The Spatiotemporal Structure of Diabatic Processes Governing the Evolution of Subantarctic Mode Water in the Southern Ocean

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ABSTRACT

A coupled ice–ocean eddy-permitting Southern Ocean State Estimate (SOSE) for 2008–10 is used to describe and quantify the processes forming and destroying water in the Subantarctic Mode Water (SAMW) density range ($\sigma_t = 26.7–27.2$ kg m$^{-3}$). All the terms in the temperature and salinity equations have been diagnosed to obtain a three-dimensional and time-varying volume budget for individual isopycnal layers. This study finds that air–sea buoyancy fluxes, diapycnal mixing, advection, and storage are all important to the SAMW volume budget. The formation and destruction of water in the SAMW density range occurs over a large latitude range because of the seasonal migration of the outcrop window. The strongest formation is by wintertime surface ocean heat loss occurring equatorward of the Subantarctic Front. Spring and summertime formation occur in the polar gyres through the freshening of water with $\sigma_t > 27.2$ kg m$^{-3}$, with an important contribution from sea ice melt. Further buoyancy gain by heating is accomplished only after these waters have already been transformed into the SAMW density range. The spatially integrated and time-averaged SAMW formation rate in the ocean surface layer is 7.9 Sverdrups (Sv; 1 Sv = $10^6$ m$^3$ s$^{-1}$) by air–sea buoyancy fluxes and 8.8 Sv by diapycnal mixing, and it is balanced by advective export into the interior ocean. Maps show that these average rates are the result of highly variable processes with strong cancellation in both space and time, revealing the complexity of water mass transformation in the three-dimensional Southern Ocean overturning circulation.

1. Introduction

The Southern Ocean (SO) plays a vital role in the global overturning circulation and global climate because of intense water mass transformation and formation taking place in this region (Macdonald and Wunsch 1996; Sloyan and Rintoul 2001a,b; Talley et al. 2003; Talley 2008). In the upper cell of the SO overturning circulation, strong westerly winds drive equatorward Ekman transport and upwelling in the Antarctic Circumpolar Current (ACC).

Once at the surface, the upwelled water gains buoyancy by freshwater input (e.g., Warren et al. 1996) and heat gain and is transported northward across the ACC (Speer et al. 2000). It subducts just to the north of the Subantarctic Front (SAF), the northernmost core of the ACC, as Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW; McCartney 1982; Rintoul 1991; Sloyan and Rintoul 2001b; Speer et al. 2000; Sallée et al. 2010). SAMW and AAIW ventilate large areas of the lower thermocline in all three Southern Hemisphere subtropical oceans (Schmitz 1996; Hanawa and Talley 2001; Herraiz-Borreguero and Rintoul 2011). The properties of these water masses show high spatial and temporal variability due to the complex mix of formation processes, including air–sea buoyancy fluxes, wind-driven Ekman flow, eddy-induced transport, diapycnal ocean mixing, and upwelling (e.g., Sallée et al. 2010; Sloyan et al. 2010; Holte et al. 2012). Understanding these
processes is fundamental for determining the oceanic transport and storage of heat, freshwater, CO₂ (Sabine et al. 2004; Ito et al. 2010), chlorofluorocarbons (Willey et al. 2004; Hartin et al. 2011), and nutrients (Sarmiento et al. 2004).

With increasing resolution of both observations and numerical model simulations, it has become clear that the transformation of water from one density class to a neighboring one and the formation/destruction within a given density class in the SO are highly variable both in space and time (Sallée et al. 2010; Naveira Garabato et al. 2014). The goal of this study is to identify and quantify the relative contribution of processes playing the most important roles in the transformation and formation of water in the SAMW density range

\[ \sigma_\theta = 26.7 - 27.2 \text{ kg m}^{-3} \]

and to provide a detailed three-dimensional and time-varying (seasonal) representation of distribution of these waters. The large seasonal meridional migration of the \( \sigma_\theta = 26.7 - 27.2 \text{ kg m}^{-3} \) outcrop suggests that a range of processes is important for the formation and destruction of waters within this density range, including freshening by sea ice melt. It is therefore necessary to consider the evolution and distribution of these processes both in time and space in order to understand the role of SAMW in the three-dimensional Southern Ocean overturning circulation.

We use the term SAMW to refer to all waters in the density range \( \sigma_\theta = 26.7 - 27.2 \text{ kg m}^{-3} \), without imposing the mode water low potential vorticity (PV) constraint. We refer to waters in this SAMW density range that also reside inside the low PV pools on the equatorward side of the SAF as “low PV SAMWs.” Thus, low PV SAMWs satisfy both density \( \sigma_\theta = 26.7 - 27.2 \text{ kg m}^{-3} \) and low PV constraints, whereas SAMWs only satisfy the density constraint. We refer to water lighter than SAMW as thermocline water (TW) and water denser than SAMW as intermediate water (IW).

We analyze the results of a data-assimilating, eddy-permitting, ocean–sea ice Southern Ocean State Estimate (SOSE; Mazloff et al. 2010) for the years 2008–10. The SOSE is particularly suitable for water mass analysis because this state estimate is constrained by numerous oceanic observations, making the large-scale SOSE solution consistent with observations (Mazloff et al. 2010). Below we show that the SAMW density field of SOSE is consistent with the optimally interpolated Argo product of Roemmich and Gilson (2009) and that the SOSE air–sea density fluxes are consistent with the fluxes from the ERA-Interim atmospheric reanalysis (Dee et al. 2014). The SOSE solution thus provides an adequate representation of the spatial and temporal distributions of the upper-ocean water masses, consistent with the findings of Cerovečki et al. (2013).

This paper expands on works that have diagnosed SAMW formation by air–sea buoyancy fluxes and inferred subsurface formation/destruction by considering the SAMW volume budget (e.g., Marsh et al. 2000; Karstensen and Quddus 2002; Badin and Williams 2010; Downes et al. 2011; Cerovečki et al. 2013). Here, we directly diagnose not only formation by surface buoyancy fluxes, but also formation by diapycnal ocean mixing. This is facilitated by the fact that the SOSE potential temperature \( \theta \) and salinity \( S \) budgets are exactly closed and that all the terms in these budgets have been archived. Following Iudicone et al. (2008b), we combined the SOSE \( \theta \) and \( S \) budgets to obtain a potential density budget that can be analyzed using the method of Walin (1982) to quantify three-dimensional and time-varying water mass transformation and formation.

The Walin (1982) analysis provides a framework within which to estimate “water mass transformation,” the diapycnal volume flux from one isopycnal layer to an adjacent one by nonadvective buoyancy flux due either to air–sea buoyancy fluxes or diapycnal ocean mixing. “Water mass formation” in each density layer is given by the divergence of transformation with respect to density.

In a seminal paper, only Iudicone et al. (2008a, hereinafter I8a) have carried out a similarly detailed diagnosis of water mass transformation in the SO. They, however, analyzed output from an ocean–sea ice model forced by monthly mean climatological atmospheric forcing and a horizontal resolution of \( 2° \times 2° \), with necessarily reduced spatiotemporal variability. A number of processes important for the formation and destruction of water in SAMW density range are strongly impacted by model resolution, such as lateral mixing in the mixed layer, the nonlinear thermodynamical processes (e.g., cabling), and the role of eddies versus the mean flow in redistributing the water mass after the formation. The present analysis of the higher-resolution SOSE thus provides new and valuable information.

The organization of the paper is as follows: The SOSE product is described in section 2, and the Walin (1982) method of estimating water mass formation is outlined in section 3. In section 4, the time-averaged distribution of SAMW in SOSE is compared with that in the gridded Argo product constructed by Roemmich and Gilson (2009). The SOSE air–sea density flux is compared with that in ERA-Interim (Dee et al. 2014) in section 5. In section 6, we describe the geographical distribution of the individual terms in the 2008–10 time-averaged volume budget of the SAMW density layer, showing formation and destruction by both air–sea buoyancy fluxes and diapycnal mixing, advective redistribution of water in SAMW density range, and storage. The seasonal evolution of these terms is discussed in section 7. In section 8, we present commonly used annually and zonally averaged transformation and formation rate estimates,
which provide a convenient summary of processes discussed in sections 6 and 7. In section 9, we discuss the role nonlinear processes play in SAMW formation, destruction, and export by assessing cabling and by partitioning between the mean and eddy contribution to SAMW advection. A summary and discussion are given in section 10. In the appendix, we give a more detailed verification of the SOSE solution used here, and we also compare the formation rate estimates in the SAMW density range obtained from SOSE with those from previous literature.

2. The Southern Ocean State Estimate

The SOSE is derived using the MITgcm ocean and sea ice model, and the assimilation software developed by the ECCO consortium (Wunsch and Heimbach 2013). It is configured in spherical coordinates with 1/6° horizontal resolution and 42 depth levels of variable thickness. Although a horizontal resolution of 1/6° is too coarse to resolve the first Rossby deformation radius poleward of the ACC, the SOSE model spontaneously develops a vigorous eddy field within the ACC and in the regions of strongest SAMW formation. A description of the SOSE and its consistency with observations is given in Mazlof et al. (2010).

To bring the forward model solution into consistency with observations, the optimization adjusts the atmospheric variables (air temperature, specific humidity, shortwave radiation, wind velocity, and precipitation) and ocean initial conditions as well as open boundary conditions at 24.7°S. Cerovečki et al. (2011) found the air–sea heat and freshwater flux estimates obtained using adjusted atmospheric variables to be consistent with other recent estimates (e.g., Large and Yeager 2009). Determining the model solution by optimizing model inputs does not require unphysical nudging terms to be introduced in the dynamics. The SOSE solution is produced using a single assimilation window and thus does not display the jumps that may arise when patching assimilation windows together in sequential optimization. That the budgets are exactly closed makes SOSE well suited to diagnose the isopycnal volume budgets.

We use the publicly available SOSE solution for the years 2008–10 (available at http://sose.ucsd.edu/sose_stateestimation_data_08.html), for which all the terms from the θ and S equations have been diagnosed and saved as 5-day averages for each individual numerical cell.

3. Methods: Water mass transformation and formation estimates using Walin analysis

The Walin (1982) analysis procedure combines conservation of volume and density providing a framework to determine diapycnal volume flux (water mass transformation) from one isopycnal layer to the other due to nonadvective density supply either by air–sea density fluxes or by diapycnal ocean mixing. We summarize only those aspects of the formalism needed to describe our results; the notation and the method used follow Marshall et al. (1999) and Iudicone at al. (2008a,b).

The conservation equation for tracer τ (potential temperature θ or salinity S) can be written as

\[
\frac{D\tau}{Dt} = d_r + f_r,
\]

where \( d_r \) is the tendency due to mixing processes, and \( f_r \) is the tendency due to external buoyancy forcing. Combining the conservation equations for θ and S yields a potential density σ equation that also has the form of Eq. (1), with \( d_\sigma = \alpha(\theta, S, p_r) d_\theta + \beta(\theta, S, p_r) d_\delta \) and \( f_\sigma = \alpha(\theta, S, p_r) f_\theta + \beta(\theta, S, p_r) f_\delta \). Here, \( \alpha \) is the thermal expansion coefficient, and \( \beta \) is the saline contraction coefficient, both evaluated at a reference pressure \( p_r \) that is taken to be surface pressure. The tendency due to thermal forcing is given by

\[
f_\theta = \frac{Q - I_0}{\rho_0 C_p \nabla z_1} + \frac{I(\delta_k)}{\rho_0 C_p \nabla z_k},
\]

and the tendency due to haline (freshwater) forcing is

\[
f_\delta = S_0 \frac{\nabla z_1}{z_1} (E - P - R).
\]

In Eq. (2), \( Q \) is the net air–sea heat flux (sum of latent heat flux, sensible heat flux, net shortwave, and net longwave radiation; positive for ocean heat gain; W m\(^{-2}\)), \( C_p \) is the specific heat of seawater, \( \rho_0 \) is a reference density, and \( I_0 \) is a fraction of the incoming shortwave radiation that is taken to penetrate below the top ocean model layer whose thickness is \( \nabla z_1 \) and to be distributed vertically as specified by a function \( I(z_k) \) over model layers close to the surface, each with thickness \( \nabla z_k \) (Paulson and Simpson 1977). In Eq. (3), \( S_0 \) is a reference salinity, \( E \) is evaporation, \( P \) is precipitation, and \( R \) is runoff (\( E, P, R \); m s\(^{-1}\)). Their effects are confined to the uppermost model layer. Positive density flux implies an ocean density increase associated with a temperature decrease by ocean heat loss or by \( E - P - R \) increase. The haline forcing importantly includes contributions from sea ice processes as modeled in SOSE.

The Walin (1982) analysis as elaborated in Iudicone et al. (2008b) considers a control volume \( V \), bounded by its intersection with the sea surface, by the two isopycnal surfaces \( \sigma_1 \) and \( \sigma \) (where \( \sigma_1 \) is a fixed reference density chosen to be less than \( \sigma \), and by an open boundary
given by the sum of interior flux across the sea surface. The transformation rate is the product of the density derivative of the volume flux across the sea surface. The transformation rate is the sum of interior volume fluxes, and the density tendency analysis). The arrows labeled \( A \) represent air–sea ice buoyancy fluxes in the following form, using volume fluxes, and flux across isopycnal surfaces (Fig. 1). As in Marshall et al. (1999), we neglect volume flux across the sea surface.

The volume budget of the control volume \( V(\sigma) \) is given by

\[
\frac{\partial V(\sigma)}{\partial t} = A(\sigma_1,t) - A(\sigma,t) - \psi(\sigma,t), \tag{4}
\]

where \( \frac{\partial V(\sigma)}{\partial t} \) is the local rate of change of volume ("storage"), \( A(\sigma_1) \) and \( A(\sigma) \) represent the diapycnal volume flux across isopycnal surfaces \( \sigma_1 \) and \( \sigma \) bounding the volume, and \( \psi(\sigma,t) \) is the volume flux out of \( V(\sigma) \). We neglect volume flux across the sea surface. The transformation rate \( A(\sigma, t) \) is the sum of interior \( \left[ \delta D(\sigma)/\delta \sigma \right] \) and surface \( \left[ F(\sigma) \right] \) components.

In this section, we compare the time-mean geographical distribution of the isopycnal layers encompassing the major SAMWs in the SOSE with that in the gridded Argo product of Roemmich and Gilson (2009). A more detailed SOSE verification is given in the appendix.

A gridded Argo product (hereafter Argo), constructed by an objective mapping of the raw Argo profiles, is available with monthly time resolution on a 0.5° × 0.5° grid (Roemmich and Gilson 2009). The SOSE assimilates the same Argo profiles but in raw, ungridded form, thus carrying out a mapping that is in principle very different from that of Roemmich and Gilson (2009). Nonetheless, the geographical distribution of thickness of the isopycnal layers encompassing the dominant Indian and Pacific SAMW, averaged over the austral winter months (July–September) in years 2008–10, agrees very well between the SOSE and gridded Argo

4. The time-averaged distribution of SAMW in SOSE and the gridded Argo product

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The location of maxima in both products also agrees very well with previously identified ventilation sites of SAMW isopycnal layers (Sallée et al. 2010). As observed, both the SOSE and Argo maxima move progressively eastward with increasing density from the onset of deep winter mixed layers in the western Indian Ocean to the southeast Pacific Ocean.

In the Indian Ocean, three distinct SAMW density ranges and ventilation sites can be identified both in the SOSE and in Argo, in agreement with Fine (1993). The region east of the Kerguelen Plateau (50°S, 70°E) is the ventilation site of the lightest variety of SAMW ($\sigma_\theta \approx 26.6 \text{ kg m}^{-3}$; Figs. 2a,b). A ventilation site of a denser variety of SAMW ($\sigma_\theta \approx 26.7 \text{ kg m}^{-3}$) is in the central Indian Ocean (Figs. 2c,d). The ventilation site of the Southeast Indian SAMW (SEISAMW), which is the densest variety of Indian SAMW with $\sigma_\theta \approx 26.8 \text{ kg m}^{-3}$, is found south of Australia and around Tasmania (Figs. 2e,f). The SEISAMW is one of the major mode water masses of the global ocean (Hanawa and Talley 2001).

Moving eastward in the Pacific Ocean, one finds the SAF located progressively farther poleward, and the thick winter mixed layers become progressively denser. Two formation sites that have been inferred from observations can be identified in both the SOSE and Argo. The central Pacific regions south of the Campbell Plateau and near the Eltanin Fracture Zone have been found to be a hotspot of downwelling for 26.9 $\sigma_\theta$ SAMW (Sallée et al. 2010; McCartney 1982; Hartin et al. 2011; Figs. 2g,h). The southeast Pacific is the formation site of Southeast Pacific Subantarctic Mode Water (SEPSAMW), which is the densest variety of SAMW (with $\sigma_\theta = 27.0$-27.1 kg m$^{-3}$).
and one of the major mode water masses of the global ocean (Hanawa and Talley 2001; Figs. 2i,j).

5. The time-averaged distribution of SOSE air–sea density flux

The SOSE 2008–10 time-averaged air–sea density fluxes are shown in Fig. 3a with the convention that positive density flux implies ocean density increase. The density flux tends to be dominated by the thermal contribution equatorward of the SAF and by the haline contribution poleward of the SAF, so that the density flux abruptly shifts from density increase (by ocean cooling) north of SAF to density decrease (by freshening) south of SAF. Close to the SAF, both processes act in a laterally inhomogeneous manner.

Compared to the corresponding ERA-Interim air–sea density fluxes (Dee et al. 2014), SOSE shows stronger ocean heat loss in the central and southeast Indian Ocean on the equatorward side of the SAF in the SAMW formation region (Figs. 3c,d), suggesting that SOSE might have overestimated SAMW formation in the Indian Ocean. In the southeast Pacific, SAMW formation region time-averaged SOSE and ERA-Interim density fluxes are very similar (Figs. 3a,b).

The difference between the SOSE and ERA-Interim haline component of density flux in the SAMW formation region is very small (Figs. 4e,f), except for a band of higher freshwater gain in SOSE, extending along the Southern ACC Front (SACCF), which coincides with the position of the northernmost wintertime sea ice extent (Fig. 3f). This freshwater gain from sea ice melt is
part of the SOSE ocean–sea ice freshwater budget but is not taken into account in ERA-Interim ocean–atmosphere flux estimates shown here.

6. The geographical distribution of terms in the time-averaged SAMW volume budget

Because of the strong spatial inhomogeneity of processes important for the SAMW volume budget, averages over ocean basins are often unrepresentative of processes at individual locations (Sallée et al. 2010; Naveira Garabato et al. 2014). We therefore consider the three-dimensional distribution of the individual time-averaged components involved in SAMW formation and destruction. The budget is obtained by volume integration of the potential density conservation equation over the SAMW density range [details of the approach are given above and in, e.g., Iudicone et al. (2008b) and Nishikawa et al. (2013)]. The budget, as given by Eqs. (4) and (5), can be written as \( \partial V(\sigma)/\partial t = -\psi(\sigma) + \Delta \sigma [\partial A(\sigma)/\partial \sigma] \) [see, e.g., Eq. (14) in Nishikawa et al. 2013], stating that the local time rate of change of volume of an isopycnal layer \( \partial V(\sigma)/\partial t \) is the integral of volume transport divergence \( \psi(\sigma) \) and of water mass formation \( \Delta \sigma [\partial A(\sigma)/\partial \sigma] \). The latter term can be further separated [using notation of Garrett...
et al. (1995) and Marshall et al. (1999)] into surface formation $\Delta \sigma \left[ \frac{\partial F(\sigma)}{\partial \sigma} \right]$ and formation by diapycnal ocean mixing $\Delta \sigma \left[ \frac{\partial^2 D(\sigma)}{\partial \sigma^2} \right]$. Maps of the geographical distribution of the volume budget per unit area are shown in Fig. 4 and have units of Sverdrups (Sv; $1 \text{Sv} = 10^6 \text{m}^3\text{s}^{-1}$) per square meter. The area integral of each term shown in Fig. 4 over the whole SOSE domain south of $30^\circ S$ yields the individual terms of the SAMW volume budget.

We separately consider the SAMW volume budget in a surface region that is directly exposed to air–sea ice buoyancy forcing and in an interior region at depths isolated from direct atmospheric influences. Following I8a, we refer to the surface region as the bowl, with the caveat that I8a defined the bowl as the maximum seasonal MLD, but we instead define the bowl by the maximum MLD at each geographical location during the entire time period 2008–10. The volume flux out of the bowl is here representative of time average permanent subduction minus obduction (Marshall et al. 1999; I8a). Following Dong et al. (2008), we define MLD by the depth at which the 5-day-averaged density has increased by 0.03 kg m$^{-3}$ from its surface value. The maximum SOSE MLD distribution here used to define the bowl (Fig. 5) agrees well with MLD distribution from Argo in September, when MLDs are deepest, as determined by Dong et al. (2008).

Formation and destruction of SAMW by surface fluxes and diapycnal ocean mixing occur almost entirely inside the bowl (cf. Figs. 4a,b with Figs. 4e,f), in agreement with the results of I8a, Marsh et al. (2000), and Sloyan and Rintoul (2001b). Outside of the bowl, significant advective redistribution of water is balanced almost entirely by storage (Figs. 4g,h), with a small contribution from ocean diapycnal mixing occurring in a few localized regions (Fig. 4f). The advective divergence of volume flux below the bowl, shown in Fig. 4h, can be integrated over the whole Southern Ocean to yield the difference between the net subduction from the bowl and the export through the SOSE open boundary.

The balance between the advective redistribution of water and storage outside of the bowl is largely a consequence of a drift, evident as a steady volume decrease in the $\sigma_0$ range 26.9–27.1 kg m$^{-3}$ (Figs. A1g,i in the appendix) and volume increase in the $\sigma_0$ range 27.1–27.3 kg m$^{-3}$ (Figs. A1k,m), predominantly in the Pacific Ocean. It occurs predominantly outside of the bowl, so that the SAMW volume variability in the bowl is consistent with the variability in the gridded Argo product, as is shown in the appendix. Because the diapycnal processes governing SAMW formation act predominantly inside the bowl (cf. Figs. 4a,b with Figs. 4e,f), we follow I8a, and for the remainder of this work consider only processes occurring inside the bowl.

a. Surface formation

The time-averaged SAMW surface formation reveals strong spatial inhomogeneity (Fig. 4a). Thus, the basin, box, or zonal averages clearly provide incomplete representations of surface formation. On account of the large seasonal shift of the SAMW isopycnal layer outcrop latitude and the wide density range considered ($\sigma_0 = 26.7$–$27.2$ kg m$^{-3}$), the time-averaged surface formation is nonzero over most of the SO poleward of about $40^\circ S$. The regions of strongest surface formation in the SAMW density range are on the equatorward side of SAF in the Indian and west

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**Fig. 5.** The maximum mixed layer depths in SOSE during years 2008–10, in m. Individual mixed layer depths are depths at which 5-day-averaged density has increased by 0.03 kg m$^{-3}$ from its surface value. The deepest of these individual mixed layer depths at each geographical location is the maximum mixed layer depth, referred to as the bowl. The color contours are climatological positions of the fronts given by Orsi et al. (1995): white is STF, red is SAF, green is PF, and yellow is SACCF.
Pacific Oceans (Fig. 4a). These are the regions of deep winter mixed layers (Fig. 5) associated with the thick SAMW density layers (Fig. 2). Surface formation is small in two regions where the band of otherwise large MLDs along the SAF becomes shallower: (i) west of the Kerguelen Plateau at 70°E, which is the region of strong upwelling of light SAMW into the mixed layer (Sallée et al. 2010), and (ii) in the region separating the Indian and Pacific deep mixed layer regions between about 180° and 200°E, where the mixed layer shoals as the flow passes around the Campbell Plateau (Sallée et al. 2010).

In the region of deep winter mixed layers in the Pacific Ocean east of about 140°W (Fig. 2), surface SAMW formation is not prominent in Fig. 4a because, integrated over the entire SAMW density range 26.7–27.2 kg m⁻³, the formation of denser SAMW is canceled out by destruction of lighter SAMW variety from which it is formed. We return to this point in section 8a when showing zonally averaged water mass formation rates as a function of density in the Pacific Ocean sector.

Surface destruction of SAMW dominates at all longitudes between the SAF and SACCF (Fig. 4a) because the SAMW outcrop window lies in this latitude band primarily during the warmer part of the year when surface buoyancy gain destroys water in the SAMW density range, transforming it into lighter water. Poleward of the SACCF, there is a complex pattern of surface formation and destruction in the SAMW density range (Fig. 4a).

b. Formation by diapycnal ocean mixing

The comparison of the formation maps shown in Figs. 4a and 4b shows that diapycnal ocean mixing plays an important role in the formation and destruction of SAMW, in agreement with Speer et al. (2000), Marsh et al. (2000), Sloyan and Rintoul (2001a, b), I8a, Iudicone et al. (2008c), Cerovečki et al. (2013), and Naveira Garabato et al. (2014). Unlike surface formation in the SAMW density range, which displays pronounced small-scale structure (Fig. 4a), formation by diapycnal ocean mixing in the bowl is far more spatially uniform (Fig. 4b). Destruction by diapycnal mixing dominates equatorward of the SAF, and formation dominates poleward of the SAF. In general, this pattern opposes the large-scale pattern of surface formation.

c. Storage and export

The time-averaged surface formation equatorward of the SAF is largely balanced by advective redistribution (Figs. 4a,d) and not by storage. The advective redistribution out of the localized regions of strong low PV SAMW formation shows that this water is either advected eastward by the ACC or northward in the subtropical gyre circulation. Interestingly, on seasonal and shorter time scales SAMW formation is predominantly balanced by storage (discussed in more detail in section 7).

7. The geographical distribution of terms in the seasonally averaged SAMW volume budget

Time-averaged maps of surface buoyancy flux (Fig. 3) have shown that surface heat flux dominates equatorward of the SAF and surface freshwater flux dominates poleward of the SACCF. Between these two fronts both fluxes act in a laterally inhomogeneous manner (Fig. 4). This forcing pattern is reflected in the pattern of time-averaged formation and destruction in the SAMW density range. Equatorward of the SAF, the strongest formation of water in the SAMW density range is by the surface ocean heat loss (Fig. 6a), while poleward of the SACCF the strongest formation is by surface freshening (Fig. 6d) and mixing with saltier water (Fig. 6f). Between these two fronts formation by heat and freshwater fluxes are comparable (Fig. 6), in agreement with Marsh et al. (2000), Speer et al. (2000), Karstensen and Quadfasel (2002), Sallée et al. (2010), Badin and Williams (2010), and I8a.

In the Southern Ocean, annually averaged air–sea buoyancy fluxes are much smaller than seasonally averaged air–sea buoyancy fluxes, on account of strong seasonal cancellation. Similarly, time-averaged formation and destruction of SAMW, such as that shown in Fig. 6, is considerably smaller than seasonally averaged formation and destruction (Figs. 7a–h). We therefore next consider individual terms of seasonally averaged volume budget of SAMW (Fig. 7). We separately consider formation and destruction in the SAMW density range by heat flux (Fig. 8) and by freshwater flux (Fig. 9).

In winter, the outcrop window in the SAMW density range is at its northernmost location, and the strongest formation is by surface ocean heat loss on the equatorward side of the SAF (Figs. 7a, 8a). These are the primary sites of SAMW subduction (Fig. 2; Sallée et al. 2010). Further poleward, in the region between the SAF and the SACCF, weak destruction of SAMW (Fig. 7a) results from transformation into denser water by surface cooling and from cooling due to diapycnal ocean mixing (Figs. 8e,m). This destruction slightly exceeds the formation of water in the SAMW density range by surface freshening and by freshening due to diapycnal mixing in the same region (Figs. 9e,m).

In spring, in the region of strongest wintertime SAMW formation on the equatorward side of SAF
surface ocean heat gain transforms water from the SAMW density range into lighter water, and the surface ocean restratifies (Figs. 7b, 8b). The resulting volume decrease (Fig. 7j) is augmented by the export of SAMW out of this wintertime formation region (Fig. 7n). Between the Polar Front (PF) and the SACCF, buoyancy gain transforms denser water into the SAMW density range (Fig. 7j). This buoyancy gain is caused by strong surface freshening (Fig. 9f), much weaker freshening by ocean diapycnal mixing (Fig. 9n), and surface ocean heat gain (Fig. 8f). The strongest surface freshening is by the melting of sea ice along and poleward of the SACCF (Fig. 9f) [cf. the position of the strongest surface formation by freshening (Fig. 9f) with the pattern in Fig. 4f showing the difference between SOSE and ERA-Interim freshwater flux, which is primarily due to freshwater flux from sea ice melt accounted for in SOSE but not in ERA-Interim].

In summer strong surface heating (Fig. 8c) and freshwater gain (Fig. 9c) transforms SAMW into lighter water in a wide latitudinal range extending from the subtropical gyre to the SACCF (Fig. 7k). At the same time, the volume of SAMW increases along and poleward of the SACCF (Fig. 7k) due to both surface
freshening (Fig. 9g) and obduction of denser water caused by freshwater mixing (Fig. 9o). Poleward of the PF, mixing with colder (Fig. 8k) and saltier water (Fig. 9k) transforms lighter water into the SAMW density range. These transformation processes include the Antarctic Surface Water (ASW), which is formed in spring by sea ice melt and which in summer becomes a warm and fresh surface layer (less than 50 m thick), overlying a cold and fresh layer (Talley et al. 2003).

In fall, strong surface cooling forms SAMW between the SAF and the SACCF (Figs. 7d, 8d). Freshwater mixing (Figs. 9l,p) further contributes to the volume increase in this region (Fig. 7i). The volume of SAMW decreases both equatorward and poleward of this formation region (Fig. 7j). Equatorward, the destruction is due to mixing with warmer thermocline water (Fig. 8l). Poleward, the destruction is due to the combined effects of surface ocean heat loss (Fig. 8h) and salinity increase. The salinity increase is partly caused by brine rejection due to sea ice formation (Figs. 9h,p).

Overall, despite strong seasonal water mass transformations between water in the SAMW density range and lighter water [Fig. 6; Table 1: Southern Ocean], the net exchange between these two water masses is small due to cancellation in the annual average. Consequently, the net formation rate of SAMW from lighter water of 4.2 Sv is weaker than the 12.5-Sv rate of formation from denser water [Table 1: Southern Ocean].

8. The annually and zonally averaged transformation and formation rates of SAMW

In the previous section, we showed that the processes governing SAMW transformation and formation rates have a complex time and space structure. Here, we present often considered annually and zonally averaged rates, which summarize the net effects of these governing processes. Overall, the SOSE SAMW formation rate estimates agree well with previous literature, although the range of values inferred from different models and
different air–sea flux products are rather wide (see the appendix).

1. Formation of SAMW by surface buoyancy fluxes

The strongest surface formation of SAMW is by surface cooling, at a rate of 6.6 Sv; formation by surface freshening is at a rate of 1.3 Sv [Fig. 10a; Tables 2: Southern Ocean and 3: Southern Ocean]. Surface cooling and freshening act over a large region and density range, acting to both form and destroy SAMW (e.g., Sloyan and Rintoul 2001b; Fig. 10a).

Considering the individual sectors of the SO, SAMW is only formed at the surface in the Indian Ocean at approximately the same rate by surface heat flux (5.5 Sv) and by freshwater flux (4.6 Sv). Thus, when considering water mass formation estimates for the whole Southern Ocean, the dominant peak in formation is centered at the SEISAMW density $\sigma_\theta = 26.8$ kg m$^{-3}$ (Fig. 11b). In the Pacific Ocean, within the SAMW density range, there is a balance between surface formation rate by heat flux (5.0 Sv) and surface destruction by freshwater flux (5.1 Sv). Although over the entire SAMW density range the surface formation in the Pacific Ocean is negligible, there is a formation peak centered at the SEPSAMW density $\sigma_\theta = 27.0$ kg m$^{-3}$ (Fig. 12b), resulting from formation by surface cooling at a rate of 5.4 Sv (Fig. 12d), which is opposed by the destruction by surface freshening at a rate of 2.7 Sv (Fig. 12f). In the Atlantic Ocean net surface destruction (at a rate of $-2.2$ Sv) is caused by heat flux destruction exceeding freshwater flux formation (Fig. 10; Table 1). This pattern of surface formation and destruction in

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**FIG. 8.** Geographical distribution of seasonally averaged SAMW formation by heat flux inside the bowl (Sv m$^{-2}$). (a)–(h) SAMW exchange driven by surface fluxes and (i)–(p) SAMW exchange driven by ocean diapycnal ocean mixing. Rows one and three show exchange between SAMW and lighter water ($\sigma_\theta < 26.7$ kg m$^{-3}$), and rows two and four show exchange between SAMW and denser water ($\sigma_\theta > 27.2$ kg m$^{-3}$). Positive values indicate SAMW formation and negative values indicate SAMW destruction.
individual ocean basins is in agreement with, for example, Sloyan and Rintoul (2001b).

**b. Formation of SAMW by diapycnal ocean mixing**

Strong formation of SAMW by diapycnal mixing of freshwater (at a rate of 11.2 Sv) is opposed by destruction by diapycnal mixing of heat [at a rate of ~2.4 Sv; Tables 3: Southern Ocean and 2: Southern Ocean]. Diapycnal mixing of freshwater forms SAMW from both lighter and denser water (Figs. 11e, f). Mixing acts to freshen water that is denser than 27.1 $\sigma_\theta$ and make saltier water that is lighter than 27.1 $\sigma_\theta$ (Fig. 11e), such that both processes add to the volume of SAMW.

Integrated over individual ocean sectors, the net formation of SAMW by mixing is strongest in the Indian Ocean, which is a combined effect of formation at a rate of 5.0 Sv from the mixing of freshwater and 2.1 Sv from the mixing of heat (Fig. 10b). Formation by diapycnal ocean mixing is small in the Pacific [1.0 Sv; Table 1: Pacific Ocean; Fig. 10c] since formation by mixing of freshwater (at a rate of 5.2 Sv) is largely opposed by destruction by mixing of heat (at a rate of ~4.2 Sv; Fig. 12d,f). Formation by mixing is also small in the Atlantic Ocean sector [0.9 Sv; Table 1: Atlantic Ocean; Fig. 10d].

**c. Overview of the formation in SAMW density layer**

Integrated over the whole SO and averaged over the 3 yr, the formation rate in the SAMW density range is 7.9 Sv from air–sea buoyancy fluxes and 8.8 Sv from diapycnal ocean mixing occurring inside the bowl. These processes, of similar magnitude, combine in a net volume export from the bowl into the ocean interior at a rate of 16.7 Sv [Table 1: Southern Ocean; Fig. 10a].

In a zonal and time average, surface freshening and freshening by diapycnal ocean mixing are the only processes that increase the buoyancy of IW with $\sigma_\theta > 27.2$ kg m $^{-3}$, converting it into SAMW (Figs. 11d,f). Only after water has already been converted into the
SAMW density range (with $\sigma_\theta < 27.2 \text{ kg m}^{-3}$), buoyancy is further increased by mixing with warmer water, in agreement with the findings of I8a.

9. Nonlinear processes important in formation, destruction, and export of SAMW

a. Cabbeling

Along-isopycnal mixing of waters of different temperature and salinity may result in the production of denser water because of the convexity of isopycnals in the temperature–salinity plane (McDougall 1984; Iudicone et al. 2008b). This mechanism of water mass transformation into denser water is called cabbeling (Hirst and McDougall 1998; Marsh 2000; Klocker and McDougall 2010; Urakawa and Hasumi 2012). Iudicone et al. (2008b) show that cabbeling is part of the density tendency term due to the mixing processes $d_a$ in Eq. (1), involving along-isopycnal gradients of $a$ and of $\beta$, and can be expressed as $-k_I \nabla \sigma \cdot \nabla \sigma a + k_I \nabla \sigma S \cdot \nabla \beta$, where $k_I$ is the isopycnal diffusion coefficient and $\nabla_I$ refers to

### Table 1.

Time-averaged (2008–10) formation rate estimates (Sv) in the $\sigma_\theta$ range 26.7–27.2 kg m$^{-3}$ due to surface buoyancy flux, ocean diapycnal mixing of buoyancy into or out of the bowl (the region bounded by the deepest mixed layers that develop at each location at any point in time during the 3 yr analyzed) for the Southern Ocean, Indian Ocean, Pacific Ocean, and the Atlantic Ocean. The individual rows show 1) SAMW formation from lighter water (with $\sigma_\theta < 26.7 \text{ kg m}^{-3}$), 2) SAMW formation from denser water (with $\sigma_\theta > 27.2 \text{ kg m}^{-3}$), 3) SAMW destruction into lighter water, 4) SAMW destruction into denser water, 5) the net exchange between SAMW and lighter water, 6) the net exchange between SAMW and denser water, 7) the net gain or loss of SAMW volume, and finally 8) the net SAMW formation given by the sum of surface formation (7S) and the formation in the bowl (Ocean diapycnal) (7OD). A schematic of processes 5–8 is presented in Fig. 10.

<table>
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<th>Ocean diapycnal</th>
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<td>31.9</td>
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<tr>
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<td>4.7</td>
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<tr>
<td>7) Net exchange = 5 + 6</td>
<td>7.9</td>
<td>8.8</td>
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<tr>
<td>8) Net SAMW formation = 7S + 7OD</td>
<td></td>
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<tr>
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<td></td>
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<td>2) Formation from denser water</td>
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<td>3) Destruction into lighter water</td>
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<td>5) Net exchange with lighter water = 1 + 3</td>
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<tr>
<td>6) Net exchange with denser water = 2 + 4</td>
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<td>8) Net SAMW formation = 7S + 7OD</td>
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differentiation along an isopycnal surface. Because SOSE is a $z$-coordinate model, we have approximated the isopycnal derivatives by taking them along a horizontal surface. The results (Figs. 13, 14) were found to be consistent with those calculated using the empirical function of McDougall (1987), justifying this approximation.

Figures 13c and 13d show that a significant fraction of the transformation due to horizontal mixing is due to cabling, with rates that are, as expected, primarily positive, indicating transformation into denser water. The strongest transformation due to cabling inside the bowl is at a rate of 2.8 Sv, at $\sigma_\theta$ approximately 26.7 kg m$^{-3}$. Outside of the bowl the greatest transformation rate is 2.3 Sv at $\sigma_\theta = 27.0$ kg m$^{-3}$. The strongest formation by cabling inside the bowl is 1.0 Sv, at $\sigma_\theta = 27.0$ kg m$^{-3}$, and outside the bowl 1.0 Sv over the $\sigma_\theta$ range 27.0–27.4 kg m$^{-3}$. Water mass formation by cabling is strongest in the density range of SAMW/AAIW, (Figs. 13b,d) in agreement with the results of modeling studies by Marsh (2000), I8a, and Urakawa and Hasumi (2012), and also with the observationally based study by Carter et al. (2014, see their Fig. 8). The spatial structure, consistent with estimates by Marsh (2000) and Urakawa and Hasumi (2012), shows the regions of greatest formation are the Brazil–Malvinas Confluence, along the PF in the Indian sector, and along the SAF in the Pacific sector (Fig. 14a).

b. The mean and eddy contributions to the export of SAMW

The integrand of the full export term $\psi(\sigma)$ of Eq. (4) has been separated into export by the mean flow and export by eddies. The eddy contribution has been defined as a deviation from a seasonal time mean. The export by eddies and the mean flow are of similar magnitude, as shown in Sallée et al. (2010), but the mean contribution is relatively smooth, while the eddy contribution occurs on short scales (Figs. 14c,d). The mean flow thus explains the large-scale structure and is the dominant mechanism transporting SAMW away from regions of strong formation.

10. Discussion and conclusions

We have analyzed the results of the eddy-permitting, data-assimilating Southern Ocean State Estimate (SOSE) for the years 2008–10, for which all terms in the heat and salt budgets have been archived as 5-day averages, allowing construction of closed and complete three-dimensional and time-varying heat, freshwater, potential density, and hence volume budgets for individual isopycnal layers. The volume budget balances
completely so that there is no need for assumptions about the estimated rates of SAMW formation and destruction.

We focus on quantifying the effects of the physical processes important in formation and destruction of water in the SAMW density range. We define SAMW as all waters in the density range $\sigma_\theta = 26.7$–27.2 kg m$^{-3}$, without imposing the mode water low PV constraint. The term SAMW thus not only describes water in this density range that reside inside the low PV pools on the equatorward side of the SAF, but it includes all waters in the SAMW and AAIW density ranges regardless of their geographical location. Moreover, the Antarctic Surface Water (ASW), which warms and freshens in spring and summer and is advected northward by Ekman transport, also enters the $\sigma_\theta = 26.7$–27.2 kg m$^{-3}$ density range (e.g., Evans et al. 2014). Therefore, the region of water mass formation in the $\sigma_\theta = 26.7$–27.2 kg m$^{-3}$ density range extends as far poleward as Antarctica (Figs. 9f,g,o).

The SOSE results for 2008–10 show that in the SAMW density range, all four terms in the isopycnal volume budget—formation and destruction by air–sea buoyancy fluxes, formation and destruction by diapycnal ocean mixing, the advective volume flux divergence, and storage—play important roles. On account of the large seasonal migration of the outcrop window of the SAMW density layer, the mix of physical processes that significantly change the volume of this layer varies in a complex manner. Averaging in time masks a strong seasonal signal, while averaging in space masks a complex spatial pattern of individual physical processes.
The most important processes for SAMW formation are surface heat loss, surface freshening, and freshening by ocean diapycnal mixing. The strongest formation is in winter by the surface ocean heat loss on the equatorward side of SAF in the region of low PV pools (Fig. 7). The SAMW outcrop migrates poleward in spring in summer, where it is strongly influenced by freshening (Figs. 9f,g). Thus, because of the choice of density range used to define the SAMW density layer, and the movement of this layer into the subpolar gyre in spring and summer, a significant formation by freshening occurs far from the region of SAMW low PV pools. Freshening is the strongest between January and April (Fig. 15c), and it is accomplished by both precipitation and sea ice melt. 

Saenko and Weaver (2001) showed that sea ice melt plays an important role in AAIW formation, and Iudicone et al. (2008c) emphasized its importance in SAMW formation. Horizontal maps of time-averaged advective redistribution of water in the SAMW density range show export out of the polar gyres (Fig. 4d). It is thus likely that some of this water will be transported equatorward by Ekman transport, in accord with Speer et al. (2000), Sloyan and Rintoul (2001b), and Rintoul and England (2002), who showed that water crossing the SAF from the south significantly contributes to SAMW formation.

Though in spring and summer sea ice melt is important for SAMW formation, in fall and winter the SAMW outcrop is located equatorward of the sea ice formation locations (Fig. 15). Thus, the combined effect of the seasonal migration of the SAMW outcrop window and the equatorward transport of sea ice during the winter
and spring leads to more SAMW being formed by freshening than is destroyed by brine rejection (Fig. 15c).

In the zonal and time average, buoyancy gain by freshening in the polar gyres is the only mechanism that transforms water denser than $\sigma_{\theta} = 27.2$ kg m$^{-3}$ into the SAMW density range. Buoyancy gain by heating is accomplished only after waters have already been transformed into the $\sigma_{\theta} = 26.7$–27.2 kg m$^{-3}$ density range by freshening, consistent with the inference of I8a.

The geographical distribution of formation and destruction of SAMW by diapycnal mixing shows that it generally opposes the surface formation but has a far more uniform large-scale structure; it tends to destroy SAMW equatorward of the SAF and to form SAMW poleward of the SAF. This mixing occurs almost entirely in the volume exposed to air–sea ice buoyancy fluxes at some point during the year. Our findings are consistent with the growing observational evidence of the importance of diapycnal mixing in the Southern Ocean for closing the upper limb of the global overturning circulation. Carter et al. (2014) inferred in the SEPSAMW formation region that observed SAMW and AAIW property distributions require diapycnal mixing with both subtropical surface water and Upper Circumpolar Deep Water (UCDW). Similarly, Meyer et al. (2015) observed enhanced mixing in the frontal regions north of the Kerguelen Plateau, thus influencing the properties of AAIW, UCDW, SAMW,
and ultimately the Southern Ocean overturning circulation.

For the entire SO, the 2008–10 averaged export of SAMW out of the bowl (i.e., out of the upper-ocean volume that is directly exposed to air–sea ice buoyancy fluxes at some point during the year) into the ocean interior is composed of water formed at the surface, almost entirely from denser water, at a rate of 7.9 Sv and water formed by diapycnal ocean mixing from both lighter (at a rate of 4.1 Sv) and denser (at a rate of 4.7 Sv) water [Fig. 10; Table 1(i)]. Thus, the export of SAMW into the ocean interior at a rate of 16.7 Sv has almost equal contributions from surface formation and from formation by diapycnal ocean mixing. Approximately 25% of the exported water forms from lighter water, and 75% forms from denser water. The exchange between SAMW and lighter water is much stronger than with denser water, but in the annual average, the net exchange rate with lighter water is almost zero [Table 1(i)].

It is possible that the partitioning of SAMW formation from lighter and denser waters is modulated by interannual variability. The major modes of Southern Hemisphere climate variability are important in modifying SAMW and AAIW properties (Naveira Garabato et al. 2009). The southern annular mode (SAM) index was high in the years analyzed here, implying stronger zonal winds and meridional Ekman transport, which in turn may have caused stronger upwelling and preferential formation from dense water by freshening. Here, we have shown the processes important for seasonal and 2008–10
time-mean evolution of water in SAMW density range. How these processes may vary interannually and in a changing climate needs to be addressed in future work.

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APPENDIX

Comparison of SAMW Formation Rate Estimates from SOSE with Literature

a. The time variability of volume in SAMW layer

To assess the time variability of SOSE water masses during the 3 yr analyzed, we compute the time series of isopycnal volume in isopycnal layers 0.1 kg m$^{-3}$ wide spanning the SAMW density range (Fig. A1). The seasonal cycle in volume variability is strong in all isopycnal layers in the SAMW density range (Fig. A1, left). Additionally, however, SOSE displays a drift over the 3 yr analyzed, affecting the denser variety of SAMW and AAIW, causing a volume decrease over the $\sigma_\theta$ range 26.9–27.3 kg m$^{-3}$ (Figs. A1g,i) and volume increase over the $\sigma_\theta$ range 27.1–27.3 kg m$^{-3}$ (Figs. A1k,m).

The processes governing water mass transformation and formation predominately occur in the surface region that is directly exposed to air–sea ice buoyancy forcing. Therefore, we separately consider the time series of isopycnal volume inside a region called the bowl, which is bounded below by the deepest mixed layer depth that occurs at each geographical location over the 3 yr analyzed (Fig. A1, right). The dominant feature of isopycnal volume variability inside the bowl is strong seasonal variability with a less significant trend. The volume drift occurs primarily below the deepest mixed layers and arises from an imbalance between the net subduction rate and export through the open northern boundary. To understand better the interannual volume variability in the bowl associated with SAMW formation and destruction, we next compare SOSE and Argo volume anomalies.

b. Interannual variability of SAMW volume in low PV pools in SOSE and the gridded Argo product

As in the entire SAMW density range, the volume anomalies in the two SAMW low PV pools associated with SEISAMW and SEPSAMW are strongly variable in time in both SOSE and Argo (Fig. A2). Both SOSE and Argo volume anomalies have been obtained as a deviation of monthly mean volume from the 2008–10 time-mean volume in each isopycnal layer 0.05 kg m$^{-3}$ wide satisfying a low PV criterion within the SAMW density range. The drift in the SOSE volume of SAMW density classes identified in Fig. A1 has been removed as a linear trend in each density bin in Fig. A2. In the Indian Ocean, the low PV pool shows a strong wintertime volume increase in the SAMW density range (Figs. A2a,c), suggesting strong SAMW formation in each of the 3 yr considered. Overall, there is a remarkable similarity between the SAMW volume anomaly in low PV pools in Argo and in SOSE, though the 2009 and 2010 wintertime volume increases in SOSE are stronger than in the Argo product.

The wintertime volume increase in the Pacific Ocean low PV pool is smaller than in the Indian Ocean low PV pool in both SOSE and Argo for each of the
3 yr considered. In both products the wintertime SEPSAMW volume increase is stronger in the years 2008/09 than in 2010 (Figs. A2b,d). Thus, some fraction of the interannual volume decrease in the $\sigma_\theta$ range 27.0–27.1 kg m$^{-3}$ over the years 2008–10 persisting in the bowl as shown in Fig. A1j may be realistic. Furthermore, both products show the density of the strongest wintertime volume increase in 2009 and 2010 was lower than in 2008. This means that the average 2008–10 SAMW formation in the Pacific sector is distributed over a wider density range ($\sigma_\theta = 26.8–27.0$ kg m$^{-3}$ range) instead of being confined to the more narrow range of SEPSAMW $\sigma_\theta = 27.0–27.1$ kg m$^{-3}$ that is often reported in the literature.

Overall, the comparison of Figs. A1 and A2 suggests that in the 3 yr considered here the SOSE may have overestimated SEISAMW formation. Although SOSE shows a drift in the SAMW density range in the Pacific Ocean sector, this drift is primarily confined to depths below the deepest mixed layers. At depths shallower than the deepest mixed layers, the volume decrease in the SEPSAMW density range $\sigma_\theta = 27.0–27.1$ kg m$^{-3}$ is much weaker (Fig. A1), and it is partly caused by a weaker SAMW formation in 2010.
c. The annually averaged SOSE SAMW formation by air–sea density flux

We next compare SAMW water mass formation rate estimates obtained using air–sea density fluxes from SOSE and from ERA-Interim (Dee et al. 2014) combined with Ishii et al. (2005) salinity, as input into the Walin (1982) framework, and relate them to the main features of the SOSE and Argo volume anomaly discussed in the previous section. We consider the Southern Ocean south of 30°S.

Considering the entire Southern Ocean, the formation rate estimate in the broad SAMW $\sigma_\theta$ range 26.7–27.2 $\text{kg m}^{-3}$ that includes both Indian and Pacific SAMWs is smaller in SOSE (13.1 Sv) than in ERA-Interim (27.4 Sv; Table A1). Considering only the Indian Ocean sector, the formation rate estimate in the broad SAMW $\sigma_\theta$ range is also smaller in SOSE (14.8 Sv) than in ERA-Interim (17.2 Sv; Fig. A3b; Table A1). Considering only the Indian Ocean sector but within the narrower density range associated only with SEI-SAMW, SOSE has high surface formation within a
narrow density range (12.2 Sv over the $\sigma_{\theta}$ range 26.75–26.85 kg m$^{-3}$), whereas ERA-Interim also has high surface formation rates but spread over a somewhat wider density range (8.3 Sv over the $\sigma_{\theta}$ range 26.75–26.85 kg m$^{-3}$ and 9.2 Sv over the $\sigma_{\theta}$ range 26.85–26.95 kg m$^{-3}$; Table A2).

Considering only the Pacific Ocean sector, the formation rate estimate in the broad SAMW $\sigma_{\theta}$ range shows net destruction both in SOSE and ERA-Interim (Fig. A3c; Table A1). However, both show formation over a narrower density range associated with the SEPSAMW; SOSE yields formation at a rate of 2.9 Sv over the $\sigma_{\theta}$ range 26.95–27.05 kg m$^{-3}$ and even stronger formation at a rate of 4.6 Sv over the $\sigma_{\theta}$ range 26.85–26.95 kg m$^{-3}$. The corresponding formation rates in ERA-Interim are smaller: 0.7 Sv over the $\sigma_{\theta}$ range 26.95–27.05 kg m$^{-3}$ and 1.5 Sv over the $\sigma_{\theta}$ range 26.85–26.95 kg m$^{-3}$.

d. Comparison of SOSE SAMW formation rates with estimates from recent literature

The SOSE SAMW formation rates agree well with estimates from recent literature, although the range of values attained from different models and different air-sea flux products is rather wide. Furthermore, the density ranges used to define SAMWs also differ between studies.

1) SOUTHERN OCEAN

In all the following comparisons care must be taken to acknowledge the density range and region that goes with each individual formation rate estimate. The SOSE

<table>
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Table A1. Time-averaged (2008–10) formation rate estimates (Sv) in the SAMW $\sigma_{\theta}$ range 26.7–27.2 kg m$^{-3}$ due to surface density flux without distributing incoming shortwave radiation vertically below ocean surface for the Southern Ocean, Indian Ocean sector, Pacific Ocean sector, and the Atlantic Ocean sector using SOSE and ERA-Interim fields to obtain surface density flux. The salinity component of ERA-Interim surface density flux was obtained using Ishii et al. (2005) monthly salinity estimates in Eq. (3). The longitude bounds are 150°E–70°W for Pacific, 20°–150°E for the Indian Ocean, and 70°W–20°E for the Atlantic Ocean sector.
formation rate estimate for the SAMW $\sigma_\theta$ range 26.7–27.2 kg m$^{-3}$ in the entire SO south of 30°S is 16.7 Sv. This is very similar to the SAMW formation rate estimates from recent literature: a 16.6-Sv formation rate estimate by Marsh et al. (2000) for the $\sigma_\theta$ range = 26.52–27.22 kg m$^{-3}$, a 19.4-Sv formation rate estimate by I8a for the neutral density $\gamma$ range 26.0–27.2 kg m$^{-3}$, and a 15–22-Sv formation rate estimate by Downes et al. (2011) for the $\gamma$ range 26.0–27.0 kg m$^{-3}$.

There are differences between the SOSE formation rate in the $\sigma_\theta$ range 26.0–27.2 kg m$^{-3}$ and the I8a formation rate in the corresponding $\gamma$ range 26.0–27.2 kg m$^{-3}$. In SOSE the formation rate is 9.5 Sv from surface freshwater flux and 6.9 Sv from surface cooling, yielding a net surface formation rate of 16.4 Sv. This is much smaller than surface formation rate of 28 Sv estimated in I8a. However, in I8a SAMW is destroyed by ocean mixing, while in the SOSE SAMW is formed by ocean mixing at a rate of 8.6 Sv, so that the total SAMW formation in SOSE in the wider $\sigma_\theta$ range 26.0–27.2 kg m$^{-3}$ is 21.5 Sv and close to the I8a estimate of 19.4 Sv in the corresponding density range. We believe that the difference in sign of the formation associated with ocean mixing between SOSE and I8a results from a different definition of the bowl. Downes et al. (2011) estimated water mass formation rates from surface buoyancy fluxes and diapycnal mixing in the SO south of 30°S in the ECCO model-based state estimate and in three free-running coupled climate models. The SAMW formation rate estimates in the $\gamma$ range 26.0–27.0 kg m$^{-3}$ were between 26 and 51 Sv, the sum of a 13–39-Sv formation rate from denser AAIW by surface warming and a 12–15-Sv formation rate from lighter water. In the interior, half of the SAMW volume was converted to adjacent water masses, while 15–22 Sv was transported out of the Southern Ocean. In SOSE in the $\sigma_\theta$ range 26.0–27.2 kg m$^{-3}$, the SAMW surface formation rate of 16.4 Sv is smaller than in Downes et al. (2011), whereas in the models analyzed in Downes et al. (2011) SAMW was destroyed by ocean mixing, in SOSE SAMW was formed by ocean mixing, so that the rate at which SAMW is exported into the ocean interior is consistent: 15–22 Sv in Downes et al. (2011) versus 21.5 Sv in SOSE.

2) INDIAN OCEAN

In SOSE the largest SAMW formation is in the Indian Ocean, with a peak centered in the SEISAMW density range $\sigma_\theta = 26.75–26.85$ kg m$^{-3}$. The net formation in the broad SAMW density range $\sigma_\theta = 26.7–27.2$ kg m$^{-3}$ is
17.2 Sv and is composed of surface formation at a rate of 10.1 Sv (formed by cooling and freshening) and of formation within the bowl at a rate of 7.1 Sv (from both lighter water at a rate of 4.0 Sv by cooling and salt gain and from denser water at a rate of 3.1 Sv by freshening and warming). This description agrees well with Sloyan and Rintoul (2001b), who estimated a surface formation of approximately 12 Sv in the $\sigma_\theta$ range 26.8–27.0 kg m$^{-3}$ from TW by cooling and freshening; 7.5 ± 1.1 Sv are formed in subtropical Indian Ocean (their box IV), and 4.4 ± 2 Sv are formed in the Southern Ocean Indian sector (their box V). They emphasize the importance of formation of SAMW from ASW and infer the largest input of ASW to SAMW occurs in the Indian Ocean, which is consistent to our diagnosis from SOSE. The SOSE formation rate is also comparable to the Marsh et al. (2000) estimate of 19.8 Sv of SAMW formation in the Indian Ocean in the density range $\sigma_\theta = 26.52–26.80$ kg m$^{-3}$.

The SAMW surface formation in the Indian Ocean in the $\sigma_\theta$ range 26.5–27.1 kg m$^{-3}$ in SOSE was 9.4 and 10.9 Sv, in years 2005 and 2006 (Cerovečki et al. 2013). Stronger surface SAMW formation in the Indian Ocean in 2008–10 than in 2005/06 may be attributed to a stronger SAM in 2008–10 than in 2005/06. The ASW/SAMW variability in the Indian Ocean is strongly correlated with the variability in the wind stress (Rintoul and England 2002). It is possible that stronger winds during the time period of enhanced SAM increased the formation of SAMW from ASW through increased equatorward Ekman transport.

3) PACIFIC OCEAN

Some of the SAMW formed in the Indian Ocean is advected eastward by the ACC into the Pacific Ocean where it is cooled and freshened (e.g., Sloyan and Rintoul 2001b). In SOSE in 2008–10, in the Pacific sector, surface SAMW formation and destruction in the broad SAMW $\sigma_\theta$ range 26.7–27.2 kg m$^{-3}$ roughly balance each other, and there is only weak formation in the interior (Fig. 10c). No net SAMW formation in the Pacific sector was also found by Sloyan and Rintoul (2001b) and by Cerovečki et al. (2013). In each case SAMW was destroyed in the Pacific sector both by surface fluxes and by ocean mixing. Marsh et al. (2000) also diagnosed strong SAMW destruction in the Pacific sector at a rate of 7.4 Sv; 15.9 Sv were destroyed in the 26.52–26.80 $\sigma_\theta$ range and 8.5 Sv were formed in the 27.03–27.22 $\sigma_\theta$ range, south of 25°S. Focusing on the SEPSAMW 26.85–27.05 $\sigma_\theta$ range, the formation rate from SOSE was 7.5 Sv. This is consistent with the estimate from Hartin et al. (2011), who used chlorofluorocarbons inventories to infer a 7.3 ± 2.1 Sv SAMW formation rate in the southeast Pacific in 26.80–27.06 $\sigma_\theta$ range.

These SOSE results also qualitatively agree with those of Sallée et al. (2010) who estimated upwelling at a rate of 10 Sv in a layer denser than 27 $\sigma_\theta$. This water is advected equatorward in lighter layers, 7 Sv were subducted into dense SAMW in the density range 26.8–27 $\sigma_\theta$, with no consistent downwelling or upwelling in the lighter SAMW in the density range 26.6–26.8 $\sigma_\theta$.

REFERENCES


Downes, S. M., A. Gnanadesikan, S. M. Griffies, and J. L. Sarmiento, 2011: Water mass exchange in the Southern Ocean

<table>
<thead>
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<th>Ocean basin</th>
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TABLE A2. Time-averaged (2008–10) formation rate estimates (Sv) due to surface density flux without distributing incoming shortwave radiation vertically below ocean surface in the SEISAMW $\sigma_\theta$ range (considering two outcrop windows, 26.75 < $\sigma_\theta$ < 26.85 kg m$^{-3}$ and 26.85 < $\sigma_\theta$ < 26.95 kg m$^{-3}$) and SEPSAMW $\sigma_\theta$ range (also considering two outcrop windows, 26.85 < $\sigma_\theta$ < 26.95 kg m$^{-3}$ and 26.95 < $\sigma_\theta$ < 27.05 kg m$^{-3}$).


