Eddy-driven meridional transport across the ACC upstream of Drake Passage

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The magnitude and vertical (density) distribution of the eddy-driven transport in the Southeastern Pacific sector of the Southern Ocean is inferred from observations and recent estimates of isopycnal diffusivity from the Diapycnal and Isopycnal Mixing Experiments in the Southern Ocean (DIMES) experiment. A vertical profile of isopycnal diffusivity derived from numerical modelling, tracer release and acoustically-tracked isopycnal floats released during DIMES is combined with historical CTD profiles. Away from the surface layer and above the depth of the sill of the Drake Passage, the eddy-driven flow is southward between $\gamma_n = 27.5$ and $\gamma_n = 28.16$ kg m$^{-3}$ before reversing to become northward where the Potential Vorticity gradient reverses. The calculated eddy velocities are small compared to other regions of the Southern Ocean where complex topography promotes instability and eddy transfer. These results lend support to the idea that local dynamics, dictated by flow-topography interactions, strongly influence the magnitude of the tracer transport and of the overturning circulation in the Southern Ocean.
1. Introduction

In the Southern Ocean, the combination of a strong eastward Antarctic Circumpolar Current (ACC) and a cross-ACC meridional circulation has far-reaching consequences for the closure of the Meridional Overturning Circulation (MOC). The role of the MOC in the global climate system has motivated, over recent decades, intense research on various aspects of its dynamics and associated water mass transformation [Rintoul and Garabato, 2013]. A combination of wind and air-sea-ice buoyancy forcing over the SubAntarctic and Antarctic domains drives the ACC and produces sharply rising isopycnals, providing an efficient isopycnal pathway to return deep water to the surface [Marshall and Speer, 2012]. Air-sea-ice interactions lead to water mass transformation of upwelling water where, north of the Antarctic Divergence (∼60 – 65°), it gains buoyancy and is transformed into Antarctic Intermediate Water (AAIW) and Sub-Antarctic Mode Water (SAMW), and continues to circulate within the thermocline. This Upper Cell of the MOC is accompanied by a Lower Cell, in which water flowing close to the Antarctic margin is transformed into a variety of dense waters on the shelf and slope, some of which becomes Antarctic Bottom Water.

Recent ideas about the MOC in the Southern Ocean emphasize the southward eddy-driven flow of deep waters, balancing to some extent the northward Ekman transport in the upper layers, the so-called eddy compensation, e.g. Marshall and Radko [2003]. Quantifying the eddy fluxes remain challenging in both observations and models due to the small spatial scales of the eddies (10-100km), their inhomogeneity and intermittency. Recent studies, for example, have identified eddy fluxes situated on the lee side of to-
ography, where the orientation of ACC fronts departs significantly from its quasi-zonal orientation, as major contributions to the cross-stream transport. In other regions, by contrast, strong ACC fronts can act as a barrier to mixing [Abernathey et al., 2010; Lu and Speer, 2010; Naveira-Garabato et al., 2011; Thompson and Sallée, 2012]. This zonal complexity emphasizes the need to evaluate regional contributions to the circumpolar meridional eddy fluxes.

In this study we focus on a quiet region where the ACC is essentially zonal, the Southeast Pacific sector of the Southern Ocean just upstream of Drake Passage (Fig. 1). Here, we seek to quantify the southward eddy-driven transport. The area comprises a zone of weak eddy-activity between more active regions close to the East Pacific Rise and Drake Passage, where the ACC passes through topographic constrictions. The Diapycnal and Isopycnal Mixing Experiments in the Southern Ocean (DIMES) project measured isopycnal tracer mixing in this region with both a tracer and subsurface floats. The tracer was confined to a narrow range of densities, while the float data was analyzed over a relatively broad range of densities and depths; both approaches required averaging over the entire sector upstream of Drake Passage, from 105 W to Drake Passage. Through a combination of the observed regional-average of eddy-diffusivity coefficient, extrapolated to a vertical profile with a high-resolution numerical model, and in situ hydrographic profiles, we estimate an average meridional eddy flux and residual flow in this area.
2. Data and Methodology

The residual momentum budget of the ACC below the surface, where momentum is input by the wind, and above topographic features that support pressure gradients is:

\[
-\rho_0 f \overline{v z_\gamma} = \rho_0 \overline{z_\gamma^2 v' Q'}.
\]  

\( \overline{() \,} \) denotes time and along-stream spatial averages, primes are departure from those averages, \( Q = (f + \zeta)/z_\gamma \) is the isentropic Ertel potential vorticity, IPV [Vallis, 2006], \( f \) is the Coriolis parameter, \( \zeta \) is the vertical component of the vorticity and \( z_\gamma = \partial z / \partial \gamma \) is the layer thickness. Assuming that the IPV flux is down its mean gradient we have

\[
\rho_0 f \overline{v z_\gamma} = \rho_0 \overline{z_\gamma^2} \kappa_{PV} \partial_y Q
\]  

where \( \kappa_{PV} \) is the IPV eddy diffusivity. If we further neglect the relative vorticity and the meridional gradient in \( f \) in Equation 2, as appropriate in the Southern Ocean [Mazloff et al., 2013], we can relate the meridional mass flux to the product of the IPV diffusivity and the meridional thickness gradient,

\[
\overline{v z_\gamma} = -\kappa_{PV} \partial_y \overline{z_\gamma}
\]  

Recognizing that the residual velocity is defined as \( v^* = \overline{v z_\gamma} / \overline{z_\gamma} \), we can also write

\[
v^* = -\kappa_{PV} \partial_y \overline{z_\gamma} / \overline{z_\gamma}
\]  

A different approach to parameterize the effect of eddies on the residual circulation was suggested by Gent and McWilliams [1990] and relies on the energy cycle of baroclinic
instability. Eddies develop in the Southern Ocean as a result of baroclinic instability of the ACC and extract energy from the mean current. An expression of the residual velocity that is guaranteed to always extract energy from the mean flow was proposed

\[ v^* = -\partial_\gamma (\kappa_{GM} \partial_y \tilde{z} / \tilde{z}_\gamma). \]  \(5\)

The reader is referred to Gent et al. [1995] for a demonstration that with this form, eddies extract energy from the mean flow regardless of the value of \(\kappa_{GM}\) diffusivity.

The IPV eddy diffusivity, \(\kappa_{PV}\), is obtained from observations and high-resolution modeling performed within the DIMES project. The observations came from 140 acoustically-tracked RAFOS floats (shown in Fig. 1, Hancock and Speer [2013]) and 3 years of tracer surveys in the Southeast Pacific and the Scotia Sea [Watson et al., 2014]. Floats [Lacasce et al., 2014] and tracer data [Tulloch et al., 2014] are used to compute the isopycnal diffusivity at the depths where data are available. State of the art numerical simulations with MITgcm are used to ENHANCE(?) the observations. [Lacasce et al., 2014] obtained a vertical diffusivity profile by releasing numerical floats at 20 different depths while [Tulloch et al., 2014] used the simulations to assess the bias caused by undersampling the tracer patch and to obtain a vertical profile of diffusivity. The diffusivity estimated from the dispersion of Lagrangian particles have been shown to match the tracer diffusivity providing that these diffusivities are calculated over sufficiently long periods [Klocker et al., 2012]. Indeed, both tracer-based and float-based diffusivity are in good agreement, giving confidence in the results. In this study, we use vertical profiles of \(\kappa_{PV}\) obtained by Tulloch et al. [2014] and Lacasce et al. [2014]. The eddy-driven meridional transport, \(v^*\), is derived from \(\kappa_{PV}\) using Eq. 4.
We derive the meridional thickness gradient, $\partial_y(z)$ across streamlines by constructing climatological sections of the thermohaline properties and pressure along isopycnals based on 783 historical full-depth CTD profiles between $80^\circ W$ and $110^\circ W$ (corresponding to the DIMES region, Figure 1) extracted from the Southern Ocean Data Base. These are complemented with additional recent profiles from the CLIVAR Repeat Hydrography program (S04P in 2011, P18 in 2008, AAIW05 and AAIW06 occupied in 2005 and 2006 respectively). We use a neutral density \cite{Jackett and McDougall, 1997}-dynamic height coordinate system in order to differentiate true along-isopycnal variability from noise introduced by isopycnal heaving and ACC meridional meandering. Dynamic height ($\Phi$) has been shown to be a useful proxy of the position of the ACC fronts delimiting the location of the ACC \cite{Sun and Watts, 2002}.

The sections are constructed following the approach used by Naveira-Garabato et al. \cite{[2011]}, that is for each CTD profile, the dynamic height at 400 m relative to 3000m, $\Phi^{3000}_{400}$, is calculated. Figure 2a shows that $\Phi^{3000}_{400}$ increases monotonically between $70^\circ S$ and $50^\circ S$ (corresponding to $\Phi^{3000}_{400} = [8 - 18] m^2 s^{-2}$) thus supporting our use of $\Phi^{3000}_{400}$ as a pseudo-geographical coordinate. Potential temperature, $\Theta(\gamma_n)$, pressure, $P(\gamma_n)$, salinity, $S(\gamma_n)$, and oxygen, $O(\gamma_n)$, profiles are mapped on discrete neutral surfaces, $\gamma_n$ at 0.02 kg m$^{-3}$ intervals. Then, we take advantage of the strong zonal coherence of the ACC water masses and make the assumption that in this region, characterized by relatively smooth topography, the along-streamline isopycnal properties are essentially homogeneous. This assumption is supported by figure 2b, which shows that the distribution of salinity along an isopycnal ($\gamma_n = 27.6$ kg m$^{-3}$) as a function of dynamic height is coherent throughout
the region irrespective of zonal location. Therefore, we define a $\Phi_{3000}^4$ grid ranging from 8 to 18 m$^2$ s$^{-2}$ with an increment $\delta\Phi = 0.4$ and for each grid point, $i$, we average together all profiles: $\Theta_{\Phi_i} = \frac{1}{m} \sum_{p=1}^{m} \Theta(\gamma_n)$, where $m$ is the number of profiles with dynamic height $\Phi_i = \Phi_i^i \pm \delta\Phi/2$.

Calculating the cross-stream gradients requires computing the distance between the climatological streamlines. Given the quasi-zonal structure of the ACC in the region, we assume that the climatological section is meridional and we calculate the climatological latitude of each streamline by fitting a cubic spline to the $\Phi_{3000}^4$ vs. latitude. Finally, the thickness gradient across the ACC is computed by fitting a cubic spline over each isopycnal pressure section (Figure 2d) in order to minimize the influence of small scale noise caused by data scarcity and the presence of small-scale features and emphasize the large-scale cross-ACC gradients. In the following section, we average this thickness gradient over the ACC region between the SAF and PF.

3. Estimates of eddy-induced transport

The climatological sections of salinity, oxygen, isopycnal thickness and pressure along-side their corresponding cross-stream gradients as well as the position of the ACC fronts [Orsi et al., 1995] are shown in Figure 3. In the deep layers, the low-oxygen signal of the Upper Circumpolar Deep Water (UCDW) at $\gamma_n = 27.9$ kg m$^{-3}$ overlies Lower Circumpolar Deep Water (LCDW), characterized by a salinity maximum centered around $\gamma_n = 28.1$ kg m$^{-3}$. At shallower levels the low salinity AAIW (above $\gamma_n = 27.5$ kg m$^{-3}$) covers the region. The core of the ACC, between the SAF and the PF, is characterized by relatively large isopycnal gradients. The salinity (oxygen) gradient (Figure 3d,e) is largely positive
(negative) and maximum above $\gamma_n = 28.00 \text{ kg m}^{-3}$. The thickness gradient (Figure 3f),
while relatively noisy, is predominantly positive, indicative of a net southward residual
flow above $\gamma_n = 28.10 \text{ kg m}^{-3}$, before reversing in the deeper water masses, suggesting a
deep northward residual velocity.

We note that, below about $\gamma_n = 28.2 \text{ kg m}^{-3}$ (corresponding to the sill depth of Drake
Passage), isopycnals intercept topography, indicating that below this depth, topographic
form stress cannot be neglected. In addition, we note that above $\gamma_n = 27.3 \text{ kg m}^{-3}$,
isopycnals are located within the top 500 m of the water column where other dynamical processes (e.g. Ekman transport, outcropping isopycnals and associated geostrophic
flow) will dominate the meridional flow. Therefore, we limit the calculation of $v^\ast$ to the
UCDW-LCDW density range ($27.5 < \gamma_n < 28.2 \text{ kg m}^{-3}$) where eddies are expected to
be the dominant contributor to the meridional transport according to the simple balance
described above.

Figure 4 shows the vertical profiles of $\kappa_{PV}$ from Lacasce et al. [2014] (solid black line)
and Tulloch et al. [2014] (dotted black line), $v^\ast$ (light gray) calculated using Equation 4,
$\kappa_{GM}$ (dark gray) calculated by numerically solving 5, salinity (red) and oxygen (brown)
against density (right panel) and depth (left panel) in the Southeastern sector of the
Southern Ocean (between $55^\circ$ and $60^\circ$S). $\kappa_{PV}$ reaches a maximum be 900 m$^2$ s$^{-1}$ between
$\gamma_n$ of 28.00 and 28.1 kg m$^{-3}$ (2000-3000m) corresponding to the salinity maximum of
LCDW. Across the oxygen minimum UCDW layer, the float-derived diffusivity quickly
decreases to about 500 m$^2$ s$^{-1}$ at 27.8 kg m$^{-3}$ while the tracer-based diffusivity gradually
decreases to reach the same value at $\gamma_n = 27.65 \text{ kg m}^{-3}$.
The corresponding eddy-driven velocity in the UCDW and LCDW layers ($\gamma_n = 27.4 - 28.10 \text{ kg m}^{-3}$), is of the order of $-1 \times 10^{-4} \text{ m s}^{-1}$ southward before changing sign (at the depth where the PV gradient also reverses) to reach $3 \times 10^{-4} \text{ m s}^{-1}$ at $\gamma_n = 28.16 \text{ kg m}^{-3}$ just below the salinity maximum. The eddy-driven velocity calculated from the tracer-based diffusivity differs by up to 40% in the UCDW density range, around $\gamma_n = 27.8 \text{ kg m}^{-3}$ corresponding to the isopycnal range where the two diffusivities differ the most. We estimate the uncertainty related to the spline fitting by varying the smoothing coefficient used. The impact on $v^*$ is less than 15% except for one isopycnal ($\gamma_n = 28.02 \text{ kg m}^{-3}$) where the error reaches 40%. This is due to the relatively weak thickness gradient (not shown) making it more sensitive to the choice of the fit. The similitude of $v^*$ for different values of $\kappa$ ($v^* \sim -1 \times 10^{-4} \text{ m s}^{-1}$ for $\kappa \sim 500 \text{ m}^2 \text{ s}^{-1}$ near $\gamma_n = 27.7 \text{ kg m}^{-3}$ and for $\kappa = 900 \text{ m}^2 \text{ s}^{-1}$ at $\gamma_n = 28.0 \text{ kg m}^{-3}$) reflects the overarching importance of the cross-stream thickness gradient (hence PV gradient) in setting the magnitude of the eddy fluxes.

The $\kappa_{GM}$ varies significantly from the float-derived $\kappa_{PV}$ both in its vertical structure and intensity (Figure 4). While $\kappa_{PV}$ peaks at $\sim 800 \text{ m}^2 \text{ s}^{-1}$ around $\gamma_n = 28.00 - 28.1 \text{ kg m}^{-3}$, $\kappa_{GM}$ is nearly vertically uniform, with a mean value of $\sim 460 \text{ m}^2 \text{ s}^{-1}$. Both estimates of $kappa_{GM}$ (from floats and tracers) are very close.

4. Discussion

Extrapolated over the whole Southern Ocean ($\sim 20000 \text{ km along } 60^\circ$), the eddy-induced transports found here of UCDW ($\gamma_n = [27.5 - 28.00] \text{ kg m}^{-3}$) and LCDW ($\gamma_n = [28.00 - 28.20] \text{ kg m}^{-3}$) equate to 1.6 Sv and 3 Sv respectively based on the float-derived diffusivity.
and 3.1 and 3.0 Sv with the tracer-derived one. These transport estimates, based on the
vertical profile of $v^*$ in the SE Pacific, are expected to be weak compared with other
sectors of the Southern Ocean. Speer et al. [2000] applying similar methodology using
hydrographic data south of Australia found values of $v^*$ roughly one order of magnitude
larger than in this study.

Assuming the eddy-driven upwelling associated with the meridional residual velocity is
predominantly oriented along isopycnals, the upwelling rate, $w^*$, can be estimated from
the slope of the isopycnals across the ACC defined as $s_\gamma = \Delta z_\gamma / \Delta y = w^* / v^*$, where
$\Delta z_\gamma$ is the difference in the isopycnal depth between the northern and southern side of
the ACC ($\sim 1000$ m within the UCDW layer) and $\Delta y$ is distance between northern and
southern limit of the ACC ($\sim 860$ km). In the UCDW layer, the average $v^* \sim -1.1 \times 10^{-4}$ m
s$^{-1}$ so that we find the rate of isopycnal upwelling $w^* \sim 110^{-7}$ m s$^{-1}$ (4 m yr$^{-1}$). Naveira-
Garabato et al. [2007] found an eddy-driven isopycnal upwelling of $330 \pm 110$ m yr$^{-1}$ ($110^{-5}$
m s$^{-1}$) in the Drake Passage/Scotia Sea region based on the cross-stream spreading rate
of mantle helium injected in the ACC near Drake Passage within the UCDW density
range. Our estimate of isopycnal upwelling just upstream of Drake Passage is about two
orders of magnitude lower than the value found by Naveira-Garabato et al. [2007]. One
possible explanation for the lowest upwelling rate might be the flatter isopycnal slope in
the SE Pacific sector compared with Drake Passage (the 28.00 isopycnal rises from 900 m
and $\sim 1500$ m respectively).

Our eddy-induced transports are significantly smaller than the $\sim 15$ Sv of deep water
geostrophically flowing southward across $32^\circ$S based on large-scale inverse models [Sloyan
and Rintoul, 2001; Lumpkin and Speer, 2007; Naveira-Garabato et al., 2013] suggesting that the SE Pacific sector of the Southern Ocean plays a relatively minor role in the ageostrophic meridional overturning. Describing the dynamical processes setting such small values of eddy transport is beyond the scope of this short paper; however, there is a growing body of evidence suggesting that zonal modulations in the ACC characteristics play a significant role in setting the strength of the meridional eddy transport (e.g. Abernathey et al. [2010]; Lu and Speer [2010]; Thompson et al. [2010]; Naveira-Garabato et al. [2011]; Sallée et al. [2011]; Thompson and Sallée [2012]. The emerging view points towards the existence of mixing "hotspots", where local interactions between the ACC fronts and the topography enhance cross-frontal mixing, separated by more quiescent regions where the ACC fronts act as mixing barriers. Thompson et al. [2010] and Thompson and Sallée [2012] based on modelling and altimetry data also support the view that the SE Pacific is a region of little cross-frontal mixing due to the absence of prominent topographical obstacle in the path of the ACC. Abernathey et al. [2010], using a Southern ocean state estimate also found reduced isopycnal mixing in the 60°W-120°W region.

Obtaining a circumpolarly integrated estimate of the eddy-driven tracer flux remains a key step toward closing the MOC in the Southern Ocean. However, there is no compelling way to extrapolate the measurements without accounting for the horizontal distribution of the PV (thickness) gradient and the zonal variations in the vertical profiles of $\kappa_{PV}$. Here, we briefly discuss the impact of variations in the PV gradient for different regions of the Southern Ocean on $v^*$. Although observations and models suggest that $\kappa_{PV}$ varies circumpolarly [Abernathey et al., 2010; Lu and Speer, 2010; Naveira-Garabato et al., 2011],
we apply a similar methodology by combining climatological sections constructed from
the Southern Ocean Data Base in the Drake Passage (Sr1b section), 30°E (corresponding
to the WOCE I6S section) and 140°E (corresponding to the WOCE SR3 line) using the
vertical profile of $\kappa_{PV}$ characteristics of the SE Pacific (Figure 4). These "back of the
envelope" estimations (not shown) reveal significant zonal variations in the intensity and
vertical structure of $v^*$ with the Drake Passage hosting the strongest residual velocity
$(-1 \times 10^{-3} \text{ m s}^{-1} \text{ just above } \gamma_n = 28.00)$ while south of Africa and Australia, the residual
velocities peak at $-4 \times 10^{-4} \text{ m s}^{-1} \text{ just above } \gamma_n = 28.00$, reflecting significant zonal
differences in the layer thickness gradient (and thus the large-scale PV gradient).

Numerical simulations have shown significant differences both in the vertical structure
and magnitude of the eddy PV and eddy buoyancy diffusivities [Treguier, 1999; Smith and
Marshall, 2009; Abernathey et al., 2013] suggesting that isopycnal diffusivity is different
from the diffusivity to be used as part of Gent-McWilliam parameterization (a difference
already noted in Gent et al. [1995]). [Smith and Marshall, 2009] provided an equation
(their equation 4.3) relating both diffusivities. Abernathey et al. [2013] suggested that it
was possible to obtain an estimate of $\kappa_{GM}$ from observations by assuming that the $v^*$
from the PV parameterization (equation 4) should be the same as the $v^*$ from the GM
parametrization (equation 5). This is the same as the ode presented in equation 4.3 from
[Smith and Marshall, 2009]. Here we follow this approach and solve the ode using the
boundary condition that the $\kappa_{GM} = \kappa_{PV}$ at 500m (corresponding to $\gamma_n = 27.5 \text{ kg m}^{-3}$).
This choice of boundary condition will impact on the magnitude of $\kappa_{GM}$ but numerical
simulations seem to suggest that in the ACC, near 500m $\sim k_{GM} \sim k_{PV}$ (c.f. for example Figure 12 of Smith and Marshall [2009] or Figure 13 of Abernathey et al. [2013]).

The profile of $k_{GM}$ obtained has a different vertical structure than the $k_{PV}$ as shown in Figure 4. $k_{GM}$ has no mid-depth maxima and just slightly increases with depth. This structure is the same as the one obtained by Smith and Marshall [2009] and Abernathey et al. [2013] from numerical simulations but the magnitudes are different (2 to 4 times smaller in our study). This is to be expected as Smith and Marshall [2009] used a QG model with mean hydrography based on a climatology of the Australian sector of the Southern Ocean and Abernathey et al. [2013] used an idealized primitive equation channel model, whereas our results are for the Southeastern Pacific sector of the Southern Ocean.

This study provides evidence inferred from observations supporting the difference in the vertical structure of $k_{GM}$ and $k_{PV}$. This profile of buoyancy diffusivity could be directly used in a numerical model of the region in order to capture both the slumping of isopycnals (characteristic of the GM parameterization) as well as providing the correct (within observational uncertainty) the residual velocity across the ACC.

5. Conclusion

We evaluate the strength of the eddy-driven fluxes in the SE Pacific sector of the Southern Ocean by combining a vertical profile of eddy-diffusivity derived from Lacasce et al. [2014] with climatological hydographic sections constructed from historical CTD profiles. Large meridional eddy fluxes were not expected in this region and the calculated eddy-driven velocity $v^*$ does appear to be small when compared with previous, indirect esti-
mates from inverse calculations [Speer et al., 2000] and natural tracer release experiment [Naveira-Garabato et al., 2007].

In addition, both thickness gradients and the vertical structure of $\kappa$ vary significantly around the Southern Ocean [Lu and Speer, 2010; Naveira-Garabato et al., 2011] and extrapolating eddy flux on the basis of either diffusivity or thickness gradient separately is not likely to be accurate since both are likely to play a significant role in the magnitude of the eddy-driven tracer fluxes and the density range over which these apply.

The relatively (compared with previous studies focusing on other sectors of the Southern Ocean) weak residual velocity found in the SE Pacific sector of the Southern Ocean lend support to the idea that other regions, where local dynamics dictated by flow-topography interactions strongly enhances the magnitude of the southward tracer transport, must play a disproportionately large role in order to close the Southern Ocean MOC. This variability needs to be incorporated in models of the Southern Ocean overturning circulation to properly reflect meridional physical and chemical transports.

Acknowledgments. LJ acknowledges financial support from NSF (OCE-1231803) and from the European Union via a Marie Curie fellowship (FP7-PEOPLE-IEF-2012 no 328416). We thank all the crew, scientists and officers who participated to the collection of the data used in this study, in particular during the DIMES cruises on board RRS James Clark Ross, RRS James Cook, RV Revelle, RV Thompson.
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Figure 1. Bathymetry (ETOPO) of the South East Pacific/Drake Passage. The trajectories of the DIMES floats are shown in green. The CTD profiles used in this study are shown as blue dots and the climatological positions [Orsi et al., 1995] of the SubAntarctic Front (SAF), Polar Front (PF) and Southern Boundary of the ACC (SnBdy) are shown in red.
Figure 2. Scatter plot of a) the CTD stations dynamic height, $\Phi$ (at 400m referenced to 3000m) versus latitude; b) salinity on $\gamma_n = 27.6$ kg m$^{-3}$ and c) thickness of the layer centered around 27.6 kg m$^{-3}$ versus latitude. In each panel, the thick line is the spline fit.
Figure 3. Climatological sections in gamma/latitude coordinates constructed from CTD profiles in the Southeast Pacific (80° – 110°W) of a) salinity, b) oxygen, c) layer thickness, d) ∇yS, e) ∇yO₂, ∇yH. The thin black lines and white line represent the 150 500 1000 2000 4000 isobars and the shallowest density interfering with the topography at each streamline respectively. The vertical dashed lines correspond to the position of the SAF (SubAntarctic Front), PF (Polar Front), and SACCF (Southern ACC Front).
Figure 4. Vertical profiles of eddy diffusivity, GM parameterization (dashed dark gray) and $v^*$ (light gray) based on the DIMES floats (solid lines) [Lacasce et al., 2014] and the tracer spreading [Tulloch et al., 2014] (dotted lines) in density coordinates (left panel) and depth coordinates (right) averaged over the 55° − 60°. Profiles of oxygen (brown) and salinity (red) are shown to relate eddy terms to the water mass structure. The zero crossing for $v^*$ and neutral density boundaries of the water masses (AAIW = $\gamma_n = [27 - 27.5]$; UCDW = $\gamma_n = [27.5 - 28]$; LCDW = $\gamma_n = [28 - 28.2]$) are shown as vertical and horizontal dotted lines respectively.