Eddies Observed by Argo Floats. Part I: Eddy Transport in the Upper 1000 dbar

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ABSTRACT

Argo floats measure horizontal current velocities at the parking depth and vertical profiles of temperature and salinity. The data are sufficient for simultaneous estimates of velocities and vertical displacements of isopycnal surfaces. More than 980,000 pairs of observations of current velocity and water column stratification were used to calculate eddy transport above 1000 dbar and its uncertainty based on the temporal-residual-mean framework. Eddy transports larger than 1.0 m² s⁻¹ were found in the North Atlantic, western North Pacific, and Southern Oceans. The eddy transport \( T \) had components perpendicular \( T_\perp \) and parallel \( T_\parallel \) to the density contours at 1000 dbar. In the midlatitude oceans, eddy transport was weaker (\(<0.5 \text{ m}^2 \text{ s}^{-1}\)), mostly perpendicular to the density contours, and equatorward. A large area of northward \( T_\perp \) was found in the south Indian Ocean; analysis of velocity and thickness perturbations suggested that this transport was a northward intrusion of Antarctic Intermediate Water. In the midlatitude oceans and in most of the southern part of the Antarctic Circumpolar Current (ACC), \( T_\perp \) was generally upgradient in density on 1000 dbar. Downgradient \( T_\parallel \) was found along the North Atlantic Current and Kuroshio Extension as well as in the northern part of the ACC. Zonally integrated meridional transport was poleward at latitudes higher than approximately 40° and equatorward at lower latitudes. The quasi-Stokes or Gent–McWilliams diffusivity coefficient was on the order of 1000 m² s⁻¹ but was associated with such large uncertainty that it was statistically indistinguishable from zero, except at midlatitudes in the Southern Hemisphere.

1. Introduction

An important role that mesoscale eddies play in global ocean circulation is to transport water and tracers. If an eddy is sufficiently strong, it traps a water mass in its core and transports the water mass as a coherent structure. Even when the eddy is not that strong, Stokes drift that results from the nonzero correlation between tracer perturbations and velocity perturbations yields a unidirectional transport of water masses. For example, salty northward flow and fresher southward return flow result in net northward salt transport, even when the net volume transport is nil.

The spatial scale of these eddies is often smaller than the grid size of numerical simulation models that run hundreds, if not thousands, of model years. Eddies are parameterized in such models. The coefficients in the parameterization are often used as tuning parameters so that the model reproduces the observed tracer distributions well. One of the reasons for this tuning procedure is that direct observations of eddy transport are difficult because they require simultaneous observations of velocity and tracer distributions at midocean depths at an eddy-resolving scale.

The aim of this paper is twofold: to examine the feasibility of plotting the eddy transport at a depth of 1000 dbar using data from Argo floats and to describe the eddy field thus estimated. As a result of the relentless effort of the international community (e.g., Riser et al. 2016), as many as about 3700 floats are working quasi globally at the time of this writing. A float records not only the temperature and salinity, but also approximate flow velocity at the parking depth (mostly 1000 dbar), which is determined by its surface positioning (Lebedev et al. 2007; Ollitrault and Rannou 2013). The data are available throughout most of the ice-free areas of the open ocean and are sufficient to estimate eddy transports, although limited to one depth.

In the absence of direct observations, eddy transports have been estimated by three different methods: First, numerical simulation models with eddy-resolving grid sizes have been used. Jayne and Marotzke (2002) used a general circulation model with an average grid size of
about 0.25° and 20 vertical levels; they found strong eddy heat transport concentrated in western boundary currents, along the Antarctic Circumpolar Current (ACC), and in the equatorial region. As discussed by Marshall and Shotts (1981), the eddy flux has two components: rotational and divergent. Jayne and Marotzke (2002) have found that the former is strong in the western boundary regions and the latter is strong in the ACC and the equatorial regions. Aoki et al. (2013) used a model with a finer grid (0.1° and 54 levels) and found a southward eddy heat transport along the Kuroshio and Gulf Stream against the background northward temperature gradient. This transport was attributed to southward advection of nonlinear warm-core eddies.

Second, numerical simulation models can also be used to estimate eddy transport by tuning the eddy transport parameters to fit the tracer distributions to observations. A systematic and physically consistent way to perform this tuning is to use the adjoint method. Ferreira et al. (2005) used the residual-mean formulation [similar to Eq. (6) below] for tracers and momentum and adjusted the eddy stress to minimize the departure of the model temperature from the observed temperature. They found vigorous eddy transport at high latitudes of the Northern Hemisphere and in the Southern Ocean along the ACC. Liu et al. (2012) used the GECCO model (1° grid with 23 vertical levels), in which the initial conditions, surface forcing field, and mixing parameters were adjusted to minimize the difference between the model and the data (sea surface height, sea surface temperature, sea surface salinity, surface drifter velocities, and subsurface temperature and salinity). Although the relative contributions of the mixing parameters to the initial and boundary conditions were small in fitting the model to the data, the best fit of the eddy transport parameters significantly reduced the difference between the simulated and observed data in the western boundary and ACC regions.

Last, it is possible to parameterize eddy transport as the fastest growing solution of the linear instability problem because these eddies are considered to originate from the baroclinic instabilities (e.g., Gill et al. 1974). With climatological data and a closure assumption in the linear instability theory, Vollmer and Eden (2013) found the horizontal diffusivity to be large in the Southern Ocean along the core of the ACC, which has a vertical maximum at depths of 2 to 3 km. This method inevitably underestimates the effects of eddies advected by background flow.

In this paper, we estimated the eddy transport directly by using the hydrographic and drift data recorded by Argo floats. We reproduced results found in these previous studies, such as enhanced eddy transport in the western boundary currents and in the ACC, but we also found differences such as strong northward eddy transport in the south Indian Ocean. In section 2, we provide a detailed explanation of how the data were processed, and we recapitulate the eddy formulation with the temporal-residual-mean framework of McDougall and McIntosh (2001). Section 3 shows our results. In section 4, we attempt to explain the mechanism responsible for some conspicuous eddy transports. We also discuss the float densities in the context of eddy analysis.

2. Data and methods

a. Velocity, temperature, and salinity at 1000 dbar

1) DATA

YoMaHa is a monthly updated compilation of quality-checked Argo trajectory data (Lebedev et al. 2007). We used the December 2015 version. All trajectory data were screened for possible groundings and velocity spikes following the method described in Katsumata and Yoshinari (2010). From a total of 982 071 trajectories, this process removed 0.07% and 4.38% as spikes and possible groundings, respectively. The latter included the trajectories collected at the parking depths we considered too shallow (<400 dbar). Those data not collected at 1000 dbar were adjusted to 1000 dbar by using geostrophic shears calculated from the World Ocean Circulation Experiment (WOCE) climatology hydrographic data (Gouretski and Koltermann 2004). We used both real-time and delayed-mode data (Carval et al. 2015) from the Argo floats. As simple quality check, profiles with salinity outside the range between 33.0 and 37.0 were removed.

2) VELOCITY

One cycle of typical Argo observation consists of descent to a parking pressure, Lagrangian drift at the parking pressure, further descent to a maximum pressure, ascent for profiling temperature–salinity, and surfacing and data communication. The duration of a cycle is usually set to 10 days. Most floats have a parking depth of 1000 dbar. We note that the Argo trajectories cannot be treated as Lagrangian trajectories at depth because of the interruption by surfacing and that the estimated velocity at the parking depth should be treated as Eulerian (Davis 1991).

Davis (1998, 2005) used a simple differencing between the last position fix before descent and the first surface position fix after the subsequent ascent to estimate the velocity at the parking depth. Errors concomitant with this method are 1) array bias (Davis 1998), 2) measurement uncertainties, and 3) sampling uncertainties. Array bias is
about 1.0 cm s\(^{-1}\) near the coast but smaller than 0.1 cm s\(^{-1}\) in open oceans (Katsumata and Yoshinari 2010, their Fig. 16). Measurement uncertainties consist of (i) positioning errors by the satellite system, (ii) clock drift, (iii) unknown surface drift before submerging and after surfacing, and (iv) unknown drifts during ascent and descent between the surface and the parking depth. Ichikawa et al. (2001) tracked four Argo floats and estimated the positioning error (point i) at about 0.1 cm s\(^{-1}\), surface drift error (point iii) at 0 to 1.2 cm s\(^{-1}\), and ascent/descent error (point iv) at 0.1 to 1.3 cm s\(^{-1}\). Clock drift (point ii) is difficult to quantify. Overall, we used an uncertainty of 0.5 cm s\(^{-1}\) to represent the array bias and measurement uncertainties in points 1 and 2. Sampling uncertainty was estimated by the bootstrap method (Efron and Gong 1983) in terms of a 95% confidence interval. To this sampling uncertainty, we added measurement uncertainties on the assumption that there was no correlation between the sampling and measurement uncertainties.

3) TEMPERATURE AND SALINITY

Some floats record temperature and pressure during parking. With these data, the approximate error in the present simple linear interpolation method can be estimated. As of December 2015, 4033 trajectories have "adjusted" (Carval et al. 2015) temperature and pressure data recorded during parking. The depths of the measurements during parking nearest in time to the midpoint between two ascending profiles were found to be 997.3 ± 5.9 dbar (plus or minus one standard deviation). Differences between the average of the temperature measured during two consecutive ascending profiles and the temperature measured during parking had a mean of 0.01°C and a standard deviation of 0.11°C. We therefore assumed a temperature error of Δ\(T\) = 0.11°C in the linear interpolation.

The root-mean-square difference of two temperatures measured during two consecutive ascending profiles was 0.088°C. Similarly, the root-mean-square difference of two salinities measured during two consecutive profiles was 0.0059. Using these numbers, we estimated the error of the salinity in the linear interpolation as Δ\(S\) = 0.11 × 0.0059/0.088 = 0.0074. Using the mean temperature (\(T\) = 4.37°C) and salinity (\(S\) = 34.467) of all profiles at 1000 dbar, we estimated the error in density in the linear interpolation as

\[
\Delta \rho = [\alpha(T, S, P)\Delta T + \beta(T, S, P)\Delta S] \times \rho(T, S, P),
\]

\[
= 0.020 \text{ kg m}^{-3}
\]

where \(P = 1000\) dbar and the Thermodynamic Equation of Seawater—2010 (TEOS-10; IOC et al. 2010) was applied at 30°C, 180° for the thermal expansion coefficient \(\alpha\) and the saline contraction coefficient \(\beta\). This measurement uncertainty in density was converted to the uncertainty in isopycnal displacement via the method depicted in Fig. 1. Sampling uncertainty was again estimated by the bootstrap method (Efron and Gong 1983) in terms of a 95% confidence interval, to which measurement uncertainty was added assuming zero correlation between the sampling and measurement uncertainties.

4) NOTATION

Throughout this paper, we use potential density referenced to 1000 dbar calculated by using version 3.03 of TEOS-10 (IOC et al. 2010). A density of \(\sigma_1\) means a potential density of (1000 + \(\chi\)) kg m\(^{-3}\) referenced to 1000 dbar. For densities around 1000 dbar, this potential density is a good local approximation to the neutral density \(\gamma\).

Most of our discussion is limited to quantities on a two-dimensional surface at 1000 dbar. The direction normal to and parallel to the density (isopycnal) contours on the 1000 dbar surface are denoted by \(\perp\) and \(\parallel\), respectively. For example, \(V_\perp\) is the horizontal velocity component perpendicular to the isopycnal contour on the 1000-dbar surface. Positive \(V_\perp\) is from low to high densities (i.e., upgradient) or

\[
V_\perp = \mathbf{V} \cdot \frac{\gamma_x}{\sqrt{\gamma_x^2 + \gamma_y^2}}, \quad V_\parallel = \mathbf{V} \cdot \frac{-\gamma_y}{\sqrt{\gamma_x^2 + \gamma_y^2}}.
\]
where horizontal velocity is \( \mathbf{V} \) and density is \( \gamma \) with spatial derivatives denoted by subscripts.

### b. Mean field

From the point measurements of density and velocity, it is necessary to define a mean field. Eddies are defined as the deviation from this mean. The Eulerian time scale, defined as the integral of the temporal Eulerian autocorrelation function, has been estimated by using, for example, a temperature time series measured by moored thermistors. Phillips and Rintoul (2000) applied this method in the Southern Ocean and found Eulerian time scales of 5 to 50 days (their Table 4). They also found that 90 days was a good cutoff period for calculating covariances between temperature and cross-stream velocity. Some trials, however, revealed that the number of Argo trajectory data was found insufficient to resolve the temporal behavior of the eddy statistics, even at an annual or biannual time interval. In this manuscript, we therefore consider no time dependence.

Eulerian spatial scales at depth are more difficult to estimate from data. Lumpkin et al. (2002) used a numerical simulation and found that the simulated Eulerian length scale decreased with increasing latitude. In the Atlantic Ocean at a depth of 1806 m, the Eulerian length scale was less than 100 km north of 30°N and increased up to 150 km near the equator. Ideally, we would use these Eulerian length scales to define the averaging area, but again we need sample size at the cost of resolution. Following Davis (1998, 2005), we used a circle with a radius of 300 km around each grid point and aggregated all data within that circle. Because the length scale was expected to be longer in the along-isobath direction than the cross-isobath direction, the radius of the circle was modified following Davis (2005):

\[
r = \sqrt{r^2_G + \mu^2 \ln^2 \left( \frac{H_1}{H_2} \right)},
\]

where \( r_G \) is the geographical distance, \( H_1 \) and \( H_2 \) are the water depths at two points (\( H_1 > H_2 \)), and \( \mu = 300 \) km. After this topographic adjustment, the “circle” is scaled so that its area is equal to a circle with a radius of 300 km.

In summary, we defined the average for each grid point as the temporal average over the period from July 1997 (first trajectory data) to December 2015 and the spatial average over a circle with a nominal radius of 300 km adjusted with isobaths. The averaging inevitably includes possible seasonal variability, which is not necessarily negligible (e.g., Hosoda et al. 2006 in North Pacific). Deviation from the average is defined to be the eddy component. For example, a circle is defined around a grid \((i, j)\) and assumes that there are \( N \) pairs of density and velocity estimates \((\gamma_k, \mathbf{V}_k)\) \((k = 1, 2, \ldots, N)\) from Argo floats within the circle. Then, the mean density is estimated as

\[
\gamma_{ij} = \frac{1}{N} \sum_{k=1}^{N} \gamma_k,
\]

and an eddy quantity such as eddy kinetic energy is estimated as

\[
\frac{1}{2N} \sum_{k=1}^{N} |\mathbf{V}_k - \gamma_{ij}|^2.
\]

Figure 2 shows the number of samples \( N \) for each grid. Grid interval is \( 2^\circ \) in latitude (85°S, 83°S, . . . ) and longitude (1°E, 3°E, . . . ). The average number of data for each grid is 602; the distribution of data is inhomogeneous. On the one hand, in regions such as the eastern North Atlantic Ocean, western North Pacific Ocean, and the South Pacific there are more than 1000 data. On the other hand, north Indian Ocean, low latitudes of the South Atlantic Ocean, and the region east of the area of intense data off Japan all suffer from relatively low data density. The Southern Ocean is covered almost to 60°S—approximately the northernmost latitude of the seasonal ice extent. Low data density appears in the results as large sampling uncertainties. If two data were sampled too close in space or in time, those data are not statistically independent. Considering Phillips and Rintoul’s (2000) estimate of Eulerian time scales of 5 to 50 days, and the estimate of Böning (1988) of the length scale of 25 to 82 km (zonal) and 13 to 54 km (meridional) in the North Atlantic, we chose a typical time scale of 10 days and length scale of 100 km and estimated the number of independent data, or degrees of freedom, assuming an exponential shape of autocorrelation functions. The results (Fig. 2) show a similar spatial distribution with average degrees of freedom of 394 for each grid.

The mean vertical density gradient \( \gamma_z \) was calculated by similarly averaging all the gradients calculated from one Argo profile between 950 and 1050 dbar. The horizontal density gradient was calculated by differencing the gridded mean density \( \overline{\gamma} \). A grid with staggered latitudes (84°S, 82°S, . . . ) was used to calculate the meridional density gradient \( \gamma_y \). The mean density field (Fig. 3) did not differ significantly from the WOCE climatology (Gouretski and Koltermann 2004). The horizontal distribution of the density at 1000 dbar reflects the horizontal circulation pattern—anticyclonic subtropical gyres and steep meridional slope in the Southern Ocean associated with the ACC. The density in the North
Atlantic shows the effect of winter convection in contrast to the North Pacific where a subpolar gyre is discernible. We optimistically assumed that these mean quantities did not introduce errors in the eddy statistics.

c. Formulation

In the averaging, we assumed that a spectral gap exists between the slowly varying mean field and the higher-frequency eddy field so that low-passed temporal and spatial derivatives were retained (McDougall and McIntosh 2001). As discussed in the previous section, it was necessary to apply spatial averaging in addition to temporal averaging. The effect of spatial averaging has been discussed in section 12 of McDougall (1998). Because the largest correction term due to spatial averaging works in the direction of density contours (e.g., $V_\parallel$) and is quadratic in grid size, the correction term is not significant except for the western boundary currents.

1) TEMPORAL RESIDUAL MEAN

Without buoyancy forcing, water masses move along constant density surfaces or isopycnals (McDougall 1987). It is therefore natural to average quantities along these isopycnals. Denoting such averaging by $\tilde{\cdot}$ and simple Eulerian averaging at a fixed height by $\cdot$, we write the conservation of density as

$$\tilde{\gamma}_t + \nabla \cdot (\tilde{U} \gamma) = \overline{Q}^\gamma - \nabla \cdot F^M$$

[McDougall and McIntosh 2001, see their Eq. (15)],

where $\overline{Q}^\gamma$ is the thickness-weighted form of the source or sink term $Q$, averaged between a pair of isopycnals [their Eq. (17)]. The eddy contribution is encapsulated in

$$F^M = \tilde{\gamma}U^\gamma + M.$$  

We have neglected small quantities in the third order of small perturbation [their $O(\alpha^3)$]. The vector $M$ is
nondivergent and vanishes in Eq. (6). The quasi-Stokes velocity $U^+$ is defined in terms of the quasi-Stokes streamfunction $\Psi = (\Psi^x, \Psi^y)$ by

$$U^+ = \begin{pmatrix} \frac{\partial \Psi^x}{\partial z} \\ \frac{\partial \Psi^y}{\partial z} \\ -\frac{\partial \Psi^x}{\partial y} - \frac{\partial \Psi^y}{\partial x} \end{pmatrix}. \tag{8}$$

The quasi-Stokes streamfunction at depth $z_a$ is defined by

$$\Psi(z_a) = \int_{z_a}^{z_a + z'_a} \mathbf{V} \, dz, \tag{9}$$

where $\mathbf{V}$ is the horizontal components of the velocity $\mathbf{U}$, and $z_a + z'_a$ is the instantaneous height of the isopycnal surface, which is located at a height of $z_a$ when averaged. For the Argo data, $z_a = 1000 \text{ m}$. We used an approximate form of Eq. (9) to correct to $O(\alpha^2)$:

$$\Psi(z_a) = \nabla z_a + \frac{1}{2} \nabla_z (z'_a)^2, \tag{10}$$

where the mean vertical shear $\nabla z$ was calculated from the horizontal gradient of mean density assuming the geostrophic balance of the mean flow. Using the same notation as Eq. (4), the first term on the right-hand side was estimated from data as

$$\nabla z_{aij} = \frac{1}{N} \sum_{k=1}^{N} (\mathbf{V}_k - \mathbf{V})(z_{ak} - z_a). \tag{11}$$

In McDougall and McIntosh (2001), a further approximation of

$$z'_a = -\frac{\gamma'}{\gamma_z} + O(\alpha^2) \tag{12}$$

was used. This approximation, however, is not accurate near the Antarctica continental shelves, where stratification
is so weak that $\gamma_r$ is tiny at 1000 dbar. Instead, we used the mean density profile to estimate the isopycnal height change as illustrated in Fig. 1. Temperature and salinity data from each Argo profile were interpolated onto a vertical grid at $z = -400, -450, \ldots, -2000$ dbar and the potential densities $\sigma_f$ were calculated on the vertical grid. The density on a depth $z_b = -1000$ dbar on Fig. 1 was used to calculate the vertical isopycnal heave $z'_b = (-150$ m on Fig. 1). This conversion from density to isopycnal heave is nonlinear; the result is that even though density perturbation $\gamma'$ at $z_b = -1000$ dbar was used to estimate $z'_b$, the average $z'_b$ was not necessarily 0, despite the fact that $\gamma' = 0$. We numerically adjusted the depth $z_b$ so that $z'_b$ lies within $\pm 10$ m.

In the Southern Ocean, the surface mixed layer sometimes reaches as deep as 400 m (Dong et al. 2008, their Fig. 5). We therefore consider the possible range of isopycnal heaving as between 400 and 2000 dbar. Isopycnal depths shallower than 400 dbar or deeper than 2000 dbar are truncated to these limits.

As discussed in section 2a(2), velocities measured by Argo floats are available at 1000 dbar. The horizontal components $V_+^i$ of the eddy advection $U_+^i$ can be written using Eq. (8), which is integrated in $z$ as

$$T = \int_{-1000}^{0} V_+^i \, dz = \left[ \Psi^i(z = 0) - \Psi^i(z = -1000) \right] - \left[ \Psi^i(z = -1000) - \Psi^i(z = -1000) \right] = -\Psi^i(z = -1000), \quad (13)$$

where the boundary condition $\Psi(z = 0) = 0$ (McDougall and McIntosh 2001, their section 8) was used. Thus, the horizontal eddy transport between the sea surface and $z = -1000$ dbar can be evaluated by using only the quantities measured at $z = -1000$ dbar. As mentioned in section 2a(4), the eddy transport in the upper 1000 dbar $T$ can be decomposed into two components, $T_\perp$ and $T_\parallel$, perpendicular and parallel to the density contours on the 1000 dbar surface, respectively.

2) GENT–MCWILLIAMS PARAMETERIZATION

In their section 10, McDougall and McIntosh (2001) emphasize that the Gent and McWilliams (1990) scheme parameterizes the quasi-Stokes velocity $U_+^i$ in Eq. (8) and that the resolved-scale velocity should be interpreted as an Eulerian-mean velocity $\overline{U}$. The Gent and McWilliams (1990) parameterization is

$$V_+^i = \partial_z \kappa \left( \frac{\Psi^i}{\Psi^i} \frac{\Psi^i}{\Psi^i} \right),$$

where $\kappa$ is the quasi-Stokes diffusivity coefficient. This velocity has only the $V_+^z$ component.

With the boundary condition $\kappa = 0$ at $z = 0$, the quasi-Stokes diffusivity $\kappa$ at 1000 dbar can be estimated from the data using (13):

$$\kappa = -T_{\perp} \frac{\Psi^i}{\sqrt{\frac{\Psi^i}{\Psi^i} + \frac{\Psi^i}{\Psi^i}}} = \left[ \Psi^i(z = -1000) - \Psi^i(z = -1000) \right] \cdot \left( \frac{\Psi^i}{\Psi^i} \frac{\Psi^i}{\Psi^i} \right) \left( \frac{\Psi^i}{\Psi^i} + \frac{\Psi^i}{\Psi^i} \right). \quad (14)$$

Note that the eddy transport $T$ can have $T_{\perp}$ and $T_{\parallel}$ components so that projection to the $\perp$ direction is needed. The other component $T_{\parallel}$ includes the rotational eddy flux (Marshall and Shutts 1981), part of the eddy flux not parameterized by the Gent–McWilliams scheme, and errors in the estimation.

3. Results

a. Eddy kinetic energy

Figure 4 shows the eddy strength at 1000 dbar in terms of the eddy kinetic energy. Strong eddies were found around the equator, in the western boundary currents, and along the ACC. Eddies were also strong in the regions where the ACC and western boundary currents interact (Agulhas retroflection region, Brazil–Malvinas confluence zone). The eastern sides of the basins were relatively quiet in the Pacific and Atlantic Oceans but not in the south Indian Ocean, probably because of the vigorous eddies originating from the Leeuwin Current (Feng et al. 2005).

The eddy kinetic energies at the surface derived from satellite altimeters (e.g., Scharfenberg and Stammer 2010, their Fig. 5) and at 1000 dbar showed some similarities, such as regions of strong eddies. A close comparison, however, revealed that the strong eddies in the western boundary currents were more confined to the western boundary at the 1000-dbar depth than at the surface. The eddies along the ACC, in contrast, showed similar distributions at the surface and 1000 dbar. This feature reflects the contrast between the weakly stratified Southern Ocean and other basins. Chiswell (2013) has reported that the ratio of deep (1000 dbar) to surface eddy kinetic energy is 0.1 to 0.4 in the Southern Ocean, whereas the ratio is much smaller (0.001 to 0.05) in other basins.

b. Isopycnal displacements (heave)

Whereas the eddy kinetic energy measures velocity perturbation by eddies, eddy potential energy measures density perturbation by eddies:

$$N^2 \frac{\sigma}{2}, \quad (15)$$

where $N^2$ is the buoyancy frequency, and $z_h$ is the vertical displacement of isopycnals or “heave.” Just the heave part of the eddy potential energy is plotted in Fig. 5. The greatest difference from the distribution of eddy kinetic energy (Fig. 4) was the lack of signal in the equatorial regions—a reflection of the shallow (<1000 dbar) and sharp thermocline there.

Large heave (>100 m) is found along the Gulf Stream and ACC. In the ACC region, a meridional section (e.g., Rintoul and Bullister 1999, their Fig. 2c) shows an isopycnal depth change of 200 m over a distance of 130 km across a zonal front at about 1000 dbar. Because a front can shift meridionally more than 2° in latitude in a few months (e.g., Sallée et al. 2008, their Fig. 8a), heaving by more than 200 m is possible in the ACC region. Similarly, there was a jump of an isopycnal across the Gulf Stream of greater than 400 m near 36°N (Koltermann et al. 2011, their section A03).

A close look at the ACC region indicated that large heaves occurred along or on the downstream (i.e., eastern) side of topographic obstacles, such as along the Southwest Indian Ridge (along 50°S between 0° and 30°E) and downstream of the Kerguelen Plateau (near 80°E), Macquarie Ridge (170°E), the Pacific–Antarctic Ridge (along 60°S between 150° and 120°W), and the Drake Passage (50°W). The distribution suggests a link between large heaving and the interaction of ACC fronts with bottom topography, as pointed out, for example, by Abernathey and Cessi (2014).

c. Eddy transport

From this section onward, we discuss eddy transport integrated from the sea surface to the isopycnal whose
average height was 1000 dbar. This transport, Eq. (13), consists of two terms [see Eq. (10)]. We note that the second term is negligibly small compared to the first term—the ratio being mostly $O(10^{-2})$—because the geostrophic shear at 1000 dbar is small. The following discussion is therefore accurate if only the first term is retained, although the second term was not neglected. McDougall (1998, section 11) provides a more formal evaluation of the smallness of the second term.

Figure 6 shows eddy transports larger than the measurement uncertainties. Large (>1.0 m$^2$s$^{-1}$) eddy transport was found along the Gulf Stream, the Kuroshio, and the ACC. These large transports often followed the density contours, although there were occasionally vectors across the contours.

In the midlatitudes oceans, in both Southern and Northern Hemispheres, except for the Gulf Stream and Kuroshio regions, eddy transport was weak (<1.0 m$^2$s$^{-1}$) and mostly perpendicular to the density contour (i.e., of the $\perp$ component), the suggestion being that eddies were generated through baroclinic instability. Figure 7a shows the $\perp$ component of eddy transport $T_\perp$. In the Southern Hemisphere, the eddy transport normal to the density contours $T_\perp$ was generally positive or upgradient in density except for a negative band along the northern edge of the ACC. Along the Northern Hemisphere western boundary currents, the positive band became negative downstream, roughly after the separation of the western boundary currents from the western boundary. Projecting the eddy transport onto the meridional direction (Fig. 7b) revealed that eddy transport along the Northern Hemisphere western boundary currents was mostly southward. The southward shoaling of the isopycnals in the Southern Ocean (Fig. 3, bottom) caused the patches of positive transport in the ACC in Fig. 7a to be associated with southward eddy transport.

In the Southern Hemisphere, large northward transport was found in midlatitudes in the Indian Ocean, larger near the Australian and African coasts. This transport was likely associated with the subduction and northward export of Antarctic Intermediate Water (about 27–27.3$\sigma_o$) and will be discussed further in section 4c.

The eddy transport of water can be compared qualitatively with previous estimates of eddy temperature transport because most of the temperature transport is carried by the upper warm water shallower than the thermocline. Figure 8a of Jayne and Marotzke (2002) and Fig. 1 of Aoki et al. (2013) show similarities with Fig. 7b with respect to southward eddy transport in the ACC, particularly the strong transport downstream of the Kerguelen Plateau and south of Africa. The southward transport in the Northern Hemisphere western boundary currents reported by Jayne and Marotzke (2002) is more confined in space and weaker than the analogous transport reported by Aoki et al. (2013). Both simulations show a strong northward transport along the coast; such transport was not found in the Argo data, which showed just a hint of northward transport in a few
grid points at about 40°N. This discrepancy might reflect the effect of spatial averaging, as discussed in McDougall (1998). The numerical simulations lack conspicuous northward eddy transport in the Southern Hemisphere Indian Ocean found in the Argo data.

The meridional projection of the $T_\perp$ transport can be integrated zonally (Fig. 7c). The integrated eddy transport showed statistically significant equatorward transport in the midlatitude (equatorward of 40°) of up to 6 Sv (1 Sv = 10⁶ m² s⁻¹). Eddy transport in the Southern Ocean was 5 to 12 Sv with relatively large uncertainty. The simulated zonally integrated eddy heat transports reported by Jayne and Marotzke (2002) and Aoki et al. (2013) are consistent in sign with our Fig. 7c. Northward transport was found only in high latitudes (>40°N) in the Northern Hemisphere and low to midlatitudes (<40°S) in the Southern Hemisphere.

Under the geostrophic balance $\rho_0 f V'' = \partial_x p'$, the zonal form stress $p' \partial_x z''_d$ can be converted to the meridional eddy transport equation [Eq. (13)], provided that the second term of the quasi-Stokes velocity equation [Eq. (10)] is negligible:

$$\rho_0 f V'' z''_d = (\partial_x p') z''_d = \partial_x (p' z''_d) - p' \partial_x z''_d. \tag{16}$$

For example, the zonal eddy stress estimated by Ferreira et al. (2005) can be compared to our results. They do not show the horizontal distribution at 1000 dbar, but zonal-mean eddy stress (their Fig. 8) at 1000 dbar shows a positive (southward eddy transport) core at latitudes of

![Fig. 7. (a) Eddy transport in the upper 1000 dbar perpendicular to the density contours $T_\perp = \int_{1000}^{0} V' dz$. (b) meridional eddy transport in the upper 1000 dbar, and (c) zonal integration of the meridional eddy transport. In (a), the positive direction is (ρ, ρ), that is, positive toward higher density. In (b) and (c), positive is northward. Only data larger than the measurement uncertainty Δh x Δt are shown. The black and white contours show where mean flow speed is 0.025 and 0.09 m s⁻¹, respectively. The thin gray contours are 3000-m isobaths. Although eddy transport with a magnitude less than the measurement uncertainty was not plotted in (b), this small transport was included in the zonal integration in (c). Setting this small transport to zero alters the meridional transport by less than 1.1 Sv.](image-url)
50° to 60°S and weakly negative (northward) eddy transport equatorward of 30°S, southward eddy transport in the Northern Hemisphere equatorward of 30°N, and northward eddy transport at higher latitudes. Their smallest contour of 0.03 N m⁻² is equal to 0.4 m² s⁻¹ at 30° latitude.

Phillips and Rintoul (2000) have reported the velocity and temperature time series near 50.5°S, 143°E from a set of four moorings. Using the Coriolis parameter at 50.5°S, we determined the interfacial form stress at 1150 dbar of 0.51 N m⁻² to be equivalent to an integrated eddy transport of 4.42 m² s⁻¹. This estimate is consistent with the present estimate at 51°S, 143°E of 2.9 m² s⁻¹ with a 95% confidence interval between 1.4 and 4.7 m² s⁻¹ (toward the south-southwest or −101°).

d. Quasi-Stokes diffusivity

Eddy transport (Fig. 7a) is often explained as a result of mass transport caused by baroclinic instability in a direction that flattens sloping isopycnals. This explanation can be tested by plotting the quasi-Stokes diffusivity or the Gent–McWilliams parameter, κ in Eq. (14), which was estimated from the data using Eq. (14). Figure 8 shows the result for the $T_L$ component, which was similar to the meridional component that can be estimated by dividing the meridional eddy transport (Fig. 7b) by the meridional isopycnal slope (Fig. 3, bottom; not shown). When κ is positive, the eddy flux is in a direction that flattens sloping isopycnals, with the result being the release of large-scale potential energy, which in turn fuels the eddies. In the midlatitude Southern Hemisphere oceans (20° to 40°S), the fact that κ was positive is consistent with the baroclinic instability mechanism.

There were also patches of positive κ in the Southern Ocean, corresponding to the southward eddy transport on Fig. 7b. In contrast, the patches of negative κ found in the Northern Hemisphere Subarctic Gyres and along the northern end of the ACC cannot be explained by baroclinic instability. We note here that although the order of magnitude of the quasi-Stokes diffusivity (1000 m² s⁻¹) agrees with other estimates (L i u et al. 2012, their Fig. 9; Ferreira et al. 2005, their Fig. 12), direct comparisons cannot be made because of the different treatments of the rotational component of eddy transport.

In contrast to the zonally integrated meridional transport (Fig. 7c), the division of eddy transport by small isopycnal slopes amplified the uncertainty in the diffusivity so that the zonal average (Fig. 8b) showed a much larger uncertainty, and the sign of the estimated zonal mean diffusivity was statistically indeterminate except for a small band around 30°S.

4. Discussion

a. Thickness and velocity perturbations

In this section, we discuss the mechanism of the eddy transport by separating the roles of velocity perturbation and isopycnal perturbation. The first term of the eddy transport equation [Eq. (10)] can be rewritten as $\nabla^2 \Phi = -\nabla^2 h$, where h is the upper-layer thickness. Under a stable stratification, we note that the upgradient (low to high) direction across the density contours on 1000 dbar is the direction into which the isopycnals become shallower or downgradient (thick to thin) in the layer thickness h. As defined in section 2a(4), positive
\( V' = V'_0 > 0 \) is upgradient in density and downgradient in upper-layer thickness. When a positive velocity anomaly \( V' > 0 \) is accompanied by a thicker upper layer \( h' > 0 \), the anomaly in the upper-layer transport is positive \( \overline{V'h'} > 0 \). A positive transport anomaly can also be a result of negative velocity anomaly \( V' < 0 \) and negative thickness anomaly \( h' < 0 \). In other words, velocity anomaly and thickness anomaly is positively correlated when \( \overline{V'h'} > 0 \). Figure 9 is an attempt to interpret the eddy transport by this correlation between \( V' \) and \( h' \); when a positive eddy transport \( \overline{V'h'} > 0 \) is found on a grid, the transport is either from the first quadrant \((V' > 0, h' > 0)\) or the third quadrant \((V' < 0, h' < 0)\) of Fig. 9c. Suppose that there is a large eddy with a transport of \( V'h' = 1.0 \text{ m}^2\text{s}^{-1} \) from the third quadrant and that there are 10 weaker eddies from the first quadrant, with each eddy transporting \( V'h' = 0.2 \text{ m}^2\text{s}^{-1} \). One could argue that the third quadrant contributes most to the transport because it is associated with the strongest eddy, but it also seems valid to argue that summed contribution to the transport is largest from the first quadrant. To avoid this complication, only those grids where the quadrant that has the largest \( \Sigma(V'h') \) as well as the largest average eddy strength \( \Sigma(V'h')/N_q \) (the sum is taken over the quadrant, and \( N_q \) is the number of samples in the quadrant) are marked in Fig. 9. In most of the colored grids where the eddy transport was larger than the measurement uncertainty, the largest contribution to transport was from a quadrant that has both the largest \( \Sigma(V'h') \) and largest \( \Sigma(V'h')/N_q \).

In the western North Pacific (Fig. 9a) off Japan, most eddy transport was negative (downgradient in \( \gamma \), upgradient in \( h \)) around and north of the Kuroshio Extension (marked by 0.09 m s\(^{-1}\) speed contour), whereas transport was positive (upgradient in \( \gamma \), downgradient in \( h \)) farther south. These transports were all southward (Fig. 7b). West of 170\(^\circ\)E, this southward transport was mostly carried by negative velocity \( (V' < 0) \) with a thicker upper layer \( (h' > 0, \) fourth quadrant) on the northern side of the Kuroshio Extension, but in the south, the thinner upper layer \( (h' < 0) \) was advected in the northward direction \( (V' > 0, \) second quadrant).

In the North Atlantic (Fig. 9b), a patch of positive (northward) transport was found in the northeastern region, and both positive and negative eddy transports were found mainly along the western boundary current. The northeastern patch was a manifestation of northward transport of thicker upper layer (first quadrant), whereas the southward transport on the western side was a combination of thicker southward (first) and thinner northward (third) transports.

In the eastern Pacific and Atlantic sectors of the Southern Ocean (Fig. 9d), there was strong positive (southward, except for the northward transport off the western coast of Africa) transport that was primarily accounted for by thick and positive (first quadrant) eddies. Negative transports on the lee side of the Mid-Atlantic Ridge (between 10\(^\circ\)W and 0\(^\circ\)W) and on the Pacific Antarctic Ridge (around 120\(^\circ\)W) was accounted for generally by positive transport of a thinner upper layer (second quadrant), possibly by topographically trapped stationary eddies.

The situation in the Indian and western Pacific sectors of the Southern Ocean (Fig. 9e) was more complicated. Strong negative transport existed on the northern flank of the Kerguelen Plateau (40\(^\circ\)S to 50\(^\circ\)S, 60\(^\circ\)E to 70\(^\circ\)E) and south of New Zealand (50\(^\circ\)S to 60\(^\circ\)S, 160\(^\circ\)E to 180\(^\circ\)E), where both \( V' < 0 \) and thicker eddies (fourth quadrant) and \( V' > 0 \) and thinner eddies (second) were found. The strong positive transports found downstream of both negative patches were also a combination of \( V' < 0 \) and thinner (third) eddies and \( V' > 0 \) and thicker (first) eddies. A patch of strong transport south of New Zealand (between 49\(^\circ\)S and 53\(^\circ\)S, 150\(^\circ\)E and 170\(^\circ\)E) might well be a trace of cold-core eddies discussed in detail by Cotroneo et al. (2013).

A conspicuous feature in the south Indian Ocean was a large area of positive eddy transport between 20\(^\circ\)S and 40\(^\circ\)S. This transport was mostly carried by positive (southward) flow of thinner upper-layer water (third quadrant).

Except for the south Indian Ocean, it was difficult to derive a simple explanation for the mechanism of eddy transport. The present analysis revealed that eddy transport was a combination of thinner and thicker perturbations. A local analysis, with more emphasis on the behavior of individual eddies (e.g., Cotroneo et al. 2013), might be more useful.

b. Southward eddy transport in the Northern Hemisphere

The southward (downgradient in \( \gamma \), upgradient in \( h \)) eddy transport found in the Northern Hemisphere western boundary currents (Figs. 9a,b) has been detected in numerical simulations (i.e., Yim et al. 2010; Liu et al. 2012). Using output from an eddy-resolving ocean general circulation model, Aoki et al. (2013) have shown that the divergent component of the eddy transport also shows this southward eddy transport; their Fig. 6 demonstrates that warm water packets are advected southward by cold-core eddies detached from the Kuroshio Extension. This explanation is consistent with our results where the region was mostly covered by the fourth quadrant eddies (Fig. 9a), but similar results were not as clear in the Gulf Stream region (Fig. 9b). The fact that there were grids with a relatively thin upper layer \( (h' < 0) \)
Fig. 9. (c) Transport across density contours $V'\tilde{h}$ was categorized into four groups. An upgradient (in density) velocity anomaly $V' = V_0 > 0$ with thicker upper-layer anomaly $\tilde{h} > 0$ is indicated by a cross. Similarly, the other three quadrants are marked by plus signs, diamonds, and squares as shown in (c). (a),(b),(d),(e) The eddy transport in color (same as Fig. 7a) marked with signs indicating the quadrant, which has the greatest contribution. See text for a detailed definition of “contribution.” The black and white thick contours show where the mean flow speed is 0.025 and 0.09 m s$^{-1}$, respectively. The thin black contours are density contours at 31.8, 31.9, \ldots, 32.4$\sigma_T$. The thin gray contours are 3000-m isobaths of ETOPO5.
might reflect a local topographic effect advecting cold-core eddies or the effect of neglected spatial variability (McDougall 1998).

c. Northward eddy transport in the south Indian Ocean

In the south Indian Ocean, Antarctic Intermediate Water subducts at a density of 27–27.3 \( \sigma_T \) (Sallée et al. 2010). On a meridional section of salinity from the WOCE Hydrography Program I8 and I9 sections (nominally 95°E), the low-salinity core of Antarctic Intermediate Water (AAIW) lies almost exactly at 1000 dbar at latitudes of 35° to 45°S. The fact that this region is part of the northwestward flow of the Indian Ocean Subtropical Gyre is consistent with box inverse calculations (e.g., Sloyan and Rintoul 2001, their Fig. 10). Northward movement of “boluses” of intermediate water pushes shallower isopycnals upward. Because the upper-layer flows northward relatively slower than the intermediate water, the eddy transport of the upper layer appears southward (negative) and thinner, that is, third quadrant in Fig. 9e. Another potential AAIW source region is the Chilean coast, but in that region the AAIW core is too shallow (at about 700 dbar) to be observed by Argo floats.

d. Float density

Figure 10 shows the spatial distribution of two components (measurement and sampling) of uncertainty in the eddy transport shown in Fig. 7a. Both components have a similar magnitude but sampling uncertainty is mostly 2 or more times measurement uncertainty. Measurement uncertainty (including array bias) in velocity was uniformly set at 0.5 cm s\(^{-1}\) [section 2a(2)]. The spatial variation in Fig. 10b thus originated from the uncertainty in heaving estimates (Fig. 1).

The larger contribution of the sampling uncertainty means that future technical advancements will not greatly reduce the overall uncertainties in the eddy

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**FIG. 10.** (a) Sampling uncertainty and (b) measurement uncertainty in the eddy transport \( T_\perp = \int_0^{1000} V_z \, dz \) (Fig. 7a). The black and white contours show where the mean flow speed is 0.025 and 0.09 m s\(^{-1}\), respectively. Note the uneven color scale. The thin gray contours are 3000-m isobaths.
statistics measured by Argo floats. As discussed in section 2a(2), the measurement uncertainty in velocity consists of array bias, positioning error, clock drift, and sampling error. Comparison of Figs. 7a and 10a suggests that sampling uncertainty is much larger than on the accuracy/precision of the sensors. Davis (1991) around the topography in the ACC.

We therefore conclude that the present distribution of Argo floats is barely adequate to study local phenomena such as eddy–topography interactions in the ACC, but it can be used to qualitatively describe the global distribution of eddy statistics (e.g., Figs. 6, 7) and to estimate zonally integrated statistics (Fig. 7c). The sampling uncertainty is large particularly where heaving is large (Fig. 5), that is, in the ACC and the Gulf Stream regions. Deployments of floats particularly in the former region is challenging, but the utility of the Argo floats in describing not only the mean field (Katsumata and Yoshinari 2010; Gray and Riser 2014) but also the eddy field encourages efforts to maintain floats in this region.

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