Eddy stirring in the Southern Ocean
A. C. Naveira Garabato¹, R. Ferrari² and K. L. Polzin³

Abstract. There is an ongoing debate concerning the distribution of eddy stirring across the Antarctic Circumpolar Current (ACC) and the nature of its controlling processes. The problem is addressed here by estimating the isentropic eddy diffusivity $\kappa$ from a collection of hydrographic and altimetric observations, analyzed in a mixing length theoretical framework. It is shown that, typically, $\kappa$ is suppressed by an order of magnitude in the upper kilometre of the ACC frontal jets relative to their surroundings. This observation is reproduced by a quasi-geostrophic theory of eddy stirring across a broad barotropic jet based on the scaling law derived by Ferrari and Nikurashin (2010). The theory interprets the observed widespread reduction of $\kappa$ in the upper layers of frontal jets as the kinematic consequence of eddy propagation relative to the mean flow within jet cores. Pronounced deviations from the prevalent regime of mixing suppression in the core of upper-ocean jets are encountered in a few special sites. Such ‘leaky jet’ segments appear to be associated with sharp stationary meanders of the mean flow that exhibit non-parallel structure on length scales comparable to those of the eddies, and that are generated by the interaction of the ACC with major topographic features. It is contended that the characteristic thermohaline structure of the Southern Ocean, consisting of multiple upper-ocean thermohaline fronts separated and underlaid by regions of homogenized properties, is largely a result of the widespread suppression of eddy stirring and mixing by parallel jets.

1. Introduction

The Southern Ocean plays a pivotal role in the global overturning circulation. The absence of continental barriers in the latitude band of Drake Passage permits the existence of the Antarctic Circumpolar Current (ACC), which is supported geostrophically by sloping isopycnals and serves as the main conduit for oceanic exchanges between the three major ocean basins. Coupled to this eastward, multi-jet flow, a meridional circulation exists in which Circumpolar Deep Water (CDW) with primordial sources in the North Atlantic, upwells along the poleward-sloping isopycnals of the ACC and is returned equatorward in a double overturning cell. This pattern of overturning can be surmised from a meridional section of almost any hydrographic or geochemical property across the Southern Ocean, such as the section of salinity shown in Figure 1a. The Upper classes of CDW are entrained into the upper-ocean mixed layer within the ACC itself and subsequently flow northward as Antarctic Surface Water, which subducts into Antarctic Intermediate Water and Subantarctic Mode Water near the current’s northern rim (the upper cell). The Lower classes of CDW are transported southward beyond the ACC and enter the system of subpolar cyclonic gyres and westward-flowing slope jets encircling the Antarctic continent. There, CDW replenishes and mixes with Antarctic surface and shelf waters, ultimately resulting in the formation and northward export of Antarctic Bottom Water (the lower cell). This meridional circulation is a key step in the vertical transfer of water masses and physical tracers required to close global overturning, and underlies the Southern Ocean’s disproportionate importance in the ventilation of the deep ocean. Rintoul et al. [2001] and Olbers et al. [2004] provide reviews of the contemporary state of knowledge on the Southern Ocean circulation, its function in the global ocean, and its essential dynamics. Beyond the above qualitative description, large quantitative uncertainties remain on the rate and structure of the Southern Ocean overturning. Conventional methods for diagnosing the regional circulation from hydrographic and velocity measurements encounter formidable challenges in the vast zonal extent, high mesoscale variability and vigorous water mass transformations that characterize the Southern Ocean. At subtropical and subpolar latitudes, these difficulties have translated into uncertainties in the overturning rate of at least a factor of 2 and significant discrepancies in the circulation’s vertical structure (e.g. Sloyan and Rintoul, 2001; Ganachaud, 2005; Lumpkin and Speer, 2007), while the overturning across the ACC band is even more uncertain. Motivated in part by the limitations of existing observation-based analyses, a large body of research [e.g. Marshall, 1997; Speer et al., 2000; Karsten and Marshall, 2002; Bryden and Cunningham, 2003; Marshall and Radko, 2003, 2006; Olbers and Visbeck, 2005] has focussed on investigating the character and driving mechanisms of the Southern Ocean overturning from a theoretical perspective. A fundamental result of those studies is that the Southern Ocean overturning may be described as the residual arising from the partial cancellation between two distinct circulations: a wind-forced Ekman overturning acting to tilt ACC isopycnals upward, and an overturning in the opposite sense induced by geostrophic eddies arising from baroclinic instability and acting to slump isopycnals. The wind-driven Ekman overturning consists of upwelling (downwelling) to the south (north) of the ACC’s axis, northward flow in a sur- face Ekman layer, and a return southward geostrophic flow below the crests of the topographic barriers in the ACC’s path. In this way, it contributes importantly to the lower cell of the Southern Ocean overturning, as well as to the northward limb of the overturning’s upper cell. The influence of the eddy-induced circulation is thought to be particularly significant above the level of the topography (roughly

¹National Oceanography Centre, Southampton, U.K.
²Massachusetts Institute of Technology, Cambridge, U.S.A.
³Woods Hole Oceanographic Institution, Woods Hole, U.S.A.

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in the top \( \sim 2500 \) m), where it sustains the upwelling of Upper CDW and promotes intermediate water subduction. Water mass conversion through diapycnal flow is often assumed to be largely confined to the upper ocean, where it is supported by air-sea buoyancy exchanges and eddy-induced diapycnal fluxes, although the extent to which this assumption holds has been questioned [e.g. Garabato et al., 2004]. The central difficulty in translating this theoretical picture of the Southern Ocean overturning’s dynamics into an accurate diagnostic of the circulation’s rate and structure pertains to the eddy-induced contribution to the overturning. In contrast to the wind-driven Ekman circulation, which can be estimated from gross knowledge of the wind stress and the topography, the eddy-induced overturning arises from transient eddy fluctuations on scales of \( O(100 \) km) that are difficult to quantify in observations. A typical approach is to relate the eddy-induced circulation to a downgradient flux of potential vorticity (PV) resulting from eddy stirring, and to parameterize this flux with a flux-gradient relation [e.g. Rhines and Young, 1982]. In terms of a generalized tracer \( \kappa \),

\[
\vec{v} \cdot \vec{c} = -\kappa \frac{\partial c}{\partial y},
\]

where the overbars indicate zonal (along-stream) and temporal averages on an isopycnal layer, the primes indicate deviations from those averages, the tracer gradient is assumed to be meridional (cross-stream), and \( \kappa \) is an isopycnal (strictly, isentropic, but taken here to be equivalent) eddy diffusivity characterizing the rate of eddy stirring. We note that the most important quantity in both dynamical and kinematic terms is the eddy tracer flux divergence \( \partial_\beta (\vec{v} \cdot \vec{c}) \), such that the spatial distribution of \( \kappa \) relative to the background tracer gradient assumes a key role. In the context of the Southern Ocean circulation, it is the relative spatial variability of \( \kappa \) and the large-scale isopycnal PV gradient that sets the rate and structure of the eddy-induced contribution to the overturning.

In the virtual absence of direct observations of eddy stirring in the Southern Ocean, attempts at determining \( \kappa \) have resorted to a wide range of approaches and yielded diverse results. The methods, assumptions and outcomes of these past studies are discussed in detail in section 4, against the backdrop of our own results. For now, it suffices to say that previous studies of eddy stirring in the Southern Ocean and beyond may be loosely grouped into two categories: those that point to an enhancement of \( \kappa \) where eddy kinetic energy (EKE) is highest, i.e. at the surface and at the core of jets [e.g. Holloway, 1986; Kuffer and Holloway, 1988; Visbeck et al., 1997; Solomon, 1998]; and others that contend that \( \kappa \) is instead enhanced along a Rossby wave critical layer (at which the phase speed of the waves approximately matches the mean zonal flow speed) lying at mid depth in the ACC core and surfacing on the current’s equatorward flank [e.g. Killworth, 1997; Treguer, 1999; Cerovečki et al., 2009; Smith and Marshall, 2009; Abernathey et al., 2009]. Both of these contrasting views can draw some support from observational analyses. Ferrari and Nikurashin [2010] addressed the controversy from a theoretical standpoint by arguing that the spatial variability of \( \kappa \) is shaped by the magnitude of the EKE and the speed of the waves relative to the mean flow. In so doing, they derived a scaling law that quantifies whether specific jets are regions of vigorous or reduced mixing depending on the relative importance of the two effects.

Here we contend that a single meridional transect of hydrographic properties (e.g., Fig. 1a) contains sufficient information to settle the debate on the spatial structure of \( \kappa \) across the Southern Ocean. To illustrate this point, we consider the potential temperature - salinity (\( \theta - S \)) diagram corresponding to the section in Figure 1a. This \( \theta - S \) diagram (Fig. 1b) displays many features that will be familiar to anyone who has examined hydrographic measurements from the Southern Ocean, e.g., well-defined clusters of \( \theta - S \) curves that characterize the ACC’s interfrontal zones above the core of CDW, gaps in thermohaline space that occur at the fronts within the same range of densities, and comparative homogeneity in the properties of the bulk of CDW. This \( \theta - S \) distribution hints at a marked suppression of eddy stirring in the upper and intermediate layers of the ACC fronts relative to deeper levels and surrounding regions, and appears to conflict with the aforementioned view that \( \kappa \) may be enhanced near the surface and at the core of jets.

In this article, we estimate the structure of \( \kappa \) in the Southern Ocean by combining a set of hydrographic sections across the ACC and spatially coincident altimetric measurements within a mixing length theoretical framework. Mixing length theory (e.g. Prandtl [1925]) asserts that \( \kappa \) may be represented as the product of an eddy velocity scale \( U_e \) and an eddy mixing length scale \( L_e \), such

\[
\kappa = c_e U_e L_e,
\]

where \( c_e \) is a constant metric of the efficiency of the stirring process. We show that, typically, \( \kappa \) is strongly suppressed in the upper \( O(1) \) km of the ACC frontal jets relative to their surroundings and, following Ferrari and Nikurashin [2010], we interpret the spatial distribution of \( \kappa \) as resulting from modulation of \( L_e \) by eddy - mean flow interactions. We show, further, that the characteristic barrier behaviour of upper-ocean jets breaks down at a few special sites. There, significant departures from parallel flow conditions and the absence of a clear separation between mean and eddy length scales appear to mark the onset of a distinct mixing regime, in which jets become leaky and mixing is no longer suppressed. It is in such leaky jet segments that high values of \( \kappa \) are seen to occur in conjunction with large isopycnal PV gradients, suggesting that these sites contribute disproportionally to the eddy-induced component of the Southern Ocean overturning.

The paper is organized as follows. Section 2 introduces the collection of data sets used in this study. In section 3, we discuss the application of our methodology to the data, and the resulting distributions of \( L_e \), \( U_e \) and \( \kappa \). Possible interpretations of these distributions are presented and assessed in section 4. Lastly, section 5 presents the conclusions of our work and considers its implications for the thermohaline structure and overturning circulation of the Southern Ocean.

2. Data

The work presented here makes use of several observational data sets. A collection of meridional high-quality hydrographic sections across the Southern Ocean is at the heart of our calculation of \( L_e \). This includes only repeat transects or sections with particularly small (< 30 km) station spacing. Experience gained by subsampling finely resolved transects suggests that, in the absence of repeats, a considerably coarser station spacing compromises the resolution of lateral structures in \( L_e \). We analyze sections conducted at five different World Ocean Circulation Experiment (WOCE) transect locations spanning a wide range of mean flow (Fig. 2a) and EKE (Fig. 2b) regimes: western (WOCE S1) and eastern (WOCE S1Rb) Drake Passage,
Our analysis of the latter three transects is restricted to the region of monotonically increasing geopotential height toward the north, for reasons that will become apparent in Section 3.

Our estimation of \( U_e \) relies primarily on the analysis of a 15-year (1992–2007) time series of weekly sea surface height anomaly fields optimally interpolated to a \( S^9 \) Merit grid by Aviso from TOPEX / POSEIDON, Jason-1, ERS-1, ERS-2 and Envisat altimetric observations [Traon et al., 2003]. This data set is used in conjunction with an estimate of the absolute sea surface dynamic topography with fine mesoscale resolution obtained by Maximenko and Niler [2005] from drifter, satellite altimetric, NCEP / NCAR reanalysis wind and GRACE gravity data. We estimate the subsurface flow by combining these sea surface topographic data sets with the hydrographic sections introduced above, in the manner described in Section 3.2.

### 3. Structure of the isentropic eddy diffusivity in the Southern Ocean diagnosed from observations

#### 3.1. Estimates of the eddy mixing length scale \( L_e \) from hydrographic sections

Our method for estimating \( L_e \) rests on the mixing length arguments of Armi and Stommel [1983] (see also Ferrari and Polzin [2005]), according to which \( L_e \) may be defined as

\[
L_e = \frac{\theta_{rms}}{\nabla_n \theta_{rms}},
\]

where \( \theta_{rms} \) is the rms potential temperature fluctuation along a neutral surface arising from eddy stirring of the large-scale potential temperature \( \theta_n \), and \( \nabla_n \) is the gradient operator on the same neutral surface. This definition is formally valid to the extent that tracer fluctuations are generated by local stirring of the large-scale tracer gradient (i.e. advection of tracer variance from regions upstream is assumed to be weak), and insofar as \( \nabla_n \theta_n \) varies slowly over the eddy mixing length \( L_e \) (i.e. a scale separation between eddy and mean flow scales is assumed). The first assumption arguably holds widely across the ACC, where cross-stream \( \theta \) gradients are observed to greatly exceed along-stream \( \theta \) gradients [Sun and Watts [2001]]. As will be seen later in this section, the second assumption might be violated in a few special areas in which jets with distinct background potential temperature gradients and eddy stirring characteristics lie approximately one eddy mixing length apart. We contend, nonetheless, that even in this case an excess of \( \theta \) variance relative to the local \( \nabla_n \theta_n \) (conducive to a large \( L_e \)) may be taken as a qualitative indicator of vigorous eddy stirring. Thus, the possible sporadic formal invalidity of our eddy mixing length definition will be overlooked in the remainder of this article.

We note that this description of lateral mixing, by assumption, distinct from that put forward by Joyce et al. [1978] in the context of a high-spatial-resolution survey of the PF that was part of the International Southern Ocean Studies (ISOS). Whereas our framework assumes that thermohaline variability on isoneutrals arises passively from the mesoscale eddy-induced filamentation of background \( \theta - S \) gradients (e.g., as found by Smith and Ferrari [2000] in the eastern subpolar North Atlantic), Joyce et al. [1978] contend that interleaving features actively enhance thermohaline variability through double diffusive processes that cause the features to slope across neutral surfaces. Their interpretation builds on the following characteristics of thermohaline filaments: (1) vertical scales of 50-100 m; (2) a typical slope of 2-4 \( \times \) 10^{-3}, much steeper than isopycnals; and (3) an increase in the density of the filaments as they cross the jets, though this last characterization requires caveats [Toole, 1981]. We argue that all these properties are consistent with the eddy stirring scenario put forward here. Smith and Ferrari [2009] show that along-isopycnal eddy advection generates filaments that are thin in the horizontal (through eddy strain) and in the vertical (through eddy shear). The thinning of filaments by mesoscale eddies is eventually arrested at vertical scales of 10-100 m (consistent with Joyce et al. [1978]) by a vertical diffusivity of \( 1 \times 10^{-5} \, \text{m}^2 \, \text{s}^{-1} \) acting equally upon temperature and salinity profiles. Smith and Ferrari [2009] further demonstrate that the resulting filaments have an aspect ratio proportional to \( f/N \). With values of \( f = 1.2 \times 10^{-4} \, \text{s}^{-1} \) and \( N = 3 \times 10^{-3} \, \text{s}^{-1} \) as observed by Joyce et al. [1978], this aspect ratio \( f/N \approx 4 \times 10^{-2} \) is consistent with the observations reported in that study. Finally, the filaments generated by eddy stirring have density ratios close to 1 and are susceptible to double diffusive instabilities that act to increase their density. Hence filaments could slope across isopycnals in response to double diffusive fluxes. We note, however, that double diffusion is not the process generating the filaments, but instead is a consequence of eddy stirring and results in a slight modification of filament densities. We thus conclude that the thermohaline filaments observed across the ACC are likely the result of isopycnal eddy stirring: there is no shortage of eddy activity in the Southern Ocean. Other interpretations would be warranted only if the observed filaments had characteristics inconsistent with the eddy stirring scenario, and this is presently not the case.

Along each hydrographic section, we calculate \( \theta_{rms} \) and \( \theta_{rms} / \nabla_n \theta_{rms} \) on discrete neutral surfaces separated by an interval of 0.02 kg m^{-3} in the neutral density variable \( R^0 \) of [Jackett and McDougall, 1997]. This involves mapping the measured \( \theta \) profiles, which are provided in a 2 dbar pressure grid, to the selected \( R^0 \) surfaces using linear interpolation. We wish to distinguish between spatial \( \theta \) anomalies caused by the meandering of streamlines, which in the ACC are aligned with horizontal contours of hydrographic properties to a very good approximation [Sun and Watts, 2001], and anomalies associated with the genuine eddy-induced translation of water parcels across streamlines, which ultimately leads to cross-stream mixing when these anomalies are eroded by small-scale mixing processes. This distinction is made by using a baroclinic streamfunction (the geopotential anomaly at 500 dbar relative to 1500 dbar, \( \phi_{500} \)), defined as \( \phi_{500} = \int_{500}^{1500} \delta dp \), where \( \delta \) is the specific volume anomaly and \( p \) is pressure) as the cross-stream coordinate in place of geographical distance. Our results are insensitive to the exact choice of baroclinic streamfunction, because the ACC streamlines are equivalent barotropic and hence very little with depth.

The distance between \( \phi_{500} = 500 \) contours at each hydrographic section is computed as the mean distance between the contours averaged over all repeats of the section. We will refer to this pseudo-distance as \( \bar{Y} \), with the origin chosen as the southernmost station in the section. Since all sections are oriented approximately perpendicular to the ACC streamlines, \( \bar{Y} \) is a reasonable estimate of mesoscale stream distance. We opted for a definition of \( \bar{Y} \) in terms of \( \phi_{500} \) (as opposed to sea level), which would give many more realizations than hydrographic section repeats) because we have greater confidence in the resolution of \( \phi_{500} \) frontal variability in a geopotanical anomaly-based reference frame rather than a sea level-based reference frame.
After mapping the θ observations along all sections to a Y - γ^2 grid, we calculate the θ_m distribution along each transect location by fitting a cubic spline to all the Y - θ data pairs sampled on each neutral surface. An illustration of the calculation is provided by Figure 3. The θ_{rms} at location Y is estimated as the one standard deviation of (θ - θ_m) for all measurements obtained within Y ± ΔY. The thermal anomalies entering the calculation of θ_{rms} have characteristic vertical scales of O(10 - 100 m), comparable to the dimensions of cross-frontal interleaving features reported elsewhere [e.g. Joyce et al., 1978; Toole, 1981]. The width of the interval ΔY is chosen to be the range 30 - 150 km, with the exact value depending on the spatial density of sampling and the width of the ACC at each transect location. Our choice is guided by the requirement to have at least 5 - 10 data points in each calculation interval to approach statistical stability. This calculation of L_e exhibits only minimal sensitivity to ΔY values on the order of 10 - 100 km.

The distributions of θ_m arising from the preceding calculations are shown in Y - γ^2 space for each of the five transect locations (Figs. 4a-5a). The main water masses and frontal features of the Southern Ocean can be immediately diagnosed. A relatively warm (θ_m ∼ 1 - 3°C) and voluminous body of CDW is seen to occupy the bulk of the sections at densities in excess of γ'' > 27.6 kg m^{-3}, overlying a layer of colder (θ_m < 0°C) and denser (γ'' > 28.27 kg m^{-3}) AABW near the southern end of the WOCE Sr1b, I6S, 18S and Sr3 transects. In the upper layers, a subsurface θ_m minimum colder than θ_m ∼ 2°C is observed in each of the sections to the south of the Polar Front, denoting the core of the wintertime variety of Antarctic Surface Water (referred to as Winter Water). Warmer (θ_m > 3°C) upper-ocean waters are found further to the north. These indicate the presence of SAMW and AAIW equatorward of the Subantarctic Front, underlining a thin layer of surface waters.

The cross-stream isoneutral gradient of θ_m at each of the transect locations is displayed in Figures 4b-8b. These reveal that the thermal anomalies upon which eddies act are largest in the upper layers of the Polar Front (particularly in Drake Passage) and of the Subantarctic and Subtropical Fronts (in WOCE I6S, 18S and Sr3), as well as more generally along the base of the pycnocline. The patterns in the distribution of θ_{rms} (Figs. 4c-8c) broadly follow those in the ∇_Nθ_m field, indicating that the local rate of thermal variance production by eddy stirring along neutral surfaces is highly dependent on the local isoneutral gradient of θ_m. However, the covariation of θ_{rms} and ∇_Nθ_m is not perfect. Rather, it exhibits substantial spatial inhomogeneity, its structure reflecting variations in the mixing length scale L_e.

Estimates of L_e as defined in (3) are shown in Figures 4d-8d. We estimate that L_e varies by at least one order of magnitude. Values on the order of 5 - 10 km are commonly found in the upper layers (uppermost 500 - 1000 m) of the major ACC frontal jets, identified by their geostrophic velocity expressions in Figures 4e-5e, whereas mixing lengths of 50 - 150 km occur in the jets’ deeper layers and in inter-frontal regions. The finding that the mixing length scale inferred from isoneutral θ fluctuations is comparable to or smaller than the horizontal eddy length scales of O(100 km) lends further support to our interpretation of these fluctuations as being the product of eddy stirring.

Specifically, we find a notable suppression of L_e in the four frontal jets in western Drake Passage (Figs. 4d-4e); the PF and SBdy in eastern Drake Passage (Figs. 5d-5e); the STF, SAF and SACCF in WOCE I6S (Figs. 6d-6e); the STF, SAF and PF in WOCE 18S (Figs. 7d-7e); and the SAF’s southern branch, the two branches of the PF, SACCF and SBdy south of Tasmania (Figs. 8d-8e). Only in three frontal jet sites do we find an obvious absence of eddy mixing length suppression: the SAF in WOCE Sr1b (Figs. 5d-5e), the PF south of Africa (Figs. 6d-6e), and the SAF’s northern branch in WOCE Sr3 (Figs. 8d-8e). The correspondence, or lack thereof, between areas of reduction in L_e and fronts is ambiguous in a few of the weaker jets, notably the SACCF in eastern Drake Passage (Figs. 5d-5e), the SBdy in WOCE I6S (Figs. 6d-6e), and the SACCF and SBdy in WOCE 18S (Figs. 7d-7e). To summarize: out of 24 frontal jet crossings, 17 exhibit evidence of suppression, such evidence is ambiguous in 4, and only 3 definite exceptions are noted.

We note that the increased magnitude of L_e in the deep ACC, below a depth of approximately 1000 m, is of dubious significance as it is associated with the tendency of ∇_Nθ_m toward zero there. This tendency could indeed reflect intense eddy stirring at depth, but it might also be a consequence of the general downward increase in the ventilation age of water masses. In spite of this caveat, at least one piece of unambiguous evidence can be uncovered that points toward a genuine intensification of eddy stirring with depth at the ACC frontal jets. Such evidence may be found in the quasi-synoptic survey of the PF conducted during ISOS [Joyce et al., 1978], which has approached only 250 km to the east of the northern end of the WOCE Sr1 transect (Fig. 2a) and was much more densely sampled than any WOCE section, thereby providing a unique view of the current’s thermohaline structure. Using this data set, we construct a cross-frontal section with a variable horizontal resolution of 3 - 15 km. The distributions of θ_m, ∇_Nθ_m, θ_{rms}, and L_e in Y - γ'' space along this section are displayed in Figures 9a-9d and exhibit many of the properties that we encounter in the WOCE transects. The ISOS section spans across the northern flank of the PF, as denoted for example by the northern terminus of the Winter Water θ_m minimum at γ'' ≈ 27.45 kg m^{-3} near Y ≈ 30 km (Fig. 9a). The strongest θ_m gradients are seen in waters lighter than γ'' ≈ 27.7 kg m^{-3} in the horizontal proximity of this terminus (Fig. 9b), and occur in association with slightly elevated values of θ_{rms} (Fig. 9c). It is in this area that eddy stirring is most strongly suppressed and L_e values under 10 km are found (Fig. 9d). These values are comparable to the cross-frontal length scales of the coherent interleaving features observed in the upper part of the PF during ISOS [Toole, 1981]. North of and below the upper layers of the front’s northern flank, eddy stirring is no longer inhibited, and L_e is enhanced by typically an order of magnitude.

We now wish to direct the reader’s attention to a feature in the ISOS section that was not immediately apparent in the WOCE transects: the boundary between the regions of weak and intense stirring in and north of the PF, respectively, is not vertical. Rather, it has a distinct slope of O(5 × 10^{-3}) that broadly parallels geostrophic isotherms (Fig. 9e). That the area of reduced stirring in the upper layers of the frontal jet is bowl-shaped and has sloping lateral boundaries is further evidenced by the distribution of \(\nabla \theta_m \approx \partial \theta_m / \partial Y^2\) along neutral surfaces in the ISOS section (Fig. 9g). A band of markedly negative values occupies the deeper part of the region of weak stirring (Fig. 9h), and mimics the shape of mean flow contours (Fig. 9e). That points toward a genuine intensification of eddy stirring anywhere curvature in \(\theta_m\) tends to be erased. Thus, we conclude that the detailed thermohaline structure of the PF in the ISOS section points to the existence of an inverse relationship between the mean flow speed and the eddy stirring rate, and is thus consistent with a genuine intensification of eddy stirring with depth at the ACC frontal jets.

3.2. Estimates of the eddy velocity \(U_e\) from altimetry
We estimate the cross-stream eddy velocity scale \( U_c \) as the one standard deviation in time of \( u \cdot \hat{n} \), where \( u \) is the horizontal velocity vector, and \( \hat{n} \) is the unit vector perpendicular to the time-mean horizontal velocity \( \mathbf{T} \). The calculation consists of five steps. First, we construct a 15-year time series of weekly maps of sea surface height \( \eta \) by interpolating along each section the combined altimetric and absolute dynamic topography products introduced in Section 2. Second, along each transect, we map the \( \gamma^2 \) and surface geopotential anomaly relative to an isobaric surface of pressure \( p \), \( \phi_0^p \), as a function of \( \eta \) at the time of \( 0 \) for each transect and \( p \). Then we fit cubic splines to the \( \gamma^2 \) \( \eta \) and \( \phi_0^p \) \( \eta \) pairs at each isobaric level. The procedure is analogous to the gravest empirical mode calculation of Sun and Watts [2001]. Third, we use these sets of spline fits to obtain weekly, two-dimensional (i.e., as a function of \( \eta \) and \( p \)) estimates of the \( \gamma^2 \) and \( u \) fields along each hydrographic section location. The surface \( u \) field is calculated from \( \eta \) using thermal wind, and projected to depth using \( \phi_0^p \)-derived geostrophic shear. Fourth, we map the along-transect distribution of \( u \), which is originally estimated on a pressure-based vertical grid, to the \( \gamma^2 \) grid used in the calculation of \( U_c \). Fifth, we compute \( U_c(\eta, \gamma^2) \) from the expression above and use the mean (of all section repeats) minimal value of \( \gamma^2 \) at which \( u \) becomes significant. This feature reflects the breadth of the zone of enhanced EKE that typifies the core of the ACC frontal jets, and that it is intense in the deeper part of the jets and in interfrontal regions. We note that the area of mixing suppression associated with each jet often extends beyond the depth range of significant isentropic PV gradients (cf. Figures 4d-8d and 4h-8h), as may be expected from a kinematic interpretation of the suppression.

In this analysis of \( U_c \), we find three unambiguous exceptions to the generalized reduction of eddy stirring at the core of the ACC frontal jets, with high values of \( \gamma^2 \) \( \kappa \) found in the upper part of the SAF in WOCE SR1b (Fig. 5g), the PF south of Africa (Fig. 6g), and the SAF’s northern branch in WOCE SR3 (Fig. 8g). In those sites, high diffusivities occur in conjunction with sizeable isentropic PV gradients (see Figs. 5h, 6h and 8h).

4. Discussion: Interpretation of the structure of the isentropic eddy diffusivity in the Southern Ocean

In this section, we examine the degree of consistency between the properties of eddy stirring across the ACC indicated by the preceding analysis and a hierarchy of dynamical ideas put forward to characterize eddy stirring across geophysical jets in the literature. These include descriptions of eddy stirring as quasi-homogeneous turbulence (section 4.1), linear waves in a parallel shear flow (section 4.2), weakly nonlinear waves in a parallel shear flow (section 4.3), aspects of wave propagation in non-parallel shear flows (section 4.4), and the law of the wall (section 4.5). Such theoretical considerations lead, in brief, to predictions of enhanced eddy stirring in jet cores (quasi-homogeneous turbulence), enhanced eddy stirring on jet flanks and below jet cores (linear waves in a parallel shear flow), suppressed eddy stirring in jet cores (weakly nonlinear waves in a parallel shear flow), potential breakdown of eddy stirring suppression in jet cores (aspects of wave propagation in non-parallel shear flows), and suppression of eddy stirring near a topographic boundary (law of the wall). We conclude that the generalized reduction of \( \kappa \) at the core of the ACC frontal jets present in our results may be best interpreted as a suppression of stirring by eddy interaction with parallel jets as (in section 4.3), and that the observed occasional deviations from this regime (the occurrence of three leaky jets) might plausibly be explained in terms of eddy interaction with narrow and twisted jets (as in section 4.4).

4.1. Eddy stirring as quasi-homogeneous turbulence

4.1.1. Concept

Descriptions of eddy stirring as quasi-homogeneous turbulence date back to the classical work of Taylor [1921], who demonstrated that particle dispersion in a homogenous, isotropic turbulent field implies an eddy diffusivity proportional to the product of the EKE and the eddy decorrelation time scale, i.e.,

\[ \kappa_{\text{Taylor}} \propto \text{EKE} \gamma^{-1}. \]  

(4)
Considering that \( U_e \propto \text{KE}^{1/2} \), this result implies that the eddy mixing length scale in a turbulent field is given by \( L_e \propto \text{KE}^{1/2} \ g^{-1} \), so as to recover the \( L_e \)-based expression for the eddy diffusivity (2).

Expressions of the form of (4) are often used to estimate eddy diffusivities in the ocean, and underlie the common perception that \( \kappa \) depends primarily on \( EKE \) and is thus highest near the surface and in the core of jets. This point of view is put forward by, for example, Holloway [1986], Keffer and Holloway [1988], Visbeck et al. [1997] and Salmon [1999], who combine a variety of theoretical ideas on the characteristics of eddies [Green, 1970; Stone, 1972; Holloway and Kristmannsson, 1984; Held and Larichev, 1996] with satellite altimetric measurements and eddy-resolving numerical simulations to contend that surface \( \kappa \) is routinely enhanced in the core of the main ocean currents. Common elements to these studies are a reliance upon mixing length theory and the assumption that \( L_e \) scales with the size of geostrophic eddies, thereby varying only gently in space. This assumption ultimately leads to the result that the spatial variability in \( \kappa \) largely follows that of \( \text{KE} \).

The idea stemming from this body of work find qualitative support in estimates of near-surface \( \kappa \) obtained from the application of Taylor’s theory (4) to drifter data, which point to an enhancement of \( \kappa \) within the ACC core relative to surrounding regions and to a prevalence of \( U_e \) in shaping the lateral variability in surface \( \kappa \) [Schäfer and Krauss, 1995; Sallée et al., 2008]. A quantification of surface eddy stirring based on finite-time Lyapunov exponents calculated from altimetric measurements yields similar findings [Waugh and Abraham, 2008]. In regard of the vertical distribution, Ferreira et al. [2005] report a surface intensification of \( \kappa \) and an exponential decay with depth, based on an estimate of the eddy stress field that minimizes the departure of a global coarse-resolution ocean model from hydrographic observations. A similar pattern of surface enhancement and downward decay is suggested by an inverse diagnostic of \( \kappa \) from observed surface forcings and hydrography ([Olers and Visbeck, 2005]), as well as by estimates of eddy temperature fluxes from moored current meter and float measurements [Phillips and Rintoul, 2000; Ferrari and Polzin, 2005]. Again, such a vertical decay of \( \kappa \) is consistent with the reduction in \( \text{KE} \), and hence \( U_e \), with depth.

4.1.2. Assessment

The assumption that \( L_e \) is proportional to the size of geostrophic eddies and the ensuing result that the structure of \( \kappa \) is primarily determined by \( U_e \) are both inconsistent with our findings. These indicate that \( L_e \) generally exhibits a marked reduction at the core of jets relative to their surroundings, and that it is the main variable shaping the \( \kappa \) distribution.

To shed light on the origin of this discrepancy, we note that estimates of \( \kappa \) from the application of Taylor [1921]’s theorem to upper-ocean drifter and altimetric data cannot in principle isolate the cross-stream component of stirring or particle dispersion from the much larger along-stream component. This deficiency likely washes out the mixing barrier signatures of jet cores, and is aggravated by the fact that convergence of the time integral in Taylor’s theory requires integration over a time scale \( \Delta t \) of \( O(100) \) days [Sallée et al., 2008], leading to an implicit spatial averaging scale of \( \sqrt{\Delta x \Delta y} \sim 100 - 300 \) km. The contribution of cross-frontal Ekman flows is also a concern in evaluating upper-ocean drifter data for evidence of a weakening of stirring in jet cores.

4.2. Eddy stirring as linear waves

4.2.1. Concept

An alternative description of eddy stirring in the Southern Ocean in terms of linear Rossby waves is provided by Marshall et al. [2006] who, in applying the ‘effective diffusivity’ technique of Nakamura [1996] to altimetric observations, suggest that surface \( \kappa \) values are amplified on the equatorward flank of the ACC relative to the current’s axis, i.e. largely outside the region of highest \( EKE \). This pattern is reminiscent of observations of eddy stirring in the atmosphere, where the intensification of mixing along jet flanks is thought to be associated with Rossby wave critical layers (e.g., Haynes and Shuckburgh [2000a, b]), and is seemingly better aligned with the definition of jets as PV jumps in the geophysical literature (e.g., Dritschel and McIntryre [2008]).

Critical layers arise in linear wave equations when the mean flow is considered. For our purpose, the lowest order incarnation of this type of equation is the barotropic quasi-geostrophic potential vorticity equation for linear eddy perturbations embedded in a large-scale and slowly-evolving zonal mean flow of speed \( U_m \):

\[
\partial_t \psi' + U_m \frac{\partial \psi'}{\partial x} + \frac{\partial \sigma}{\partial z} \frac{\partial \psi'}{\partial y} = 0, \tag{5}
\]

where \( \psi' \) denotes the eddy geostrophic streamfunction and \( \sigma = \nabla^2 \psi' \) the eddy potential vorticity. With an assumption about the functional representation of \( \psi(y, z) \), and consequently \( U_m \), \( \psi(y, z) \), and \( \sigma \) plane wave solutions to (5) of the form \( \psi' \propto e^{i(kx-ct)} \) exhibit large-amplitude crossfrontal displacements as \( \sigma \) and \( \kappa \) and the corresponding phase speed of the wave, \( \sigma \) its intrinsic frequency, and \( k \) its zonal wavenumber (e.g., Pratt et al. [1995]). Such large amplitudes imply markedly enhanced stirring rates relative to near-zero background values. In the linear wave limit, cross-jet mixing is absent away from critical layers. This characterization of eddy stirring is consonant with the linear instability theory of Killworth [1997] and the arguments of Tregruer et al. [1999], Cerovecki et al. [2009], Smith and Marshall [2009] and Abernathy et al. [2009] based on numerical simulations of varying complexity. These studies point to an enhancement of \( \kappa \) at depth, allegedly associated with Rossby wave critical layers. It has also been suggested that the critical layer phenomenon may underlie the observation of inhibited (enhanced) exchange of floats in the upper ocean (at mid depth) across the Gulf Stream jet (Bower et al., 1985; Bower and Rossby, 1989; Bower and Lévié, 1994; Lévié et al., 1997; Rogerson et al., 1999).

4.2.2. Assessment

The results of our analysis in section 3 bear much qualitative resemblance to the preceding description of eddy stirring in terms of linear wave ideas, in that they indicate that \( \kappa \) does not depend strongly on \( EKE \), and that stirring rates are enhanced outside of, rather than within, jet cores. However, a subtle discrepancy between our results and the linear wave interpretation is that we find no clear evidence of an enhancement of stirring rates in the upper ocean along jet flanks and at mid depth below jet cores relative to off-jet regions. Rather, \( L_e \) and \( \kappa \) values in off-jet regions are comparable in magnitude to values enveloping jet cores.

4.3. Eddy stirring as weakly nonlinear waves

4.3.1. Concept

A description of eddy stirring as weakly nonlinear waves is put forward by Ferrari and Nikurashin [2010], who investigate eddy stirring in a baroclinic quasi-geostrophic model. Their key result may be captured by considering the simpler problem of eddy-induced mixing across a barotropic jet. The barotropic quasi-geostrophic potential vorticity equation for weakly nonlinear eddy perturbations embedded in a large-scale and slowly-evolving zonal mean flow (cf. expression 5) is given by

\[
\frac{\partial \psi'}{\partial t} + U_m \frac{\partial \psi'}{\partial x} + \frac{\partial}{\partial z} \left( \frac{\partial \psi'}{\partial y} \right) = \nabla^2 \psi' \tag{6}
\]
\[ \dot{q} + U_m \partial_q + \beta \partial \psi' + J(\psi', q) = 0, \] (6)

where \( J \) the Jacobian operator and \( \beta = df/dy \) the constant planetary potential vorticity gradient. For simplicity, we assume that the mean flow is zonal. Curve terms associated with bending of the mean flow are neglected, consistent with the assumption that the mean flow varies on scales much larger than the eddies.

As in Ferrari and Nikurashin [2010], the nonlinear term \( J(\psi, q) \) is represented with a fluctuation-dissipation stochastic model [Landahl, 1975; Farrell and Ioannou, 1993; Flierl and McGillicuddy, 2002]. The stochastic model is monochromatic to keep the problem linear and can be thought of as representing the excitation of waves by instability at horizontal wavenumber \((k, l)\). Dissipation in a fluctuation-dissipation model is through linear damping at a rate \( \gamma \) which sets the eddy decorrelation time scale as shown below.

Following Ferrari and Nikurashin [2010], the streamfunction solution of the stochastic model (see Ferrari and Nikurashin [2010] for details of the derivation) can be used to advect a tracer \( \theta \) and compute the mixing length defined as the typical length of the resulting tracer filaments,

\[ L_e = \frac{k^2 EKE^{1/2} \gamma^{-1}}{k^2 + \beta^2 1 + \gamma^{-2} k^2 (U_m - c)^2}. \] (7)

where \( c = U_m - \beta / (k^2 + \beta^2) \) is the phase speed of barotropic Rossby waves propagating in the mean flow \( U_m \), \( k \) is the zonal eddy wavenumber, and \( \gamma^{-1} \) the eddy decorrelation (damping) time scale. Using (2), the reduction in \( L_e \) results in a suppression of the eddy diffusivity,

\[ \kappa \propto U_m L_e = \frac{EKE \gamma^{-1}}{1 + \gamma^{-2} k^2 (U_m - c)^2}. \] (8)

This expression synthesizes the key result of a characterization of eddy stirring in terms of weakly nonlinear waves. It suggests that the mixing length is proportional to the mean displacement induced by an eddy, \( EKE^{1/2} \gamma^{-1} \), also that the presence of a mean flow acts to reduce \( L_e \). Thus, for eddies propagating at the same speed as the jet, \( c = U_m \), \( L_e \propto EKE^{1/2} \gamma^{-1} \) and the mean flow has no effect on mixing. However, when the eddy phase speed does not match the jet velocity, eddy propagation relative to the mean flow results in a temporal oscillation of cross-frontal displacements. If the oscillation period \( k^{-1} (U_m - c)^{-1} \) is shorter than the eddy decorrelation time scale \( \gamma^{-1} \), cross-frontal filamentation is arrested. This mixing length suppression is captured by the denominator in (7), which is proportional to the ratio of the advective and decorrelation time scales, and is purely a kinematic effect of coherent phase propagation.

The eddy stirring suppression effected by the presence of a mean flow is brought out most clearly by formulating the eddy diffusivity in (8) as a modification to Taylor's expression,

\[ \kappa = \frac{k^{2 \gamma^{-1}}}{4 \gamma^{-2} k^2 (U_m - c)^2}. \] (9)

where \( k^{2 \gamma^{-1}} = k^2 (k^2 + \beta^2)^{-1} EKE \gamma^{-1} \) is Taylor's definition of the cross-jet (meridional) eddy diffusivity. The expression in (9) predicts that the diffusivity is reduced in the core of (broad) jets relative to the jets' flanks, as long as \( U_m \geq c \) as is the case near strong ocean jets.

It is important to note that, despite qualitatively endorsing the basic characterization of oceanic jet cores as mixing barriers, the expressions for \( \kappa \) in (8)-(9) differ in a significant way from the bulk of the oceanographic literature adopting that description. This literature (e.g., [Marshall et al., 2006; Smith and Marshall, 2009; Abernathy et al., 2009]) explains the spatial distribution of \( \kappa \) in the vicinity of a jet in terms of an enhancement of mixing at critical layers girdling the jet core. This stems from the widely accepted paradigm in the atmospheric community (e.g., [Taylor, 2007, 2009], the nonlinear stochastic model [2010], the streamfunction solution of the stochastic model [see Ferrari and Nikurashin [2010] for details of the derivation] can be used to advect a tracer \( \theta \) and compute the mixing length defined as the typical length of the resulting tracer filaments,

\[ L_e = \frac{k^2 EKE^{1/2} \gamma^{-1}}{k^2 + \beta^2 1 + \gamma^{-2} k^2 (U_m - c)^2}. \] (7)

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The emerging picture of widespread mixing suppression along the ACC jets is suggested by observations of the propagation of coherent eddy features in the ACC. Hughes et al. [1998] applied a Radon transform technique to altimetric measurements to estimate the zonal phase speeds of these eddying motions. An updated illustration (courtesy of C. Hughes, personal communication) of their most fundamental result is shown in Figure 2d, where the observed zonal speed of eddy propagation is displayed along with mean dynamic topography. Zonal phase speeds are positive (eastward) in a broad band following the path of the ACC, whereas they are negative (westward) elsewhere, as expected from Rossby wave propagation in the absence of a mean flow. It may be readily concluded (Hughes, 1995, 1996; Hughes et al., 1998; Smith and Marshall, 2009) that eastward propagation of eddy features in the ACC region is the result of advection by the eastward mean flow. This is reiterated by a detailed comparison of the zonal phase speed and mean dynamic topography fields; there is a remarkable spatial coincidence between the highest eastward phase speeds (surpassing 0.06 m s⁻¹, see Figure 2d) and the fastest jets (exceeding 0.2 m s⁻¹, see Figure 2a). Note, however, that the difference between the eddy and phase speeds (or \( U_m - c \)) is larger in the main jets than in regions of weak mean flow. This observation implies, following (8), that \( \kappa \) is suppressed in the jets, i.e., water parcels are advected rapidly past eddying motions and so the extent to which eddies stir tracers is reduced locally.
The eddy field, though, is comprised of more than a single dominant frequency $\sigma$ and phase speed $c$. Unlike the monochromatic version of the stochastic fluctuation-dissipation model used to obtain (8), the eddy field is multichromatic and has both barotropic and baroclinic components. Rather than formulating a representation that includes such refinements, Ferrari and Nikurashin (2010) conduct a comprehensive test of the scaling law in (8) with altimetric data and find a suppression in $\kappa$ of up to a factor of 2-3 in the core of the ACC jets. They show that an accurate approximation to (8) is given by

$$\kappa = \frac{K_{\text{Taylor}}}{1 + 4U_m^2 \cdot EKE^{-1}}. \quad (10)$$

Combining typical values of $U_m$ and EKE (cf. Figs. 2a-2b) in (10) predicts a reduction in $\kappa$ of 20%-70% in the ACC core with hardly any impact on the current’s flanks (cf. Fig. 2c), in broad agreement with the findings of studies discussed in Section 4.2. We contend, however, that as these studies rely exclusively on relatively coarse [O(100 km)] altimetric observations or eddy-permitting model output, the suppression of $\kappa$ is likely to be underestimated.

Indeed, the in situ measurements analyzed in Section 3 suggest that the ACC jets are both narrower and faster than estimated from those coarse data sets, which would result in a larger term in the denominator of (8) or (10) and a predicted stronger suppression of $\kappa$.

In order to more rigorously assess the extent to which the weakly nonlinear eddy ideas are consistent with our analysis of the thermonaline finestucture of hydrographic sections, the locations of the transects analyzed in section 3 are overlaid on a map of the stirring suppression factor, defined as $(1 + 4U_m^2 \cdot EKE^{-1})^{-1}$ (i.e. the inverse of the denominator in (10)) and calculated using the mean surface flow of Maximenko and Niiler [2005] (Fig. 2a) and an altimetric estimate of EKE (Fig. 2b). Squares along the sections denote the mean dynamic topography values of the jets at which mixing length suppression is indicated by the hydrography, while circles mark jets where no suppression is observed in the in situ measurements. Overall, there is a good correspondence between the theoretical predictions and in situ estimates of suppression, but for two ambiguous cases (the PF in WOCE I6S and WOCE I8S) and two obvious exceptions (the SAF in WOCE SR1b and the northern branch of the SAF in WOCE SR3). Significantly, three of these four ambiguous or exceptional jets were identified as leaky by our analysis of the hydrography. While the lack of mesoscale detail in the map in Figure 2c prevents us from categorizing these three jets as being clearly inconsistent with the weakly nonlinear eddy theory, the hydrographic analysis of those jet regions reveals the presence of strong and narrow mean flows with no mixing length suppression. This pattern is incompatible with the prediction of equations (8) and (10), and will serve to motivate our subsequent discussion of eddy stirring across a narrow and twisted jet.

4.4. Eddy stirring across a narrow, twisted jet

4.4.1. Concept

The theory underpinning expressions (7)-(9) relies on two major assumptions that are not always satisfied in the ACC: (1) that a separation between the length scales of the jet and the eddies exists; and (2) that the mean flow is zonal or, more generally, a parallel shear flow with velocity gradients in the vertical and only one horizontal direction. Studies of mixing across a narrow parallel Bickley jet (Rypina et al. [2007]) show that, for parallel shear flows, mixing suppression continues to hold even in the absence of scale separation, such that the first assumption above is not critical. In contrast, there are firm indications of the theory’s predictions being shaped decisively by the second assumption.

Passive scalar dynamics is greatly enriched when a jet is twisted (non-parallel). Key illustrations of how such richness comes about may be found in past studies of the roles of the (isentropic) rate of strain (Okubo, 1970; Weiss, 1991) and Lagrangian acceleration ([Hua and Klein, 1998] in shaping the stirring properties of a 2-d turbulent fluid.

A finding of those investigations is that non-parallel flows host an exponential separation of particle pairs with time, rather than the (asymptotically much slower) linear separation with time that is characteristic of parallel flows. The rapid increase in particle separation is associated with twisting and folding of tracer filaments and thereby promotes a local intensification of cross-stream tracer fluxes.

This translates into an enhancement of the eddy diffusivity ($\kappa$) relating the magnitude of the tracer flux to the background tracer gradient in the region of the straining mean flow. The behavior of linear waves in non-parallel, horizontally non-divergent background flows is, in the limit of an extreme scale separation between the waves and the background, analogous to that of passive scalars [Bühler and McIntyre, 2005; Polzin, 2008]. While the complicated aspects of non-parallel flows render the problem analytically intractable, a recent highly idealized numerical study by Thompson [2009] shows that cross-jet mixing indeed ceases to be suppressed if the background PV gradient has small-scale non-parallel structure. Thompson generated such a background PV gradient with topographic hills, but we may expect to see a similar behavior of eddy stirring regardless of the specific mechanism that maintains sharp and non-parallel PV gradients. Thus, following this argument, we anticipate that a breakdown of mixing suppression may occur in narrow, non-parallel jets.

Although no generalized theory yet exists to predict the detailed distribution of $\kappa$ in non-parallel flows, the Okubo-Weiss parameter ($D$) has been shown to be a useful indicator of regions where particle separation increases exponentially and jets are likely to become leaky (e.g., [Haller and Yuan, 2006]). The Okubo-Weiss parameter is defined as

$$D = \frac{S_2^2 + S_1^2 - \Gamma^2}{4}, \quad (11)$$

where $S_n = \partial u/\partial x - \partial v/\partial y$ is the normal component of the rate of strain, $S_z = \partial w/\partial x + \partial u/\partial y$ is the shear component of the rate of strain, and $\Gamma = \partial v/\partial x - \partial u/\partial y$ is the vertical component of the relative vorticity [Okubo, 1970; Weiss, 1991]. In the limit of small particle pair separation in a horizontally non-divergent, 2-d turbulent velocity field, $D^{-1/2}$ is a characteristic time scale of exponential particle pair separation in strain elements of the flow (for which $D > 0$), whereas it is a characteristic time scale of the variation in the relative orientation of particle pairs in vortical elements of the flow (for which $D < 0$). Parallel flows fall into a special category of $D \equiv 0$ and linear (asymptotically slow) particle pair separation.

4.4.2. Assessment

In order to assess whether the absence of eddy stirring suppression observed in a small subset of the jets in the hydrographic sections may conceivably be interpreted in terms of the preceding ideas, we calculate the Okubo-Weiss parameter of the mean surface flow from Maximenko and Niiler [2005]’s mean dynamic topography. Given the small number of jets in which stirring suppression is absent, this assessment is necessarily exploratory. The result of the calculation is shown in Figure 2e. We find...
that the mean streamlines associated with the three exceptional, high $\kappa$ upper-ocean jets (the SAF in WOCE SR1b, the PF in WOCE I6S, and the SAFs northern branch in WOCE SR3) meander through alternating patches of large strain ($D \gtrsim 10^{-13} \text{ s}^{-2}$) and vorticity ($D \ll -10^{-13} \text{ s}^{-2}$) closely upstream of the sections at which they are sampled, whereas conventional straining barrier jets are widely accentuated by small values of $D$ indicative of parallel flow (with one obvious exception, the SAF in WOCE S1, which will be discussed in the following section). This suggests that the breakdown of straining suppression in certain segments of the ACC jets may be plausibly attributed to the occurrence of non-parallel conditions in those segments. Following from this, we hypothesize that ACC jets become leaky in regions of large mean flow strain (such as when a jet is narrow and twisted), while they act as barriers to eddy stirring elsewhere.

If our hypothesis regarding the breakdown of suppression in non-parallel flows is valid, some insight into the dynamical mechanisms controlling the degree of leakiness of ACC jets may be obtained by close examination of the spatial distribution of the straining and vortical elements in the time mean velocity field, as revealed by the Okubo - Weiss parameter (Figure 2e). Two features of this distribution are most readily apparent. First, straining and vortical elements of the flow commonly occur in sets of alternating patches associated with meanders in the mean flow that have characteristic wavelengths of 300 - 500 km. Second, the occurrence of these patches is neither ubiquitous nor seemingly random, but it is focussed on relatively confined regions of strong mean flow in the vicinity of complex topography. See, in particular, the isolated areas of positive and negative $D$ over the Southwest Indian Ridge near WOCE I6S, the trains of straining and vortical patches around the northern edges of the Del Caño Rise and the Kerguelen Plateau between 40°E and 80°E, the region around the northern flank of the Southeast Indian Ridge between 95°E and 120°E, the region of Macquarie Island and Campbell Plateau, south of Tasmania and New Zealand, the Udintsev and Eltanin Fracture Zones, around 140°W, and finally the northern Scotia Sea and neighboring basins to the north.

The character of the meander trains was investigated by Hughes [2005] in the course of a study of the ACC’s nearsurface vorticity balance. He found the meanders to have properties akin to those of short (compared to $2\pi L_e$, where $L_e \approx 2000 \text{ km}$ is the barotropic Rossby radius of deformation) westward-propagating barotropic Rossby waves held stationary by the mean flow, as illustrated most clearly by the meanders’ nonlinear vorticity balance. Following his findings, this nonlinearity may be used as a precise indicator of the relevance of short barotropic Rossby wave dynamics to explaining the approximation of mean flow and eddy length scales and the departure from parallel flow behaviour that are seemingly associated with the breakdown of eddy stirring suppression. To this effect, two contours of $\bar{u} \cdot \nabla \bar{T}$ (the advection of the mean relative vorticity by the current flow, which approximately balances the advection of planetary vorticity by the mean flow within the waves) are displayed in Figure 2e. It may be readily appreciated that, for the most part, areas of positive and negative $D$ appear in association and approximately $90^\circ$ out of phase with patches of high $\bar{u} \cdot \nabla \bar{T}$ magnitude and alternating sign, thereby indicating that the breakdown of mixing suppression could be associated with the presence of stationary short barotropic Rossby waves in the ACC’s path. The clustering of the waves around areas of complex bathymetry indicates that they are ‘jet waves’, generated by the interaction of the ACC flow with topographic obstructions in the manner characterized by e.g., Tansley and Marshall [2001] and Rhines [2007]. The topographic localization of eddy stirring suggested by our analysis is reminiscent of the findings of MacCready and Rhines [2001] in numerical simulations of an idealized channel flow over topography, although the extent to which the processes discussed here operate in their model is unclear.

To conclude, we must caution that our estimate of $L_e$, based on equation (3), formally assumes a separation in scale between the eddy mixing length and the large-scale potential temperature gradient. The above analysis suggests that the three regions in which suppression of $L_e$ was not found at jets are characterized by the existence of sharp meanders in the mean flow, which may violate the scale separation assumption. Hence, careful analysis of numerical simulations is likely required to confirm the breakdown of eddy stirring suppression in non-parallel jets, and to settle the dynamics at work.

### 4.5. The law of the wall

#### 4.5.1. Concept

A final theoretical ingredient is required to complete our interpretation of the observed patterns of eddy stirring in the hydrographic transects. This is commonly referred to as the “law of the wall” (e.g., Tennekes and Lumley [1972]) and states that in the presence of a solid boundary, the physical scales of eddies are limited by the distance to the boundary. It is thus reasonable to expect that the eddy mixing length $L_e$, as a root-mean-square measure of cross-stream particle displacements, and therefore $\kappa$ to be on the order of a few thousand km. Second, the occurrence of these patches is neither ubiquitous nor seemingly random, but it is focussed on relatively confined regions of strong mean flow in the vicinity of complex topography. See, in particular, the isolated areas of positive and negative $D$ over the Southwest Indian Ridge near WOCE I6S, the trains of straining and vortical patches around the northern edges of the Del Caño Rise and the Kerguelen Plateau between 40°E and 80°E, the region around the northern flank of the Southeast Indian Ridge between 95°E and 120°E, the region of Macquarie Island and Campbell Plateau, south of Tasmania and New Zealand, the Udintsev and Eltanin Fracture Zones, around 140°W, and finally the northern Scotia Sea and neighboring basins to the north.

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To conclude, we must caution that our estimate of $L_e$, based on equation (3), formally assumes a separation in scale between the eddy mixing length and the large-scale potential temperature gradient. The above analysis suggests that the three regions in which suppression of $L_e$ was not found at jets are characterized by the existence of sharp meanders in the mean flow, which may violate the scale separation assumption. Hence, careful analysis of numerical simulations is likely required to confirm the breakdown of eddy stirring suppression in non-parallel jets, and to settle the dynamics at work.

#### 4.5.2. Assessment

In section 4.4.2, we found that one of the jets in the hydrographic sections (the SAF in WOCE S1) meanders through alternating patches of large strain and vorticity closely upstream of the S1 transect (Fig. 2e), yet the hydrographic analysis reveals that it is not leaky. This is at odds with the discussion of eddy stirring across non-parallel jets in section 4.4. We contend that the absence of a strain-related breakdown of mixing suppression in this jet can be readily explained by invoking the law of the wall. As shown by Garabato et al. [2003] and hinted at by Figure 4e, the SAF jet in western Drake Passage overlies a very steep segment of the South American continental slope, suggesting that the suppression of eddy stirring across the jet is simply the result of a reduction of $L_e$ in the proximity of a boundary. The downstream evolution of the jet’s mixing characteristics appears to support this explanation. By the time it crosses the WOCE SR1b section in eastern Drake Passage, the SAF jet has moved off the continental slope (Fig. 5e) and, with increasing distance from the boundary, becomes leaky (Fig. 5d).

### 5. Conclusions

The character of eddy stirring in the Southern Ocean has been investigated by analyzing a collection of hydrographic transects and spatially coincident altimetric measurements within a mixing length theoretical framework. The outcomes of this analysis provide significant insight into the long-standing debate concerning the distribution of eddy stirring across the ACC and the nature of its controlling processes.

We have shown that, typically, the isentropic eddy diffusion of $\kappa$ is reduced by an order of magnitude in the upper $O(1 \text{ km})$ of the ACC frontal jets relative to their surroundings. The spatial structure of $\kappa$ primarily results from variability in the eddy mixing length: $L_e$ is not simply proportional to the physical scale of eddies, as it is often assumed in the oceanographic literature (e.g., Hollovery, 1986; Visbeck et al., 1997; Salmon, 1998), rather it is modified by eddy - mean flow interactions. This character of eddy stirring is reproduced and illuminated
by a quasi-geostrophic theory of eddy stirring across a broad barotropic jet founded on the work of Ferrari and Nikurashin [2010]. The theory indicates that the observed widespread reduction of $L_e$ and $\kappa$ in the upper layers of frontal jets is the kinematic consequence of eddy propagation relative to the mean flow within jet cores. As shown by other investigators in the past (e.g., Hughes et al. [1998], eddying motions in the ACC propagate eastward considerably more slowly than the surface mean flow does. The observed cessation of mixing suppression outside and in the deep ($O(\sim 1 \text{ km})$) layers of jet centers may thus be attributed to the tendency of the eddy propagation relative to the mean flow to become small there.

While the inferred spatial distribution of $\kappa$ across the ACC shares several basic features with patterns highlighted by the model-based studies of Smith and Marshall [2009] and Abernathy et al. [2009], amongst others, our interpretation of these features differs. Based on both observational and theoretical leads, we suggest that the prevalent control on the $\kappa$ distribution is the suppression of eddy stirring at the centre of the jets, rather than the enhancement of $\kappa$ above background values at Rossby wave critical layers girdling the jets. Most fundamentally, we contend that the characteristic thermohaline structure of the ACC, consisting of multiple upper-ocean thermohaline fronts separating and underlaying by regions of homogenized properties, is largely a result of suppression of eddy stirring by eddy-mean flow interactions outlined above.

As widely valid as this characterization of eddy stirring is likely to be, our analysis reveals that it is not universally applicable across the Southern Ocean. Pronounced deviations from the prevalent mixing regime are encountered in a few special sites. There, mixing suppression is observed to break down at the core of upper-ocean jets, allowing intense cross-stream exchange of thermohaline properties to ensue.

The key condition associated with the emergence of such leaky jet segments is tentatively identified as the occurrence of non-parallel structure in the mean flow on length scales comparable to those of the eddies. In the light of the limited available evidence, the Okubo-Weiss parameter of the mean flow has some skill as a qualitative indicator of the satisfaction of this condition. While no generalized theory of mixing in a non-parallel mean flow yet exists, a number of past studies have shown that passive scalar and linear wave dynamics is greatly enriched in straining flows, as a result of their hosting a much faster along-stream particle pair separation and sharpening of cross-stream tracer gradients than parallel flows. In the case of the ACC, the mean flow strain that likely underpins the breakdown of mixing suppression occurs within sharp stationary meanders of 300-500 km wavelength. These can be characterized as short westward-propagating barotropic Rossby waves that are generated by the interaction of the ACC with topography and held stationary by the mean flow. Such associations between stationary meanders and prominent bathymetric features provides a plausible explanation for the relatively rare occurrence of leaky jet segments in the ACC, and suggests that complex eddy-mean flow interaction dynamics are at play in regions of strong topographic steering and form drag.

We conjecture that these sparse and localized disruptions to the regular ‘mixing barrier’ behaviour of upper-ocean jets may underlie the as-yet-unexplained occurrence of abrupt along-stream changes in the properties of Southern Ocean mode and intermediate waters as the ACC negotiates major topographic obstacles (Sallée et al. [2010]). To illustrate this point, we consider the mass transport within a neutral layer of thickness $h$, given by

$$\bar{\nabla} \mathbf{h} = \vec{K} \beta \mathbf{f} - \frac{\bar{\nabla} \mathbf{h}}{\rho} + \frac{1}{\rho} \mathbf{E},$$

(12)

where we use the downgradient closure for the PV flux in (12). Below the surface Ekman layer and above the depth of topography, $\mathbf{E} = 0$ and cross-jet transport can only be supported through eddy stirring. Typically, $\kappa$ is small across ACC jets within the upper pycnocline, and so little cross-ACC transport can ensue. However, if $\kappa$ is much larger at leaky jet segments, it follows that cross-jet frontal subduction of mode and intermediate waters must be greatly enhanced in these regions.

Taking a vertical integral of equation (12), the left hand side vanishes and one is left with the well known balance between wind- and eddy-induced transports. This relationship constrains the density structure of the Southern Ocean, represented by $\vec{h}$, as a function of wind stress, planetary vorticity gradient and $\kappa$. Extant estimates of the residual overturning and associated stratification assume that $\kappa$ is constant or enhanced at jets (e.g., Olbers and Visbeck [2005]; Marshall and Radko [2006]). Our results show that these assumptions are inconsistent with the eddy-mean flow interactions that so profoundly shape the thermohaline structure of the Southern Ocean. In future papers, we intend to investigate the dynamics underpinning the observed spatial variability in $\kappa$, and the impact of these variations on the stratification and overturning circulation of the Southern Ocean.

Acknowledgments. This study was conducted during ACNG’s stay at MIT, which was supported jointly by MIT and the U.K. Natural Environment Research Council (NERC) through a NERC Advanced Research Fellowship (NE/C517633/1). RF acknowledges the support of NSF award OCE-0825376. KP’s participation in this work was supported by WHOI bridge support funds. We are grateful to David Marshall, Alan Plumb, Peter Rhines, Bernadette Sloyan and Andy Thompson for helpful discussions; to Rob Scott for advice with the analysis of altimetric observations; and to Chris Hughes for providing the altimetric estimate of zonal wave speed reproduced in Figure 2d.

Notes

1. A description of the calculation may be found in Smith and Marshall [2009].

References


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Raffaele Ferrari, Massachusetts Institute of Technology, Cambridge, U.S.A.

Alberto C. Naveira Garabato, National Oceanography Centre, Southampton, U.K. (acng@noc.soton.ac.uk)

Kurt L. Pulz, Woods Hole Oceanographic Institution, Woods Hole, U.S.A.
Figure 1. (a) Meridional section of salinity (in color) across eastern Drake Passage (section location indicated as WOCE SR1b in Figure 2a). Neutral density contours (contour interval = 0.1 kg m$^{-3}$) are shown in black, and the major water masses labeled (AABW = Antarctic Bottom Water; AAIW = Antarctic Intermediate Water; AASW = Antarctic Surface Water; LCDW = Lower Circumpolar Deep Water; SAMW = Subantarctic Mode Water; UCDW = Upper Circumpolar Deep Water). Station positions are indicated by the white tick marks at the base of the figure. The locations of the ACC fronts (PF = Polar Front; SACCF = Southern ACC Front; SAF = Subantarctic Front) and its Southern Boundary (SBdy) are shown by the arrows at the top of the figure. (b) $\theta$ - $S$ diagram corresponding to Figure 1a. $\theta$ - $S$ curves are colored according to the interfrontal zone in which they lie. Selected isoneutral contours are displayed in black. Water masses and fronts are labeled as in Figure 1a.
Figure 2. (a) Surface flow speed (in color) estimated from the mean dynamic topography of Maximenko and Niiler [2005]. Mean dynamic topography contours at intervals of 0.1 m are shown by the thin black lines in all panels. The locations of the hydrographic sections analyzed in this study are indicated by the thick black lines in all panels and labelled in (a). The thick grey line in Drake Passage marks the position of the ISOS section discussed in section 4.1. The white lines in (a)-(d) show segments (extending over 20° of longitude upstream of each section) of the mean dynamic topography contours associated with the cores of frontal jets at each of the hydrographic transects, and are labelled in (a) with the standard frontal terminology (STF = Subtropical Front; see caption of Figure 1 for other definitions). These lines are displayed in black in (e) for clarity. The black arrows in all panels indicate the positions of the leaky jets sampled by the hydrographic sections. (b) Eddy kinetic energy (in color) calculated from the Aviso gridded altimetry product. (c) Inverse suppression factor $[1 + 4U_0^2 EKE^{-1}]^{-1}$ (in color), calculated as described in section 5. The jets in the hydrographic sections that exhibit mixing suppression are indicated by open squares, and leaky jets are marked by open circles. (d) Zonal speed of propagating eddies (in color) estimated from the Aviso gridded altimetry product by Chris Hughes. (e) Okubo-Weiss parameter $D$ (in color) and advection of the mean relative vorticity by the mean flow (grey lines; the $-3 \times 10^{-13}$ and $3 \times 10^{-13}$ s$^{-2}$ contours are respectively shown by dotted and solid lines), calculated from Maximenko and Niiler’s mean dynamic topography.
Figure 3. $\theta$ measurements (shaded circles) on the 27.7 kg m$^{-3}$ isoneutral as a function of along-transect pseudo-distance $Y$ (defined in section 4.1) for ten occupations of the WOCE SR1b section in eastern Drake Passage. Light- (dark-) shaded circles indicate measurements at pressures smaller (larger) than 150 db, typical of the winter mixed layer base across much of the ACC. The cubic spline fit defining $\theta_m$ is shown by the solid line.
Figure 4. Distributions of (a) $\theta_m$; (b) $\nabla_m \theta_m$; (c) $\theta_{rms}$; (d) $L_e$; (e) $U_{geos}$, the geostrophic velocity relative to the deepest common level; (f) $U_e$; (g) $c_0^{-1} \kappa$; and (h) the natural logarithm of (planetary) potential vorticity $q$, as a function of $Y$ and $\gamma$ along the WOCE S1 section. The black lines denote selected pressure contours, labeled in dbar. The white lines in (d) and (g) show the upper boundary of the area of the deep Southern Ocean where $\nabla_m \theta_m \approx 0$, and so $L_e$ and $c_0^{-1} \kappa$ are ill-defined. The positions of hydrographic fronts are indicated at the top of each set of panels and labeled as in Figures 1 and 2. Fronts interpreted to exhibit (lack) mixing suppression in their upper and intermediate layers are labeled in blue (red), with ambiguous cases indicated in black.
Figure 5. As in Figure 4, but for the WOCE SR1b section.
Figure 6. As in Figure 4, but for the WOCE I6S section.
Figure 7. As in Figure 4, but for the WOCE I8S section.
Figure 8. As in Figure 4, but for the WOCE SR3 section.
Figure 9. Distributions of (a) $\theta_m$; (b) $\nabla_n \theta_m$; (c) $\theta_{rms}$; (d) $L_e$; (e) $U_{geo}$, the geostrophic velocity relative to the deepest common level; (f) $\log_{10}(q)$; and (g) $\partial^2 \theta_m / \partial Y^2$, as a function of $Y$ and $\gamma^n$ along the ISOS section. The black lines denote selected pressure contours, labeled in dbar. The white line in (d) shows the upper boundary of the area where $\nabla_n \theta_m \approx 0$ and $L_e$ is ill-defined.
Table 1. Meridional high-quality CTD sections analyzed in this study. The nominal location, vessel and dates of each occupation are listed.

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<th>Nominal location</th>
<th>Vessel</th>
<th>Dates of occupation</th>
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<td>15 - 20 November 1996</td>
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<td>RRS James Clark Ross</td>
<td>22 - 28 November 2000</td>
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<td>27 December 2002 - 1 January 2003</td>
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