Spiciness Anomalies of Subantarctic Mode Water in the South Indian Ocean

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ABSTRACT: This study investigates spreading and generation of spiciness anomalies of the Subantarctic Mode Water (SAMW) located on 26.6 to 26.8 $\sigma_T$ in the south Indian Ocean, using in situ hydrographic observations, satellite measurements, reanalysis datasets, and numerical model output. The amplitude of spiciness anomalies is about 0.03 psu or 0.13°C and tends to be large along the streamline of the subtropical gyre, whose upstream end is the outcrop region south of Australia. The speed of spreading is comparable to that of the mean current, and it takes about a decade for a spiciness anomaly in the outcrop region to spread into the interior up to Madagascar. In the outcrop region, interannual variability in mixed layer temperature and salinity tends to be density compensating, which indicates that Eulerian temperature or salinity changes account for the generation of isopycnal spiciness anomalies. It is known that wintertime temperature and salinity in the surface mixed layer determine the temperature and salinity relationship of a subducted water mass. Considering this, the mixed layer heat budget in the outcrop region is estimated based on the concept of effective mixed layer depth, the result of which shows the primary contribution from horizontal advection. The contributions from Ekman and geostrophic currents are comparable. Ekman flow advection is caused by zonal wind stress anomalies and the resulting meridional Ekman current anomalies, as is pointed out by a previous study. Geostrophic velocity is decomposed into large-scale and mesoscale variability, both of which significantly contribute to horizontal advection.

KEYWORDS: Indian Ocean; Oceanic mixed layer; Sea surface temperature; Water masses/storage; Heat budgets/fluxes; Interannual variability

1. Introduction

Subantarctic Mode Water (SAMW) is a thick water mass generated in the Southern Ocean (McCartney 1977, 1982; Hanawa and Talley 2001), which plays a major role in global mass (Sloyan and Rintoul 2001), heat (Talley 2003), and material transport (Sabine et al. 2004; Ito et al. 2010). One of the generation regions of SAMW is south of Australia between the Subtropical and Subantarctic Fronts (Sloyan and Rintoul 2001; Cerovecki et al. 2013), where mixed layer temperature and salinity increase equatorward and tend to be density compensating, forming a region with almost uniform density between about 40° and 50°S (Figs. 1a,b). A low-potential-vorticity (PV) mode water is generated there in the wintertime thick mixed layer driven by surface cooling (Fig. 1c; McCartney 1977; Dong et al. 2008), subducts into the interior (Karsten and Quadfasel 2002; Sallée et al. 2010b), and is transported toward the tropics via the advection of the subtropical gyre (Fig. 2a; McCarthy and Talley 1999; Herraiz-Borrego and Rintoul 2011; Jones et al. 2016; Nagura and McPhaden 2018). SAMW in the south Indian Ocean is located on isopycnals between 26.6 and 26.9 $\sigma_T$ (Fig. 2b; Fine 1993; Cerovecki et al. 2013) sandwiched between high-PV waters of the south Indian subtropical water in the main pycnocline (Rochford 1964; Talley and Baringer 1997; Wijffels et al. 2002) and of the Antarctic Intermediate Water (AAIW) between 26.9 and 27.2 $\sigma_T$ (Fine 1993; Toole and Warren 1993; McCarthy and Talley 1999).

It is known that the temperature and salinity ($T/S$) relationship of SAMW varies in time, which is related to ocean heat uptake. Variation in the $T/S$ relationship of a water mass is referred to as spiciness anomaly, which can be measured by salinity variability on isopycnals and considered to be a dynamically passive tracer (Veronis 1972; Munk 1981; detailed discussions are presented in section 5d). Wong et al. (1999), Bindoff and McDougall (2000), and Bryden et al. (2003) examined repeated ship observations along 32°S and found spiciness anomalies of SAMW. Wong et al. and Bindoff and McDougall reported freshening and cooling of SAMW from 1962 to 1987, and Bryden et al. detected the increase of salinity from 1987 to 2002. Kołodziejczyk et al. (2019) also reported isopycnal temperature anomalies in the south Indian Ocean in the density range of SAMW using Argo float observations. For a 50-yr period, subsurface temperature tends to warm almost globally due to thermocline heave (Roemmich et al. 2015; Häkkinen et al. 2016). Spiciness anomalies in the density range of SAMW decrease in the southern Indian Ocean, whose cooling tendency counteracts the warming trend owing to thermocline heave (Durack and Wijffels 2010; Häkkinen et al. 2016). The climatic importance of spiciness anomalies of SAMW is further emphasized by Banks et al. (2000), who reported based on numerical model experiments that spiciness anomalies of SAMW are sensitive to anthropogenic carbon dioxide emissions. Previous studies also reported variability in thickness and volume of SAMW (Gao et al. 2018; Kołodziejczyk et al. 2019; Hong et al. 2020; Portela et al. 2020), whereas we focus on spiciness variations in this study.

Spiciness anomalies were observed globally in midlatitude regions of the Pacific (e.g., Deser et al. 1996; Li et al. 2012; Kołodziejczyk and Gaillard 2012), Indian (Li and Wang 2015), and Atlantic Oceans (Qu et al. 2016), and their generation mechanisms have been discussed. One possible mechanism is the excursion of a $T/S$ front (Schneider 2000; Li and Wang 2015). When two currents are confluent in the interior of the ocean, different water masses are located next to each other, which leads to a formation of a sharp $T/S$ front on an isopycnal surface. Spiciness variations can be generated if variability in...
currents leads to the excursion of the front. Another possible cause is diapycnal forcing near the sea surface. In most of the regions, spiciness anomalies are generated in the formation region of mode water, where temperature and salinity near the surface tend to be density compensating in the meridional and/or vertical direction. In such a region, a water of higher (or lower) spiciness can be replaced with lower (or higher) spiciness virtually without changing density, giving rise to a spiciness anomaly in a subducted water mass. For example, interannual variability in surface heat flux leads to the meridional migration of the outcrop line of a density surface, which generates a spiciness anomaly when temperature and salinity at the sea surface tend to be density compensating in the meridional direction (Nonaka and Sasaki 2007; Laurian et al. 2009). In the region where temperature and salinity tend to be density compensating in the vertical direction, enhancement of vertical mixing near the surface injects saline water to subsurface and generates a positive spiciness anomaly, which is referred to as “spiciness injection” (Yeager and Large 2004, 2007; Kolodziejczyk and Gaillard 2012, 2013; Kolodziejczyk et al. 2015; Wang and Luo 2020). Interannual variability in surface heat and freshwater flux can generate a spiciness anomaly even if surface temperature and salinity are not density compensating (Nagura and Kouketsu 2018).

On the other hand, previous studies emphasized the importance of horizontal advection near the surface in the generation of spiciness anomalies in the formation region of SAMW. Mixed layer temperature and salinity tend to be density compensating in this region (Figs. 1a,b), and westerly winds drive equatorward Ekman currents. Rintoul and England (2002) proposed that variability in meridional Ekman transport in this region alters the $T/S$ relationship of subducted water mass virtually without changing density, which results in spiciness anomalies. The Subantarctic Front is located just south of the formation region of SAMW (Nagata et al. 1988; Orsi et al. 1995; Belkin and Gordon 1996), where the eastward Antarctic Circumpolar Current (ACC) is intense (Nowlin and Klinck 1986; Stramma 1992), and eddies are vigorous (Fig. 1d; Morrow et al. 2003; Frenger et al. 2015). Sallée et al. (2008a) and Herraiz-Borreguero and Rintoul (2010) pointed out that eddy advection and mixing can cause variability in the $T/S$ relationship of the surface and subducted water mass. In addition, Wong et al. (1999) and Banks and Bindoff (2003) claimed that spiciness anomalies of intermediate water masses are attributable to variability in surface air–sea fluxes of heat and freshwater. These previous studies focused only on an aspect of the mixed layer $T/S$ budget, and a comprehensive view has not been provided. Wong et al. (1999) and Banks and Bindoff (2003) focused on surface air–sea fluxes and did not compare their effect with other terms, such as horizontal advection. Rintoul and England (2002) discussed the mixed layer heat budget using the typical values for mixed layer depth (MLD), horizontal gradient of temperature, and ocean currents because observational data were not available to them. Sallée et al. (2008a) focused on temperature variability tracked by one Argo float and did not relate results to the $T/S$ relationship of a subducted water mass. Herraiz-Borreguero and Rintoul (2010) investigated eddy advection using observations along a ship section south of Tasmania. It is uncertain if their hypothesis is valid for variability in a larger domain. Also,

FIG. 1. Climatological mixed layer (a) temperature, (b) salinity, and (c) depth in September. Mixed layer density in September is shown by contours. (d) Standard deviations (colors) and the mean (contours) of absolute satellite sea surface height. Contour intervals are 0.2 kg m$^{-3}$ for (a)–(c) and 15 cm for (d). Boxes are the analysis region for the mixed layer heat budget (38°–48°S, 100°–150°E; see section 4). Mixed layer temperature, salinity, depth, and density were obtained from in situ hydrographic observations provided by the World Ocean Database (Boyer et al. 2018; Garcia et al. 2018) and objective mapping. Satellite sea surface height was provided by the Copernicus Marine and Environment Monitoring Service (Ducet et al. 2000; Rio and Hernandez 2004).
Fig. 2. (a) The mean planetary potential vorticity (PV; colors), streamlines (contour lines), and geostrophic velocity (vectors) on 26.85 \( \sigma_T \). Hatching shows the outcrop region, where the maximum of mixed layer density exceeds 26.85 \( \sigma_T \). (b) The mean planetary PV (colors), potential density (thin solid contours), and the maximum of mixed layer depth (thick solid line) along 100°E. PV is computed as \( [f(\rho, \beta, z)]^{-1} \); following McCarthy and Talley (1999). PV, potential density, and mixed layer depth were estimated from in situ hydrographic observations obtained from the World Ocean Database (Boyer et al. 2018; Garcia et al. 2018). Streamlines and geostrophic velocity were obtained from Nagura and McPhaden (2018).


2. Data

Spiciness anomalies were estimated using in situ hydrographic observations provided by the World Ocean Database (WOD; Boyer et al. 2018; Garcia et al. 2018). We used 7/5 profiles labeled as conductivity–temperature–depth (CTD) or profiling float (PFL) data in the region from 60°S to the equator and from 40° to 160°E for the period from 2004 to 2018. We first checked the quality flag of observations and discarded any that were not “accepted value” or “accepted cast.” Data provided by WOD underwent various tests including statistical tests, by which data that deviated from the mean by more than 4 standard deviations in a 5° × 5° box were flagged (Garcia et al. 2018). Data that did not pass this test were removed by our flag check. Still, we found apparently erroneous data in the remaining profiles. To remove them, we further conducted a statistical check using the mean and standard deviations on a 5° × 5° grid obtained from the World Ocean Atlas 2013 (Locarnini et al. 2013; Zweng et al. 2013) and eliminated data if they deviated from the mean by more than 3 standard deviations. In this study we focus on large-scale variability, and the use of the mean and standard deviation fields mapped on a 5° × 5° grid matches our purpose. The flag and standard deviation checks removed 18% and 0.7% of the total data, respectively. We used 246 431 profiles in this study that passed the checks. About 97% of observations were from PFL data, and almost all of PFL data were obtained from Argo float observations (Boyer et al. 2018). Note that WOD provides data that passed delayed-mode quality controls (Boyer et al. 2018), in which adjustments for various measurement errors, including salinity drifts, were conducted (Wong et al. 2020). In situ observations cover the basin, although observations are relatively sparse in the Indonesian Seas, coastal regions near Australia, west of Madagascar Island, and the Antarctic region south of about 50°S (Fig. 3a). The number of observations increased since the implementation of Argo float program, and observations are abundant in our analysis period (2004–18; Fig. 3b).

We also used the mean absolute velocity data provided by Nagura and McPhaden (2018). In this dataset, mean velocity...
on isopycnal levels was computed from in situ hydrographic observations provided by WOD and Argo float trajectories at the parking depth (nominally 1000 m depth) provided by Lebedev et al. (2007). Hydrographic observations were objectively mapped on isopycnal levels, and vertical shear of geostrophic velocity was computed at the reference depth of 1000 m. Absolute velocity at 1000 m depth was estimated from Argo float trajectories using the method of Katsumata and Yoshinari (2010) and added to geostrophic velocity shear, resulting in absolute velocity on each isopycnal level.

We also used satellite measurements. Daily averages of sea surface temperature (SST) were obtained from the Remote Sensing Systems (RSS; http://www.remss.com/measurements/sea-surface-temperature/oisst-description/). This dataset uses SST measured by satellite microwave sensors and is constructed by optimal mapping onto a 0.25° × 0.25° grid. We chose microwave SST rather than infrared SST, because the former has better sampling than the latter in the southern subpolar region (Chelton and Wentz 2005). Sea surface height (SSH) was obtained from the Copernicus Marine and Environment Monitoring Service (CMEMS; Ducet et al. 2000; Rio and Hernandez 2004; CMEMS 2020). We also used surface geostrophic velocity data provided by CMEMS computed from satellite altimetry. SSH and geostrophic velocity data are on a 0.25° × 0.25° grid and daily averages.

We obtained surface wind stress, sea level pressure (SLP), and shortwave and longwave radiation data from the ECMWF interim reanalysis (ERA-Interim; Dee et al. 2011). These data are daily averages on a 0.75° × 0.75° grid. Turbulent heat fluxes were obtained from Objectively Analyzed Air–Sea Heat Fluxes (OAFlux; Yu and Weller 2007), which are daily averages mapped on a 1° × 1° grid. Renfrew et al. (2002) compared surface turbulent heat flux obtained from shipboard in situ observations with that from reanalysis products and pointed out that reanalysis turbulent heat flux is erroneous due to the inappropriate formulation of the bulk flux algorithm. They recommended to recalculate turbulent heat flux using meteorological fields obtained from a reanalysis dataset and an appropriate bulk algorithm. Dong et al. (2007) adopted this approach and examined the mixed layer heat budget in the Southern Ocean on the seasonal time scale. Turbulent heat flux of OAFlux is computed from meteorological data obtained from atmospheric reanalyses and satellite measurements using the bulk flux algorithm version 3.0 developed from the Coupled Ocean–Atmosphere Response Experiment (COARE3.0; Fairall et al. 2003), and we use this product following Dong et al. (2007). However, it is highly likely that the estimate of surface heat flux includes large uncertainty in the Southern Ocean, where observations are sparse (Bourassa et al. 2013). We discuss the accuracy of surface heat flux in section 5c, using in situ observations obtained from a moored buoy.

We examine the mixed layer heat budget using model output in section 5a. We used output from the OGCM for the Earth Simulator version 2 (OFES2; Sasaki et al. 2020). OFES2 is based on the Modular Ocean Model version 3 (Pacanowski and Griffies 1999). The model domain is from 76°S to 76°N. The bottom topography was obtained from ETOP01 (Amante and Eakins 2009). The horizontal grid intervals are 0.1°, and the number of vertical levels is 105. Vertical mixing near the surface was parameterized by the scheme of Noh and Kim (1999). Surface heat flux was computed from meteorological data obtained from Japanese 55-year atmospheric reanalysis (Tsujino et al. 2018), simulated SST and the bulk formulas of Large and Yeager (2004). No data assimilation was applied. We used daily averages of output for the period from 2004 to 2016.

We also used output from the Estimating the Circulation and Climate of the Ocean version 4 (ECCO v4; Forget et al. 2015). ECCO v4 is based on the Massachusetts Institute of Technology general circulation model (Marshall et al. 1997; Adcroft et al. 2004) and assimilates various in situ and satellite observations by an adjoint method without introducing artificial sources of heat and buoyancy. The model domain is global. The horizontal grid intervals range from 0.5° to 1° in the tropical and midlatitude regions. The number of vertical levels is 50. We used output on the native grid. The turbulent closure scheme of Gaspar et al. (1990) and a simple convective adjustment scheme were adopted. Surface heat flux was computed meteorological fields of ERA-Interim and the bulk formula of Large and Yeager (2004). State estimation accounts for uncertainties of reanalysis meteorological fields. The ECCO v4 dataset provides monthly snapshots of temperature and salinity at the first day of each month and monthly averages of temperature, salinity, and advective and diffusive heat flux for the period from 1992 to 2017.

3. Methods

a. Gridding of in situ hydrographic observations

The T/S profiles obtained from in situ observations were interpolated onto 33 potential density levels from 21.5 to 27.74 σθ. Potential temperature and potential density were computed.
using the equation of the state of the International Thermodynamic Equation of Seawater 2010 (McDougall and Barker 2011). The resulting $T/S$ data were objectively mapped on a $1^\circ \times 1^\circ$ grid on each density surface, assuming a Gaussian covariance function with decorrelation scales of $6^\circ$ in longitude and $3^\circ$ in latitude. This mapping was done for each calendar year, which provides yearly estimates of $T/S$ on isopycnals.

Using $T/S$ profiles obtained from in situ observations, MLD was computed as the depth where density is larger than the value at 10 m depth by 0.03 kg m$^{-3}$, following de Boyer Montégut et al. (2004). Averages of temperature, salinity, and density from the sea surface to the base of the mixed layer were also computed using obtained MLD. These variables were objectively mapped onto a $1^\circ \times 1^\circ$ grid in the following two manners. First, MLD, mixed layer temperature, mixed layer salinity, and mixed layer density were mapped on a monthly basis from 2004 to 2018, results of which include interannual variability. Second, MLD was mapped on a monthly basis ignoring years, the results of which provided monthly climatologies. We refer to the first estimate as the “monthly estimate” and to the second as the “monthly climatological estimate.” For both, a Gaussian covariance function was assumed in objective mapping with decorrelation scales of $6^\circ$ in longitude, $3^\circ$ in latitude, and 45 days in time. The reason why we conducted the second estimate is sparseness of observations. In monthly estimates, data points are sometimes sparse, which leads to an erroneous estimate, such as negative MLD in summer. Such an erroneous estimate does not occur in monthly climatological estimate of MLD. For the monthly estimate of MLD, we replaced values less than 10 m with a constant value of 10 m.

Error for gridded fields was computed as standard deviation. First, nondimensional error for objective mapping was computed from assumed covariance function and the distance between data points and the grid point, following Bretherton et al. (1976). Nondimensional error ranges from 0 to unity depending on data density. Then, nondimensional error was converted to dimensional by multiplying standard deviations of data values around each grid point.

After salinity drifts are adjusted in delayed mode quality check, error for salinity of each profile is about 0.01 psu or less (Wong et al. 2003, 2020). It is considered that this is a random measurement error and reduced by gridding (e.g., Kouketsu et al. 2017). Typically, about 50 profiles are located within $6^\circ \times 3^\circ$ decorrelation scales around each grid point in a year. The resulting measurement error for the gridded salinity field can be roughly estimated using the propagation of error formula (Emery and Thomson 2004) as $\sqrt{50 \times (0.01 \text{ psu})^2/50} \approx 0.001 \text{ psu}$. This is smaller than the typical error obtained from standard deviations (about 0.004 psu). For monthly climatological estimates (or monthly estimates), about 70 (4) profiles are located within the range of the decorrelation scale around each grid, and the estimated measurement error for gridded mixed layer salinity is 0.001 (0.005) psu, which is smaller than standard deviation error, 0.01 (0.16) psu. We believe that measurement error for salinity is not crucial in this study.

In this study, isopycnal salinity and mixed layer variables were first computed from each $T/S$ profile and then objectively mapped. This is to avoid artificial variability in the $T/S$ characteristic of a water mass, which is caused by isobaric averaging (Lozier et al. 1994), and to avoid errors for mixed layer variables, which occur when MLD is computed from spatially and temporally smoothed gridded fields (de Boyer Montégut et al. 2004; Toyoda et al. 2017). Widely distributed datasets that are based on Argo float observations, such as Roemmich and Gilson’s (2009) dataset, the Grid Point Value of the Monthly Objective Analysis using Argo float data (MOAA-GPV; Hosoda et al. 2008) and the In Situ Analysis System (ISAS; Gaillard et al. 2016; Kolodziejczyk et al. 2017), are mapped onto isobaric surfaces, and thus we chose to construct our own gridded dataset. However, these distributed datasets are constructed by more sophisticated methods of quality controls and gridding procedure than those adopted here. For check, we repeated the analysis using Roemmich and Gilson’s (2009) dataset, MOAA-GPV, and ISAS and found that the gross features presented in this study were obtained from these datasets, although details differed from dataset to dataset. This shows that results presented here are insensitive to adopted methods. Below we show results obtained from the dataset gridded in this study.

b. Mixed layer heat budget

The heat budget of the surface mixed layer is written as (Dong et al. 2007; Nagura et al. 2015)

$$\frac{\partial T}{\partial t} = \frac{Q}{\rho_0 c_p h} - \mathbf{u} \cdot \nabla_h T - \frac{T - T_b}{h} \left( w_b + \frac{\partial h}{\partial t} + \mathbf{u} \cdot \nabla_h h \right) + R,$$

(1)

where $T$ denotes temperature; $h$ is the MLD; $\mathbf{u} = (u, v)$ and $w$ are horizontal and vertical velocities, respectively; $Q = Q_0 - q(-h)$ is surface heat flux ($Q_0$) minus the penetration of shortwave radiation at the base of the mixed layer [$q(-h)$]; $R$ is the residual; $t$ is time; $\nabla_h$ is the horizontal gradient operator; $\rho_0 = 1025 \text{ kg m}^{-3}$ is the mean seawater density, and $c_p = 3994 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat of the ocean. The overbar denotes the average over the mixed layer, and the subscript $b$ denotes the value at the mixed layer base. Here, horizontal velocity averaged over the mixed layer $\overline{\mathbf{u}}$ was computed as the sum of geostrophic velocity ($\mathbf{u}_g$) and Ekman currents ($\mathbf{u}_e$). Ekman currents were computed as Ekman transport divided by MLD following Sallée et al. (2006) and Dong et al. (2007), that is, $\mathbf{u}_e = \tau \times \mathbf{k}/(\rho_0 h)$, where $\tau$ denotes the surface wind stress vector, $\mathbf{k}$ is the unit vertical vector, and $f$ is the Coriolis parameter. This formulation is based on the fact that surface Ekman currents tend to be trapped in the surface mixed layer (Chereskin and Roemmich 1991). The term on the left-hand side of Eq. (1) represents temporal variability in mixed layer temperature. The first term on the right-hand side of Eq. (1) is warming/cooling due to surface heat flux. The second term on the right-hand side is horizontal advection. The third term on the right-hand side is the entrainment term and the advection term across the base of the mixed layer, representing heat exchange at the base of the mixed layer. We ignored the effects of currents at the base of the mixed layer ($u_b = w_b = 0$) because reliable estimates are unavailable. The examination of output of OFES2 shows that the term related to $u_b$ and $w_b$ is small in magnitude (section 5a).
The most straightforward way to compute the right-hand side terms in Eq. (1) is to obtain monthly estimates of $h$ and $T_h$ from in situ observations and calculate the entrainment term, with computing the surface heat and horizontal advection terms using $Q$, $\mathbf{u}$, and $T$ obtained from other datasets. We tried this method, but results are erroneous and did not give a closed budget. A possible error in this method is in the estimation of the entrainment term, owing to the fact that in situ observations are not dense enough to resolve entrainment processes. Then, we adopted the concept of effective MLD proposed by Deser et al. (2003), which circumvents the explicit calculation of the entrainment term.

The concept of effective MLD deals with the impact of surface heat forcing on wintertime mixed layer temperature. At midlatitudes, the surface mixed layer is thin in summer and thick in winter. If the mixed layer is anomalously warmed by surface heat flux in summer, a positive mixed layer temperature anomaly is generated, which is large in amplitude, because the mixed layer is thin. The amplitude of the temperature anomaly decreases during fall, when subsurface water entrains into the mixed layer, and the volume of the surface mixed layer increases. As a result, we can compute a response of wintertime mixed layer temperature to a surface heat flux anomaly as $Q_t/(\rho_0 g h^{\text{eff}})$ irrespective of the season when a surface heat flux anomaly is applied. Here, the prime denotes the anomaly defined as the deviation from the mean seasonal cycle, and $h^{\text{eff}}$ denotes the wintertime MLD, called effective MLD. The validity of this idea was confirmed by numerical experiments with a one-dimensional mixed layer model by Nagura and Kouketsu (2018). As temperature in the wintertime mixed layer determines that of the subducted water mass (Stommel 1979), we can estimate temperature anomalies of a subducted water mass using effective MLD. Similarly, Ekman currents are computed as $\tau \times \mathbf{k}/(\rho_0 g h)$; they are large in magnitude in summer because of small $h$ and can cause large response in summertime mixed layer temperature via horizontal heat advection. The amplitude of the resulting mixed layer temperature anomalies is reduced during fall owing to entrainment of subsurface water. The response of wintertime mixed layer temperature to Ekman currents can be computed using effective MLD, too. As a result, year-to-year variation in wintertime mixed layer temperature can be computed as (see appendix A for details)

$$
\mathcal{T}(t_1) - \mathcal{T}(t_0) = \int_{t_0}^{t_1} \frac{Q_t}{\rho_0 g h^{\text{eff}}} dt - \int_{t_0}^{t_1} \left[ \mathbf{u}_g + \frac{\tau \times \mathbf{k}}{\rho_0 g h^{\text{eff}}} \right] \cdot \nabla \cdot \mathcal{T} \right| \cdot dt + R, \quad (2)
$$

where $t_0$ denotes time at winter in a year, and $t_1$ is time at the next winter. Note that we assumed that geostrophic currents are vertically uniform from the sea surface to effective MLD.

We computed each term of Eq. (2) as follows. The temporal tendency of mixed layer temperature anomalies [left-hand side of Eq. (2)] was computed from monthly estimates of mixed layer temperature obtained from in situ hydrographic observations. Surface heat fluxes were obtained from OAFlux and ERA-Interim. The penetration of shortwave radiation was computed using the formula of Paulson and Simpson (1977) for water type IB. Effective MLD was obtained as the maximum of monthly climatological estimate of MLD at each grid point. Geostrophic velocity and wind stress data were obtained from CMEMS and ERA-Interim, respectively. To resolve the Subtropical and Subpolar Front in the Antarctic region, we used RSS’s microwave SST to compute horizontal gradient of mixed layer temperature. Each term on the right-hand side of Eq. (2) was integrated from September in a year to August in the next year (i.e., $t_0$ is September in a year, and $t_1$ is August in the next).

The residual term $R$ includes the effects of processes neglected in our estimate. One of them is horizontal diffusion. We computed this term using RSS’s SST with an eddy diffusivity of 500 m$^2$ s$^{-1}$ following Dong et al. (2007), but contributions were negligibly small, which is consistent with results of Dong et al. (2007) and Tamsitt et al. (2016). The horizontal diffusion term is omitted to show in the rest of the paper. Another process neglected in this study is the effects of interannual variability in wintertime MLD, which are discussed in detail in section 5b. To validate our approach, we computed the mixed layer heat budget based on Eq. (1) using output from OFES2 and ECCO v4 and compared results with those based on Eq. (2). Results of this examination are presented in section 5a.

4. Results

We first describe the spreading patterns of spiciness anomalies and specify their generation region. Then, the mixed layer heat budget in the generation region is presented. Here we use salinity anomalies on isopycnal surfaces as a measure of spiciness variability. The validity of this choice is discussed in section 5d.

On 26.2 and 26.4 $\sigma_g$, spiciness anomalies are large in amplitude along 15°S from the northwestern coast of Australia to about 60°E (Figs. 4a,c). The mean salinity on these isopycnals is high south of 15°S and low north of 15°S, forming a sharp front at about 15°S (Figs. 4b,d). These patterns suggest that spiciness anomalies along 15°S are generated by meridional excursions of the salinity front. The salinity front becomes weaker (Fig. 4f) or absent (Fig. 4h) on density surfaces related to SAMW. On 26.6 $\sigma_g$, spiciness anomalies are large in amplitude in the southeastern Indian Ocean (20°–35°S, 80°–110°E), and the amplitude is relatively small along 15°S (Fig. 4e). On 26.8 $\sigma_g$, a large amplitude of spiciness anomalies is in the south Indian Ocean along the streamline of the mean flow, whose upstream end is the outcrop region south of Australia (Fig. 4g). These patterns suggest that spiciness anomalies of SAMW are not the consequence of meridional excursions of a salinity front, but generated in the outcrop region and spread toward the interior via the advection of the mean current. The amplitude of spiciness anomalies is small all over the basin on 27.0 $\sigma_g$ (Figs. 4i,j), which is the density range of AAIW.

In our analysis period, positive spiciness anomalies are prevalent from 2004 to 2009, and negative anomalies occupy the basin after 2012 (Fig. 5). The decreasing trend of spiciness anomalies in the south Indian Ocean was also reported by Kołodziejczyk et al. (2019, their Fig. 5i). Negative anomalies first appear south of Australia in 2007 and 2008 (Figs. 5d,e) and then spread over the south Indian Ocean (Figs. 5g–o).
To see if the spreading speed of spiciness anomalies is consistent with the mean flow speed, we deployed 23 artificial particles in the region of negative anomalies on 1 January 2011 and advected them by the mean current provided by Nagura and McPhaden (2018). Seasonal variability in meridional velocity is small south of 20°S (Nagura 2018), and the amplitude of interannual variability in subtropical gyre transport in the south Indian Ocean is about 17% of the mean (Nagura 2020). We expect that the use of the mean velocity is a good approximation. Note that Nagura and McPhaden’s (2018) velocity dataset is constructed with objective mapping with a zonal decorrelation scale of 6° and does not resolve detailed structure of coastal currents. Here, if a particle hits the coast, we stopped the computation for the particle. This computation ignores the exchange of water between coastal currents and interior circulation, although such an exchange is found by a numerical study (Domingues et al. 2007).

Results show that particles migrate northwestward from the southeastern Indian Ocean, which is roughly consistent with the spreading of negative spiciness anomalies (Figs. 5h–o). This result suggests that spiciness anomalies are advected by the...
Fig. 5. Yearly estimates of salinity anomalies on 26.8 $\sigma_\theta$ from 2004 to 2018 obtained from in situ observations. Anomalies are masked if they are smaller than analysis error in magnitude. Symbols illustrate positions of artificial particles, which are deployed on 1 Jan 2011 and advected by the mean absolute velocity on 26.8 $\sigma_\theta$ obtained from Nagura and McPhaden (2018). Hatching indicates the outcrop region, where the maximum of mixed layer density for respective year exceeds 26.8 $\sigma_\theta$. 
mean current as a passive tracer, rather than generated by the meridional excursion of a salinity front. A discrepancy is found in 2017 and 2018 (Figs. 5n,o), in which negative spiciness anomalies spread within 15°–35°S, 55°–110°E, but artificial particles are not present east of 90°E between 15° and 30°S. This is likely because of the lack of detailed representation of coastal currents and the resulting absence of migration of particles from south of Australia to west of Australia. Except for this discrepancy, the spreading of spiciness anomalies is consistent with migration of artificial particles.

Note that there is a decadal time lag between spiciness anomalies in the interior and those in the generation region. Negative anomalies south of Australia in 2007–08 reach the interior of the south Indian Ocean in 2013–18. This time scale is consistent with the estimate of Jones et al. (2016), who tracked passive tracers in an OGCM. It is expected that positive anomalies south of Australia in 2016–18 will spread to the interior in the next decade, resulting in a warming of the interior water. Håkkinen et al. (2016) reported that the cooling tendency of spiciness anomalies in the Southern Ocean counteracts the warming tendency due to thermocline heave in the last 50 years. It is possible that warming signals of spiciness anomalies will reinforce the warming tendency in coming years.

Mixed layer salinity, temperature, and density anomalies in the outcrop region are shown in Fig. 6. We chose 38°–48°S, 100°–150°E as the analysis region because the surface density in this region is 26.6 to 26.8 σθ (Figs. 1a–c), which is the typical density range of SAMW (Fig. 2b), and also this region is the upstream end of interior spiciness anomalies (Fig. 4g). We plot values in September because the mixed layer is thickest in this month, and the T/S relationship of subducted SAMW is determined by the surface water in this month. We confirmed that results presented below were not sensitive to the choice of the domain.

Mixed layer salinity is anomalously low in the former half of the analysis period and high in the latter half (Fig. 6a). To compare area-averaged salinity anomalies with isopycnal salinity anomalies, we interpolated mixed layer salinity onto density surfaces and averaged over 100°–150°E and 26.6 to 26.8 σθ. Analysis error for area-averaged values is shown by vertical bars. The temporal means for the period from 2004 to 2018 were subtracted from the time series in (a) and (b). A three-point running-mean filter was applied.

Fig. 6. Mixed layer (a) salinity, (b) temperature, and (c) density in September averaged over 38°–48°S, 100°–150°E obtained from in situ observations (black lines). The red line in (a) shows anomalies of mixed layer salinity on isopycnals averaged over 100° to 150°E and 26.6 to 26.8 σθ. Analysis error for area-averaged values is shown by vertical bars. The temporal means for the period from 2004 to 2018 were subtracted from the time series in (a) and (b). A three-point running-mean filter was applied.

Mixed layer salinity is anomalously low in the former half of the analysis period and high in the latter half (Fig. 6a). To compare area-averaged salinity anomalies with isopycnal salinity anomalies, we interpolated mixed layer salinity onto density surfaces and averaged over 100°–150°E and 26.6–26.8 σθ. The resulting isopycnal salinity anomalies (Fig. 6a, red line) show a consistent time evolution with area-averaged mixed layer salinity anomalies. Note that negative isopycnal salinity anomalies in the former half of the analysis period spread into the interior in subsequent years, as shown in Fig. 5. Mixed layer temperature anomalies averaged over the analysis region are negative from 2004 to 2009 and positive from 2011 to 2015 (Fig. 6b), nearly in phase with mixed layer salinity anomalies, and tend to be density compensating. Consistently, mixed layer density is almost constant and remains to be within the density range of SAMW (Fig. 6c). Results show that the mixed layer heat or salinity budget analysis in a fixed domain can be used for the discussion of the generation mechanism of spiciness anomalies, as is done by Schneider et al. (1999) and Katsura et al. (2013) for the subtropical Pacific Ocean. This is in contrast to Nagura and Kouketsu (2018), who investigated the region where variability in SST and sea surface salinity (SSS) is not density compensating and thus conducted a Lagrangian analysis.

Here we chose to examine the mixed layer heat budget, rather than the mixed layer salinity budget. The calculation of the horizontal advection term in the budget requires satellite measurements of SST or SSS to resolve the narrow Subantarctic Front and mesoscale variability. Satellite SST measurements obtained from RSS is available for the period from 1998 to the present, whereas the longest records of satellite SSS can be obtained from the European Space Agency’s Soil Moisture and Ocean Salinity mission but available only for the period from May 2010 to March 2016, which does not cover the analysis period of this study.
Results of the mixed layer heat budget computed using the method described in section 3b are shown in Fig. 7a. The temperature tendency term is close to zero in 2005–07, positive from 2008 to 2012 and negative after 2013 (thick black line). Positive tendency in 2008–12 corresponds to warming and the generation of positive spiciness anomalies. The sum of the surface heat flux and horizontal advection terms compares well with the temperature tendency term both in phase and magnitude (thick gray line). The sum of the surface heat flux and horizontal advection terms tends to be smaller than the temperature tendency term in 2008–09 and larger in 2011–15, but the discrepancy is minor.

The comparison of the surface heat flux term and the horizontal advection term indicates that the latter is the main contributor (red and blue lines in Fig. 7a). In particular, the advection term contributes to warming of mixed layer temperature from 2008 to 2012, which shows that positive spiciness anomalies are mainly generated by horizontal advection. The decomposition of the advection term indicates that both Ekman and geostrophic flow contribute to warming (Fig. 7b). Below we examine these advection terms.

Meridional advection dominates the Ekman flow advection term, and advection due to zonal Ekman currents is negligible in amplitude (Fig. 8a). This is owing to the fact that the isolines of the mean mixed layer temperature are almost zonal (Fig. 1a). Zonal wind stress anomalies are negative from 2007 to 2012 and positive after 2013 (Fig. 8b). The meridional advection term due to Ekman flow is roughly out of phase with zonal wind stress anomalies. In this region, the mean wind is westerly, which excites northward Ekman flow and equatorward advection of polar cool water. A weaker westerly wind leads to a weaker northward Ekman flow and anomalous warming. This result supports the idea of Rintoul and England (2002).

A climate mode that can be influential in surface wind variability in the Antarctic region is the southern annular mode (SAM; Hall and Visbeck 2002; Lovenduski and Gruber 2005; Verdy et al. 2006; Ciasto and Thompson 2008; Sallée et al. 2010a; Meijers et al. 2019). We computed the SAM index following Limpasuvan and Hartmann (1999), Gong and Wang (1999), and Thompson and Wallace (2000) by applying empirical orthogonal function analysis to SLP from 2004 to 2018 in the region poleward of 20°S. We used the time series of the first mode as the SAM index. We defined the sign of the SAM index such that pressures over Antarctica are relatively low (or high) compared to those at midlatitudes during its positive (negative) phase. Results were almost the same if we used the SAM index of Marshall (2003). Anomalies of the obtained SAM index tend to be negative from 2005 to 2013 and positive after 2014. The correlation between the SAM index and zonal wind stress anomalies is not apparent (Figs. 8b,c). Sallée et al. (2010a) showed Ekman heat flux anomalies regressed onto the SAM index, which are relatively weak in magnitude in the region south of Australia (their Fig. 3b). It is suggested that the influence of SAM is weak in this region.

Another climate mode that affects wind variability in the Southern Ocean is ENSO (Karoly 1989; Fogt and Bromwich 2006; Verdy et al. 2006; Ciasto and Thompson 2008; Sallée et al. 2008b; Meijers et al. 2019). Niño-3.4 SST anomalies tend to be negative from 2006 to 2012 and positive afterward (Fig. 8d), which is in phase with zonal wind stress anomalies. However, the correlation coefficient with the two indices for a longer period (1982–2018) is 0.41, which is not significant even at the 70% level. We also conducted the multiple regression analysis using the SAM index and Niño-3.4 SST anomalies as the independent variables and zonal wind stress anomalies averaged over the analysis region as the dependent variable. Note that the correlation between the SAM index and Niño-3.4 SST anomalies is 0.08, and these two can be treated as independent variables. Results explain only 32% of the variance, which further suggests weak influences of these climate modes. More elaborated study is necessary to clarify the factor that controls zonal wind stress variability south of Australia.

It is expected that variability in geostrophic flow can be decomposed into large-scale and mesoscale variability, the former of which is variability in the strength of the mean flow,
and the latter is due to eddies, meanders, and rings. The spatial scale of mesoscale eddies is about 80 km (Frenger et al. 2015), and that of meanders is about 500 km (Phillips and Rintoul 2000). The time scale of mesoscale variability is from 20 to 120 days (Phillips and Rintoul 2000). We define the large-scale field (denoted by $u^L$ and $v^L$) by averaging $u_g$ and $v_g$ over the analysis region (38°–48°S, 100°–150°E) and smoothed the resulting time series by a 1-yr running mean filter. The mesoscale field ($u^M$ and $v^M$) was computed by subtracting the large-scale field from the total. The advection term due to geostrophic flow was decomposed using this filtering (Fig. 9). The total geostrophic advection term (thick black line) is negative before 2008 and positive in 2011–14. The contribution from large-scale fields ($-u^L \cdot \nabla_h T^L$, red line) is negative in 2005 and 2006, positive from 2008 to 2014, and negative afterward. That from mesoscale fields ($-u^M \cdot \nabla_h T^M$, blue line) is negative in the former half of the analysis period, increases after 2009, and is positive in 2011–16. In 2008–10, the advection terms due to large-scale and mesoscale fields compensate for each other, and the total is close to zero. In 2011–14, the terms due to the two components are both positive. This result shows that mesoscale variability related to eddies contributes to the mixed layer heat budget, which supports Sallée et al. (2008a) and Herraiz-Borreguero and Rintoul (2010). Also, large-scale variability has a major contribution. The cross product ($-u^M \cdot \nabla_h T^M$ and $-u^L \cdot \nabla_h T^L$) is small in magnitude (green line).

Meridional advection is the primary contributor to advection due to large-scale geostrophic currents (Fig. 10a) because the isolines of the mean mixed layer temperature are almost zonal (Fig. 1a). Variability in the large-scale component of geostrophic meridional velocity ($u^L_y$) is out of phase with $-u^L_y \cdot \nabla_h T^L$ (Fig. 10b), which indicates that an anomalous southward current advects a warm water to the

![Fig. 8](image_url)

![Fig. 9](image_url)
south, resulting in anomalous warming. Wind stress curl anomalies averaged over the region tend to be negative in 2008 and 2012–13 (Fig. 10c), which coincides with southward anomalies of large-scale geostrophic currents. This relationship is suggestive of the Sverdrup balance, which holds in most of longitudes of the Southern Ocean (excluding the region near the Drake Passage) in the steady state (Wells and De Cuevas 1995). However, the time series of wind stress curl anomalies does not agree well with that of $L_g$ anomalies. For example, the magnitude of negative peak is smaller in 2008 than in 2012–13 in $L_g$ anomalies, but it is larger in 2008 than in 2012–13 in wind stress curl anomalies. A possible cause of this discrepancy is the contribution from bottom pressure torque (Wells and De Cuevas 1995) or advection (Hughes 2005) to the vorticity balance. Also, we examine interannual variability, and the steady vorticity balance may not be a good approximation.

Finally, we discuss advection due to mesoscale fields of geostrophic currents. It is known that poleward transport of eddy-induced velocity counteracts equatorward Ekman transport near the surface in the ACC region (Karsten and Marshall 2002; Marshall and Radko 2003; Sallée et al. 2010b). Considering that eddy-induced velocity advects large-scale tracer field (e.g., Gent et al. 1995), it is expected that eddy-induced velocity advects low-latitude warm water to the south and results in warming. Consistently, the mean of the advection term due to the mesoscale fields ($-\mathbf{u}_M^V \cdot \nabla T$) is positive (about 0.49°C yr$^{-1}$). We measured activity of mesoscale variability by area averages of mesoscale geostrophic velocity squared and mesoscale SST gradient squared. Both indices tend to increase from 2009 to 2014 (Fig. 11). As the mesoscale fields act as warming in the mixed layer heat budget in the mean, the strengthened activity can result in anomalous warming. The mesoscale advection term shows an increasing trend after 2008 (blue line in Fig. 9), which is consistent with this idea.

5. Discussion

a. Mixed layer heat budget estimated from output of OFES2 and ECCO v4

In this study we computed the mixed layer heat budget using the effective MLD approach. We discuss the validity of this approach using output from OFES2 and ECCO v4 in this subsection. We computed the mixed layer heat budget using daily averages of OFES2 output and the discrete version of Eq. (1), which is presented in Kim et al. (2006) and Nagura et al. (2015). OFES2 computed vertical diffusion using an implicit scheme, and it is impossible to retrieve the vertical diffusion term from daily averages of output. Here, the vertical diffusion term is estimated as the residual of the budget. We computed the mixed layer heat budget using ECCO v4 output using the method described in the appendix B. For both OFES2 and ECCO v4, MLD is computed as the depth where density is larger than the value at the topmost level by 0.03 kg m$^{-3}$, which is consistent with the definition for in situ observations. Results were integrated from September in a year to August in the next year.

Surface density obtained from OFES2 and ECCO v4 shows a region of uniform density between 38°S and 48°S (Figs. 12a,b). The wintertime mixed layer obtained from these
products is thick between 40° and 55°S with the maximum of MLD being about 550 m (Figs. 12c,d). These patterns are similar to observations (Fig. 1a), but a contour of 26.8 σρ between 120° and 130°E (Fig. 12a). A similar northward bend is seen in observations (Fig. 1a), but a contour of 26.8 σρ between 120° and 140°E in the simulation. OFES2 and ECCO v4 reproduce the increase of mixed layer temperature anomalies in the analysis region between 2008 and 2012 (Figs. 12e,f), which compares well with observations (Fig. 6b), although OFES2 tends to underestimate the magnitude of mixed layer temperature anomalies.

OFES2 and ECCO v4 have their own advantages and disadvantages. OFES2 simulates the mean field and variability near the surface well (Figs. 12a,c,e), but it is not able to simulate spiciness anomalies in the interior of the south Indian Ocean in the density range of SAMW. Interior spiciness anomalies simulated by OFES2 tend to be out of phase with observed anomalies (figure not shown). This discrepancy indicates an error in generation, subduction, and/or spreading processes of spiciness anomalies in the model. However, daily averages are available as output of OFES2, from which the terms in Eq. (1) can be computed directly. ECCO v4 reproduces the surface fields (Figs. 12b,d,f) and interior spiciness anomalies well, as it assimilates observational data. However, we need to compute MLD from monthly averages of temperature and salinity, which likely leads to nonnegligible errors for the mixed layer heat budget terms, in particular for the entrainment term (appendix B). Here, we present results from both of OFES2 and ECCO v4.

The mixed layer heat budget estimated from OFES2 output shows positive anomalies of the temperature tendency term in 2008–11 (Fig. 13a, thick black line), which corresponds to the increase of mixed layer temperature anomalies shown in Fig. 12e. The surface heat flux term shows anomalous cooling in 2008–10 (red line). The entrainment term (green line) and the diffusion term (orange line) show anomalous warming in 2008 and 2009, which tend to offset anomalous cooling due to the surface heat flux term. This is consistent with the idea of effective MLD, which claims that heat given to the surface mixed layer is redistributed to the depth of effective MLD by entrainment, and thus the effect of surface heat flux is counteracted (appendix A). Also, as discussed in the appendix A, vertical diffusion likely plays a similar role to that of entrainment, which is consistent with results in Fig. 13a. Anomalies of the entrainment (diffusion) term tend to be out of phase with those of the surface heat flux term in 2005–09 and 2012–13 (2008–09 and 2013). The horizontal advection term shows anomalous warming in 2006–11 (blue line), which contributes to the increase of mixed layer temperature anomalies. This supports the conclusion derived in section 4 that horizontal advection is the main driver of anomalous warming of the mixed layer. Warming due to the horizontal advection term is larger in magnitude than that of the temperature tendency term, which is likely owing to the fact that Ekman flow is trapped in a thin mixed layer and swift in summer. Meridional velocity averaged over the surface mixed layer and 38°–48°S, 100°–150°E obtained from OFES2 output tends to be larger in magnitude in summer than in winter (figure not shown). The advection term across the base of the mixed layer is relatively small in amplitude (gray line).

Results from ECCO v4 also show the out-of-phase relationship between the surface heat flux term and the entrainment (diffusion) term in 2005, 2009, and 2011–15 (2008–10 and 2013) (Fig. 13b). The horizontal heat advection term shows anomalous warming in 2006–11, which contributes to the increase of mixed layer temperature anomalies. The surface heat flux term obtained from OFES2 shows anomalous cooling in 2008–10 and anomalous warming in 2012, but that from ECCO v4 shows anomalous warming in 2008–10 and cooling in 2011–15. This discrepancy may be due to the difference in the method of surface heat flux computation. ECCO v4 corrects surface heat flux using ocean state estimate, whereas given meteorological data were used to compute surface heat flux in the integration of OFES2.

Finally, we estimated the mixed layer heat budget using model/reanalysis output and the effective MLD approach. The MLD, mixed layer temperature, surface heat flux, and surface wind stress were obtained from model and reanalysis output. Surface geostrophic velocity was computed from output of SSH. We used daily averages of OFES2 output and monthly averages of ECCO v4 output. Results show the contribution of horizontal advection to anomalous warming in 2008–12...
which is consistent with observational results (Fig. 7a) and the heat budget estimate based on Eq. (1) (Figs. 13a,b). The surface heat flux term computed from OFES2 output shows weak anomalous cooling from 2007 to 2014 (red line in Fig. 13c), which compares well with observational results in Fig. 7a. On the other hand, the surface heat flux term computed from ECCO v4 output shows anomalous warming before 2009 (red line in Fig. 13d), which partly contributes to the increase of mixed layer temperature anomalies. It is difficult to determine which result is correct. We confirmed that this discrepancy is due to the differences in surface heat flux data. The sum of the surface heat flux and horizontal advection terms computed using the effective MLD approach agrees only roughly with the temperature tendency term (black and gray lines in Figs. 13c and 13d), which shows that the effective MLD approach is valid only qualitatively.

b. Effects of interannual variability in wintertime MLD on the mixed layer heat budget

As discussed in Appendix A, we assumed that MLD can be replaced with its seasonal cycle plus the mean in the computation of the mixed layer heat budget terms. However, wintertime MLD averaged over the analysis region varies from about 220 to 320 m, which is comparable to seasonal variability in MLD in amplitude (Fig. 14). The following errors can be caused by ignoring interannual variability in wintertime MLD. First, we estimated \( h_{\text{eff}} \) from monthly climatological estimates of MLD, but wintertime MLD and thus \( h_{\text{eff}} \) change from year to year. This can affect the magnitude of the surface heat flux term and the Ekman flow advection term in Eq. (2). Second, if wintertime MLD is larger in a year than in the previous year, anomalously more subsurface water is entrained into the surface, which can cause interannual variability in mixed layer temperature. Third, interannual variability in MLD and the resulting variability in vertical mixing can generate spiciness anomalies by spiciness injection. In this subsection, we discuss these three issues.

1) INTERANNUAL VARIABILITY IN EFFECTIVE MLD

As described in Appendix A, the formulation of Eq. (2) is defined as the time integral of the forcing terms from winter in a year \( (t = t_0) \) to that in the next \( (t_1) \). Thus, it is possible to define effective MLD for each year and compute the right-hand side terms of Eq. (2). Here we define the maximum of MLD in each calendar year as effective MLD in the year and refer to it as “yearly” effective MLD. We refer to effective MLD obtained from monthly climatological MLD as “climatological” effective MLD. Yearly effective MLD defined for 2010 was used to compute the right-hand side terms of Eq. (2) from September in 2009 to August in 2010, for example. The mixed layer heat budget was computed using yearly effective MLD using observations and output of OFES2 and ECCO v4. Results obtained from OFES2 and ECCO v4 show that the surface heat flux term estimated from yearly effective MLD (Figs. 15b,c, red line) is almost identical to that from climatological effective MLD (Figs. 13c,d, red line). The horizontal
advection term estimated from yearly effective MLD compares well with that from climatological MLD, both of which show anomalous warming in most years after 2007 (Figs. 15b,c, blue line). This result indicates that interannual variability in effective MLD has a small effect on the estimate of the mixed layer heat budget.

On the other hand, some discrepancy can be found in results obtained from observations. The surface heat flux term computed from observations and yearly effective MLD significantly contributes to anomalous warming of mixed layer temperature in 2009–11 (Fig. 15a), which is in contrast to results obtained from climatological effective MLD (Fig. 7a). This discrepancy likely reflects errors for yearly effective MLD obtained from in situ observations, as discussed below.

We compared the surface heat flux term computed with climatological effective MLD with that computed with yearly effective MLD obtained from observations (Fig. 16a). The difference is largest between about 115° and 135°E north of about 43°S, where the wintertime mixed layer is thinnest in the analysis region (Fig. 16b). The surface heat flux term is proportional to the reciprocal of $h_{\text{eff}}$, and it is sensitive to variability in $h_{\text{eff}}$ if the mean $h_{\text{eff}}$ is small. The analysis error for monthly estimates of wintertime MLD (defined in section 3a) ranges from about 10 to 40 m (Fig. 16c), and the ratio of error to the mean is largest between 120° and 140°E north of about 44°S (Fig. 16d). The region with large noise-to-signal ratio approximately coincides with the region of the large difference between the two estimates of the surface heat flux term. Standard deviations of yearly effective MLD in this region are about 40 m (Fig. 16c), which is comparable to analysis error. The amplitude of surface heat flux anomalies ($Q^\prime$) is almost the same in the analysis region (figure not shown), and it is likely that errors for MLD in the region of a thin mixed layer give rise to the difference between the two estimates of the surface heat flux term computed from observations. The noise-to-signal ratio is much smaller in monthly climatological estimate (Fig. 16f), which indicates that results obtained from climatological effective MLD suffer less from error. The horizontal

**FIG. 13.** (a) The mixed layer heat budget terms obtained from OFES2 based on Eq. (1) and averaged over the analysis region ($38^\circ$–$48^\circ$S, $100^\circ$–$150^\circ$E). The temperature tendency term ($\partial T/\partial t$; black line), the surface heat flux term [$Q/(\rho c_p h)$; red line], the horizontal advection term ($-\mathbf{w} \cdot \nabla T$; blue line), the entrainment term [$-(T - T_b) h^{-1} \partial T/\partial h$; green line], the diffusion term (orange line), and the term related to advection across the base of the mixed layer [$-(T - T_b) h^{-1}(w_b + \mathbf{v}_b \cdot \nabla h)$; gray line]. (c) The mixed layer heat budget terms obtained from OFES2 using the effective MLD approach [Eq. (2)] and averaged over the analysis region. The temperature tendency term (black line), the sum of the surface heat flux term and the horizontal advection term (gray line), the surface heat flux term (red line), and the horizontal advection term (blue line). (b),(d) As in (a) and (c), respectively, but obtained from ECCO v4. The temporal mean was subtracted, and a three-point running-mean filter was applied.

**FIG. 14.** Monthly estimate of MLD obtained from in situ observations and averaged over $38^\circ$–$48^\circ$S, $100^\circ$–$150^\circ$E. Values in September are marked by asterisks.
advection term obtained from observations is less sensitive to these errors for MLD (Fig. 15a, blue line). These results show that yearly effective MLD obtained from in situ observations is erroneous. Considering this, we estimated the mixed layer heat budget from observations using climatological effective MLD.

2) ENTRAINMENT

The mixed layer in the analysis region is seasonally thinnest in January and thickest in September (Fig. 17a). Subsurface water is entrained into the surface mixed layer from summer to winter. If the wintertime mixed layer is deeper than climatology, subsurface water is entrained more than normal to the surface, which can cause interannual variability in mixed layer temperature. This effect can be computed as

$$T(t_1) - T(t_0) = \int_{t_0}^{t_1} \frac{\Delta T_{clm}}{h_{clm}} \frac{\partial h'}{\partial t} dt,$$

(3)

where $$\Delta T = T - T_b$$ denotes the difference of temperature between in the mixed layer and at the base of the mixed layer. Other symbols are defined in the appendix A. As entrainment does not affect the mixed layer heat budget, the time integral in the right-hand side of Eq. (3) can be replaced with the time integral for the entrainment season. MLD is about 50 m in every summer and largest in September in almost all years (Fig. 14). Considering this, we estimated $$\Delta T_{clm}/h_{clm}$$ as the average from June to August (JJA), when the base of the mixed layer deepens most rapidly (Fig. 17a). Equation (3) is computed as

$$\bar{T}(t_1) - \bar{T}(t_0) = \frac{\Delta T_{clm}}{h_{clm}} \int_{JJA} [h'(t_{1,JJA}) - h'(t_{0,JJA})],$$

(4)

which indicates that more (or less) subsurface water is entrained if wintertime MLD increases (decreases) from a year to the next. We carried out this computation using observations. We repeated computation with estimating $$\Delta T_{clm}/h_{clm}$$ as the average for the whole entrainment season (i.e., January to August), but results were essentially the same. Results show that the amplitude of the estimated entrainment term is typically 0.01°C yr⁻¹ (Fig. 17b), which is small in magnitude compared to the other terms (Fig. 7a). Note that the entrainment terms shown in Figs. 13a and 13b represent the effect of MLD and $$\Delta T$$ variability on interannual, seasonal, and shorter time scales, whereas that in Fig. 17b represents only the effect of interannual variability in wintertime MLD.

The average of $$\Delta T_{clm}$$ for JJA is positive north of 40°S and negative in a narrow band between 40° and 55°S (Fig. 16e), where the climatological wintertime mixed layer is thickest (Fig. 16b). The region with large variability in wintertime MLD coincides with the boundary between positive $$\Delta T_{clm}$$ and negative $$\Delta T_{clm}$$, where the difference of temperature between the mixed layer and entrained water is small. As a consequence, entrainment caused by interannual variability in wintertime MLD has only a weak effect on the mixed layer heat budget.

3) SPICINESS INJECTION

If a warm salty water overlays a cool freshwater, anomalously vigorous vertical mixing injects high-salinity water near the surface to subsurface virtually without changing density, which results in a spiciness anomaly (Yeager and Large 2004, 2008).

**FIG. 15.** (a)–(c) As in Fig. 7a, Fig. 13c, and Fig. 13d, respectively, but computed with yearly effective MLD. See text for the definition of yearly effective MLD.
the base of the mixed layer is smaller than 45° in this region in late winter except east of 140°E (Fig. 18). This is clearly smaller than the typical Turner angle in the subtropical regions, where previous studies reported the occurrence of spiciness injection (e.g., Fig. 12 in Kolodziejczyk et al. 2015). In the southwestern part of the analysis region, the Turner angle tends to be smaller than −45°, which indicates that both $\alpha_T$ and $\alpha_S$ are negative. This region coincides with the region with negative $\Delta T_{\text{clm}}$ in Fig. 17c. The minimum of the Turner angle in the analysis region is about −61°, and thus temperature and salinity are not strictly density compensating. These results show that the necessary conditions for spiciness injection are not satisfied in most parts of the analysis region.

The Turner angle is larger than 75° south of Tasmania, where interannual variability in MLD is also large in amplitude. Spiciness injection likely occurs in this limited region. Considering zonally wide patterns of spiciness anomalies (Figs. 4 and 5), we speculate that spiciness injection is not the primary generation mechanism, but it may play a secondary role.

c. Accuracy of surface heat flux data

An issue about the accuracy of our analysis is the quality of surface heat flux data. Surface heat flux data obtained from a reanalysis dataset can be problematic in the Southern Ocean, owing to sparse observations assimilated to the reanalysis system (Bourassa et al. 2013). Here we compare net surface heat flux obtained from OAFlux and ERA-Interim with that from in situ observations. We used observations obtained from the Southern Ocean Flux Station (SOFS; Schulz et al. 2012) moored buoy. The SOFS buoy was deployed at 47°S, 142°E from March 2010 to March 2019 and measured various atmospheric and oceanic variables, including shortwave and longwave radiation. Turbulent heat flux computed using the COARE 3.5 bulk flux algorithm (Edson et al. 2013) was provided, too. Here we used the time series up to November 2017, which was available to us. We bi-linearly interpolated surface heat flux obtained from OAFlux and ERA-Interim onto the buoy location, the result of which compares well with that from the SOFS buoy (Fig. 19a). The phase and amplitude of the seasonal cycle agree well between the two. However, noticeable discrepancies can be found, for example in 2010 and 2013, in which OAFlux and ERA-Interim overestimate the magnitude of wintertime cooling compared to in situ observations. The root-mean-square (RMS) difference is about 20 W m$^{-2}$. Tamsitt et al. (2020) reported statistics and mechanisms of surface heat flux variability obtained from the SOFS buoy in detail.

It is difficult to specify error for our heat budget analysis based on the comparison with moored buoy observations. Surface heat flux anomalies used in this study are deviations from the climatological mean, which are averaged yearly and smoothed by a three-point running mean filter. The longest continuous time records of buoy observations are 12 months long, from which we cannot compute yearly averages of anomalies smoothed by a three-point filter. Yearly-averaged surface heat flux anomalies averaged over the analysis region are about 3 W m$^{-2}$ in magnitude (Fig. 19b). Note that the analysis region used in this study roughly agrees with the east Indian Ocean box defined by Tamsitt et al. (2020), within which

\begin{equation}
Tu = \tan^{-1}\left(\frac{\alpha T + \beta S}{\alpha T - \beta S}\right).
\end{equation}

is close to 90° (or −90°) at the base of the mixed layer in winter, in which temperature and salinity are density compensating in the vertical direction (Yeager and Large 2004, 2007). Also, interannual variability in MLD needs to be large in amplitude, which indicates that the strength of vertical mixing varies interannually.

In the analysis region, interannual variability in MLD is large in amplitude south of 40°S (Fig. 16c).
surface heat flux anomalies tend to be in phase. An RMS error of 20 W m$^{-2}$ obtained above is for monthly averages, which include the mean bias and errors for seasonal and shorter-period variability and cannot be considered as error for interannual anomalies. More continuous time series of in situ observations are necessary to precisely quantify the accuracy of surface heat flux anomalies.

As described in section 2, we obtained shortwave and longwave radiation from ERA-Interim and turbulent heat flux from OAFlux, following the method adopted by Dong et al. (2007). On the other hand, Cerovečki et al. (2011) reported that the estimate of the net surface heat flux is erroneous in the Southern Ocean if the net surface heat flux is obtained from surface radiative heat flux of reanalysis data and turbulent heat flux is computed from reanalysis output and a bulk algorithm, although turbulent heat flux computed with a bulk algorithm is more accurate than reanalysis output. Considering this possible error, we computed the surface heat flux term in Eq. (2) using turbulent and radiative surface heat flux obtained from ERA-Interim. Results show that the surface heat flux term does not contribute to the increase of mixed layer temperature anomalies in 2008–12 (Fig. 19c), which is consistent with results obtained from OAFlux and ERA-Interim.

d. Definition of spiciness

In this study we use variability in salinity on isopycnal surfaces as a measure of spiciness anomalies. Its validity is discussed in this subsection. Spiciness $\tau$ is defined as a variable whose contours are orthogonal to potential density contours in the potential temperature–salinity ($\theta$–$S$) space (Stommel 1962; Veronis 1972; Munk 1981) as

$$d\tau = \beta dS + \alpha d\theta,$$  

where $\alpha = -\rho^{-1}(\partial\rho/\partial\theta)$ is the thermal expansion rate and $\beta = \rho^{-1}(\partial\rho/\partial S)$ is the salinity contraction rate. The defined $d\tau$ measures the distance between two water particles in the $\theta$–$S$
space, whose temperature and salinity differ by $d \theta$ and $d S$, respectively (e.g., Wang and Luo 2020). Equation (5) is not a total derivative, because the partial derivative of $\beta$ with respect to $\theta$ is not equal to the partial derivative of $\alpha$ with respect to $S$. This means that Eq. (5) cannot be satisfied in all the directions in the $\theta - S$ space, and the line integral of $\tau$ depends on integral path. If we enforce Eq. (5) along the line of constant density following Jackett and McDougall (1985) and McDougall and Krzysik (2015), spiciness variation between two water particles can be defined as $\int d \tau = \int 2 \beta d S$, where integral is computed along the line of constant density in the $\theta - S$ space from one water particle to another. (Note that $\beta d S = \alpha d \theta$ along an isopycnal surface.) Variation in $S$ on an isopycnal surface is approximately proportional to those in $\tau$, because the salinity contraction rate $\beta$ does not change largely in the oceanographic range of $\theta$ and $S$. In our case, $\beta$ varies only by about 1% in the density range of SAMW in the south Indian Ocean. This justifies the use of isopycnal salinity anomalies as a measure of spiciness anomalies. Alternatively, $\int d \tau$ can be computed as $\int 2 \alpha d \theta$, but $\alpha$ changes by about 14% in the density range of SAMW in the south Indian Ocean, and potential temperature anomalies are a less accurate measure of spiciness anomalies.

As is mentioned in the previous paragraph, Eq. (5) is not a total derivative, and the line integral of $d \tau$ depends on integral path. Recently, Huang et al. (2018) circumvented this arbitrariness by using a least squares method and were able to compute $\tau$ such that its contours are approximately orthogonal to potential density contours everywhere in the $\theta - S$ space. They referred to the resulting variable as “spiciness.” To check consistency between spicity and isopycnal salinity anomalies, we computed spicity using the subroutine provided by Huang et al. (2018) and each profile of in situ hydrographic observations with the reference pressure of 0 dbar and objectively mapped them as is described in section 3a. The resulting spiciness anomalies are illustrated in Fig. 20, which shows almost the same pattern as isopycnal salinity anomalies illustrated in Fig. 5. This agreement shows that isopycnal salinity anomalies are a good substitute for spiciness anomalies.

6. Summary

This study revisited the issue of spreading and generation of spiciness anomalies related to SAMW in the south Indian Ocean, using in situ hydrographic observations, various satellite and reanalysis datasets, and an OGCM output. Results showed that spiciness anomalies in the density range of SAMW (26.6–26.8 $\sigma_T$) are generated in the outcrop region south of Australia and spread into the interior of the south Indian Ocean being advected by the subtropical gyre. This is expected by previous studies but explicitly shown by this study. It takes about a decade for spiciness anomalies to fully spread from the outcrop region to the interior up to Madagascar Island, which is consistent with results of the numerical model study by Jones et al. (2016).

In the outcrop region (38°–48°S, 100°–150°E), variability in wintertime mixed layer temperature and salinity tends to be density compensating, resulting in isopycnal $T/S$ anomalies. To examine the possible cause of variability in mixed layer temperature, we estimated the mixed layer heat budget using the concept of effective MLD proposed by Deser et al. (2003). The sum of the estimated surface heat flux and horizontal advection terms compares well with the temperature tendency term. The contribution from the surface heat flux term is relatively small, if we compute this term using surface heat flux data obtained from OAFlux and ERA-Interim. Wintertime mixed layer temperature responds to cumulative heat loss at the sea surface.
during fall and winter. The small contribution from the surface heat flux term is partly owing to the thick mixed layer in these seasons. However, results obtained from other dataset (ECCO v4) showed that the surface heat flux term contributes to mixed layer warming mainly in 2009. The accuracy of surface heat flux data in the Southern Ocean is questionable owing to sparse observations, and our conclusion about the role of surface heat flux is not decisive.

The budget obtained from the effective MLD approach shows that horizontal advection is the main contributor to the generation of mixed layer temperature anomalies. The decomposition of the advection term showed that both of Ekman flow and geostrophic currents significantly contribute to horizontal advection. Advection due to Ekman flow is caused by meridional Ekman flow anomalies, which is highly correlated with zonal wind stress anomalies. Advection due to geostrophic flow was further decomposed into contributions from large-scale and mesoscale fields. Here, the large-scale field was defined as variability averaged over the analysis region, whose time scale is longer than a year. The mesoscale field includes eddies, meanders, and rings. These two components have comparable contributions to horizontal advection. The contribution from the large-scale field is caused by variability in meridional geostrophic velocity, and that from the mesoscale field is owing to the temporal changes in activity of mesoscale variability.

The uniqueness of our results is in the estimation of relative importance of the forcing terms in the generation of spiciness anomalies. Our results showed that it is less likely for surface air–sea flux anomalies to be a primary factor to the generation of spiciness anomalies, although there is an issue of the accuracy of surface heat flux data. Both Ekman flow and eddy-related geostrophic currents contribute to the generation of spiciness anomalies, which supports results of Rintoul and England (2002), Sallée et al. (2008a), and Herraiz-Borreguero and Rintoul (2010). In addition, our results showed that variability in large-scale meridional geostrophic velocity has a nonnegligible contribution, which has not been pointed out before to our knowledge. Results also showed that contributions from Ekman flow and large-scale and mesoscale geostrophic currents are comparable in magnitude. Results of this study suggest that it is unlikely that spiciness injection effectively works in the formation region of SAMW. It is also suggested that the contribution from entrainment caused by interannual variability in wintertime MLD is small, which is owing to the small difference in temperature between the mixed layer and subsurface in the entrainment season. We confirmed the validity of the effective MLD approach using output from an OGCM and a reanalysis dataset.

Finally, the analysis period of this study is from 2004 to 2018, and it is unclear how results of this study are related to long-term trends reported by Durack and Wijffels (2010) and Håkkinen et al. (2016). Such a work will require observational data or model output for a longer period and remains to be a future study.

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**APPENDIX A**

**Formulation of the Mixed Layer Heat Budget Based on the Concept of Effective MLD**

If mixed layer temperature is only caused by surface heat flux, Ekman flow advection and entrainment, the heat budget is written as
We have assumed that winter in a year (mixed layer temperature. Integrating the above equation from the result by wintertime MLD, we obtain

\[ \frac{dT}{dt} = \frac{Q}{\rho_s c_p h} - \left( \frac{k}{\rho_s f c_p} \right) \nabla_h \mathbf{T} - \frac{T_b}{h} \frac{\partial h}{\partial t} \]  

(A1)

We have assumed that \( u_b = w_b = 0 \). We replace MLD (\( h \)) with its climatology (\( h^{\text{clim}} \)), which is defined as the mean plus the seasonal cycle. The validity of this assumption is discussed in section 5b. The resulting budget equation is written as

\[ \frac{dT}{dt} = \frac{Q}{\rho_s c_p h^{\text{clim}}} \left( \frac{k}{\rho_s f c_p} \right) \nabla_h \mathbf{T} - \frac{T_b}{h^{\text{clim}}} \frac{\partial h^{\text{clim}}}{\partial t}. \]  

(A2)

The equation for anomalies (denoted by prime) is written as

\[ \frac{dT'}{dt'} = \frac{Q'}{\rho_s c_p h^{\text{clim}}} - \left( \frac{k}{\rho_s f c_p} \right) \nabla_h \mathbf{T}' - \frac{T_b'}{h^{\text{clim}}} \frac{\partial h^{\text{clim}}}{\partial t}'. \]  

(A3)

Note that anomaly is defined as the deviation from its climatology (\( q' = q - q^{\text{clim}} \) for an arbitrary variable \( q \)). We further assume that the anomaly of entrained water temperature is zero (i.e., \( T_b' = 0 \)). This is equivalent to focusing on temperature anomalies that are generated by surface forcings. The budget equation is now

\[ \frac{dT'}{dt'} = \frac{Q'}{\rho_s c_p h^{\text{clim}}} - \left( \frac{k}{\rho_s f c_p} \right) \nabla_h \mathbf{T}' - \frac{T_b'}{h^{\text{clim}}} \frac{\partial h^{\text{clim}}}{\partial t}'. \]  

(A4)

which is written as

\[ \frac{\partial (h^{\text{clim}} \mathbf{T})}{\partial t} = \frac{Q'}{\rho_s c_p} - \left[ \frac{k}{\rho_s f c_p} \right] \nabla_h \mathbf{T}'. \]  

(A5)

We are interested in year-to-year variability in wintertime mixed layer temperature. Integrating the above equation from winter in a year (\( t = t_0 \)) to that in the next year (\( t_1 \)) and dividing the result by wintertime MLD, we obtain

\[ T'(t_1) - T'(t_0) = \int_{t_0}^{t_1} \frac{Q'}{\rho_s c_p h^{\text{clim}}} dt - \int_{t_0}^{t_1} \left[ \frac{k}{\rho_s f c_p} \right] \nabla_h \mathbf{T} \right] dt, \]  

(A6)

where \( h^{\text{eff}} = h^{\text{clim}}(t_0) = h^{\text{clim}}(t_1) \) denotes effective MLD. Thus, the entrainment term can be included in the surface heat flux and Ekman advection terms by replacing \( h \) with \( h^{\text{eff}} \). This is equivalent to examining temperature variability in the surface layer, whose thickness is the same as wintertime MLD.

Equation (A4) indicates that entrainment, represented by the third term on the right-hand side, works as a damping for \( T \). Considering this, Vivier et al. (2010) included a Newtonian damping term in their diagnostic model as a parameterization of entrainment, but it required them to use an arbitrarily chosen damping coefficient. In this study, such an arbitrariness is avoided by the use of the effective MLD concept.

If we include horizontal advection due to geostrophic currents in Eq. (A1), the equation for the mixed layer heat budget is written as

\[ \frac{dT}{dt} = \frac{Q}{\rho_s c_p h} - \left( \frac{k}{\rho_s f c_p} \right) \nabla_h T - \frac{T_b}{h} \frac{\partial h}{\partial t}. \]  

(A7)

A calculation similar to the above yields

\[ T'(t_1) - T'(t_0) = \int_{t_0}^{t_1} \frac{Q'}{\rho_s c_p h^{\text{clim}}} dt - \int_{t_0}^{t_1} \left[ \frac{k}{\rho_s f c_p} \right] \nabla_h T \right] dt. \]  

(A8)

Results computed with Eq. (A8) are shown in Fig. A1. The summation of the right-hand-side terms mostly explains warming tendency from 2008 to 2011 and cooling tendency in 2014–16, but the magnitude of warming and cooling is underestimated. The derivation of Eq. (A8) assumes that geostrophic currents are trapped in the surface mixed layer, as Ekman currents are. This is unlikely according to previous studies, which reported that currents related to ACC and mesoscale eddies are deeper than the typical MLD (Killworth 1992; Schodlok and Tomczak 1997; Phillips and Rintoul 2000; Frenger et al. 2015). Thus, we included the advection due to geostrophic currents as follows:

\[ T'(t_1) - T'(t_0) = \int_{t_0}^{t_1} \frac{Q'}{\rho_s c_p h^{\text{clim}}} dt - \int_{t_0}^{t_1} \left[ \frac{k}{\rho_s f c_p} \right] \nabla_h T \right] dt. \]  

(A9)

This formulation is equivalent to assuming that \( u_g \) is vertically uniform from the surface to \( z = -h^{\text{eff}} \). Results (Fig. 7a) show a better agreement between the left-hand side and the right-hand side terms. Also, similarity between Figs. 7a and A1 indicates that results do not essentially change due to the ambiguity of the formulation of the geostrophic current advection term.

In addition to entrainment, vertical diffusion can contribute to redistribution of heat between the surface and subsurface layers. Intense warming/cooling caused by surface heat flux or Ekman flow advection is likely diffused to subsurface across the base of the summertime thin surface mixed layer. The mixed layer heat budget computed from output of OFES2 and ECCO v4 shows a significant contribution from vertical diffusion (section 5a). Vertical diffusion is not included in the above formulation, but it is expected that its role is essentially the same as that of
The conservation equation of heat is written as

$$\frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T = \frac{\partial}{\partial z} \left( K \frac{\partial T}{\partial z} \right) + D_h(T), \quad \text{(B1)}$$

where $\mathbf{v} = (u, v, w)$ denotes the velocity vector, $\nabla$ is the three-dimensional gradient operator, $K$ is the vertical diffusivity coefficient, and $D_h$ is the horizontal diffusion operator. The equation of the mixed layer heat budget is obtained by vertically averaging Eq. (B1) from the sea surface to the base of the mixed layer:

$$\frac{1}{h} \int_{-h}^{0} \frac{\partial T}{\partial t} \, dz = \frac{\partial T}{\partial t} \bigg|_{z=0} + \frac{T - T_b}{h} \frac{\partial h}{\partial t}, \quad \text{(B2)}$$

$$\frac{1}{h} \int_{-h}^{0} \mathbf{v} \cdot \nabla T \, dz = -\mathbf{u} \cdot \nabla T + \frac{T - T_b}{h} \left( w + u_b \cdot \nabla h \right), \quad \text{(B3)}$$

The Leibnitz integral rule and the conservation of volume have been used. We have omitted the term related to deviations of $u$, $v$, and $T$ from mixed layer averages, because this term is usually negligible [this term is retained in Eq. (B3) in Nagura et al. (2015)]. Setting vertical diffusion of heat at the surface identical to air–sea heat flux $[K \delta_z T|_{z=0} = Q/(\rho c_p h)]$, we obtain Eq. (1).

Snapshots of $T$ and salinity ($S$) at the first day of each month and monthly averages of $T$, $S$, $-\mathbf{v} \cdot \nabla T$, and $\partial_x (K \delta_x T) + D_h(T)$ are provided as ECCO v4 output. The heat budget computed from ECCO v4 output is exactly closed at each grid. The temperature tendency term ($\partial_t T$) in a month was computed from the snapshot of $T$ at the first day of the month and that at the first day of the next month. MLD was computed from monthly averages of $T$ and $S$, and $\partial_t T$ was vertically averaged from the sea surface to MLD. Using snapshots of $T$ and $S$, mixed layer temperature at the first day of each month was calculated, from which the tendency of mixed layer temperature ($\partial_t T$) was obtained. The entrainment term $[\delta_T (T - T_b)]$ was computed as the difference between $h^{-1} \int_{-h}^{0} \partial_t T \, dz$ and $\partial_t \bar{T}$ [see Eq. (B2)]. The horizontal advection term ($h^{-1} \int_{-h}^{0} \mathbf{v} \cdot \nabla T \, dz$) and the vertical diffusion term ($h^{-1} \int_{-h}^{0} \partial_z (K \delta_z T) + D_h(T) \, dz$) were computed using MLD obtained from monthly averages of $T$ and $S$. We do not distinguish the first and second terms on the right-hand side.
of Eq. (B3) for results of ECCO v4. Results of OFES2 show that the second term on the right-hand side of Eq. (B3) is smaller in magnitude than the first term (gray line in Fig. 13a). The surface heat flux term \[ Q / (\rho_c c_h) \] was computed using surface heat flux and MLD obtained from monthly averages of \( T \) and \( S \). The computed mixed layer heat budget is exactly closed.

Below we estimated the error caused by the use of MLD calculated from monthly averages of \( T \) and \( S \). We computed the surface heat flux term, the entrainment term, and the vertical average of the horizontal advection term using OFES2 output, obtaining MLD from daily averages of \( T \) and \( S \) and from monthly averages of \( T \) and \( S \) (Fig. B1). Error for the diffusion term cannot be computed, because an implicit scheme was adopted in the integration of OFES2, and the vertical diffusion term cannot be computed from daily averages of output. The horizontal advection term is robust against this error (Fig. B1a). Results computed from monthly MLD tend to follow those from daily MLD. The surface heat flux term is more susceptible to this error than the horizontal advection term (Fig. B1b). In 2006, 2007, and 2010, anomalies of the surface heat flux term obtained from monthly MLD are opposite in sign to those from daily MLD. However, anomalies of the surface heat flux term obtained from monthly MLD roughly follow those from daily MLD, and the amplitude of the former is similar to that of the latter. The entrainment term suffers most from this error (Fig. B1c). Results obtained from monthly MLD tend to be smaller in magnitude than those from daily MLD, and the two results are opposite in sign in 2005, 2010, 2013, and 2014. Note that the mixed layer heat budget computed from ECCO v4 output is exactly closed, and the sum of the forcing term is identical to the heat budget computed from ECCO v4 output is exactly closed, and the sum of the forcing term is identical to the heat budget computed from ECCO v4. Results of OFES2 show that the first term (gray line in Fig. 13a).

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References


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