

The observed spatiotemporal variability of Antarctic Winter Water

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Key Points:

- Seasonal climatologies of Antarctic Winter Water (WW) and its properties are mapped using 18 years of in-situ observations.
- Observations reveal the distinct seasonal and regional characteristics of WW and their connections to processes such as sea-ice formation.
- Localized redistribution of WW properties equatorward is steered by large topographic features.

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Abstract

The Southern Ocean is central to the global overturning circulation. South of the Antarctic Polar Front, Antarctic Winter Water (WW) forms in the wintertime mixed layer below sea ice and becomes a subsurface layer following summertime restratification of the mixed layer, overlaying upwelled deep waters. Model simulations show that WW acts as a conduit to seasonally transform upwelled deep waters into intermediate waters. Yet, there remains little observational evidence of the distribution and seasonal characteristics of WW. Using 18 years of in situ observations, we show seasonal climatologies of WW thickness, depth, core temperature and salinity. This study reveals, for the first time, the distinct regionality and seasonality of WW. The seasonal cycle of WW characteristics is tied to the annual sea ice evolution, whilst the spatial distribution is impacted by the main topographic features in the Southern Ocean driving an equatorward flux of WW. Through the identification of these localized northward export regions of WW, this study provides further evidence suggesting an alternative view from the conventional ‘zonal mean’ perspective of the overturning circulation. We show that specific overturning pathways connecting the subpolar ocean to the global ocean can be explained by ocean-topography interactions.

Plain Language Summary

The Southern Ocean around Antarctica is central to the global ocean circulation system. The cold wintertime atmosphere drives ocean cooling and sea ice formation, which causes surface waters to become denser, mixing with deep waters that rise to the surface from the deep ocean. As the ocean surface layer warms in summer, there remains a cold layer below the surface known as Antarctic Winter Water (WW). This layer warms throughout the summer, thinning the WW layer. However, the properties of WW (temperature, salinity, thickness, depth) vary in space around the Southern Ocean and in erosion rate. By compiling 18 years of ocean observations, we investigate the physical dynamics that determine how WW changes in space and over the average annual cycle. We find that there are localized regions across the Southern Ocean where WW properties are transported northwards as part of the ocean circulation system, which typically align with large topographic features and act to connect Southern Ocean water masses to the global ocean.

1 Introduction

The Southern Ocean (SO) is characterized by the zonally unbounded energetic flow of the Antarctic Circumpolar Current (ACC), which consists of several baroclinic jets that delineate hydrographic properties (Orsi et al., 1995). South of the Antarctic Polar Front, salinity is the dominant property of density in the upper ocean due to a low thermal expansion coefficient (Stewart & Haine, 2016; Roquet et al., 2022). Consequently, this allows for the formation of a cold subsurface layer known as Antarctic Winter Water (AAWW; henceforth WW) that resides below the summertime mixed layer (Toole, 1981).

WW lies above the salty and relatively warm Circumpolar Deep Water (CDW), which has been upwelled along isopycnals from the deep ocean (Tamsitt et al., 2017). CDW gets transformed to become either dense bottom water in the southern upwelling limb, sinking back to the abyssal ocean (Ganachaud & Wunsch, 2000; Jacobs, 2004; van Sebille et al., 2013), or transformed to intermediate water north of the ACC (Munk, 1966; Whitworth et al., 1994). Evans et al. (2018) showed, using a general circulation model, that WW acts as a seasonal conduit to connect CDW to Antarctic Intermediate Water in summertime which, when subducted, forms the downwelling limb of the SO meridional overturning circulation (Speer et al., 2000; Marshall & Speer, 2012; Talley, 2013). Water mass transformation in the upper SO is associated with net freshening due to sea

63 ice export and melt (Abernathey et al., 2016; Pellichero et al., 2018), and Tamsitt et al.
64 (2018) showed that upwelled CDW becomes fresher and colder before transforming into
65 intermediate waters, further implying the seasonal role of WW in the overturning cir-
66 culation. Drake et al. (2018) show through investigation of Lagrangian pathways that
67 CDW upwelling is not well represented via the time mean overturning framework, and
68 thus, by implication, a similar argument likely applies to the representation of WW in
69 the same zonally integrated overturning framework. Given these findings, the evolution
70 and characteristics of WW are important to the overall SO's role in the global overturn-
71 ing circulation.

72 WW forms in the wintertime mixed layer (ML; Figure 1a), which deepens through
73 buoyancy loss via cooling and sea ice formation. Brine rejection deepens the under-ice
74 ML, driving entrainment of the warm subsurface CDW and thus increases ML heat and
75 salinity (Gordon & Huber, 1984). In turn, this melts sea ice and freshens the ML to re-
76 duce its buoyancy; thus, through this feedback system, sea ice thickness is maintained
77 as relatively thin (~ 0.5 m) (Shaw & Stanton, 2014) and the under-ice ML is generally
78 confined to less than 200 m depth (Martinson, 1990; Biddle & Swart, 2020; Wilson et
79 al., 2019). Following a change in sign of the atmospheric heat flux, the summertime ML
80 shoals becoming warm and fresh with a depth of only tens of meters (Pellichero et al.,
81 2017), leaving a residual subsurface cold wintertime ML below the summer-warmed sur-
82 face layer (Figure 1b) and thus creating a warm-cold-warm layering of ML-WW-CDW
83 (Park, Charriaud, & Fieux, 1998).

84 Eventually, WW erodes over the annual cycle or is re-entrained into the following
85 winter ML. Giddy et al. (2023) found that WW erosion rates are primarily driven by en-
86 trainment of WW into the mixed layer through mechanical mixing, with further warm-
87 ing due to double diffusive convection from below. However, these findings are specific
88 to the Weddell Sea where there is typically a sharp thermocline yet a weak pycnocline
89 (Wilson et al., 2019), which impacts the temperature gradient as well as the rate of mix-
90 ing that takes place. Further, variability of WW properties have been observed in re-
91 gions up- and downstream of large topographic features that interact with the ACC, en-
92 hancing the downstream mesoscale field, such as that found by Sabu et al. (2020). Lund
93 et al. (2021) show similar spatial variability across the circumpolar SO via climatolog-
94 ical WW temperatures. Sabu et al. (2020) also observe the WW layer responding to large-
95 scale atmospheric variability, finding enhanced warming in years of positive Southern An-
96 nular Mode, likely due to elevated wind-driven mixing from stronger winds. The vari-
97 ability of WW observed is tied to various mechanisms that impact upper ocean struc-
98 ture, such as varying eddy kinetic energy (Dove et al., 2022; Nikurashin et al., 2013), dif-
99 fering wind stress fields (Abernathey et al., 2011; Lin et al., 2018), and CDW outcrop-
100 ping pathways (Narayanan et al., 2023; Tamsitt et al., 2021).

101 In this study we provide revised criteria for defining the vertical distribution of WW.
102 We apply this to 18 years of in-situ observations over the SO to examine the spatiotem-
103 poral distribution of WW, its properties using seasonal climatologies and the dynam-
104 ics governing these distributions. A seasonally evolving heat budget for WW is estimated
105 to shed light on the dominating processes which impact the evolution and erosion of WW
106 and to better understand its connection to the surface layer of the ocean and meridional
107 circulation. Finally, we reveal linkages between WW distribution and its properties to
108 the underlying topography of the SO. WW variability is directly tied to rates of over-
109 turning, so determining possible pathways of enhanced overturning is critically impor-
110 tant in understanding our global climate and the redistribution of oceanic properties.

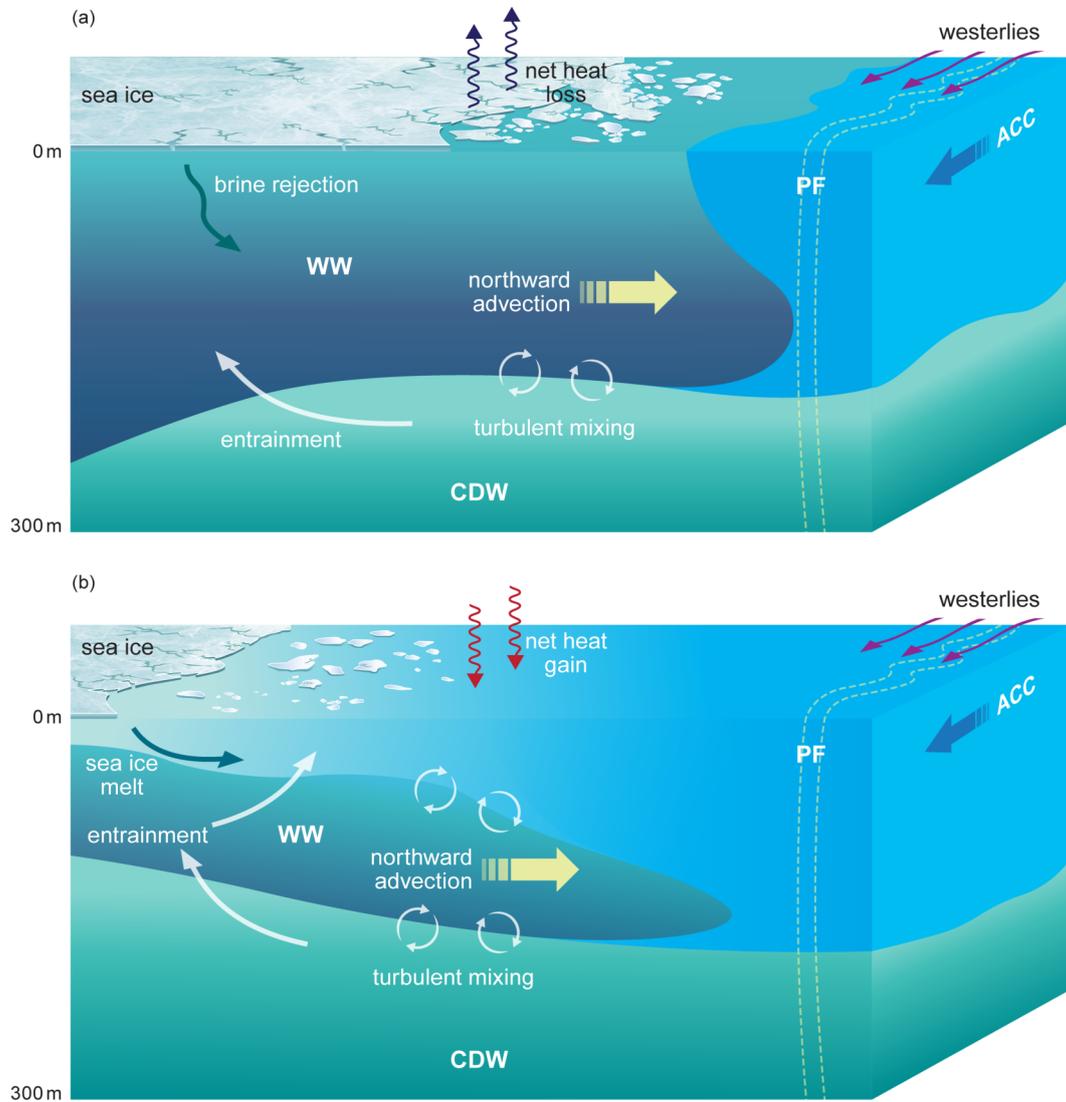


Figure 1. Conceptual view of Antarctic Winter Water. (a) WW_{ML} in wintertime and (b) WW_{SS} in summertime following summertime mixed layer restratification. Shown are the various physical mechanisms that impact the formation, distribution and erosion of WW.

2 Data and Methods

2.1 Hydrographic Profiles

This study uses hydrographical profiles south of 40°S from the period 2004 to 2021 compiled using the Argo float dataset (Wong et al., 2020), tagged marine seals dataset (MEOP: Marine Mammals Exploring the Oceans Pole to Pole) (Treasure et al., 2017), Southern Ocean biogeochemical floats (SOCCOM: Southern Ocean Carbon and Climate Observations and Modeling) (SOCCOM, 2019), as well as ship-based CTD casts and glider profiles (from WOD18: The World Ocean Database 2018) (Boyer et al., 2018). Data are relatively well distributed across the SO due to sampling from the Argo and MEOP programmes (Figures 2e-2h), which is extensive compared to limited historical CTD ship-based data collection (Brett et al., 2020). The Pacific sector is mainly sampled throughout the annual cycle via the Argo programme, whereas MEOP data cover a great extent south of the Polar Front, particularly in the Atlantic and Indian Ocean sectors, as well as below sea ice and across the circumpolar continental shelf (Narayanan et al., 2023). MEOP seals are tagged in the summertime, so provide least coverage in spring after CTD sensors have dropped off the seals from molting their winter fur, whilst their habitat location and foraging behavior largely dictate their spatial distribution (McMahon et al., 2019).

Additionally, the location of float sampling is initially spatially biased to deployment location due to ship-based deployments, and subsequently follow quasi-Lagrangian trajectories. Therefore, float data are heterogeneously distributed in space and time (Figures 2a-2d) with different regions sampled in different time periods. There are fewer total profiles per month from 2020 and onwards (Figure 2i) because of limited scientific cruises during the global pandemic (Boyer et al., 2023; Sarmiento et al., 2023). Consequently, there are limited MEOP data in 2020 and 2021, but the remaining operating floats continued to sample the SO.

In sea-ice-covered regions, ship-based and seal-based measurements are geographically accurate as they contain GPS fixes; profiling floats, however, use a temperature-based sea ice avoidance algorithm and cannot transmit their location when under sea ice. Profiling floats park at depth when not profiling where the currents are assumed to be low, so the profile locations are linearly interpolated. For this study, we take it as appropriate for use since we are interested in broadscale findings, where the positional errors from linear interpolation have little effect when gridding to spatial bins of 1°, as shown by Wong and Riser (2011) who find a mean distance between profiles to be 32 km.

Float-based measurements of temperature and salinity have been corrected for any sensor biases in real-time and undergo delayed mode quality control testing (Wong et al., 2020). We only use data flagged as “good” and “under sea-ice” for all data sources. Following Wilson et al. (2019), we further quality control the data such that each profile contains at least five measurements in the top 300 dbar with at least one data point in the top 25 dbar to ensure representation of the upper ocean processes of interest, ensuring both the ML and majority of WW properties are captured. Profiles containing unrealistic values of potential temperature and practical salinity are removed such that $-2.5 < T < 16$ °C and $30 < S < 36$. These thresholds are obtained from the observed range of properties south of the Subantarctic Front. Consequently, in this study, we analyze 620,293 quality-controlled CTD profiles.

The profiles are linearly interpolated to a standard two dbar vertical grid followed by a one-dimensional Gaussian filter with a standard deviation of four to smooth the data but retain key features. Absolute salinity and conservative temperature are derived from these gridded, quality controlled and smoothed profiles. The WW detection algorithm (Section 2.2) is applied to each profile, with all depth-related variables converted to meters, before gridding the data onto a median 1°×1° latitude-longitude seasonal climatol-

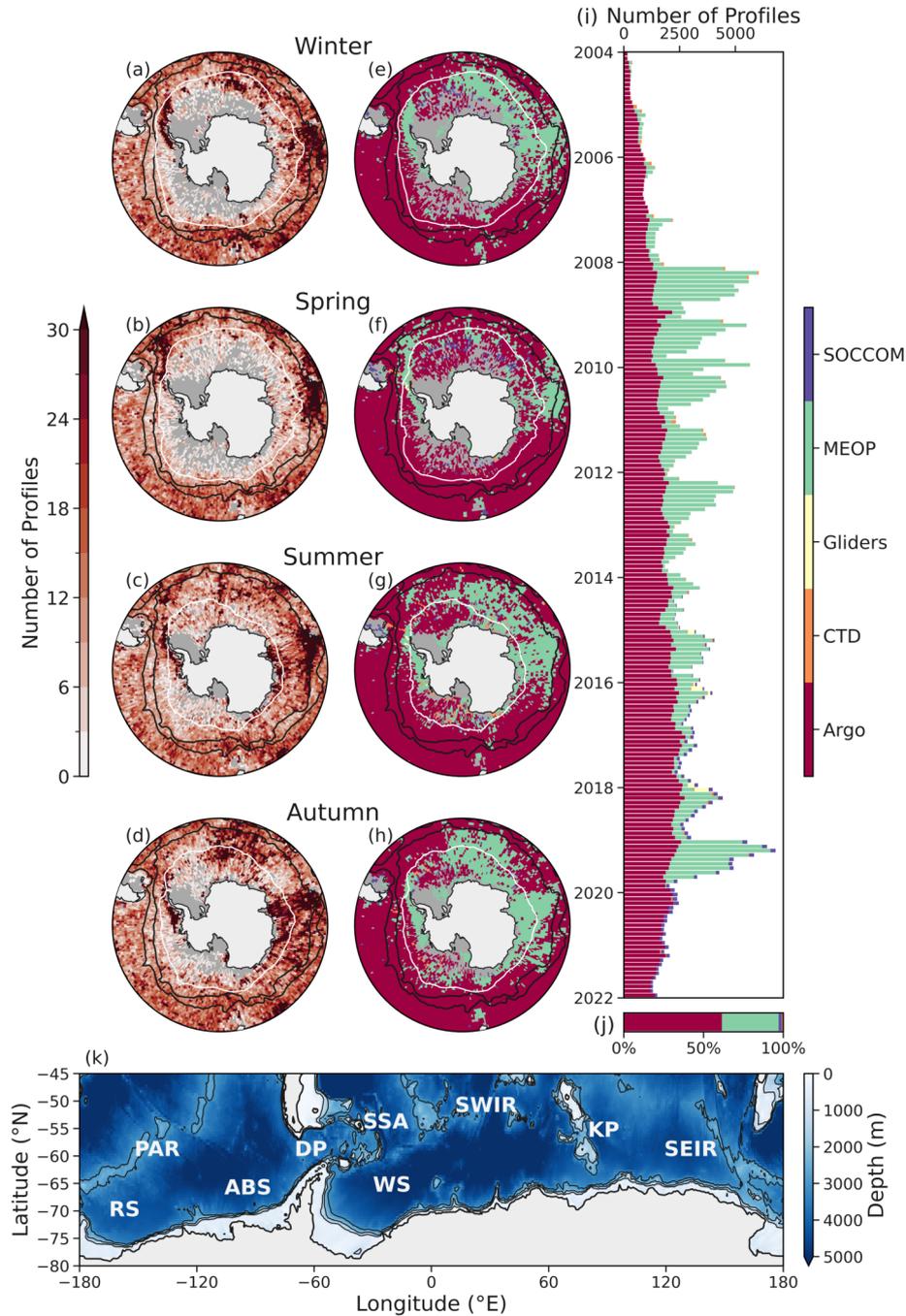


Figure 2. Hydrographic data distributions. (a-d) Spatial distribution of the total number of hydrographic profiles per season from winter through to autumn, respectively. (e-h) Spatial distribution of the mode data source per grid cell. The black contours indicate the Polar Front and Subantarctic Front, and the white lines are the mean 15% sea ice concentration for the relevant season in (a-h). (i) Monthly time series of data sources (stacked) across the Southern Ocean. (j) Total proportion of data per data source as a percentage. (k) Southern Ocean bathymetry map with bathymetric features and seas. Thin black lines denote 1, 2 and 3 km bathymetric contours. Abbreviations are as follows: RS=Ross Sea, PAR=Pacific-Antarctic Ridge, ABS=Amundsen and Bellingshausen Seas, DP=Drake Passage, SSA=South Sandwich Arch, WS=Weddell Sea, SWIR=Southwest Indian Ridge, KP=Kerguelen Plateau, SEIR=Southeast Indian Ridge.

162 ogy. Since the formation and annual cycle of WW is tied to sea ice evolution, we define
 163 the climatological seasons based on the annual cycle of sea ice formation and melt such
 164 that January, February, and March represent summer, and so on (Holland, 2014; Goosse
 165 et al., 2023).

166 2.2 Antarctic Winter Water Algorithm

167 WW definitions vary in present literature. Park, Charriaud, Pino, and Jeandel (1998)
 168 define WW as a relatively homogenous subsurface layer south of the PF with a subsur-
 169 face temperature minimum, defining the upper boundary with a strong hydrographic gra-
 170 dient (thermo-, halo-, pycnocline) and the lower boundary as strong temperature and
 171 salinity gradients. Evans et al. (2018) define WW via a temperature threshold such that
 172 $\Theta < 0.5$. Sabu et al. (2020) define WW as the depth of the coldest temperature inver-
 173 sion observed below the ML. Giddy et al. (2023) use a more comprehensive definition:
 174 the upper boundary as the mixed layer depth (MLD), core depth as the temperature min-
 175 imum below the MLD, and lower boundary as the max temperature gradient below the
 176 WW core, marking the boundary between WW and CDW.

177 We characterize WW into two classifications: mixed layer WW (WW_{ML} , Figures
 178 1a and 3e) and subsurface WW (WW_{SS} , Figures 1b and 3f-3h). WW_{ML} is defined as con-
 179 strained to mixed layer (ML), taking the upper bound as the ML reference depth and
 180 the lower bound as the mixed layer depth (MLD); we assume the ML is well-mixed, and
 181 therefore take the mid-depth of MLD as the WW core depth, whilst the core properties
 182 are defined as the mean properties across the ML. WW_{SS} only exists if there is a tem-
 183 perature inversion below the ML. That is, WW_{SS} only exists if there are at least 10 in-
 184 versions, equivalent to 20 m, between the mean ML temperature and subsurface tem-
 185 peratures. We describe WW_{SS} through its core properties (the temperature minimum,
 186 the depth at which the temperature minimum occurs, as well as the salinity and den-
 187 sity at that depth) and thickness (difference in depth between upper and lower bound-
 188 ary). We summarize the property definitions of the WW classifications in Table 1. Note
 189 that we define the MLD following the density difference criterion of 0.03 kg m^{-3} from
 190 a reference density taken from the surface at 10 dbar, as per de Boyer Montégut (2004).

191 WW only exists if the core temperature is below $2 \text{ }^\circ\text{C}$ since this temperature de-
 192 notes the boundary for the PF at 200 m depth (Belkin & Gordon, 1996; Morrow et al.,
 193 2008; Pollard et al., 2002). The PF is a baroclinic barrier to WW, marking the transi-
 194 tion from beta- to alpha-dominating regimes in the upper ocean, which determines the
 195 existence of temperature inversions (Orsi et al., 1995; Stewart & Haine, 2016; Caneill et
 196 al., 2023). If the temperature at the lower bound is warmer than $2 \text{ }^\circ\text{C}$, the $2 \text{ }^\circ\text{C}$ isotherm
 197 depth is used as the WW lower boundary instead. Further, if the core depth is deeper
 198 than the lower bound depth, we conclude no WW exists in that profile. Similarly, if the
 199 core temperature is warmer than $2 \text{ }^\circ\text{C}$ then we conclude no WW exists in that profile.
 200 The lower boundary and, or core depth of WW can feasibly reach depths below 300 m,
 201 for example via deep convection cells as a result of polynya driven processes, which typ-
 202 ically are found close to the Antarctic continent (Morales Maqueda et al., 2004; Tamura
 203 et al., 2008; Mohrmann et al., 2021; Whitworth et al., 1994). However, these processes
 204 are associated with deep water formation (Jacobs, 2004; Johnson, 2008; Kushara et al.,
 205 2017; Meredith, 2013), and therefore associated with dynamics outside the scope of this
 206 study. WW is otherwise typically found in the bounds of 50-200 m (Lund et al., 2021;
 207 Sabu et al., 2020) (Figure 3). Hence, WW profiles with a lower boundary deeper than
 208 300 m are removed.

209 In total, 43% of all profiles (266,856 profiles) contain WW: 105,588 WW_{ML} pro-
 210 files and 161,268 WW_{SS} profiles (Figures 3a-3d). We then compute the climatologies,
 211 which are smoothed using a $3^\circ \times 3^\circ$ latitude-longitude rolling mean and interpolated across
 212 a maximum gap of three missing grid cells to the maximum WW extent.

Table 1. Summary of Antarctic Winter Water properties: upper boundary, core and lower boundary depth definitions. We take depth, z to be positive and therefore increasing with depth.

Property	WW _{ML}	WW _{SS}
Upper boundary (z_{ub})	10 m	MLD or 2°C isotherm if $T_{MLD} > 2^\circ\text{C}$
Core	Core depth: mid-depth of ML Core T, S, density: average across ML	Depth of temperature minimum such that $z_{ub} < z_{core} < z_{lb}$ and $T_{min} < 2^\circ\text{C}$
Lower boundary (z_{lb})	MLD	$z_{core} < z \left(\frac{dT}{dz} \right)_{max} < 300$ dbar or 2°C isotherm if $T_{core} > 2^\circ\text{C}$ such that $z_{core} < 300$ dbar

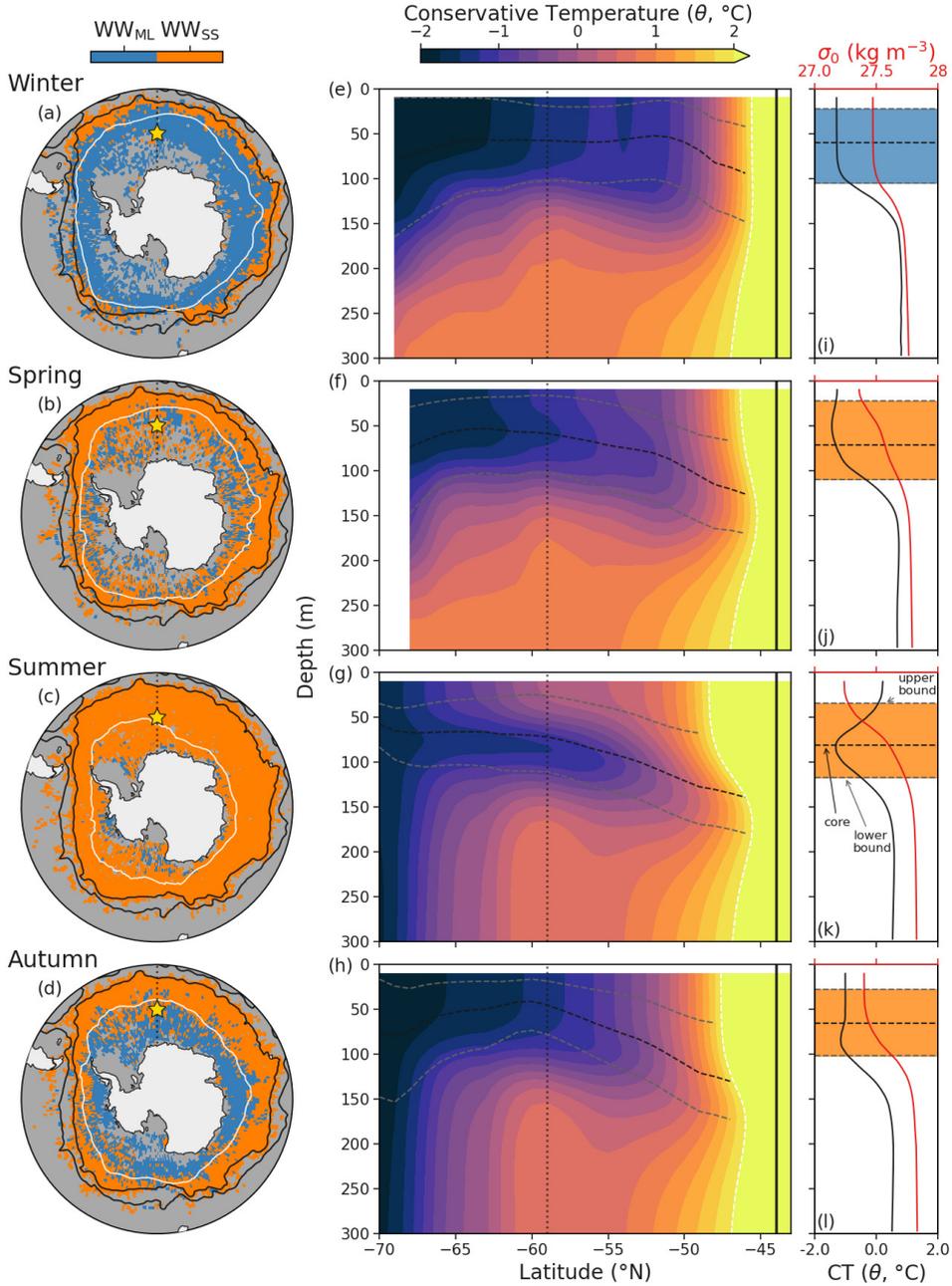


Figure 3. Antarctic Winter Water classification. (a-d) Spatial distribution of the mode classification of WW for each grid cell from winter through to autumn, respectively, where blue (orange) represents WW_{ML} (WW_{SS}) profiles. The black lines denote the PF and SAF of the ACC, and the white lines show the seasonal mean 15% sea ice concentration; the dotted black line shows the cross section location in (e-h) and the star denotes the profiles in (i-l). (e-h) Smoothed meridional cross section along 0°E for each season. The solid black vertical lines denote the PF; the light gray dashed lines denote the upper and lower bounds of WW, the black dashed line denotes the WW core, and the white dashed line denotes the 2 °C isotherm. The dotted black lines show the location of the star in (a-d) and are the profiles shown in (i-l). (i-l) show profiles of conservative temperature (black) and potential density (red) from the location 59°S, 0°E throughout the seasonal cycle. The light gray dashed lines denote the upper and lower bounds of WW, and the black dashed line denotes the WW core as in (e-h). The classification is shown by shading: blue (orange) shading represents WW_{ML} (WW_{SS}) profiles, as in (a-d).

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2.3 Temperature Tendency / Heat Budget

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Adapting the temperature tendency equation of Giddy et al. (2023) to Equation 1, we compute a temperature tendency of WW_{SS} to investigate the dynamics that impact how WW evolves and erodes over the annual cycle to better understand the connection between the subpolar surface ocean and the meridional circulation system.

$$\frac{\overbrace{\delta T_{WW}}^{(I)}}{\delta t} = \frac{1}{\rho_0 C_p h_{WW}} \left(\underbrace{\Delta Q_{SW}}^{(II)} + \underbrace{Q_{mixing}}^{(III)} + \underbrace{Q_{entrainment}}^{(IV)} + \underbrace{Q_{formation/erosion}}^{(V)} + \underbrace{Q_{residual}}^{(VI)} \right). \quad (1)$$

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Term (I) is the temperature tendency of the WW_{SS} system, which can be calculated by taking the time derivative of the WW integrated heat content, $\frac{\delta T_{WW}}{\delta t} = \frac{\delta}{\delta t} \left(\frac{1}{h_{WW}} \int \text{thickness} \Theta \delta z \right)$ (Huguenin et al., 2022), after adjusting Θ by the freezing point of seawater to negate sign errors.

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Term (II) represents the penetrative shortwave radiation absorbed by the WW layer, which is the difference in shortwave entering at the upper boundary and leaving at the lower boundary. This term is approximated via exponential decay of shortwave solar radiation from the surface using Jerlov water type II as per Paulson and Simpson (1977):

$$R_s(z) = R_s(0) \left[0.77 \exp\left(\frac{-|z|}{1.4}\right) + 0.23 \exp\left(\frac{-|z|}{14}\right) \right].$$

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Term (III) is the down-gradient turbulent heat flux term across the ML-WW interface due to wind-driven mechanical mixing. Nicholson et al. (2022) show a close correlation between theoretical and observed dissipation rates in Southern Ocean summertime ($r^2 = 0.75$, see their Figure S5 in Supporting Information). Hence, we calculate the diffusivity, κ , such that:

$$\kappa = \Gamma \frac{\varepsilon}{N^2}$$

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where the mixing coefficient is a constant, which we take as the modified Osborn number for a transitional regime, $\Gamma = 0.2$ (Osborn, 1980; Bouffard & Boegman, 2013; Gregg et al., 2018). Assuming winds is the dominant driving factor, we approximate turbulence dissipation, (ε) at the base of the mixed layer (i.e., the upper boundary of the WW) via law of the wall (von Kármán, 1931) such that $\varepsilon = \frac{u_*^3}{kz}$, using frictional velocity, $u_*^3 = \sqrt{\frac{\tau}{\rho_{sw}}}$, where τ is the climatological mean wind stress and $\rho_{sw} = 1035 \text{ kg m}^{-3}$ is the reference density of seawater (Huguenin et al., 2022). Further, we use von Kármán's constant, $k = 0.41$, and z as the MLD. N^2 is the stratification across the upper boundary of the WW_{SS}.

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Term (IV) measures the change in WW heat from entrainment and detrainment, which results in a flux of temperature, at both the upper (ML-WW) and lower (WW-CDW) boundaries. Term (IV) is calculated using the general formula, $Q_{ent} = \frac{H}{h} w_{eIF} \Delta T_{IF}$, where H is the Heaviside function of vertical velocities such that

$$H(w_{eIF}) = \begin{cases} 0 & \text{if } w_{eIF} < 0, \\ 1 & \text{if } w_{eIF} > 0. \end{cases}$$

244 where $w_{e_{IF}} = \frac{dh}{dt}$ is the vertical velocity and IF is the interface between water
 245 masses. ΔT is taken across the ML-WW interface as the difference between the ML tem-
 246 perature and 15 m below the MLD (Pellichero et al., 2017; du Plessis et al., 2022). Across
 247 the WW-CDW layer, we use the average WW temperature minus the temperature 5 m
 248 below the WW, as per Giddy et al. (2023).

249 Term (V) represents temperature due to the formation or erosion of WW_{SS}. WW_{SS}
 250 is created following the restratification of the wintertime ML, resulting in a warm and
 251 fresh summertime ML that caps a cold layer of remnant wintertime ML that we have
 252 defined as WW_{SS}. WW_{SS} is formed if there exists no WW_{SS} in t_{i-1} and WW_{SS} exists
 253 in t_i , which results in an input of heat into the WW_{SS} system, thereby warming WW_{SS}.
 254 Conversely, we assume WW_{SS} has been eroded if there is WW_{SS} in t_{i-1} and no WW_{SS}
 255 in t_i , which causes a cooling of WW_{SS}. This assumes that WW_{SS} totally erodes over the
 256 annual cycle and that WW_{ML} formation includes remaining WW_{SS} re-entrained into the
 257 wintertime ML. However, given the temporal resolution of the seasonal climatology, it
 258 does not allow for direct observation of total erosion WW_{SS}.

259 Lastly, term (VI) incorporates any processes not accounted for in the other terms,
 260 which includes (but is not limited to) background mixing, diffusive convection, small-
 261 scale processes, uncertainties in the calculated terms, and horizontal advection term. Dif-
 262 fusive staircases perpetuate across large regions of the SO (van der Boog et al., 2021).
 263 Whilst the resultant diffusive mixing is important for water mass transformation (Evans
 264 et al., 2018), the heat flux contribution remains small (Bebieva & Speer, 2019; Giddy
 265 et al., 2023; Wilson et al., 2019). Further, vertical processes, such as submesoscale mo-
 266 tions, are not resolvable given the geostrophic scale of the gridded dataset, meaning their
 267 attributed heat fluxes are not directly computable (Section 4.2), but are incorporated
 268 as part of the geostrophic mean field flux. Given the numerical findings of Morrison et
 269 al. (2016), who show a substantial northward heat flux by the mean flow, we therefore
 270 assume that lateral advection dominates the residual term, similar to Pellichero et al.
 271 (2018).

272 2.4 Other Data

273 We determine the sea ice edge using gridded daily sea ice data (Spreen et al., 2008).
 274 Since WW evolution is tied to sea ice dynamics, we often distinguish the SO into two
 275 domains: the under-ice zone, which we define as the region south of the 15% sea ice con-
 276 centration line for the respective period that is typically associated with sea ice cover,
 277 which includes a transitional region from an area of mixed sea ice and open ocean con-
 278 ditions to the sea ice capped ocean; and the open-ocean zone refers to north of the 15%
 279 sea ice concentration line, which is not necessarily ice-free but is an area of mixed sea
 280 ice conditions that transitions from the sea-ice capped ocean to open ocean.

281 ACC boundaries of the PF and SAF were defined using absolute dynamic topog-
 282 raphy at -0.58 m and -0.1 m, respectively (Park et al., 2019), using monthly altimetry
 283 data via AVISO. Shortwave radiative heat flux (Q_{sw}) and wind stress, τ , at monthly and
 284 1° resolution from ERA5 reanalysis of the European Centre for Medium Range Weather
 285 Forecasts (Hersbach et al., 2023) are used for the calculation of term (I) and term (III)
 286 of Equation 1, respectively, and are gridded before use in computation.

287 3 Results and Discussion

288 3.1 The Seasonal Cycle of Antarctic Winter Water

289 3.1.1 Antarctic Winter Water Spatial Extent

290 The annual cycle of WW is intricately linked to the sea ice cycle (Figure 3a-3d):
 291 WW forms during the sea ice formation periods of autumn and winter in the under-ice

292 zone in the ML (WW_{ML}) as a cold and homogenous water mass (Figures 1a, 3a and 3d).
 293 WW_{ML} also forms north of the 15% sea ice concentration line from ocean surface layer
 294 cooling. In winter, WW_{ML} forms a mean maximum extent of 400 ± 280 km (approx-
 295 imately 3.4 ± 2.8 degrees of latitude) north of the sea ice concentration line in the open-
 296 ocean zone. 78% of WW profiles in winter are classified as WW_{ML} , and the remainder
 297 of profiles are WW_{SS} (profiles containing a subsurface temperature minimum) in the open-
 298 ocean zone at the northern extent of WW. The WW_{SS} profiles are on average north of
 299 the Southern Boundary of the ACC, as shown by Jones et al. (2023), which delineates
 300 regions of differing oceanic properties.

301 During periods of sea ice melt (spring and summer), the ML restratifies and warms
 302 from solar heating, which overlays the remnant cold ML. Consequently, there exists a
 303 subsurface temperature minimum, WW_{SS} (Figures 1b, 3f-h and 3j-l), which is sustained
 304 via beta ocean properties (Stewart & Haine, 2016). The spring season has a heteroge-
 305 neous mix of WW_{ML} and WW_{SS} classified profiles, with 76% of spring WW profiles be-
 306 ing WW_{SS} (Figure 3b). The WW_{ML} extent retreats at different rates around the circum-
 307 polar SO during spring; across and to the west of KP remains as a mix of WW_{ML} and
 308 WW_{SS} in the open-ocean zone, whereas the rest of the open-ocean zone is largely WW_{SS} .

309 The ocean transitions to a summer ocean that is dominated by WW_{SS} (93% of WW
 310 profiles; Figure 3c). WW_{ML} is only found near the Antarctic continent in summer, which
 311 are likely regions that do not form any WW_{SS} throughout the entire annual cycle. An
 312 example of such a region is Pine Island Bay in the Amundsen and Bellingshausen Seas
 313 (ABS), where coastal modified CDW maintains a warm subsurface ocean and the Thwaites
 314 glacier concurrently drives a cold surface layer (Dotto et al., 2022). Similarly, WW_{SS} per-
 315 sists throughout the entire annual cycle at the northernmost extent of WW domain (Fig-
 316 ures 3a-3d and S1), which can only be sustained through cold subsurface WW being trans-
 317 ported northward to the regions where WW_{ML} do not form.

318 Surface ocean cooling and sea ice growth during autumn lead to an increase of WW_{ML}
 319 (Figure 3d), with the under-ice zone almost totally WW_{ML} and localized regions form-
 320 ing WW_{ML} north of the 15% sea ice concentration line. Northern portions of the under-
 321 ice zone in the Weddell Sea and Ross Sea have a mix of WW_{ML} and WW_{SS} , which is
 322 likely related to gyre circulation: cold WWs in the ML are maintained in the gyre core
 323 and edge, whilst mixing drives warmer and deeper (subsurface) WWs at gyre periph-
 324 eries where the ocean transitions to ACC dynamics (Jones et al., 2023). Similarly, the
 325 northern under-ice region of the ABS has a mixed WW composition of WW_{ML} and WW_{SS} .

326 WW profiles are largely constrained to the extent of the PF, as per the WW def-
 327 inition (Section 2.2; Table 1), but there are regions with WW detected northward of the
 328 PF. In the Pacific sector of the SO, WW is consistently detected north of the PF through-
 329 out the seasonal cycle (Figures 3a-3d); SAMW, similar to WW, forms via buoyancy loss
 330 and drives ML deepening, which takes place in the vicinity of the ACC and northwards
 331 (Boland et al., 2021; Cerovečki et al., 2013; Meijers et al., 2019; Wang et al., 2022). Hence,
 332 the detection of WW north of the PF in the Pacific sector where Shallow Salinity Min-
 333 imum Water (Karstensen, 2004) is observed in a region transitioning from a beta den-
 334 sity regime to an alpha density regime (Roquet et al., 2022) (see their Figure 2b). There-
 335 fore, WW may potentially be aliased with the cold Subantarctic Mode Water of the Pa-
 336 cific sector, which is also a low stratification water mass formed through surface buoy-
 337 ancy loss in the ML (Li et al., 2021; Herraiz-Borreguero & Rintoul, 2011; Xia et al., 2022).
 338 These WW profiles are removed from the analysis. Further, there is likely very little WW
 339 identified between the PF and SAF due to the ACC acting as a strong baroclinic bar-
 340 rier.

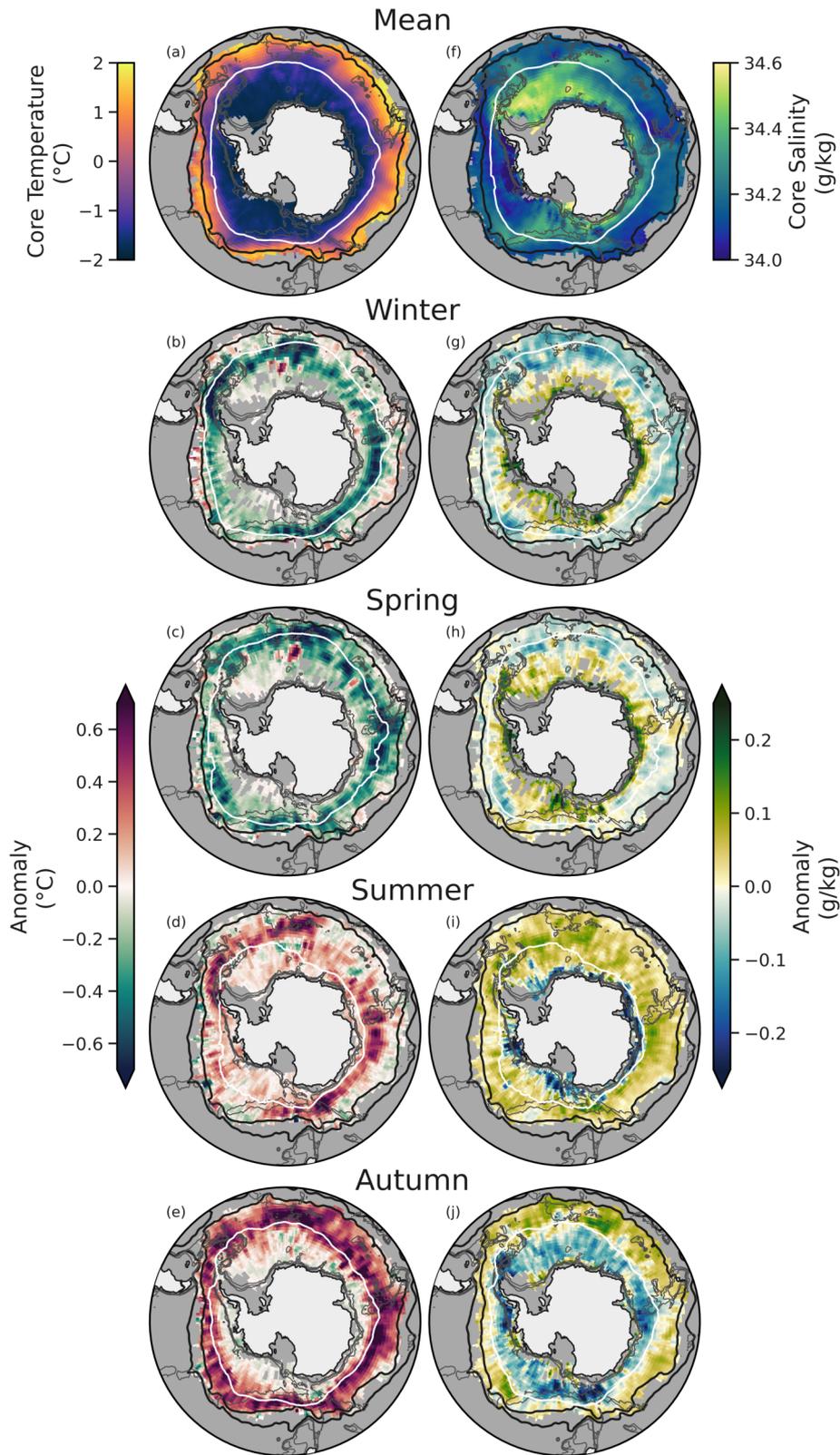


Figure 4. Annual mean and seasonal anomalies of Antarctic Winter Water: (a-e) core conservative temperature, and (f-j) core absolute salinity. A negative (positive) anomaly implies a decrease (increase) from the annual mean. Black lines indicate the PF and SAF, white lines indicate the mean 15% sea ice concentration for the time period, and gray lines indicate 1km, 2km and 3km isobaths.

3.1.2 Antarctic Winter Water Core Thermohaline Properties

Core temperature and salinity of WW exhibit distinct seasonal cycles (Figures 4); nonetheless, both core properties exhibit large-scale spatial homogeneity in the under-ice and open-ocean zones (Figures 4a, 4f, 5b and 5d). WW forms below sea ice from a cold ocean nearing freezing point, so is largely homogenous in its mean temperature across the SO (Figures 5a), particularly in the under-ice zone (Figures 4a and 5b). Brine rejection from sea ice formation, and surface cooling drives buoyancy loss that results in a deepening of the ML (Wilson et al., 2019). Consequently, the coldest and saltiest WW is consistently observed during under-ice conditions throughout the annual cycle (Figures 4a, 4f, 5b and 5d).

Spatially, WW core temperature in the under-ice zone remains largely homogeneous with a mean temperature of $-1.4 \pm 0.5^\circ\text{C}$ in the under-ice zone, whilst core salinity exhibits a circumpolar mean of 34.26 ± 0.18 g/kg (Figures 5b and 5d). WW cores are substantially fresher (34.11 ± 0.12 g/kg) in the Amundsen-Bellinghousen Seas (ABS): the proximity of the ACC fronts and the eastward limb of the Ross Gyre transport CDW southwards, maintaining the region as a warm shelf sea and facilitating elevated sea ice melt rates, which is likely responsible for the fresher WW core salinity as well as freshening the under-ice zone mean and increasing the standard deviation (Figure 5c) (Nakayama et al., 2018; Narayanan et al., 2023; Tamsitt et al., 2021; Thompson et al., 2018). The coldest and most saline WW cores are observed within the polar gyres (approximately -1.5°C , 34.6 g/kg; Figures 4a, 4f, 5a and 5c), which are maintained through their southern proximity and elevated sea ice production. On the other hand, there is distinct variability displayed in all core properties (Figures 4-7) in the open-ocean zone when comparing properties up- and downstream of large topographic features. Our findings agree with Sabu et al. (2020), who show distinctly warmer core temperatures downstream of Kerguelen Plateau (KP); similar distinctions can be made when considering the differences in core properties up- and downstream of the Southwest Indian Ridge (SWIR) and the Southeast Indian Ridge (SEIR). These differences in properties may be a consequence of topographic features stirring the downstream water column, which enhances mixing rates (Nikurashin et al., 2013; Rosso et al., 2015; Mashayek et al., 2017; Tamsitt et al., 2017) and alters the spatial gradients of WW core temperature and salinity. Nonetheless, meridional mean WW core temperatures (Figure 5a) do not differ substantially across the circumpolar SO since the formation of the temperature minimum is determined during WW formation in WW_{ML} from wintertime net ocean heat loss, and consistently approaches freezing point (Figures 4b-4e and 5a).

Core temperature and salinity in the open-ocean zone have a strong tie to latitudinal extent, getting warmer and fresher as latitudes extend northward with a mean warming rate of 0.16°C per degree of latitude and decreasing in salinity at a rate of 0.01 g/kg per degree of latitude in the open-ocean zone (Figures 5b and 5d). Note that the latitudinal mean core salinity peaks in freshness with a large standard deviation at approximately 70°S (34.19 ± 0.22 g/kg; Figure 5d); this is influenced by the fresh WW core salinities in the ABS, which is the only region that has a meridional mean salinity below 34.2 g/kg (Figure 5c). The mean open-ocean zone core temperature and salinity are $0.6 \pm 0.8^\circ\text{C}$ and 34.23 ± 0.10 g/kg, respectively, with the warmest and freshest cores found in the northernmost domain of WW extent ($\sim 50^\circ\text{S}$; Figures 5b and 5d). WW core temperature and salinity properties that are typically associated with under-ice conditions (i.e. colder, more saline) extend northwards into the open-ocean zone in regions that align with large topographic features; that is: Pacific-Antarctic Ridge (PAR), South Sandwich Arch (SSA), SWIR, KP and the SEIR (Figures 4a, 4f, 5a, and 5c). This agrees with the findings of Park et al. (2009), who observed a subsurface cold water tongue advected northwards across the KP.

WW core temperature exhibits large changes in temperature throughout the annual cycle, which are spatially heterogeneous (Figures 4b-4e). In winter, a net ocean-atmosphere

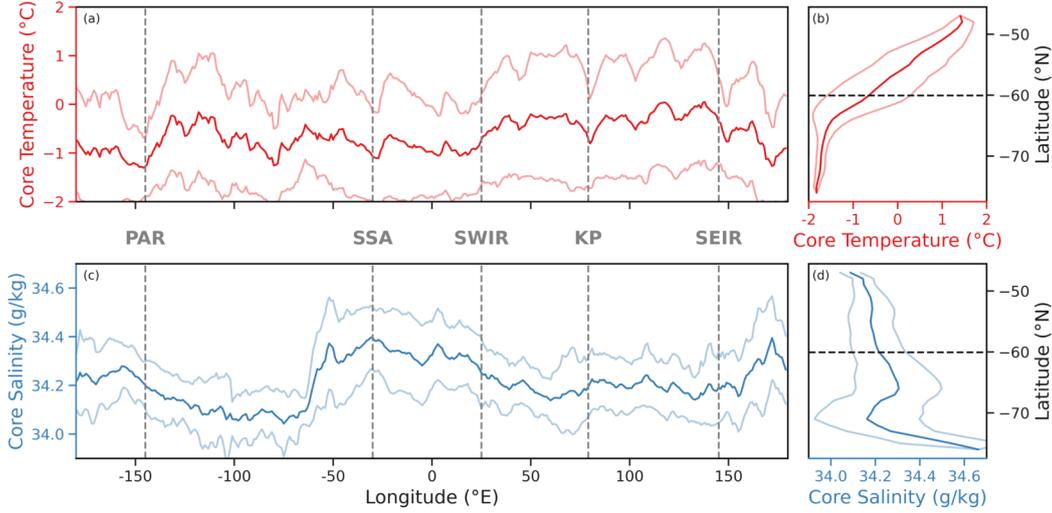


Figure 5. Meridional and zonal means of Antarctic Winter Water core properties.

Meridional and standard deviation of (a) core temperature and (c) core salinity. Zonal mean and standard deviation of (b) core temperature and (d) core salinity. Dashed light gray vertical lines in (a) & (c) indicate locations of topographic features and are labeled as in Figure 2. Dashed black horizontal lines in (b) & (d) indicate the location of the mean 15% sea ice concentration.

394 heat loss cools the entire SO forming cold WW_{ML} below sea ice (Figure 3a). WW dis-
 395 plays the largest seasonal temperature anomaly in winter ($-0.5\text{ }^{\circ}\text{C}$ anomaly) close to the
 396 sea ice edge, where there is the greatest change in WW core temperature throughout the
 397 annual cycle. Further, large wintertime temperature anomalies are observed in the Wed-
 398 dell Sea and Indian Ocean sector. The largest anomalies align with or are downstream
 399 of topographic features. These regions have elevated mixing rates (Dove et al., 2022; Mashayek
 400 et al., 2017; Mohrmann et al., 2022) so have warmer summertime WW temperatures.
 401 This thereby skews the annual mean temperature, which results in a larger magnitude
 402 anomaly signal in wintertime. Therefore, the large magnitude anomaly regions in win-
 403 tertime may be representative of regions with elevated mixing rates. The ABS exhibits
 404 a smaller magnitude of temperature anomaly ($-0.3\text{ }^{\circ}\text{C}$ anomaly) in winter. Similar low
 405 magnitude temperature anomalies are observed (but positive) in the ABS during the warm-
 406 ing season in summer (Figure 4d).

407 WW core temperature is at its coldest mean seasonal state in spring at -0.7 ± 1.2
 408 $^{\circ}\text{C}$ (Figure 4c). The sea ice begins to recede as the net heat flux becomes positive be-
 409 tween September-October (Tamsitt et al., 2016), so the SO transitions to a WW_{SS} -dominated
 410 ocean (Figure 3b and 3c). Localized warming of WW cores initiates downstream of the
 411 Drake Passage (DP), as shown by the lower magnitude temperature anomaly (Figures
 412 4c and S2c in Supporting Information). This region is highly energetic, resulting in a higher
 413 rate of mixing and, thus, a faster onset of warming due to mixing with subsurface CDW
 414 and the warm surface ML (Stephenson Jr. et al., 2012). Large-scale warming of the WW
 415 core follows in summer (Figure 4d and S2d), initiated in the Atlantic and Indian Ocean
 416 sectors downstream of topographic features: DP, SSA, KP, and PAR. In the regions of
 417 deeper bathymetry and fewer topographic features (such as west of KP and in the Pa-
 418 cific abyssal plain), mixing rates are lower and therefore warming of WW onsets in au-
 419 tumn, later in the seasonal cycle (Figure 4e) (Frants et al., 2013; Rosso et al., 2015; Wa-
 420 terhouse et al., 2014). WW across the SO in autumn is at its warmest mean core tem-

421 perature at -0.5 ± 1.2 °C due to the erosion of WW_{SS} in the open-ocean zone (Figure
422 4e).

423 Note that there is a localized patch in the eastern Weddell Sea that has a warm
424 mean core temperature (59°S, 7°E, Figures 4b-4e) so appears warmer throughout the win-
425 ter and spring (Figures 4b and 4c). This patch aligns with a region of thin sea ice in the
426 Weddell Gyre associated with southward CDW penetration (Holland et al., 2014; Ver-
427 net et al., 2019).

428 WW core salinity (Figures 4g-4j) has a clear seasonal cycle: WW cores are fresher
429 in winter with a seasonal mean of 34.22 ± 0.16 g/kg. In winter, the WW is largely clas-
430 sified as residing in the ML (WW_{ML}), which is observed as fresher than the mean core
431 salinity since the ML is largely composed of the fresh remnant summertime ML. The south-
432 ern section of the winter and spring under-ice zone has a large negative anomaly signal
433 due to elevated sea ice formation driving brine rejection and salinification of WW_{ML}. Salin-
434 ity increases from the Antarctic continent northward in springtime (Figures 4h and S2h
435 in Supporting Information) to the most saline period of the annual cycle with a seasonal
436 mean WW core salinity of 34.25 ± 0.16 g/kg. In summer, the circumpolar SO WW core
437 salinity has a positive anomaly. This is driven by WW_{SS} mixing with the underlying salty
438 CDW, which necessarily densifies WW through cabbelling via salinification (Figure 4i)
439 (Evans et al., 2018; Groeskamp et al., 2016; Jones et al., 2023), as well as a component
440 of salinity redistribution via advection as shown by (Pellichero et al., 2018). Regions close
441 to topographic features in summer have a greater magnitude in core salinity anomaly,
442 for example across the KP and along the PAR, because mixing rates are elevated and
443 thereby increase salinification through mixing with CDW. Further, the Weddell Gyre east-
444 ward limb has an elevated summertime core salinity anomaly due to mixing with south-
445 ward intruding CDW (Vernet et al., 2019; Jones et al., 2023). Around the Antarctic coast-
446 line, sea ice melt continues into summer leading to the freshening of WW_{ML} (anomaly
447 of approximately -0.2 g/kg). These localized coastal regions return to a positive anomaly
448 in autumn (Figures 4j and S2j) from an increase in salinity via sea ice formation, driv-
449 ing the formation of WW_{ML} in the under-ice zone. The autumn under-ice zone other-
450 wise experiences a mass freshening, which is indicative of the change from the salty and
451 capped WW_{SS} to the fresh remnant summertime ML becoming WW_{ML}. The greatest
452 changes in core salinity anomaly in autumn are observed in the under-ice zone, along the
453 ABS to downstream of DP, as well as along the SEIR, where there is a large change from
454 salty WW_{SS} via mixing with CDW to fresher WW_{ML}. In the open-ocean zone, no sea
455 ice formation takes place and WW remains classified as WW_{SS}, so WW core salinity con-
456 tinues to increase.

457 *3.1.3 Antarctic Winter Water Core Depth and Thickness*

458 Over the circumpolar extent, WW has a mean core depth of 88 ± 42 m and a mean
459 thickness of 99 ± 30 m (Figures 6a, 6f, 7a and 7c). The mean WW core depth is shal-
460 lower under-ice, with a mean depth in the under-ice zone of 60 ± 21 m (Figure 7b). Sim-
461 ilarly, the WW layer is thinner in the under-ice zone with a mean thickness of 87 ± 27
462 m (Figure 7d). Coastal polynya formation near the Antarctic continent results in deep
463 WW cores and thick WW layers due to extreme heat loss and are also synonymous with
464 more persistent WW_{ML} conditions (Figures 3a-3d). Consequently, the standard devi-
465 ation magnitudes of WW core depth and thickness are large for the under-ice zone, which
466 is otherwise relatively spatially homogeneous (Figures 7b and 7d). Below sea ice sub-
467 mesoscale fluxes are heightened in the wintertime due to deeper mixed layers and stronger
468 lateral gradients (Biddle & Swart, 2020). Because submesoscale fluxes are associated with
469 enhanced vertical fluxes, which, in this region would transport warm and salty CDW wa-
470 ter into the WW_{ML}, enhanced submesoscale fluxes in winter may contribute to main-
471 taining a the observed consistent WW depth and thickness in the under-ice zone. WW
472 core depth and thickness in the open-ocean zone have a mean of 127 ± 35 m and $117 \pm$

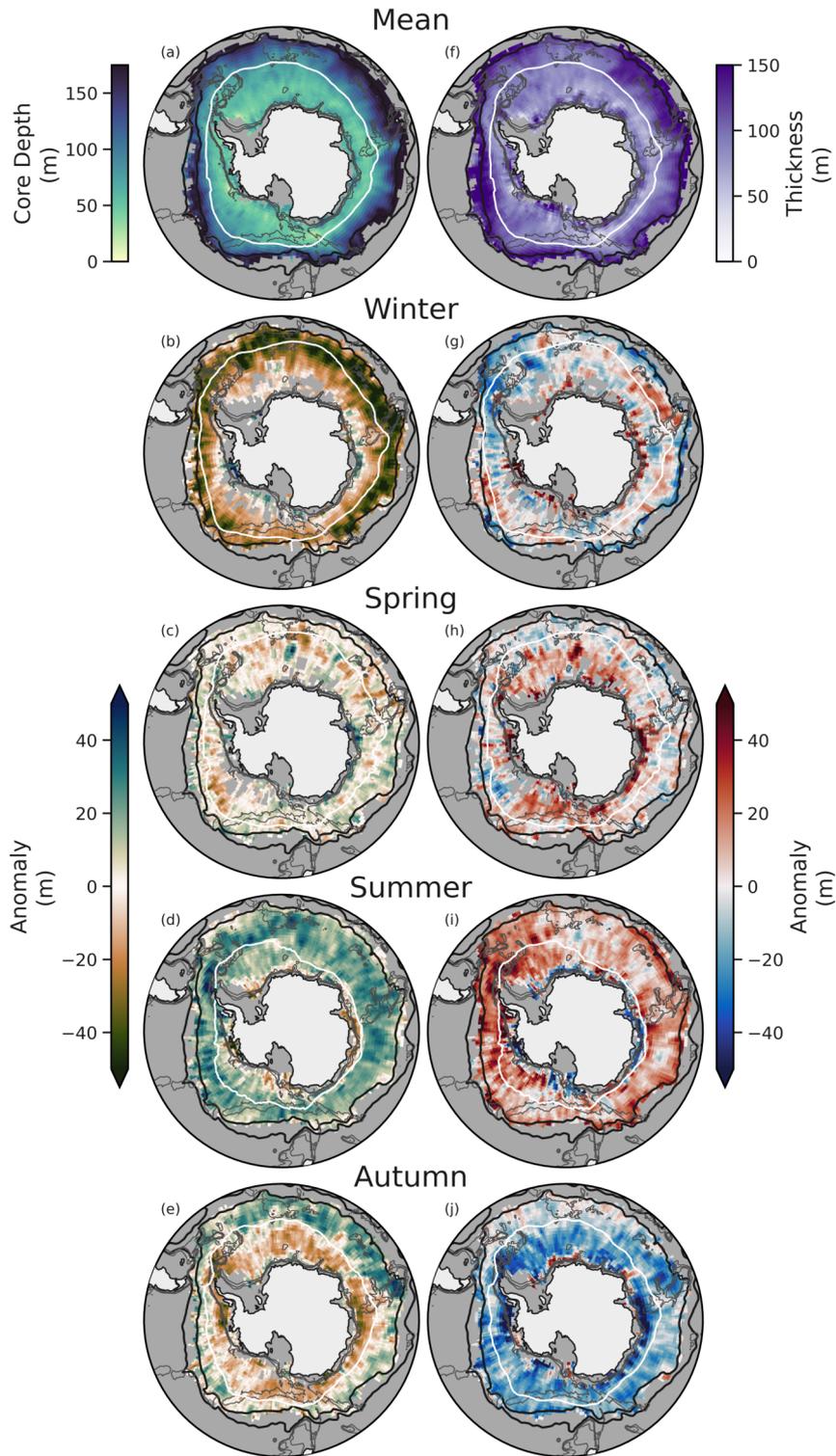


Figure 6. Annual mean and seasonal anomalies of Antarctic Winter Water: (a-e) core depth, and (f-j) thickness. (b-e) Core depth seasonal anomaly: positive (negative) indicates deeper (shallower) than the annual mean; (g-j) thickness anomaly: positive (negative) indicates thicker (thinner) than the annual mean. Black lines indicate the PF and SAF, white lines indicate the mean 15% sea ice concentration for the time period, and gray lines indicate 1km, 2km and 3km isobaths.

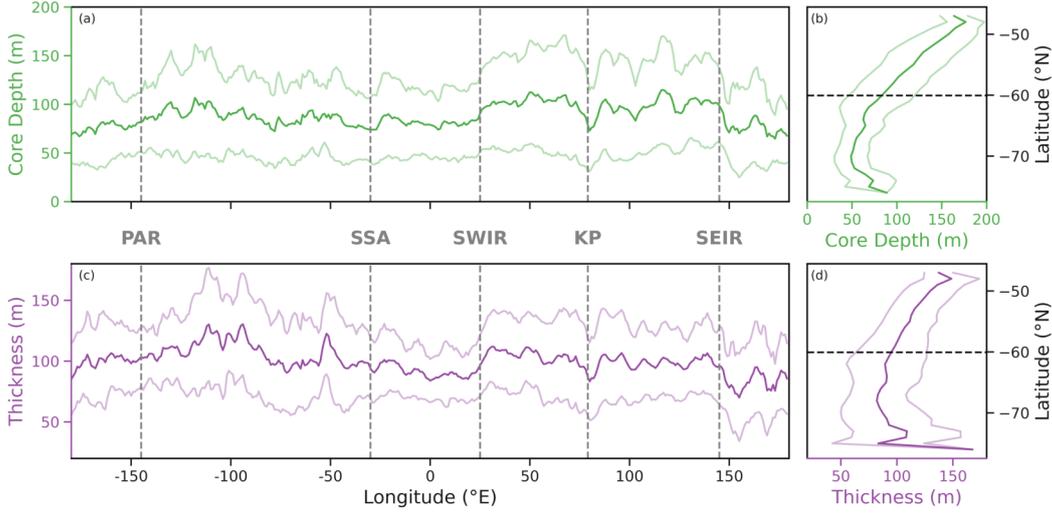


Figure 7. Meridional and zonal means of Antarctic Winter Water core properties. Meridional mean and standard deviation of (a) core depth (green) and (c) thickness (purple). Zonal mean and standard deviation of (b) core depth and (d) thickness. Dashed light gray vertical lines in (a) & (c) indicate locations of topographic features and are labeled as in Figure 2. Dashed black horizontal lines in (b) & (d) indicate the location of the mean 15% sea ice concentration.

473 25 m, respectively, and both have a northward latitudinal gradient in the open-ocean zone,
 474 with an average deepening rate of 6 m per degree of latitude northward and an average
 475 thickening rate of 3 m per degree of latitude northward. The deepest cores are found close
 476 to the PF (Figures 6a and 7b), whilst the thickest layer of WW are observed close to the
 477 Antarctic continent (Figures 6f and 7d). The average deepening and thickening of WW
 478 towards the north largely follows the deepening extent of CDW equatorward of the Antarc-
 479 tic shelf and can be associated with the reduced stratification and deep mixed layers as-
 480 sociated with the PF region of the ACC (du Plessis et al., 2022). Other dominating pat-
 481 terns are shallower and thinner WWs near open ocean topographic features while under
 482 sea ice. These characteristics can be observed, for example, near the KP, SWIR, as
 483 well as in the Atlantic sector near the SSA and SEIR (Figures 6a, 6f, 7a and 7c). Deep
 484 and shallow MLs are observed up- and downstream of topographic features, respectively
 485 (Figures S3a-S3e), which directly impact WW thickness and core depth.

486 One potential explanation for up- and downstream discrepancies of WW thickness
 487 is that regions downstream of topographic features are greater in EKE (Dove et al., 2022;
 488 Rosso et al., 2014), which results in a heightened submesoscale field due to mesoscale strain
 489 via frontogenesis (McWilliams, 2021). This increases vertical mixing rates that act to
 490 increase MLD variance in space and time and alter the depth and thickness character-
 491 istics of WW (Viglione et al., 2018) from the initial stages of WW_{ML} , particularly in the
 492 presence of topographic features or in regions of heightened EKE in the ACC (e.g. at
 493 KP in Figure 6f). Since summertime MLs are ubiquitously shallow across the SO (mean
 494 summertime MLD is 41 ± 18 m) (Caneill et al., 2023; Dong et al., 2008; Pellichero et
 495 al., 2017), submesoscale ML restratification takes particular effect in winter when MLs
 496 are deep (114 ± 38 m) leading to more ‘patchy’ spatial and temporal variability in MLD
 497 (Mahadevan & Campbell, 2002; du Plessis et al., 2017).

498 WW core depth (Figures 6b-6e) varies strongly seasonally (annual range of 88 m).
 499 In winter, the core depth is shallowest by definition because WW is largely classified as

500 WW_{ML} (Figure 3a), with a seasonal circumpolar mean of 71 ± 34 m and the largest win-
 501 tertime anomaly north of the sea ice edge where the core depth has shoaled (Figure 6b).
 502 Wintertime WW cores are only observed to deepen in the under-ice zone (Figure S2l),
 503 particularly close to the Antarctic shelf (e.g., Eastern Antarctica). Otherwise, winter-
 504 time core depth in the under-ice zone is relatively shallow and more spatially homoge-
 505 neous with a seasonal under-ice zone mean of 55 ± 18 m, whereas the open-ocean zone
 506 is nearly double that with a mean winter core depth of 109 ± 33 m. Spring is a tran-
 507 sition period from a WW_{ML}-dominated ocean to a WW_{SS}-dominated ocean (Figure 3b)
 508 as per our WW classifications (Table 1). As a result, the springtime SO is a heteroge-
 509 neous mix of WW_{ML} and WW_{SS}, particularly in the under-ice zone. Thus, the core depth
 510 deepens as WW_{SS} becomes capped by the newly formed ML (Figures 6c and S2m) to
 511 a circumpolar mean of 93 ± 43 m. WW core depth continues to deepen in summer to
 512 the deepest seasonal mean of 100 ± 43 m, with a greater magnitude anomaly downstream
 513 of the DP and through the Weddell Gyre. In autumn, convection drives the formation
 514 of WW_{ML} following sea ice formation (Figure 3d), mixing the remaining WW_{SS} into the
 515 mixed layer and shoaling the WW core depth in the under-ice zone (Figure 6e). This shoal-
 516 ing leads to the largest discrepancy between mean core depth in the under-ice zone and
 517 open-ocean zone, which are 52 ± 22 m and 133 ± 38 m, respectively.

518 WW thickness displays patchiness in its seasonal cycle, thickening and thinning at
 519 different rates over the year (Figures 6g-6j and S2q-t). WW thickness in winter (Figure
 520 6g) has a circumpolar mean of 100 ± 29 m, with a positive anomaly (thicker than the
 521 mean) close to the Antarctic continent where deep MLs form. WW is thicker close to
 522 the sea ice edge, such as in the Pacific and Indian sectors, which is driven by coastal polynya
 523 convective mixing processes. Furthermore, the low EKE upstream from the KP allows
 524 for the formation of thicker WW_{ML}. Downstream of the DP is a region of high EKE so
 525 WW_{ML} remains thin. In spring, the SO begins to transition to a WW_{SS} ocean and the
 526 mean circumpolar thickness increases to 108 ± 28 m.

527 Summer WW displays the thickest seasonal mean of 109 ± 33 m (Figure 6i). Thick-
 528 ening of WW_{SS} is indicative of a deeper subsurface maximum temperature gradient, i.e.
 529 lower boundary (Figure S3n in Supporting Information), due to CDW mixing with WW_{SS}
 530 which blurs the lower boundary. Close to the Antarctic continent, summertime WW thins
 531 due to elevated mixing along the Antarctic Slope Front through the WW layer, as shown
 532 by Hirano et al. (2010) in East Antarctica. Hirano et al. (2010) also computed an up-
 533 ward heat flux in East Antarctica that is more than double the upward heat fluxes ob-
 534 served in both the WS and western Peninsula shelf region (in ABS). Our results of an
 535 elevated magnitude of shallowing of the WW lower boundary depth along the coastline
 536 of the Indian sector (Figure S3n) agrees with these findings of increased heat fluxes. Fur-
 537 ther, WW is observed to reduce in thickness in the Ross Gyre in summer, which may
 538 be attributed to the seasonality in Ross gyre circulation, with a reduced summertime gyre
 539 strength driving increased heat flux from subsurface CDW (Dotto et al., 2018). A greater
 540 magnitude anomaly of summertime WW thickening is observed in the ABS and WS. These
 541 appear to thicken due deepening of the lower boundary (Figure S3n), which is a signal
 542 of WW mixing with underlying CDW resulting in the max temperature gradient (that
 543 is, the lower boundary; Table 1) observed at a deeper depth. This posit is supported by
 544 the core temperature increasing in summer (Figure 4d).

545 The seasonal mean WW thickness is observed at its thinnest in autumn with a mean
 546 of 81 ± 29 m (Figure 6j). Whilst WW close to the Antarctic continent is observed to
 547 thicken due to the formation of convective cells that contain WW_{ML}, WW otherwise thins
 548 in the under-ice zone due to the formation of WW_{ML}, mixing remaining WW_{SS} with the
 549 ML (Figure 3d). In the open-ocean zone, WW remains as WW_{SS} and erodes in thick-
 550 ness from continued mixing with underlying CDW as well as via deepening of the over-
 551 laying surface ML (Figure 3l, and Figures S3j and S3o).

3.1.4 Seasonal Heat Budget of Antarctic Winter Water

Given that temperature is the determining characteristic of the existence and classification of WW, we investigate a heat budget of WW_{SS} following Equation 1 (Section 2.3) to understand the mechanisms driving the distribution of WW properties and to further examine the potential link in the overturning circulation system. The heat budget describes the integrated heat change across the WW layer normalized by WW thickness from season to season, which we decompose into its constituent terms (Equation 1; Section 2.3). Note, in this section we refer to the transition from autumn to winter as winter (first row in Figure 8), which represents autumn subtracted from winter, and so on.

There are several physical processes that may impact the WW_{SS} heat budget (Figure 1), including penetrative shortwave radiation, vertical mixing, entrainment and detrainment, WW_{SS} formation (via summertime restratification of the ML, capping the subsurface remnant wintertime ML) as well as WW_{SS} erosion (via the onset of winter convection). WW_{SS} formation and erosion is the emergence and disappearance of WW_{SS} from one season to the next, and accounts for physical processes that may not be entirely captured by this dataset due its temporal coarseness. The former terms can be computed following Equation 1. The residual term consists of processes not accounted for by the other terms, such as diffusive convection, background mixing, and lateral advection (Section 2.3). We take advection to dominate the residual signal since diffusive and background mixing both make small contributions to heat fluxes (Bebieva & Speer, 2019; Giddy et al., 2023; van der Boog et al., 2021), whilst lateral advection is known to play a significant role in SO heat fluxes (Morrison et al., 2016).

The temperature tendency of WW_{SS} (Figure 8) and its constituent terms (Figure S4 in Supporting Information) show clear seasonality. WW_{SS} experiences an annual tendency to lose heat with an annual mean of $-0.02 \pm 1.67 \text{ }^\circ\text{C season}^{-1}$ across the circumpolar SO. The largest mean cooling is observed in winter with a mean temperature tendency of $-0.9 \pm 1.80 \text{ }^\circ\text{C season}^{-1}$ (Figure 8a), which is spatially heterogeneous. The greatest magnitude of wintertime cooling takes place downstream of the DP ($-1.89 \pm 1.12 \text{ }^\circ\text{C season}^{-1}$), which is driven by the seasonal cycle of heat flux forcing, as shown by Stephenson Jr. et al. (2012). Conversely, spring has the largest mean warming of WW_{SS}, warming across the season $0.82 \pm 2.17 \text{ }^\circ\text{C season}^{-1}$ and greater warming rates observed in regions in close proximity to the PF. Summer continues to warm WW_{SS} at a circumpolar mean rate of $0.54 \pm 0.74 \text{ }^\circ\text{C season}^{-1}$. du Plessis et al. (2022) observe WW warming by $\sim 1^\circ\text{C}$ over the three months December through February at 60°S , 0°E with the largest changes in temperature observed across December and January (their Figure 7f), which agrees with our findings of greatest warming in spring (October to December) and continued warming in summer (January to March). In autumn, WW_{SS} begins to cool again at a mean rate of $-0.73 \pm 1.39 \text{ }^\circ\text{C season}^{-1}$, cooling at greater magnitudes southeast of KP, along the SEIR and in the DP.

The formation and erosion of WW_{SS} is a leading order term of temperature tendency, accounting for $\sim 40\%$ of the annual temperature tendency budget. WW_{SS} formation represents the restratification of the ML in summertime to form a subsurface remnant wintertime ML. This addition of water and heat into the WW_{SS} system thus appears as a warming of WW_{SS} (red in Figures 8e-8h). Conversely, the erosion of WW_{SS} results in the removal of heat from WW_{SS} (blue in Figures 8e-8h), which can represent the erosion of WW_{SS} or re-entrainment into the wintertime ML. During wintertime, 37% of WW profiles change from WW_{SS} to WW_{ML} resulting in the removal of heat from the WW_{SS} system by a mean of $-1.16 \pm 0.79 \text{ }^\circ\text{C season}^{-1}$, accounting for 61% of the seasonal temperature tendency. In spring, the surface ocean begins to warm such that 41% of WW_{ML} profiles become (Figures 3 and S1b), resulting in a positive temperature tendency with a circumpolar mean of $0.87 \pm 2.77 \text{ }^\circ\text{C season}^{-1}$ and accounting for 54% of the season's temperature change. The minimum contribution (17%) from WW_{SS} ero-

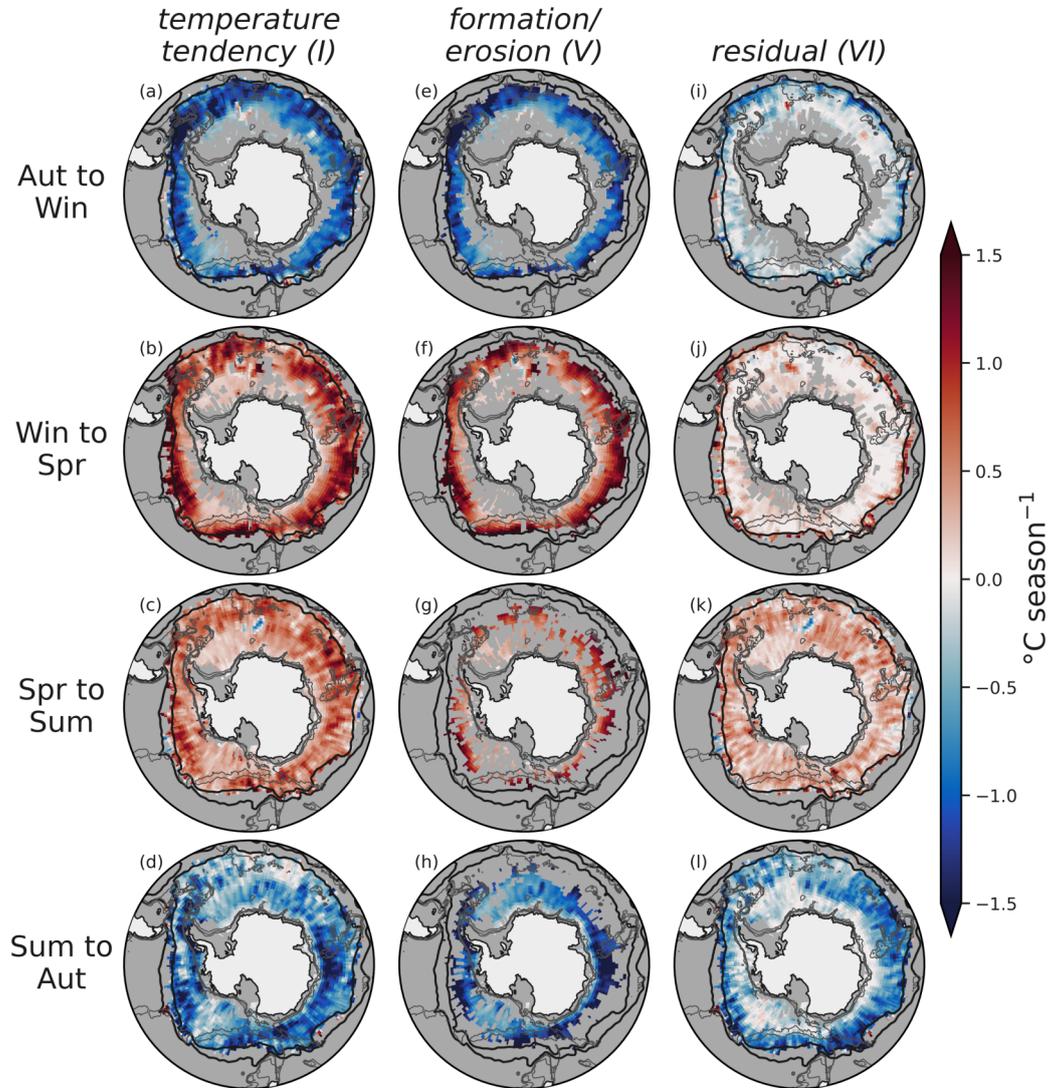


Figure 8. Seasonal heat budget of subsurface Antarctic Winter Water. (a-d) Temperature tendency where red (blue) indicates warming (cooling); (e-h) WW_{SS} formation (blue) and erosion (red) terms; (i-l) residual where red (blue) indicates warming (cooling) via advection. The labels on the top of each row indicate the term from Equation 1. Black lines indicate the PF and SAF, and gray lines indicate 1km, 2km and 3km isobaths. All terms from Equation 1 are plotted in Figure S4 in Supporting Information.

605 sion takes place in the summer season, warming by 0.54 ± 0.74 °C season⁻¹ with smaller
 606 magnitude temperature tendency close to the Antarctic continent, such as in the Wed-
 607 dell Sea and Indian sector. Autumntime erosion of WW_{SS} drives cooling of the WW_{SS}
 608 system, especially close to the Antarctic continent in the regions where WW_{ML} forms,
 609 for example southeast of KP (Figure 8h). Autumn has the largest removal of heat at -
 610 0.47 ± 1.12 °C season⁻¹ but only accounts for 37% of the temperature tendency.

611 The residual, which we assume is representative of advection and refer to it as such
 612 throughout this section, is of second order importance to the temperature tendency (Fig-
 613 ures 8i-8l). In autumn and winter, WW_{ML} and WW_{SS} co-exist such that deep winter-
 614 time MLs, which experience a net heat loss, reside alongside WW_{SS}. Consequently, cold
 615 WW_{ML} waters are advected into the WW_{SS} and therefore cool the WW_{SS}. Conversely,
 616 in spring and summer the upper ocean is no longer cooling, and so WW_{ML} is no longer
 617 forming. Therefore, advection drives warming of WW_{SS} during spring and summer. In
 618 winter, surface cooling initiates the formation of WW_{ML} and the subsequent equator-
 619 ward advection, which results in a cooling of circumpolar WW_{SS} by -0.33 ± 1.97 °C season⁻¹.
 620 This advection-driven cooling is largest close to the PF at the northern extent of WW
 621 where perennial WW_{SS} exists throughout the entire annual cycle (Figures S1). Spring
 622 WW_{SS} warms by 0.31 ± 0.92 °C season⁻¹, which largely takes place close to the PF. Sum-
 623 mer has the largest magnitude mean warming via advection at 0.37 ± 0.69 °C season⁻¹
 624 and is spatially homogeneous (Figure 8k). Advection in autumn drives the largest mag-
 625 nitude cooling with a seasonal mean of -0.47 ± 1.12 °C season⁻¹ (Figure 8l), but is low
 626 near the Antarctic continent due to erosion of WW_{SS} (as shown by the formation term;
 627 Figure 8h). Spatial heterogeneity is observed in the autumntime residual term, with large
 628 magnitude signatures near KP, SEIR and downstream of PAR.

629 The additional terms (II, III, and IV) of Equation 1 are presented in Figure S4 in
 630 Supporting Information. Overall, these terms explain the remainder of the seasonal tem-
 631 perature tendency. Wind-driven mixing consistently warms WW_{SS} across the circum-
 632 polar SO throughout the seasons in agreement with the findings of Giddy et al. (2023)
 633 (Figures S4e-S4h). Heightened wind-driven mixing is particularly evident across SEIR,
 634 where high winds are known to impart an enhanced wind-stress (Abernathy et al., 2011;
 635 Lin et al., 2018). The largest magnitude of WW_{SS} warming due to wind-driven mixing
 636 occurs in autumn where there are the increased temperature gradients across the ML-
 637 WW boundary. Entrainment/detrainment varies based on the balance between warm
 638 CDW entrained into WW from below, driving warming, and heat removed from the WW
 639 layer through entrainment into the ML, thus cooling WW_{SS}. Substantial heterogeneous
 640 cooling of WW_{SS} due to entrainment takes place in autumn in regions void of topographic
 641 features. This is coherent with deeper ML formation by ocean surface layer heat loss dur-
 642 ing autumn, particularly in the subantarctic ocean (Pellichero et al., 2017), driving en-
 643 trainment of WW into the ML and the loss of heat from WW_{SS}.

644 The heat budget (Figure 8) provides key information to understand the seasonal
 645 change of WW_{SS}. We show that the cooling of WW_{SS} takes place in autumn and win-
 646 ter whilst warming and mixing with the subsurface takes place in spring and summer,
 647 indicating the seasonal nature of heat fluxes in the upper SO and is potentially signif-
 648 icant in water mass transformation (Abernathy et al., 2016; Evans et al., 2018; Pellichero
 649 et al., 2018).

650 **3.2 Mechanisms Governing the Spatial Distribution of Antarctic Win-** 651 **ter Water**

652 Mean WW properties (Sections 3.1.2 and 3.1.3; Figures 4-7) exhibit broadscale char-
 653 acteristics that are relatively consistent throughout all properties across the SO: that is,
 654 quasi-homogeneous properties in the under-ice zone with meridional gradients in the open-
 655 ocean zone that, typically, increase in the northward direction. However, there exist re-

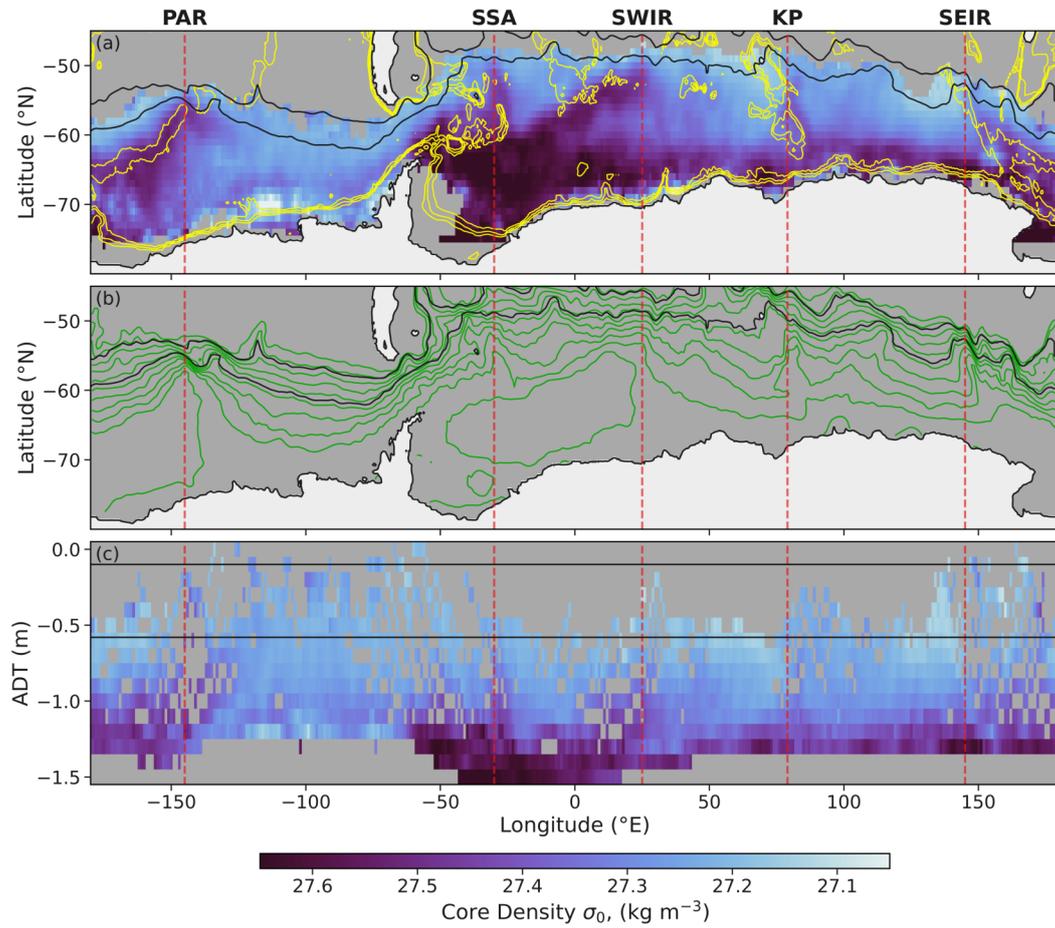


Figure 9. Antarctic Winter Water core density. (a) Annual mean WW core density. Yellow contours indicate 1km, 2km and 3km isobaths. (b) Absolute dynamic topography (ADT) streamlines are depicted in green. (c) Annual mean WW core density mapped onto ADT-Longitude coordinate space. (a-c) Black lines indicate the PF and SAF, and vertical dashed red lines show regions of topographic features, which are labeled as in Figure 2.

656 regions where properties associated with under-ice conditions appear to be transported north-
 657 ward, towards the PF. These regions of conserved under-ice properties in the open-ocean
 658 zone typically align with topographic features, and can be observed particularly clearly
 659 in the mean WW core density (Figure 9a) with dense WW cores in the open-ocean tightly
 660 constrained to the following topographic features: PAR, SSA, SWIR, KP, SEIR (west
 661 to east). In order to further investigate the spatial drivers of the large-scale properties
 662 of WW, we transform WW core density from latitude-longitude coordinate space to ab-
 663 solute dynamic topography (ADT)-longitude space (Figures 9c).

664 The distribution of WW is largely constrained to along-stream transport (referring
 665 to along-ADT streamlines). East of the PAR to the DP, in the open ocean ABS, WW
 666 is lighter in density (27.24 ± 0.10 kg m⁻³) and tightly constrained to ADT contours -
 667 a region where there is deep ocean bathymetry and few topographic features (the Pa-
 668 cific abyssal plane). Therefore, ocean interior mixing rates are lower than other parts of
 669 the SO (Frants et al., 2013; Ledwell et al., 2011), arresting densification of WW via re-
 670 duced mixing with underlying CDW. Furthermore, CDW intrusions onto the ABS con-
 671 tinental shelf results in a warm shelf basin (Narayanan et al., 2023; Tamsitt et al., 2021).
 672 Consequently, melt rates of sea ice are elevated in the region, as shown by the region-
 673 ally fresh WW cores (see Section 3.2.2; Figures 4f and 7c). In the SO, salinity is the do-
 674 minating contributor towards density (Roquet et al., 2022; Stewart & Haine, 2016). Thus,
 675 fresh WW cores lead to light WW core densities in the ABS.

676 Conversely, the Weddell Sea region is characterized by denser WW cores ($27.43 \pm$
 677 0.17 kg m⁻³, Figures 9) that advect northwards out of the Weddell Sea and across stream-
 678 lines of SSH, following the southward edge of the SSA. Locations of across-stream trans-
 679 port align with topographic features throughout the SO, with the strongest across-stream
 680 signal observed at the SWIR (Figures 9c). The locations of elevated WW density extend-
 681 ing equatorward reflect the pathways of northwards sea ice export (Haumann et al., 2016)
 682 and can potentially play a critical role in redistributing CDW and sea ice-driven brine
 683 properties. Further, WW core density reflects the northward advection of surface wa-
 684 ters from the Weddell Sea where the SSA is a region known for high rates of surface ad-
 685 vection, as is reflected by iceberg pathways (Budge & Long, 2018), and understood in
 686 observational studies via the use of the geostrophic thermal wind relation (Katsumata
 687 & Yoshinari, 2010; Oke et al., 2022).

688 Dense WW extending northward is observed close to the SWIR, which is likely a
 689 result of the Weddell Gyre northward circulation advecting WW northwards. Similarly,
 690 dense WW is transported northwards across the PAR, which is facilitated by the Ross
 691 Gyre; Zilberman et al. (2017) show similar findings of equatorward transport of ocean
 692 interior water masses such as Subantarctic Mode Water north of the ACC. Moreover,
 693 the regions of across-stream transport of WW coincide with the bathymetrically steered
 694 export pathways of Antarctic Bottom Water, as detailed by Solodoch et al. (2022) as well
 695 as others (Kusahara et al., 2017; Meredith, 2013; van Sebille et al., 2013). This also agrees
 696 with property transport preferentially facilitated by transient eddies, which are more prob-
 697 able to be present in regions of large bathymetric features and high EKE (Naveira Gara-
 698 bato et al., 2011; Thompson & Sallée, 2012; Wilson et al., 2022).

699 4 Assumptions and Limitations

700 4.1 Data Limitations

701 Hydrographic data in the Southern Ocean is prone to patchy spatiotemporal dis-
 702 tributions (Figures 2a-2d and 2i). Certain regions have low data coverage, such as the
 703 Pacific sector and within the seasonal ice zone. Many factors impact the distribution of
 704 data, for example, ship transects determine ship-based CTD cast time and location as
 705 well as float deployments. Also, floats being largely Lagrangian are steered and entrained

706 by the dominant ACC fronts (Figures 2e-2h), which leads to a somewhat reduced sam-
 707 pling of inter-frontal zones subject to reduced current strengths or shallower areas due
 708 to topographic steering (Kamenkovich et al., 2017). Nonetheless, profiling floats provide
 709 impressively broad coverage of the circumpolar SO (Figure 2j). Furthermore, MEOP is
 710 an important dataset that provides near full annual coverage in the southern domains
 711 of the SO and sea ice regions (Figures 2a-2h). However, seal behavior restricts broad and
 712 unbiased geographically coverage, resulting in observation hotspots (for example, the West-
 713 ern Antarctic Peninsula in ABS and KP). Further biases are introduced due to forag-
 714 ing preferences since seals favor frontal peripheries and regions that experience less in-
 715 tense surface warming (Siegelman et al., 2019; Azarian et al., 2024). The data included
 716 coincide with the Weddell Sea polynya years of 2016 and 2017 (Figures 2g and 2i), which
 717 results in a localized sampling bias of the warm water column over Maud Rise ($\sim 65^{\circ}\text{S}$,
 718 3°E) (Gülk et al., 2023; Mohrmann et al., 2022).

719 The ocean is also temporally variable, as noted by the differences in WW charac-
 720 teristics shown in the varying seasons. Certain regions, particularly the sea ice impacted
 721 ocean, have locations of limited profiles (e.g. less than 20) in the seasonal climatology
 722 of 18 years. An implication of this is that should any interannual variability exist in the
 723 WW characteristics, there is likely to be a bias towards the characteristics of the par-
 724 ticular period sampled. Furthermore, should there be a bias in a given month sampled,
 725 e.g., more profiles existed in February during the summer climatology for a given region,
 726 then there would likely be a bias towards thinner, warmer WW_{SS} , as indicated by the
 727 seasonal evolution shown in this study. Such aspects around spatiotemporal data cov-
 728 erage means that the characteristics of WW are not always consistently described in space
 729 and time, particularly in the permanent sea ice regions of the Weddell and Ross Seas and
 730 on parts of continental shelves. Nonetheless, to date, data coverage is at a status which
 731 allows robust descriptions and heat budget calculations over the vast majority of circum-
 732 polar SO and at a seasonal temporal resolution. The expected increase in observations
 733 and spatial coverage, particularly due to the onset or growth of various programmes, such
 734 as SOCCOM and MEOP (Figures 2i and 2j) will further narrow these observational gaps
 735 in the coming years to better resolve aspects around interannual variability or finer scale
 736 processes modulating WW.

737 4.2 Heat Budget Methodology

738 The heat budget work we have conducted provides information, for the first time,
 739 of the role WW_{SS} plays in redistributing heat in the upper SO and how this varies sea-
 740 sonally (Section 3.3). Below, we cover the assumptions made in the heat budget com-
 741 putation and methodology (Section 2.3). When calculating the down-ward turbulent heat
 742 flux (Term (III), Equation 1), we use the correlation found by Nicholson et al. (2022)
 743 between theoretical dissipation and observed summertime dissipation. Thus, we make
 744 the assumption that this correlation holds for the circumpolar SO across the annual cy-
 745 cle. The seasonal period is important because buoyancy plays a minor role in internal
 746 mixing in the summer period due to stratifying the water column whereas buoyancy de-
 747 stratifies the ML and drives convection in winter, resulting in both convection and shear
 748 driving the dissipation of turbulent kinetic energy (Sohail et al., 2018). Since we only
 749 apply our diffusivity parameterization to the open-ocean zone (north of the 15% sea ice
 750 concentration) and to WW_{SS} , the wintertime mixing terms contributes a small portion
 751 of the heat budget ($<10\%$ for winter; Figure S4). Nonetheless, we find our diffusivity re-
 752 sults to be comparable across the SO, which is of the $\text{O}(10^{-5})$ (Wu et al., 2011). Fur-
 753 thermore, localized results agree with observations up- and downstream of DP, where
 754 mixing regimes differ due to topographically enhanced mixing (Frants et al., 2013; Mer-
 755 rifield et al., 2016; Whalen et al., 2012). We further cross-referenced our heat budget find-
 756 ings with that of Giddy et al. (2023), whose mean January heat flux of 15 W m^{-2} in (60
 757 $^{\circ}\text{S}$, 0°E) from observations agrees with our estimates. Thus, whilst this methodology is
 758 based on assumptions with propagating uncertainties, it is a development from existing

759 methodologies (Pellichero et al., 2017), who use a constant diffusivity term, whereas we
 760 have used a spatiotemporally varying term that, to a first approximation, takes into ac-
 761 count spatiotemporal variation in wind forcing and ocean stratification.

762 5 Conclusion

763 This study investigates Antarctic Winter Water (WW) across the circumpolar SO
 764 using over 600,000 in-situ hydrographic observations across 18 years (Figure 2). We eval-
 765 uate the various physical mechanisms that impact the spatiotemporal distribution of WW,
 766 as conceptually portrayed in Figure 1. We have shown that the annual cycle of WW varies
 767 seasonally with sea ice growth such that WW_{ML} forms below sea ice during sea ice for-
 768 mation periods and is capped following ML restratification to form WW_{SS} as sea ice melts
 769 (Figure 3). Furthermore, we found that the properties of WW (core temperature, core
 770 salinity, core depth, and thickness) vary seasonally with distinct annual cycles (Figures
 771 4 and 6). However, annual means exhibit similar spatial distributions such that WW prop-
 772 erties are largely homogenous below sea ice (south of the 15% sea ice concentration ext-
 773 tent; Figures 4-6), with meridional gradients when north of sea ice. We find that WW
 774 properties associated with under-ice characteristics are transported equatorwards near
 775 large topographic features, such as the Pacific Antarctic Ridge, South Sandwich Trench,
 776 Southwest Indian Ridge, Kerguelen Plateau and Southeast Indian Ridge. We find that
 777 these regions show dense WW that flow across SSH streamlines (Figures 9), which is in-
 778 dicative of elevated velocities in the equatorward direction (through the thermal wind
 779 relation), transporting subpolar properties at different rates seasonally (Figure 8). Hence,
 780 equatorward redistribution of under-ice associated WW properties is largely driven by
 781 topographic steering, agreeing with findings of bathymetrically steered export pathways
 782 of Antarctic Bottom Water (Solodoch et al., 2022) to the global ocean.

783 This work provides a basis for future WW studies and raises many questions re-
 784 garding the role WW plays in the upper ocean and overturning circulation. For exam-
 785 ple, whether subpolar properties are transported to the global ocean through the ener-
 786 getic ACC via eddies and whether the properties are retained or mixed away. Further
 787 studies on CDW-WW interactions will help to gain an understanding of how ocean in-
 788 terior properties interact with the surface ocean. Critically, this work identifies poten-
 789 tial meridional pathways for the upper limb of the overturning circulation and provides
 790 additional evidence to support heterogeneous and seasonal localized overturning path-
 791 ways, developing from the zonal mean overturning framework.

792 Open Research

793 All computer code and the quality controlled, vertically gridded hydrographic pro-
 794 files are available in open-access repositories (Spira et al., 2023; Spira, 2024). The var-
 795 ious hydrographic data were collated from the following programmes: the International
 796 Argo Program provided the Argo floats hydrographic data ([https://portal.aodn.org](https://portal.aodn.org.au)
 797 [.au](https://portal.aodn.org.au)); the marine mammal hydrographic data were sourced from the International MEOP
 798 Consortium and the national programs that contribute to it (<http://meop.net>); the South-
 799 ern Ocean biogeochemical float hydrographic data were provided by SOCCOM ([https://](https://socom.princeton.edu/)
 800 socom.princeton.edu/); the hydrographic ship-based CTD casts and glider profiles
 801 were provided by NOAA WOD ([https://www.nodc.noaa.gov/OC5/SELECT/dbsearch/](https://www.nodc.noaa.gov/OC5/SELECT/dbsearch/dbsearch.html)
 802 [dbsearch.html](https://www.nodc.noaa.gov/OC5/SELECT/dbsearch/dbsearch.html)). ERA5 data are generated using Copernicus Climate Change Service
 803 Information, available online at [www.ecmwf.int/en/forecasts/datasets/archive-datasets/reanalysis-](http://www.ecmwf.int/en/forecasts/datasets/archive-datasets/reanalysis-datasets/era5)
 804 [datasets/era5](http://www.ecmwf.int/en/forecasts/datasets/archive-datasets/reanalysis-datasets/era5). The sea-ice data are made available via [http://data.meereisportal.de/](http://data.meereisportal.de/data/iup/hdf/s/)
 805 [data/iup/hdf/s/](http://data.meereisportal.de/data/iup/hdf/s/). The ADT data used to determine the fronts are generated by the Archiv-
 806 ing, Validation, and Interpretation of Satellite Oceanographic data service of Centre Na-
 807 tional D'Etudes Spatiales, available online www.avisio.altimetry.fr/en/data/data-access/

808 gridded-data-extraction-tool.html. Bathymetry data is available by GEBCO Com-
809 pilation Group (GEBCO, 2023).

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Supporting Information for “The observed spatiotemporal variability of Antarctic Winter Water”

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Contents of this file

1. Figures S1 to Sx

Introduction Supporting Information contains 4 additional figures that provide further details on the seasonal cycle of AAWW. Figure S1 shows the seasonal change of WW classification. Figure S2 shows an alternative approach from anomaly plots, as shown in Figures 4 and 6 in main text, to show seasonal temperature change (ie, $d/dt(\text{var})$), which provides further nuances to the seasonal cycle. We provide component-wise information to thickness variability by presenting the upper and lower bound depths as well as the MLD in S3. Lastly, we present all the component terms of the heat budget equation (Equation 1 in main) in S4.

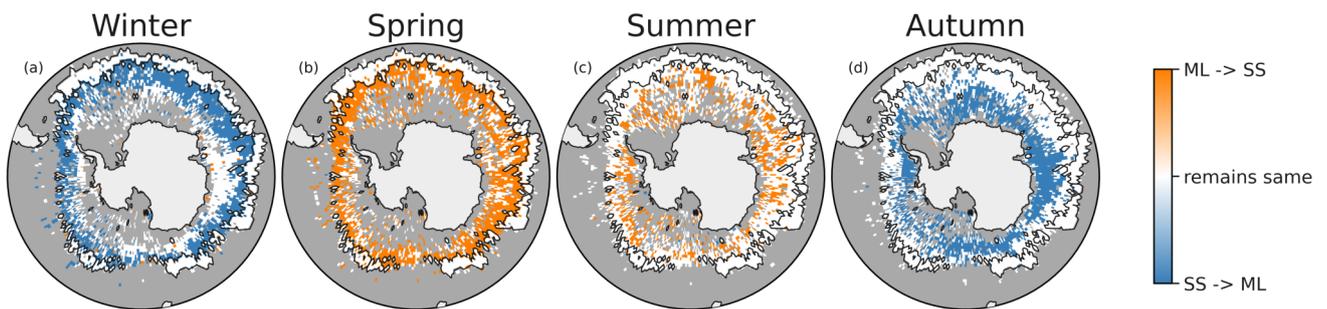


Figure S1. Seasonal change of WW classification. Orange indicates a change from WW_{ML} to WW_{SS} , white indicates the WW remains the same, blue indicates a change WW_{SS} to WW_{ML} , and black contours indicate WW that remains unchanged throughout the entire annual cycle. Black lines indicate the PF and SAF, white lines indicate the mean 15% sea ice concentration for the time period, and gray lines indicate 1km, 2km and 3km isobaths.



Figure S2. Annual mean and seasonal change of WW: (a-e) core temperature, (f-j) core salinity, (k-o) core depth, and (f-j) thickness. Positive (negative) indicates an increase (decrease) from the previous season. Black lines indicate the PF and SAF, white lines indicate the mean 15% sea ice concentration for the time period, and grey lines indicate 1km, 2km and 3km isobaths.

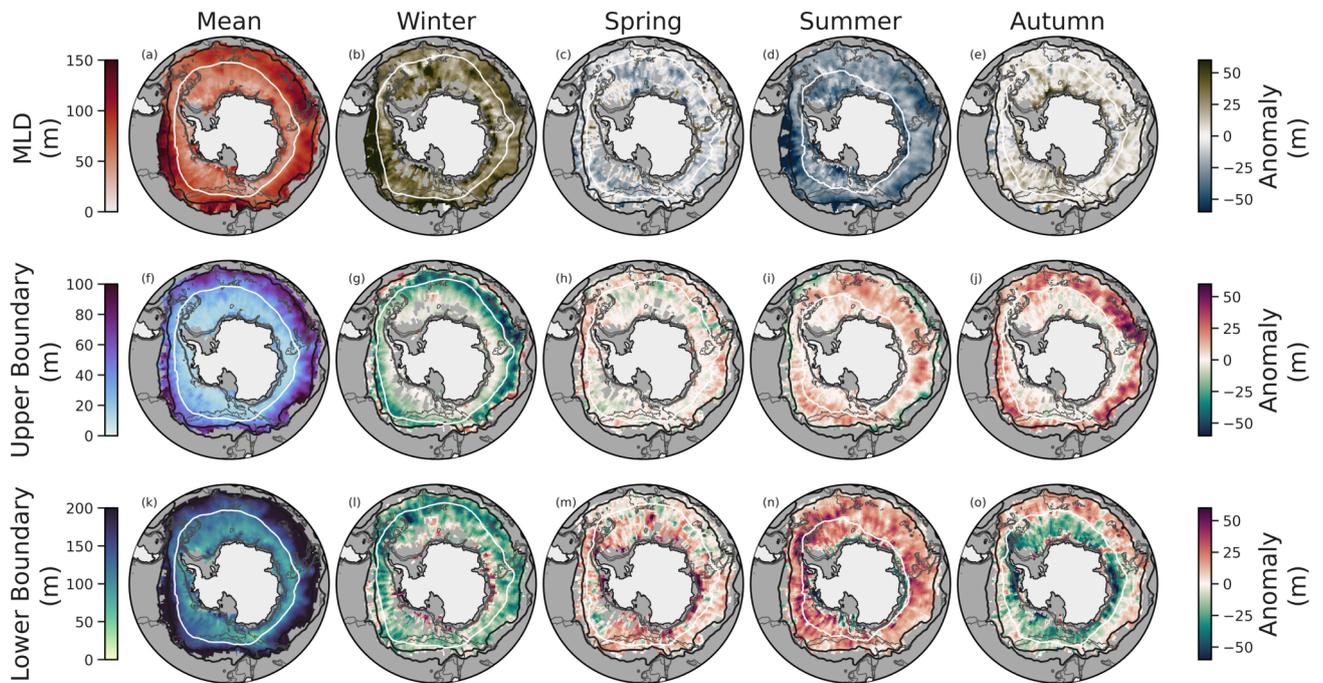


Figure S3. Annual mean and seasonal change of WW: (a-e) core temperature, (f-j) core salinity, (k-o) core depth, and (f-j) thickness. Positive (negative) indicates an increase (decrease) from the previous season. Black lines indicate the PF and SAF, white lines indicate the mean 15% sea ice concentration for the time period, and grey lines indicate 1km, 2km and 3km isobaths.

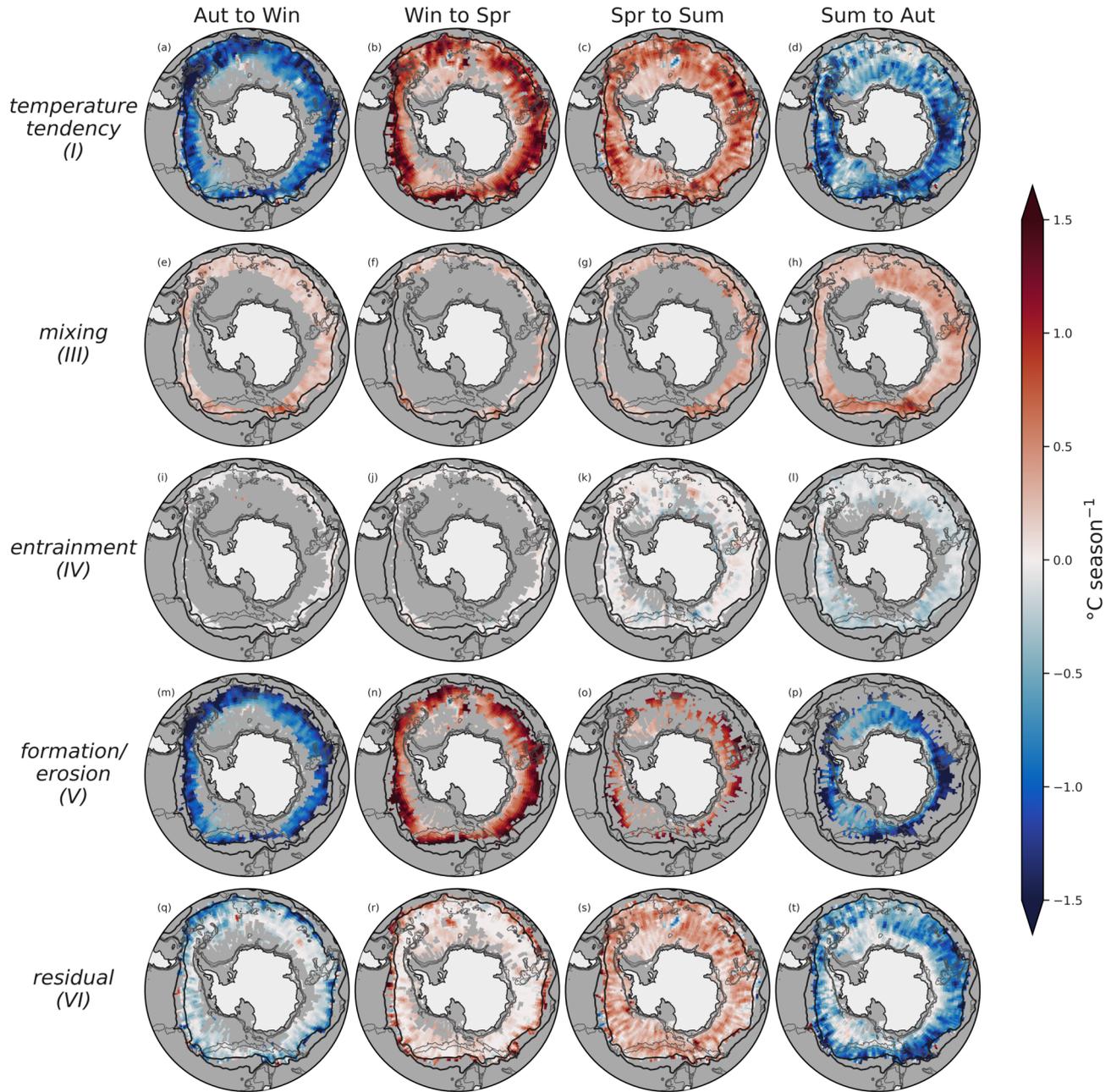


Figure S4. Full seasonal heat budget of WW_{SS}. (a-d) temperature tendency; (e-h) turbulent vertical mixing; (i-l) entrainment/detrainment; (m-p) WW subduction/erosion; (q-t) residual. The Roman numerals on the right hand side represent the term of the row from Equation 1. Black lines indicate the PF and SAF, and gray lines indicate 1km, 2km and 3km isobaths.