

S1. Numerical model

The numerical model is an updated version of the ACCESS-OM2-01 coupled ocean and sea-ice model presented in (Kiss et al., 2020). An ice-shelf model is not included, and ice shelf cavities are masked (marked as land regions) in the model grid. Unlike Kiss et al. (2020) the initial condition for the first cycle is World Ocean Atlas 2013 v2 (WOA13), rather than a spinup under repeat-year forcing. This reduces the initial bias. For the initialization we use the WOA13 temperature and salinity data (Locarnini et al., 2013; Zweng et al., 2013) decav product (average of decadal averages between 1955–2012), using boreal winter conditions under 1500m depth, and January conditions above. The model
is forced with the 1958-2018 JRA55-do version 1.4.0 atmospheric reanalysis, the updated version of (Tsujino et al., 2018), for three consecutive 61 year cycles. We analyze the 3rd cycle.

Additional differences from the version in (Kiss et al., 2020) include the following. The model prognostic temperature variable is conservative temperature (rather than potential temperature as in Kiss et al. (2020). WOA13 potential temperature is converted to conservative temperature via TEOS-10 (using the gsw python package) for the initial condition. As in Kiss et al. (2020) the prognostic salinity variable is practical salinity and the Jackett, McDougall, Feistel, Wright, and Griffies (2006) “pre-TEOS10” equation of state and freezing temperature are used. The relative wind velocity is used for stress calculation over both ice and ocean (in Kiss et al. (2020) it was absolute velocity over ice, relative over ocean). The WOA13 sea surface salinity restoring data is more smoothly interpolated onto the model grid. We now use Large and Yeager (2009) latitude-dependent ocean albedo rather than a constant 0.1. The topography has had small improvements: removal of terraces in shallow water, removal of very small ocean grid cells near the Arctic tripoles, and removal of a few seamounts in the Arctic for numerical stability. We now use explicit vertical background diffusivity of $10^{-6}$ m$^2$s$^{-1}$, spatially uniform (in Kiss et al. (2020) this was zero).

The passive tracers analyzed in the paper are initialized at the beginning of the 3rd cycle with zero concentrations everywhere, and are passively advected and mixed by the circulation velocity (online). Source terms are included within masks on the Antarctic continental shelves (one mask per tracer, highlighted in red in each panel of main text)
figure 1) by linear relaxation, at the surface only, to a value=1, with a time constant of 1000 seconds. Tracer destruction occurs at the surface everywhere outside of the masks, by relaxation to a zero value with a time constant of 1 day.

The tracer concentrations at the simulation end are shown in figure S1 over a larger latitudinal range (south of 10°S) than shown in main text figure 1, to highlight which pathways are blocked by topography. More details are given in the figure caption.

S2. Meridional Overturning Circulation

Here we provide diagnostics and validation of the meridional overturning circulation (MOC), and especially its lower cell in the Southern Ocean, since export of the abyssal water formed around Antarctica is described by this cell. The MOC streamfunction $\Psi_\rho$ is calculated in density coordinates as follows:

$$\Psi_\rho(j,k) = \sum_{k'}^{k} \sum_{i} T_{y,\rho}(i,j,k'),$$  \hfill (S1)

where $T_{y,\rho}(i,j,k')$ is meridional transport calculated online at longitude and latitude ordinals $i$ and $j$, and within the density bin $k'$, which is bounded above by density value $\sigma_2 = \sigma_2(k')$; and where $k = 1$ corresponds to the highest density valued bin.

The MOC streamfunction $\Psi_\rho$ is shown in figure S2. The peak value of the model deep MOC cell is 21 Sv. The peak deep MOC value at 60°S, 20 Sv, occurs within density classes denser than $\sigma_2 = 37.125$ kg/m$^3$. These values are at the center of the range of observational-based estimates, i.e., 10-30 Sv (Lumpkin & Speer, 2007; Talley, 2013; Kunze, 2017; De Lavergne et al., 2016; Cessi, 2019). An inter-comparison of deep MOC estimate values near 30°S is also made by Cessi (2019), with values between 15-30 Sv in Lumpkin and Speer (2007), Talley (2013) and Cessi (2019). While in figure S2 the
MOC value at 30°S is only about 9 Sv, it is an artifact of the coarse density bins used (see below), since the core of the deep MOC cell occurs at slightly lighter density values at 30°S relative to further south. A calculation of the Eulerian MOC\textsuperscript{1} shows that 17.6 Sv travels northward under the mean depth of the $\sigma_2 = 37.08$ kg m$^{-3}$ isopycnal, i.e., within the above-mentioned range of estimations. We verified that the Eulerian MOC value is a good approximation of the Isopycnal MOC value (at 30°S) under the $\sigma_2 = 37.0$ or $\sigma_2 = 37.125$ kg m$^{-3}$ density surfaces (the top of density bins used in the online $T_y$ computation), where both isopycnal and Eulerian quantities can be calculated exactly.

The transport $T_{y,\rho}$ (in equation S1) is calculated online within density bins of width $\Delta \sigma_2 = 0.125$ kg/m$^3$. We also calculated the MOC, $\Psi_\rho$, offline at 60°S with higher density bin resolution ($\Delta \sigma_2 = 0.01$ kg/m$^3$), and verified that $\sigma_2 = 37.125$ kg m$^{-3}$ is a good approximation of the interface between northward and southward parts of the deep MOC: the maximum MOC value is $\sim 2$ Sv (10%) larger with the finer density resolution. The offline $T_y$ calculation is possible due to averages of temperature, salinity, and meridional transport (on the model vertical $z^*$ coordinate), that were saved with high temporal resolution (daily, rather than monthly) only south of 59°S.

We also evaluate model deep MOC transports into individual ocean basins in table S1. The deep transport near 30°S (details in table caption) is compared between our model and several inverse models (Ganachaud & Wunsch, 2000; Lumpkin & Speer, 2007; Kunze, 2017; De Lavergne et al., 2016), and with a geostrophic calculation modified by Ekman fluxes and depth-average constraints (Talley, 2013). We use the model Eulerian mean, calculated under the time-mean density surfaces, since we can do so exactly and with
higher density resolution offline (see discussion above in this section). We find (table S1) that the model basin transports are all within the ranges of at least some of these previous estimates. The Atlantic fraction is near the center of these estimates and within one standard deviation of three of the four models\(^2\) (and within 1.5 standard deviations of the fourth); the Pacific fraction is in the lower half of the estimates, and within one standard deviation of three of the four models; and the Indian fraction is on the low end, and within one standard deviation of two of the four models.

**S3. AABW Definition**

In the model analysis, AABW refers to water with 2km-referenced potential density values \(\sigma_2 \geq 37.125\ \text{kg/m}^3\), since zonal-mean meridional transport in the Southern Ocean is northward (southward) beneath (above) this density surface (dashed line in figure S2b). This isopycnal also corresponds approximately to the neutral density surface \(\gamma_n = 28.2\ \text{kg/m}^3\), which is the minimal density threshold used in the study of Van Sebille et al. (2013) (see main text discussion section). To test the neutrality of the \(\sigma_2 = 37.125\ \text{kg/m}^3\) surface, we have compared it with the \(\gamma_n = 28.2\ \text{kg/m}^3\) surface both at 60\(^\circ\)S, within the center of the Southern Ocean, and north of it at 30\(^\circ\)S. We used the neutral surfaces computation code of Jackett and McDougall (1997), obtained from [https://www.teos-10.org/preteos10_software/neutral_density.html](https://www.teos-10.org/preteos10_software/neutral_density.html). We find that at both of these latitudes the two density surfaces stay within about 100m of one another, hence providing a rough test of the neutrality of this \(\sigma_2\) surface in the region, and its suitability for analysis of transport.
Multiple AABW definitions have been used in the past (e.g., see table 3 in Naveira Gabartó, McDonagh, Stevens, Heywood, and Sanders (2002)). In Orsi, Johnson, and Bullister (1999), a neutral density threshold of $\gamma_n = 28.27 \text{ kg/m}^3$ was defined for AABW, based on the highest density which may cross over the sill at Drake Passage. The value of Orsi et al. (1999) corresponds to $\sigma_2 \approx 37.16 \text{ kg/m}^3$, while the $\sigma_2$ threshold we use here corresponds to a union of the AABW of Orsi et al. (1999) with the layer defined by them as ACC bottom water (ACCbw). The ACCbw is formed from mixing of the abyssal AABW with the overlying Circumpolar Deep Water, according to Orsi et al. (1999). The $\gamma_n = 28.27 \text{ kg/m}^3$ threshold is useful for examining the affect of surface fluxes around Antarctica on ventilation of the area south of the ACC. Yet the definition is limiting for examining the export of AABW and its derivatives to the north of the ACC, since most of the AABW-sourced material exported northward is actually in the ACCbw density range (Orsi et al., 1999). The high density of strict-sense AABW prohibits it from crossing the ridge systems separating the Southern Ocean from the basins to its north. The more inclusive AABW definition used here is qualitatively on par with studies of AABW property trends outside of the Southern Ocean, e.g., (Johnson et al., 2008; Johnson, 2008; Gebbie & Huybers, 2011; Purkey et al., 2018).

S4. Tracer Transport Calculations

Here we describe the offline calculation of tracer transport. Time-mean meridional transport of a tracer species $i$ (i.e., the Weddell Sea tracer) within an isopycnal layer at a specific longitude-latitude point, is given by the following expression:

$$T_{y,i} = \int_{z_1}^{z_2} T_y A_i dz. \quad (S2)$$
Here $T_y$ is northward volume flux, $A_i$ is the tracer concentration (with values 0-1), an overline denotes a time-average, and $z_1$ ($z_2$) is the depth of the bottom (top) of an isopycnal layer. Since isopycnal averages of $T_y$ (i.e., $T_{y,\rho}$, section S2) were saved online, but not those of $A_i$ nor $T_y \cdot A_i$, we approximate $T_{y,i}$ by replacing the tracer concentration $A_i$ by its Eulerian average (the expression inside the parentheses in the following equation):

$$
T_{y,i} = \left( \frac{1}{z_2 - z_1} \int_{z_1}^{z_2} A_i \, dz \right) \int_{z_1}^{z_2} T \, dz.
$$

This approximation is used in calculating the tracer transports in main text figures 2d and 4.

In this section we also estimate tracer transport by the rate of change of tracer volume north of a latitude circle:

$$
F_{i,\text{basin}}(n_0) = \sum_{m \in \text{basin}} T_{y,i}(n_0) \approx \sum_{n > n_0} \sum_{m \in \text{basin}} \sum_k \frac{\partial A_i}{\partial t} \Delta x \Delta y \Delta z,
$$

where $\frac{\partial A_i}{\partial t}$ is the time-average rate of change in concentration of a specific AABW tracer, $(m,n,k)$ are model zonal, latitudinal, and vertical level indexes; $n_0$ is a latitude index specifying the latitude circle of interest, and $n$ increases to the north; $\Delta z$ is model cell thickness at level $k$ (for cells intersecting the seabed, thickness is calculated above the seabed), $\Delta x$ is the zonal cell length, and $\Delta y$ is the meridional cell length. The summation $\sum_{m \in \text{basin}}$ is taken over all zonal indices within one ocean basin at the latitude defined by $n_0$ (cf. text S3). The quantity $F_{i,\text{basin}}$ avoids the need for the approximation in equation S3. However, tracer dissipation or detrainment processes are included in the volume rate of change, and cannot be distinguished from the actual flux across 30$^\circ$S. Nevertheless, the results of both estimators (S3) and (S4) are similar: values in the connectivity matrix shown in main text figure 4 change by less than 6% using the estimator (S4) (not shown).
The transport of all AABW tracers north across 30°S increases in magnitude since model initialization (figure S3). The volume fluxes rise from near-zero values 10 years or more after initialization, i.e., model AABW propagates northwards at an effective propagation speed of order 1 cm s⁻¹, likely representing the joint effects of northward propagation with higher velocities within DWBCs, and of motion within recirculations. However, by the last few decades, the rate of increase of the volume transport of AABW sourced from the Weddell and Ross Seas has decreased and becomes difficult to distinguish from multi-annual oscillations. The rate of export of Adelie-sourced AABW into the Pacific (its main destination) has also decreased substantially by the early 2000s, but its export into the Atlantic appears to steadily rise until the simulation end. The rate of increase of export of Prydz-sourced AABW (mainly into the Atlantic) does not seem to diminish by simulation end. The steady increase in Prydz AABW transport is consistent with its longer route around Antarctica before merger with the Weddell AABW routes; the increase is also consistent with the higher rate of Prydz AABW production and its lower rate of export in the model relative to, e.g., Weddell AABW.

S5. Pathway and blending diagnostics details

The horizontal blending metric $B$ (main text eq. 1 and figure 2) is shown here (figure S4) for tracers which are not principal (high concentration) members of the same conduit.

For purposes of calculations of tracer density profiles within specific oceans (main text figure 3), the boundaries between ocean basins were defined as follows: at 30°S the boundaries between the Atlantic, Indian, and Pacific Oceans were defined by the coastlines of the American, African, and Australian continents. At 58°S, three zonal sections are defined
by boundaries based on topographic barriers: Drake Passage at 65°W (a main ridge in the passage occurs at 65°W, 58°S), Kerguelen Plateau at 80°E, and the Pacific-Antarctic Ridge at 150°W.

The export and import fractions appearing in the connectivity matrix, i.e., figure 4 in the main text, are reported with higher precision in tables S3-S4.

S6. Validation

*Upper Ocean Circulation*

Sea Surface Height (SSH) in the Southern Ocean is compared in figure S5 between the simulation and reprocessed altimetric observations distributed by Copernicus (https://catalogue.marine.copernicus.eu/documents/PUM/CMEMS-SL-PUM-008-032-062.pdf), over the period 1993-2010. The ACC appears in the closely-spaced contours in the Southern Ocean in the model SSH, which closely tracks the observed SSH. Observed SSH meanders appear in the model as well, e.g., near Drake Passage, Scotia Arc, Mid-Atlantic Ridge, Kerguelen Plateau, and Campbell Plateau. Some differences in SSH contours appear further to the North, and near 30°S in the East Pacific and East Atlantic. These differences largely reflect differences in the meridional SSH gradient in the Southern Ocean, mainly between 48-52 S, which is not obvious in the coarse contours, but may be noticed in the color shading. The total transport through Drake Passage in the model is 138.0 ± 3.7 Sv, where the second value is the standard deviation of annual-mean values. In contrast, recent observations (Donohue et al., 2016) find ACC transport through Drake Passage of magnitude 170 ± 10.7 Sv, considerably higher than the model ACC. We note that a recent modeling study (Xu et al., 2020) has suggested that the
observations are biased high (due to undersampling of a deep recirculation), and that the real transport is closer to 150 Sv.

The model geostrophic surface velocity (derived from model SSH) is compared more directly with the altimetric observations in figure S5c-d. The main ACC fronts are reproduced by the model in roughly the right places, although the model velocity magnitude $|\mathbf{v}_m|$ is generally weaker than the observed $|\mathbf{v}_o|$ value. The mean value of their difference $|\mathbf{v}_m| - |\mathbf{v}_o|$ is equal to $-2\%$, in regions where the velocity magnitude is $\geq 0.1 \text{ m}^2\text{s}^{-1}$.

We also examine the fidelity of the subpolar gyres’ full-depth transport. The model Weddell Gyre position and large-scale spatial pattern agree well with observations (Reeve et al., 2019) (not shown). We diagnose a Weddell Gyre peak strength of $40 \pm 5$ Sv full-depth, or $24 \pm 3$ Sv in the upper 2 km. Here, peak strength refers to the largest cross-stream-cumulative transport within a closed streamline of the gyre. Using Argo floats, Reeve et al. (2019) estimated the upper 2 km peak transport at $34 \pm 9$ Sv in the Weddell Gyre, somewhat higher than the model result, albeit with relatively high uncertainty. Some recent full-depth estimates include the estimate of Reeve et al. (2019) by vertical extrapolation ($83 \pm 22$ Sv); by Jullion et al. (2014) using inverse methods ($52 \pm 15$ Sv); and in the state estimate of Mazloff, Heimbach, and Wunsch (2010) ($40 \pm 8$ Sv, see previous estimates herein). Reeve et al. (2019) also estimated a gyre-mean strength of $17 \pm 4$ Sv in the upper 2 km. Defining the gyre to include all streamlines with cross-stream-cumulative transport higher than values between $3–5$ Sv (which seems comparable to the definition in (Reeve et al., 2019), e.g., their figure 6), we find $13.5–14.2$ Sv, i.e., close to
the observational range. A detailed examination of seasonal and interannual variability of the Weddell Gyre in the ACCESS-OM2-O1 model has been conducted by Neme, England, and Hogg (2021).

There are fewer observational estimates of the Ross Gyre. In a total transport map appearing in Reid (1986) (their figure 73), a 20 Sv closed contour appears around the Ross Gyre region. Dotto et al. (2018) estimated 23 ± 8 Sv using radar altimetry, and assuming zero vertical shear. Mazloff et al. (2010) found, within their Southern Ocean State Estimate, a Ross Gyre strength of 20 ± 6 Sv. We diagnose a Ross Gyre strength of 25 ± 4 Sv, similar to the aforementioned values.

Geostrophic surface Eddy Kinetic Energy (EKE) is computed from the model SSH (EKE$_m$) and the altimetry ADT anomalies (EKE$_o$), and compared in figure S6. The locations of high EKE values in the model generally agree with the observations; these regions include: along the ACC fronts, especially at the north ACC flank, and further north over the continental margins to the east of the continents, i.e., in Western Boundary Current regions. However, the model EKE values are on average lower. For example, in regions where \( \frac{EKE_m + EKE_o}{2} > 0.1 \text{ m}^2\text{s}^{-2} \), the model EKE is, on average, more than 40% lower than the observed EKE.

**AABW Properties, distribution, pathways**

The model AABW tracer distributions can be qualitatively compared with CFC observations in the AABW layer. Indeed, horizontal distributions of AABW tracers (figures 1 and 4 in the main text) are similar to the AABW CFC-11 concentration map presented in Orsi et al. (1999) (their figure 7), i.e., high CFC concentrations north of the Southern
Ocean occur in the West Atlantic, and the same two zones leading into the West Indian Ocean. AABW is not exported to the Pacific in the Orsi et al. (1999) maps, which plots water with a higher minimal density threshold (\(\gamma_n > 28.27 \text{ kg m}^{-3}\)) than used here. These denser waters are too deep to cross the Australian-Antarctic ridge, which is overflowed only by the water mass formed above them by mixing with Lower Circumpolar Deep Water (see section S3).

We make a second comparison with the oceanic CFC-11 distribution, along the 32\(^\circ\)S section presented by Purkey et al. (2018) (their figure 6a). The model tracer concentrations (figure S11) peak at similar depths and longitudes to those of the CFC-11 section, i.e., the west sides of the Atlantic, Indian, and Pacific Oceans. The model tracer has lower concentrations in two locations in the Indian Ocean: in the Natal Basin around 39\(^\circ\)E and the Southwest Indian Ridge around 60\(^\circ\)E. From 35\(^\circ\)S, the Natal Basin shoals northward by about 1km within 10 degrees. The Prydz and Weddell model tracer concentrations decrease abruptly along this shoaling region (main text figure 1), perhaps due to spurious numerical mixing. Further north within the Mozambique Channel, the passage shoals even more to just 2.5 km deep. Therefore, we suggest that this route is likely not significant in terms of AABW transport. Near the SWIR, the local peak (in both CFC observations and model tracers) is likely due to along-ridge rather than across-ridge flow, and is not necessarily indicative of bias in across-ridge transport. The apparent bias here is qualitatively consistent with the fact that the total AABW transport into the Indian Ocean is low compared with results from inverse models (section 1).
In figure S7 we compare potential temperature and salinity distributions at the seafloor, averaged over the last 10 years of the simulation, with the World Ocean Atlas (WOA) (Locarnini et al., 2013; Zweng et al., 2013). For both cases, points with potential temperature greater than 2 degrees Celsius are masked. WOA data is interpolated horizontally and vertically to the model grid. For visual comparison, we mask WOA data on mask points of the model, e.g., land and ice shelves. The spatial patterns in general look quite similar, demonstrating fidelity of representation of AABW routes in the simulation. Model bottom water is on average warmer (by $0.16 \pm 0.55^\circ$C) and fresher ($0.033 \pm 0.07$ psu) than WOA data (calculated in areas colder than $2^\circ$C, south of $30^\circ$S), where the uncertainty range is the spatial standard deviation of the difference. These model-observations differences reflect a temporal adjustment occurring in the model, since the WOA presented here was used in the model initialization, i.e., $\sim$178 years earlier.

The difference maps represent evolution over three cycles of model evolution through 61 years of 1958-2018 atmospheric forcing. Therefore we expect a signature of anthropogenic climate change to be present, albeit modified relative to the real ocean (due to the three cycles of 1958-2018 atmospheric forcing, and due to cycle 1 initial conditions at 1958 representing a 1959-2013 decadal average among other reasons). We therefore make only qualitative comparisons of the difference maps with recent observed trends of AABW properties Purkey et al. (2018). Like in the observations, model AABW becomes fresher in most Southern Ocean regions. Additionally, the warming trend in the model occurs where warming most clearly and most intensely occurs in the observations, e.g., south of the main quasi-zonal ridges in the Southern Ocean, in the Western South Atlantic, and
between the Kerguelen Plateau and Australia. Likewise, cooling occurs in both the model and observations in the eastern South-Atlantic and in the western Indian Ocean. Cooling in the model additionally occurs north of the Australian-Antarctic Ridge, whereas mixed warming and cooling trends are seen in different locations in this region in observations. The magnitudes of warming and freshening in the model are also of similar order of magnitude to the observations, but we do not conduct an exact comparison, for the reasons mentioned above.

Although the first model forcing cycle was initialized with WOA temperature and salinity values, the comparison is made 178 years later. This time scale is considerably longer than the duration of a direct route between 70°S and (say) 30°S in a boundary current with a hypothetical speed of 0.01–0.1 m/s, i.e., about 1.5-15 years. 178 years is also a considerable amount of time for AABW changes in the Southern Ocean to occur, based on its flushing time \( t_f = V_{\text{AABW}}/T_{\text{AABW}} \), where \( T_{\text{AABW}} \sim 20 \text{ Sv} \) is the total meridional AABW volume flux in the Southern Ocean, and \( V_{\text{AABW}} \) is the AABW volume between 70°S and 30°S, \( V_{\text{AABW}} \sim 2\pi R_e^2 H_{\text{AABW}} (\sin(70) - \sin(30)) \), with \( R_e \) being Earth’s radius, and \( H_{\text{AABW}} \sim 2 \text{ km} \) is the thickness of AABW in the Southern Ocean. This gives \( t_f \sim 100 \) years, i.e., shorter than the integration time.

Finally, in figure S8, we compare AABW (\( \sigma_2 \geq 37.125 \text{ kgm}^{-3} \)) thickness in the model with the same WOA product described above. The main AABW thickness patterns and values are similar between the observations and model. We summarize statistics of the relative thickness differences. The spatial median absolute magnitude (spatial standard deviation) is 16% (36%) and 12% (17%) in areas in which the WOA AABW thickness is
larger than 500m and 1500m, respectively. As before, these values are calculated in areas colder than 2°C, south of 30°S.

Deep Western Boundary Currents

We next compare the model with several observational datasets of southern hemisphere boundary currents which transport AABW or its derivatives northwards. In the Atlantic sector of the Southern Ocean (at 60.7°S for specificity of transport values), we find (not shown) that the majority of northward AABW transport occurs in the model within narrow ≤ 1 degree wide topographic channels (fracture zones): ∼ 5 Sv in the Orkney Passage region (40°W), and ∼ 4 Sv at the East Scotia Arc (27W). These pathways are consistent with observations (Abrahamsen et al., 2019). Further equatorward in the South Atlantic, the DWBC was monitored at 34.5°S, near the northern boundary of the Argentine Basin, by a mooring array and ship sections across the continental slope (Valla et al., 2019). In figure S9 we present the model mean cross-section (northward) velocities and potential temperature contours, for comparison with an average of eight LADCP sections between 2011 and 2018 by Valla et al (their figure 1c). Three of the sections contained LADCP data only west of 47.5°W. The positions of DWBC cores are qualitatively similar in the model and observations, as are the order of magnitude of the AABW-layer mean velocities, i.e., ∼ 1 cm s^{-1}. The total AABW northward transport based on five of the sections in Valla et al. (2019) is reported to be 0.3 ± 1.6 Sv. In the model, the total AABW northward transport at this section is quite similar, 0.5 ± 1.2 Sv, where the latter number (S) is the standard-deviation of monthly-mean values. Assuming 3-month periods are statistically independent, the standard error in the mean is \( S/\sqrt{N_m/3} = 0.0768 \) Sv (with
the number of months $N_m = 61 \cdot 12 = 732$), making the 0.5 Sv model northward transport highly significant statistically. The observational estimate has a higher uncertainty, and the model results are not statistically distinguishable from them.

On the eastern slope of the Kerguelen plateau, Fukamachi et al. (2010) deployed a mooring array, and measured the velocity profile and flux of the DWBC for two years (February 2003 –January 2005). Model velocity is averaged between 1980-2018. The model DWBC core appears at roughly the same location as in the observations (S10). However, the model DWBC peak velocity (0.073 ms$^{-1}$) is considerably lower than the observational peak (0.236 ms$^{-1}$), and the model DWBC appears wider, i.e., it extends further down-slope. The model Eulerian DWBC transport of AABW in the section is $5.49 \pm 5.33$ Sv in the model, and $\approx 10.4 \pm 5$ Sv in the observations$^4$. The standard deviation values given are based on monthly and (Fukamachi et al., 2010) two-hourly bins, respectively. In summary, in comparison with the observations, the model Kerguelen DWBC appears in the same location as in the observations, but both the peak velocity and total Eulerian transport are roughly 50% smaller, while the model standard deviation of DWBC transport is somewhat higher.

In the western South Pacific, Whitworth III, Nowlin Jr, Pillsbury, Moore, and Weiss (1991) have deployed a 1000 km wide (179W–168W) mooring array at 32.5S, east of the Tonga–Kermadec Ridge (north of Campbell Plateau). Whitworth et al found a mean northwards transport of magnitude $16 \pm 11.9$ Sv under 2 km depth. The moorings had point velocity sensors at three depth levels. In comparison, Lele et al. (2021) find a transport of 10 Sv across the same section under the $\gamma_n = 28.1$ neutral density surface,
averaged over four LADCP-referenced CTD-surveys. We find that the spatial pattern of meridional velocity distribution is very similar to observations in the model (not shown). The 61-year mean amplitude is considerably lower than the observational mean (which does have a large relative uncertainty), with total transport of magnitude $5.1 \pm 2.1$ Sv. However, annual-mean transport in individual years can be as high as $10$ Sv in the model, within the range of the observations.

**S7. Water mass transformations**

We calculate the watermass transformations (WMT) using daily surface heat and salt fluxes on the Antarctic shelves, everywhere south of the 1 km isobath, as well as within each of the tracer masks (each mask terminates offshore at or inshore of the 1 km isobath). For description of the methodology see, e.g., Abernathey et al. (2016). We report on WMT into densities greater than $\sigma_2 = 37.125 \text{ kgm}^{-3}$. This value corresponds roughly to densities of AABW in the Weddell Sea.

The transformation into densities greater than $\sigma_2 = 37.125 \text{ kgm}^{-3}$, averaged over the entire simulated period (61 years), are given in table S2 within each mask, and also for all Antarctic shelf regions shallower than 1 km. The sum of transformations within the four masks ($11.14 \pm 1.09$ Sv, with the latter number being the temporal standard deviation) accounts for $84 \pm 6\%$ of the total transformations on the shelves ($12.93 \pm 1.40$ Sv), qualitatively supporting the fidelity of the model WMT in the observed regions of AABW formation. Salt fluxes (not shown) account for the majority ($\approx 95\%$) of the WMT into AABW densities on the shelves, again qualitatively consistent with observations.
A limitation is that WMT due to mixing processes are not included, since these can only be assessed online. Mixing terms have previously been shown to be significant in the mixed layer for the upper MOC branch-WMT in the Southern Ocean (Abernathey et al., 2016). Hence, we present a second proxy of the formation rate of AABW in the model over the tracer mask regions, given by the rate of change of tracer-weighted AABW volume (anywhere) in the model domain, which we call Tracer Volume Rate of Change (TVRC). In steady state, the average TVRC should be zero, as formation becomes balanced elsewhere by transformation to lighter classes. However, abyssal water lightening (to densities lighter than the threshold we employ here, $\sigma_2 = 37.125$ kg m$^{-3}$) is expected to occur mainly within the Pacific and Indian Oceans (Talley, 2013), and these regions are filled by AABW over a time scale much longer than the 61 year integration, i.e., $\sim 1000$ years (Gebbie & Huybers, 2012). Based on the long equilibration time necessary before significant tracer lightening takes place, and since there is little trend in the rate of change of the tracer-weighted volumes over the 61 year model integration period, we argue that TVRC is a useful estimator of model AABW formation. Unlike our WMT calculation, this estimate does include mixing effects within the formation region, including sub-surface mixing. It should not include subsurface entrainment along the down-slope AABW path.

The mean TVRC, averaged over the entire simulated period (61 years), is given in table S2. The TVRC of each tracer species is on average smaller by 25-50% relative to the WMT occurring within the mask region of the same tracer, which we interpret (see above) to be mainly due to local mixing processes near the surface.
Based on a CFC volume budget, Orsi et al. (1999) estimated AABW production rates in the range of $8 - 9.5$ Sv (with an error estimate of 33%). Previous circumpolar AABW production estimates cited in (Orsi et al., 1999) are in the range $5 - 15$ Sv. Orsi et al. (1999) also estimated about 60% (ratio of CFC-11 saturations of 35% and 60% estimated in AABW and in shelf water, respectively) of AABW is sourced from surface shelf water, and the rest from Circumpolar Deep Water, which brings the estimated surface contribution from Antarctic shelves to $4.7 - 5.5$ Sv. The range here and above is due to different assumptions on the level of vertical mixing of CFCs within AABW. The model total TVRC, $6.3 \pm 0.66$ Sv, is thus quite close to the observations-based values, $4.7 - 5.5$ Sv ($\pm 33\%)$.

**Movie S1.**

Animation of the Weddell-sourced tracer values at the sea bottom. Each frame shows an annual average. Presented tracer values are saturated at 0.1 (maximal values are 1). Values smaller than 0.001 are masked.

**Movie S2.**

Animation of the Prydz-sourced tracer values at the sea bottom. Production details identical to movie S1.

**Movie S3.**

Animation of the Adelie-sourced tracer values at the sea bottom. Production details identical to movie S1.

**Movie S4.**
Animation of the Ross-sourced tracer values at the sea bottom. Production details identical to movie S1.

References


Fukamachi, Y., Rintoul, S., Church, J., Aoki, S., Sokolov, S., Rosenberg, M., & Wakat-


Jullion, L., Garabato, A. C. N., Bacon, S., Meredith, M. P., Brown, P. J., Torres-Valdés, S., … others (2014). The contribution of the Weddell Gyre to the lower limb of the


Tsujino, H., Urakawa, S., Nakano, H., Small, R. J., Kim, W. M., Yeager, S. G., ... others

*Ocean Modelling, 130*, 79–139.


**Notes**

1. The Eulerian MOC is defined similarly to equation S1, except $T_{y,ρ}(j)$ is replaced by the time mean transport within model depth level $j$, $T_y$.
2. Here we assume that the results from (Talley, 2013) have uncertainty similar to the largest uncertainty between the inverse models in the same basin.
3. We define AABW here by $θ \leq 0°$ C, which is almost identical at this section with the neutral density surface AABW definition used by Valla et al. (2019), i.e., $γ = 28.27$ kg m$^{-3}$.

February 17, 2022, 6:56pm
4. For consistency with the definition in Fukamachi et al. (2010), AABW is defined here according to potential temperature
\[ \theta \leq 0^\circ \text{C}. \]
Figure S1. AABW tracer concentrations after 61 years of model integration (i.e., at model date 2018-12-31). This figure is identical to main text figure 1, except that here the latitudinal extent continues 20 degrees further north, and that tracers are vertically averaged over a larger range of densities (37.0 rather than 37.125 kg/m$^3$). Tracer values less than 0.0001 are not displayed, and the colormap is saturated at concentration=0.1. Tracer source region masks are shown in red in each panel. 1 and 3 km depth bathymetric contours are shown in gray.

February 17, 2022, 6:56pm
Figure S2. Model residual Meridional Overturning Circulation (MOC, equation S1), averaged over 2009-2018. The bottom panel is an expanded view of higher density values. The lower MOC cell is seen in $\sigma_2 \gtrsim 36.9$ kg m$^{-3}$. Zonal-mean northward transport in the lower MOC branch, i.e., export of Antarctic Bottom Water, occurs in $\sigma_2 \gtrsim 37.125$ kg m$^{-3}$, marked by the dashed line in the bottom panel. Contours, excluding the zero contour, are shown for every multiple of 2 Sv.
Figure S3. Time series of (annual averages of) tracer transport northwards across 30°S into each ocean basin. Each panel shows results for a different AABW tracer: (a) Weddell-sourced AABW; (b) Prydz-sourced AABW; (c) Ross-sourced AABW; (d) Adelie-sourced AABW. The transport values in this plot are estimated based on the rate of change of tracer volume north of 30°S. Note that tracer transport in main text figures 2b and 4 is calculated differently (see text S2).
**Figure S4.** The horizontal blending metric (main text eq. 1) is shown here for the four tracer pairs that are not shown in main text figure 2. Tracer values lower than 0.001 are masked. Tracer masks are shown in red. The 1 and 3 km bathymetric contours are shown in black.
Figure S5. Caption appears on next page
Figure S5. (Continued from previous page.) Mean SSH in the model (a) and in observations (b, Absolute Dynamic Topography). (c) Difference of (a) and (b). Both model and observational data are averaged over the period 1993-2018. ADT data is downloaded from the Copernicus Marine Service website (https://resources.marine.copernicus.eu/). The model SSH has been adjusted by an additive constant, for visual comparison purposes, to match ADT at 60°S, 0E. (e) Model and (f) observational surface geostrophic velocity magnitude. (d) Magnitude of the vector difference of surface geostrophic velocity in the model vs observations. In all panels, SSH contours are shown for (-1) to (+1) m values, at 0.4m intervals; and gray dashed lines show latitudes 40, 60, and 80°S.
Figure S6. Geostrophic surface EKE in the model (panel a) and observations (panel b, from Sea Level Anomaly). Both data are averaged over the period 1993-2018. The difference between model and altimetric observations is shown in panel c. Gray and black contours are as described in figure S5. ADT data is provided by the Copernicus Marine Service and is freely available online (https://resources.marine.copernicus.eu/)
Figure S7. Caption appears on next page
Figure S7. (Continued from previous page.) Bottom potential temperature (θ) and salinity (S) comparison: (a) model bottom temperature θ_m; (b) World Ocean Atlas (WOA) bottom temperature θ_w (Locarnini et al., 2013); (c) θ_m − θ_w; (d) S_m − S_w (definitions follow); (e) model bottom salinity S_m; (f) World Ocean Atlas (WOA) bottom salinity S_w (Zweng et al., 2013); Model fields are averaged over the last model decade (2009–2018). Points with potential temperature greater than 2 degrees Celsius are masked in each panel. WOA data is interpolated to the model grid and masked at the mask points of the model, e.g., land and ice shelves.

Figure S8. AABW (σ_2 ≥ 37.125 kg m⁻³) thickness map in the model (panel a) and in the World Ocean Atlas (WOA, panel b). Panel c shows the difference between model and WOA AABW thicknesses. Model fields are averaged over the last model decade (2009–2018). Points with potential temperature greater than 2 degrees Celsius are masked in each panel. WOA data is interpolated to the model grid and masked at the mask points of the model, e.g., land and ice shelves.
Figure S9. Model meridional velocity (colors) and potential temperature section at 34.5S, near the northern end of the Argentine Basin. Model data is averaged over the entire model integration period, 1958-2018. Compare with the LADCP observations of Valla et al. (2019) (their figure 1c).
Figure S10. Kerguelen plateau along-slope velocity (colors) section in the model (top, 1980-2018) and in mooring observations (February 2003 - January 2005) by Fukamachi et al. (2010) (bottom). Potential temperature is shown in black contours. Section position is shown in Fukamachi et al. (2010). The external moorings were at positions (82.6175E, 58.1238S) and (84.7888E, 57.0469S). The observational data was provided by Fukamachi Yasushi, who we gratefully acknowledge.
Figure S11. Depth-longitude passive tracer distributions at 32 S: (a) Weddell Sea tracer; (b) Prydz Bay tracer; (c) Adelie Land tracer; (d) Ross Sea tracer. The $\sigma_2 = 37, 37.1, \text{and} 37.125 \text{ kg m}^{-3}$ isopycnals are shown in gray contours. Compare with CFC distribution at the same latitude in Purkey et al. (2018) figure 6a.
Table S1. Abyssal water export to each ocean based on the present numerical model, on published inverse models (Ganachaud & Wunsch, 2000; Lumpkin & Speer, 2007; Kunze, 2017; De Lavergne et al., 2016), and on geostrophic calculations (Talley, 2013). The analysis latitude is $\approx 30^\circ$S (exactly so in our model, see details in the cited publications), with differences of more than a few degrees latitude in the following: (1) $\approx 16^\circ$S for the Pacific in Ganachaud and Wunsch (2000), and (2) basin-averages in Kunze (2017). The top-bounding density surface for the abyssal transport in each model is given in column 2. The approximate density surface value in Talley (2013) is deduced from Talley (2008), e.g., their table 14.

<table>
<thead>
<tr>
<th>Model</th>
<th>Density boundary To Atlantic</th>
<th>Density boundary To Pacific</th>
<th>Density boundary To Indian</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACCESS-OM2-01 $\sigma_2 = 37.08$ kg m$^{-3}$</td>
<td>4.26 Sv</td>
<td>8.77 Sv</td>
<td>4.10 Sv</td>
<td>17.62 Sv</td>
</tr>
<tr>
<td>Ganachaud and Wunsch 2000 $\gamma_n = 28.11$ kg m$^{-3}$</td>
<td>$6 \pm 1.3$ Sv</td>
<td>$7 \pm 2$ Sv</td>
<td>$8 \pm 4$ Sv</td>
<td>21 Sv</td>
</tr>
<tr>
<td>Lumpkin &amp; Speer 2007 $\gamma_n = 28.15$ kg m$^{-3}$</td>
<td>$5.6 \pm 3.0$ Sv</td>
<td>$11.0 \pm 5.1$ Sv</td>
<td>$9.2 \pm 2.7$ Sv</td>
<td>25.8 Sv</td>
</tr>
<tr>
<td>Talley 2013 $\sigma_4 \approx 45.85$ kg m$^{-3}$</td>
<td>$3.85$ Sv</td>
<td>12 Sv</td>
<td>14 Sv</td>
<td>30 Sv</td>
</tr>
<tr>
<td>Kunze 2017 $\gamma_n \approx 28.1$ kg m$^{-3}$</td>
<td>$4 \pm 5$ Sv</td>
<td>$10 \pm 14$ Sv</td>
<td>$4 \pm 6$ Sv</td>
<td>$18 \pm 25$ Sv</td>
</tr>
</tbody>
</table>

Table S2. Water mass transformation rates (WMT) within tracer masks and in all Antarctic shelf regions shallower than 1 km, and tracer volume rate of change (TVRC) for each tracer species. The WMT to TVRC ratio is given in the last column. Values are given in form $m \pm s$, where $m$ is the mean value, and $s$ is the standard deviation of annual-mean values over the 61-year model integration period.

<table>
<thead>
<tr>
<th>Region/tracer</th>
<th>Water mass transformation rate of change [Sv]</th>
<th>Tracer volume rate of change [Sv]</th>
<th>Ratio [-]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weddell</td>
<td>$3.35 \pm 0.73$</td>
<td>$1.42 \pm 0.36$</td>
<td>0.47</td>
</tr>
<tr>
<td>Prydz</td>
<td>$3.84 \pm 0.62$</td>
<td>$1.99 \pm 0.35$</td>
<td>0.54</td>
</tr>
<tr>
<td>Adelie</td>
<td>$1.46 \pm 0.23$</td>
<td>$1.04 \pm 0.26$</td>
<td>0.76</td>
</tr>
<tr>
<td>Ross</td>
<td>$2.50 \pm 0.34$</td>
<td>$1.85 \pm 0.35$</td>
<td>0.75</td>
</tr>
<tr>
<td>All masks</td>
<td>$11.14 \pm 1.09$</td>
<td>$6.3 \pm 0.66$</td>
<td>0.57</td>
</tr>
<tr>
<td>All shelves</td>
<td>$12.93 \pm 1.40$</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
Table S3. AABW tracer export distributions, for each source region. This table provides all percentages appearing (rounded to integer values there) in the green-colored pie charts of main text figure 4. Values in each row do not add exactly to 100% due to rounding error.

<table>
<thead>
<tr>
<th>Origin/tracer</th>
<th>To Atlantic</th>
<th>To Pacific</th>
<th>To Indian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weddell Sea</td>
<td>88.24%</td>
<td>3.04%</td>
<td>8.72%</td>
</tr>
<tr>
<td>Prydz Bay</td>
<td>78.72%</td>
<td>5.72%</td>
<td>15.56%</td>
</tr>
<tr>
<td>Adelie</td>
<td>8.66%</td>
<td>89.78%</td>
<td>1.56%</td>
</tr>
<tr>
<td>Ross</td>
<td>3.03%</td>
<td>96.49%</td>
<td>0.48%</td>
</tr>
</tbody>
</table>

Table S4. Distributions of AABW tracer sources transport to 30°S, for each ocean basin. This table provides all percentages appearing (rounded to integer values there) in the red-colored pie charts of main text figure 4. Values in each row do not add exactly to 100% due to rounding error.

<table>
<thead>
<tr>
<th>Ocean basin</th>
<th>From Weddell</th>
<th>From Prydz</th>
<th>From Adelie</th>
<th>From Ross</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atlantic Ocean</td>
<td>60.00%</td>
<td>32.39%</td>
<td>4.37%</td>
<td>3.25%</td>
</tr>
<tr>
<td>Indian Ocean</td>
<td>43.49%</td>
<td>46.97%</td>
<td>5.78%</td>
<td>3.76%</td>
</tr>
<tr>
<td>Pacific Ocean</td>
<td>1.35%</td>
<td>1.54%</td>
<td>29.59%</td>
<td>67.53%</td>
</tr>
</tbody>
</table>