Baroclinic Control of Southern Ocean Eddy Upwelling Near Topography

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Abstract In the Southern Ocean, mesoscale eddies contribute to the upwelling of deep waters along sloping isopycnals, helping to close the upper branch of the meridional overturning circulation. Eddy energy (EE) is not uniformly distributed along the Antarctic Circumpolar Current (ACC). Instead, “hotspots” of EE that are associated with enhanced eddy-induced upwelling exist downstream of topographic features. This study shows that, in idealized eddy-resolved simulations, a topographic feature in the ACC path can enhance and localize eddy-induced upwelling. However, the upwelling systematically occurs in regions where eddies grow through baroclinic instability, rather than in regions where EE is large. Across a range of parameters, along-stream eddy growth rate is a more reliable indicator of eddy upwelling than traditional parameterizations such as eddy kinetic energy, eddy potential energy, or isopycnal slope. Ocean eddy parameterizations should consider metrics specific to the growth of baroclinic instability to accurately model eddy upwelling near topography.

Plain Language Summary The Southern Ocean plays an essential role in redistributing heat, salt, and biogeochemical tracers of importance in the climate system. In particular, locations in which strong ocean currents interact with large topographic features are hotspots for eddy-driven upward transport, and are crucial pathways to bring deep, carbon- and nutrient-rich waters to the surface. The processes which set the location and magnitude of this eddy “upwelling” remain challenging to understand. This study uses a series of high-resolution idealized simulations in which an ocean jet encounters a piece of topography to investigate what controls the eddy upwelling near topography. We find that the upwelling due to eddies occurs in regions where the eddies are growing through a mechanism called “baroclinic instability,” rather than in regions where eddies are highly energetic or energized by other mechanisms. Regions of growing eddy energy are a simple, first-order indicator of regions of eddy upwelling, but future parameterizations of transport should consider the mechanism of instability to be more accurate.

1. Introduction

The Southern Ocean is an essential component of the global overturning circulation, which redistributes heat, salt, and biogeochemical tracers of importance in the climate system (J. Marshall & Speer, 2012). In particular, sloping density surfaces (isopycnals) in the Southern Ocean provide an adiabatic route for deep waters to be upwelled to the surface. This along-isopycnal transport brings cold, carbon-rich waters to the surface (Le Quéré et al., 2007), imposing an important control on the Southern Ocean CO₂ sink and contributing to delayed warming of Southern Ocean surface waters (Armour et al., 2016).

Mesoscale eddies, which are particularly energetic in the Southern Ocean (Fu et al., 2010), play a dominant role in this along-isopycnal transport and therefore can have a critical influence on the associated mass, carbon, and heat transports. Eddy activity in the Southern Ocean is not uniform in time or space. Zonal variations along the path of the Antarctic Circumpolar Current (ACC) are punctuated by regions of elevated EE downstream of where the ACC interacts with major topographic features (Foppert et al., 2017; Frenger et al., 2015; Sokolov & Rintoul, 2009; Thompson, 2010), visible both at the surface (e.g., Fu et al., 2010) and at depth (e.g., Thompson & Naveira Garabato, 2014). These hotspots of EE are favorable to stronger cross-jet exchange (Dufour et al., 2015; Thompson & Sallée, 2012) and enhanced upwelling of deep and intermediate waters (Foppert et al., 2017; Tamsitt et al., 2017; Viglione & Thompson, 2016).

Regions of elevated EE are typically colocated with stationary meanders downstream of a topographic obstacle. The presence of these meanders, which are formed by arrested Rossby waves (Hughes & Ash, 2001), introduces...
non-zonal velocities, which lead to departures from the traditionally assumed dynamical balances derived from a zonally integrated view. The stationary meanders play an essential role in balancing zonal momentum and provide a mechanism for rapid barotropic adjustment of the flow to changes in forcing (Thompson & Naveira Garabato, 2014). These meanders appear to dominate the meridional heat transport (Dufour et al., 2012), but such heat transport predominantly occurs through transient eddies acting along the meander structure (Abernathey & Cessi, 2014). The essential role of transient eddies in this heat transport is visible when the transport is calculated in density-depth space (Zika et al., 2013) or following streamlines (Abernathey & Cessi, 2014).

The strength of eddy-induced transport in the Southern Ocean is often assumed to scale linearly with eddy kinetic energy (EKE) along the lines of the classical mixing length hypothesis (Holloway, 1986; Prandtl, 1925). For example, studies investigating the response of Southern Ocean circulation to changes in forcing often examine the response of EKE (e.g., Hogg et al., 2015; Meredith & Hogg, 2006; Patara et al., 2016), but few studies diagnose eddy-induced transport. Dufour et al. (2012) noted the increased southward transport due to transient and stationary eddies under increased wind forcing, but did not relate its response to that of EKE. However, there are no direct observations or modeling studies which support a direct, local proportionality between EKE and eddy-induced upwelling. On the contrary, Tamsitt et al. (2017) reports enhanced upwelling upstream of EKE maxima, but does not provide a dynamical explanation for this spatial separation. Likewise, Foppert et al. (2017) noted an offset between eddy heat fluxes and EKE in the Drake Passage, and suggest that the sea surface height deviation (a proxy for eddy potential energy, EPE) is a better indicator of the divergent eddy heat flux and, by extension, eddy upwelling owing to a direct connection to baroclinic instability (Watts et al., 2016). This offset is also found in the idealized simulations of Bischoff and Thompson (2014), which notes that EKE is not colocated with the steepest isopycnal slopes. An examination of how topography modulates eddy-induced upwelling and, further, an identification of the relationship between EE and the mechanisms controlling upwelling location and magnitude are needed, in particular to inform our design of eddy upwelling proxies.

This study focuses on how a single unstable jet in a 2-layer system supports intense, localized, isopycnal upwelling associated with transient eddies. This jet is an analog for a single frontal jet of the ACC; the simplicity of this system allows unambiguous definition of cross-jet volume transport to quantify eddy upwelling, revealing insights that are not possible in a more comprehensive model. In particular, the question of whether local EE, or one of its constituents (EKE or EPE), is a good indicator of local eddy-induced upwelling is examined. Lastly, we show that a simple parameterization of eddy-induced upwelling based on the zonal evolution of EE provides a better representation of the zonal variability of upwelling around topography, compared with proposed parameterizations based on EKE, EPE, or time-mean isopycnal slope.

2. Methods

2.1. Model Configuration

Our model simulations are designed to represent the interactions between a baroclinic ocean jet and an isolated topographic feature, in a configuration relevant to the Southern Ocean (see Figure 1). The set-up used is identical to that of Barthel et al. (2017). The model configuration is a channel on a β-plane, with dimensions of 9,600 km × 1,600 km and a horizontal resolution of 4 km. It consists of two isopycnal layers with a free surface. We use MOM6 (Adcroft et al., 2019) to solve the hydrostatic thickness-weighted primitive equations under the Boussinesq approximation. The background horizontal viscosity is parameterized with a biharmonic horizontal viscosity of \( A_\theta = 1.5 \times 10^6 m^2 s^{-1} \) to ensure numerical stability, while bottom friction is modeled by a weak quadratic bottom drag (with \( C_{drag} = 5 \times 10^{-4} \)). The dynamics in the interior of the channel are purely adiabatic.

The channel is forced to sustain an eastward-flowing jet (Figure 1) by restoring the stratification (including sea surface height) at the western boundary. The jet characteristics are representative of a typical frontal jet observed in the Southern Ocean, with a 50- to 150-km-wide jet core containing peak velocities of 0.5–1 m s⁻¹ (in the upper 1,000 m), while velocity below 1,000 m is of order 0.1 m s⁻¹ (Sheen et al., 2014; Waterman et al., 2013). The eastern boundary also features a “sponge” region where isopycnal heights are restored to allow the flow to readjust to the inflowing conditions. This boundary forcing of the flow provides a direct control of the jet structure at the inflow, as well as prescribing the total zonal transport. In this regard, this study differs from wind-driven channel studies which rely on a wind-friction equilibration (e.g., Bischoff & Thompson, 2014; Chapman et al., 2015) and can feature significantly different zonal transports depending on the presence of bottom topography (see...
Abernathey & Cessi, 2014, their Figure 8). Stratification is also restored at the northern and southern boundaries, thus sustaining a large-scale meridional isopycnal slope, with the upper layer shoaling southward. This combination of forcing in the “sponge” regions allows a nonzero residual overturning circulation to emerge in the domain, as it does in the Southern Ocean (e.g., G. J. Marshall, 2003; Lumpkin & Speer, 2007).

To explore topographic control of eddy-driven isopycnal upwelling, we compare flat-bottom simulations with cases which include either a circular seamount, or a meridional ridge, with a range of heights (0–500 m). The range of topographic heights is small (compared with the Southern Ocean) because the topography has a disproportionately large effect in a two-layer system. Definitions of energy reservoirs (EKE and EPE) and exchanges in this layered system are presented in the Supplemental Information.

2.2. Quantifying Meridional Transport

We diagnose the eddy-driven upwelling by quantifying the southward volume transport due to transient eddies in the upper, southward-shoaling isopycnal layer. Importantly, we account for the presence of stationary meanders downstream of topography by isolating the volume transport across the time-mean jet axis (hereafter the cross-jet transport). As our simulations have only one southward-shoaling layer, we sidestep the difficulties of defining a depth-dependent jet axis and focus on the transport by eddies across the contour of maximum upper-layer time-mean velocity. The transport, \( T \), perpendicular to the time-mean velocity field is written as

\[
T(x, y) = \frac{h_1}{|u_1|} \times \frac{u_1'}{|u_1'|} = \frac{h_1}{|u_1|} \times u_1' \times |u_1| = \frac{h_1'}{|u_1'|} \times \frac{u_1}{|u_1|},
\]

where \( h_1 \) is the thickness and \( u_1 \) is the horizontal velocity in the upper layer. The overbar indicates the time-mean of a quantity, and the prime is the deviation from that mean (i.e., the eddy component). By construction, only the eddy quantities contribute to the net transport across the time-mean velocity field. The cross-jet transport, \( X_{jt} \), is defined on the jet axis:

\[
X_{jt} = T(x, y_w(x)) ,
\]

Figure 1. Model domain. A prescribed 2-layer jet flows eastward over topography, leading to stationary meanders downstream of topography. The maximum inflow velocities at section A are 0.7 m.s\(^{-1}\) for the upper layer and 0.3 m.s\(^{-1}\) for the lower layer.
where $y_m$ is the value of $y$ for which $|\mathbf{u}_1|$ is maximal. We count the transport as positive when it is southward, that is, when it is associated with upwelling along the isopycnal layer.

The advantage of the above definition is that it allows us to robustly compare the net transport by eddies in the presence of stationary meanders that form downstream of topographic obstacles. As stationary meanders have significant time-mean meridional velocities, they would have an alternating signal in southward and northward transport across a fixed latitude line (see Hallberg & Gnanadesikan, 2001, for a discussion on transport across streamlines vs. fixed contours). Calculating the cross-jet transport at the jet axis allows a more meaningful comparison of net cross-jet transport between cases with and without jet meanders.

3. Results

3.1. Eddy-Driven Upwelling

The mean flow state from three selected runs is presented in Figure 2. In each case the inflowing jet becomes unstable as it evolves eastward. The EE (sum of EKE and EPE; indicated by colors in the upper panel of each subplot) has a distinct spatial pattern, growing with $x$ as the flow evolves, with an along-stream maximum (highlighted by the red vertical bar in the lower panel). Eddy energy then remains constant or decays with further distance downstream. The qualitative evolution of EE is similar in all three cases, although the zonal extent of the EE growth region and the magnitude of the EE depends on the nature of the topography. Similar results are obtained if we examine EKE and EPE individually (not shown).

The transient motions lead to an eddy-induced upwelling, quantified by the eddy-induced thickness transport across the time-mean jet axis (cyan arrows in Figure 2). This transport also has zonal variations along the jet axis. In the case of the jet evolving over a flat bottom (Figure 2a), the transport is southward, and preferentially takes place in a limited region ($1,500 \text{ km} < x < 3,100 \text{ km}$). Further downstream, both southward and northward flux can occur locally, but these fluxes contribute little to the net transport. The bulk of the eddy-induced southward transport is localized in the region of EE growth, with a 99% correlation between the zonal variations in the zonally integrated cumulative transport and local EE (Figure 2a, lower subpanel).

In the presence of topography (illustrated by the 150 m ridge case; Figure 2b), localized regions of enhanced eddy-induced cross-jet transport persist. The signature of the stationary meanders is visible in the eddy-induced cross-jet transport variability (manifested as alternating regions of southward and northward transport), making it difficult to distinguish the net effect of eddies. It is therefore especially helpful in this case to consider zonally integrated cumulative transport (cyan line, Figure 2b, lower subpanel). This metric shows that the region immediately downstream of the ridge ($x = 1,500–2,000 \text{ km}$) contributes significantly to the net southward transport relative to the regions further downstream. Around $x \approx 2,000 \text{ km}$ (highlighted by the cyan vertical bar), there is a transition between a region of net southward transport ($x < 2,000 \text{ km}$) and a region of net northward transport ($x > 2,000 \text{ km}$). In some cases, the cumulative transport at $x = 6,000 \text{ km}$ is northward, which may be due to the lack of disturbances to break down the meanders downstream. The close relationship between zonal growth of EE and southward cross-jet transport, seen in the flat bottom case, also holds in the 150 m ridge case (71% correlation).

Most of the simulations with topography conducted in this study provided results that are consistent with the 150 m ridge case: the transport values vary but the relationship between southward transport and eddy growth holds (cf. Supporting Information S1). However, the third case presented in Figure 2, that with a 300 m high seamount, showcases a different regime. This case is consistent with the results above in that it shows a qualitatively similar zonal evolution of EE, and regions of preferential cross-jet transport immediately downstream of topography, but differs in the lack of correlation between along-stream eddy growth and cumulative southward transport. The break-down in this relationship provides insights into the underlying dynamics at play, and is explored in more detail in the next section.

In summary, two main points emerge from examination of the along-stream variations of EE and transport in these idealized simulations. First, the presence of topography leads to enhanced eddy-induced cross-jet transport localized immediately downstream of the topographic obstacle, relative to the same jet evolving over a flat bottom. The magnitude and location of the eddy-induced transport depend on the properties of the topography.
present. Second, EE and eddy-induced cross-jet transport ("eddy upwelling") have distinct zonal distributions. This transport tends to be localized in the region of along-stream eddy growth, but exceptions can occur where eddy-induced transport occurs in a region of smaller zonal extent than EE growth.

### 3.2. Mechanism for Topographic Control

We probe the dynamics underpinning the differing spatial distributions of EE and eddy-induced transport by looking at the two instability mechanisms that energize the eddy field in the 300 m seamount case (Figure 3b), and comparing them with the flat-bottom case (Figure 3a). Following the thickness-weighted energetics approach
used in Aiki and Richards (2008) and Barthel et al. (2017), we diagnose the eddy-mean flow energy conversions due to (a) the work of interfacial form stress, \(-\bar{u}_1 \cdot \nabla \phi_1\) (where \(\phi_1\) is the Montgomery potential and ' denotes the anomaly from the time mean), responsible for the generation of EE in baroclinic instability (dark blue lines) and (b) the work of Reynolds stress associated with horizontal convergence of momentum in the upper layer, associated with barotropic instability (black lines). The latter energy conversions terms are both calculated for the upper layer, and are positive when energy is fluxed from the mean into the eddy field. The outer product of two vectors is denoted by \(\otimes\), and \(\rho_0\) is the reference density of the Boussinesq approximation (see Barthel et al., 2017, for the full derivation).

The 300 m seamount is a helpful case to disentangle the contributions of form stress and Reynolds stress because they have distinct zonal patterns (Figure 3b). These patterns indicate that the eddy-induced transport is associated exclusively with baroclinic instability (i.e., positive conversion of energy into the eddy field via form stress). This relationship is consistent with our conceptual understanding that baroclinic instability contributes to flattening isopycnals, and with observations in Drake Passage that indicate the eddy heat flux is best aligned with the production of EPE (Foppert et al., 2017; Watts et al., 2016). These results further suggest that the zonally averaged link between interfacial form stress and meridional thickness flux (e.g., Olbers et al., 2004) may apply at the local scale. Understanding that the mechanism for eddy-induced transport is baroclinic instability acting as a source of EE is consistent with the alignment of the region of eddy upwelling with the region of along-stream eddy growth, rather than with regions of elevated EE.

The relationship between eddy upwelling and the action of eddy form stress in energizing the eddy field is robust across all simulations, both with and without topography. In most cases, the region of southward eddy transport extends over the entire region of along-stream eddy growth because both energy conversion terms have the same zonal patterns, as illustrated by the flat bottom case (Figure 3a). Nevertheless, it is important to keep in mind that
baroclinic instability alone provides the dynamical mechanism to generate cross-jet transport and eddy upwelling. As such, it is possible that along-stream eddy growth can occur in regions without net southward eddy transport (i.e., without active baroclinic instability) when, for instance, horizontal shear instability is responsible for EE growth. This scenario is nicely illustrated by the 300 m seamount case (Figure 3b).

These examples speak to the method by which topography influences the eddy-induced transport. We infer that the topographic obstacles affect the flow in such a way that either baroclinic or barotropic instability, or both, are enhanced. In some, but not all, cases there is a strong correspondence between these two different instability mechanisms. However, southward eddy-induced transport is only dependent on the action of baroclinic instability, where isopycnal interfaces slump to release available potential energy into the eddy field.

4. Implications for Eddy Parameterizations

Our results suggest that energy conversion terms are an unambiguous indicator of eddy-induced cross-jet transport, however, we recognize that these are unlikely to be practical indicators of eddy upwelling in coarse resolution models. Our analysis also indicates that the along-stream growth of EE may be a valuable predictor of eddy upwelling in many cases, and hence may inspire new eddy parameterizations for coarsely resolved models. Thus, in this section, we assess whether a coarsely resolved zonal pattern of EE may be used to estimate the cross-jet transport occurring in that region. For that purpose, we compare transport estimates obtained from assuming transport is proportional to the rate of along-stream EE growth to those employing other common parameterizations for eddy upwelling based on large-scale variables, such as the mean isopycnal slope, EKE and EPE.

Specifically we consider parameterizations based on the following relationships:

1. Transport can be parameterized as a constant diffusivity applied to the time-mean isopycnal slope ($\bar{S}$):
   \[ X^{GM}_h(x) = \kappa \bar{S} + B, \text{ with constant } \kappa = A, \text{ inspired by Gent and McWilliams (1990)} \]

2. The diffusivity $\kappa$ is proportional to EKE: $X^{EPE}_h(x) = A.EKE \cdot \bar{S} + B$, inspired by Jansen et al. (2015)

3. Eddy transport is proportional to the barotropic EPE: $X^{EPEbt}_h(x) = A.EPEbt + B$, with $EPEbt = \frac{\alpha}{2} g \eta'^2 (\rho_0)$:

4. Eddy transport is proportional to the rate of along-stream EE growth: $X^{dEE}_h(x) = A.\frac{d}{dx} EE + B$ where EE denotes total EE.

5. Eddy transport is proportional to the local EFS: $X^{EFS}_h(x) = A.EFS + B$, used as a reference in this exercise.

To compare the relative performance of these parameterizations in our model configuration, the large-scale variable on which each parameterization is based ($\bar{S}$, EKE, EPEbt, EE, or EFS) was smoothed and sub-sampled to a 80 km horizontal resolution, roughly equivalent to output from a 1° ocean model. In each case, a least squares fit was performed to determine the parameters $A$ and $B$ that minimize the total error in transport over the domain. The parameterized transport is then compared to the modeled transport in four different cases (Figure 4). Note that (a) the list of parameterizations evaluated is not exhaustive, (b) the comparison is advantageous to parameterizations: it relies on “model-perfect” input fields (coarse-grained), and the parameters are optimized for each case. A realistic implementation may rely on biased fields and constant parameters.

Results from this exercise confirm the conclusions from the previous section. Local values of energy reservoirs, such as EPE, are not a good indicator of cross-jet transport (e.g., Figure 4a3 for the flat-bottom case). The zonal variations in EPE and transport are so different that minimizing the total error leads to applying a small southward transport almost uniformly over the whole domain, leading to compensating underestimated transport upstream (light gray) and underestimated transport further downstream (dark gray). Similarly, the other relationships based on time-mean isopycnal slope and EKE (Figures 4a1–4a2) fail to capture the zonal pattern of transport, with the best parameterization being an almost uniform transport of small magnitude.

In contrast, the zonal growth rate of total EE, $\frac{d}{dx} EE$, is able to reproduce the zonal variations in eddy transport, producing a parameterized transport which adequately portrays regions of little to no transport, and regions of localized, enhanced transport. Local EFS is overall the best indicator for eddy transport, but is unlikely to be readily available output from climate models or observations. In the absence of EFS, the zonal growth of EE may be a valuable indicator of where eddy-induced transport occurs, and outperforms commonly used parameterizations of eddy upwelling, in most cases considered in this study (see Supporting Information). One exception is the...
300 m seamount case (Figures 4C1–4C4) where the relationship between cross-jet transport and the along-stream rate of change of total EE breaks down (Figure 3b) due to the influence of barotropic instability in generating EE.

5. Discussion

This study highlights that eddy-driven cross-jet transport within a shoaling isopycnal occurs in regions of EE growth through baroclinic instability. The presence of topography leads to enhanced eddy upwelling in the region immediately downstream of the obstacle (especially in the first meander) because it modifies the growth of baroclinic instability. The idealized set-up allows exact calculations of quantities not usually diagnosed in global climate models, and the simulations performed in this study provide a plausible mechanism explaining the location of the upwelling pathways from Tamsitt et al. (2017) which occurs in regions upstream of EKE maxima, and further the offset between the divergent eddy heat flux and EKE discussed by Foppert et al. (2017). Simple parameterizations based on mean isopycnal slope, EKE, and EPE fail to reproduce this strong, localized, eddy transport near topography. In most cases, the along-stream growth of EE is a good indicator for southward transport, with the exception of cases where barotropic instability and baroclinic instability have distinct growth regions (e.g., a steep isolated seamount; Figure 4c4).

The benefit of an idealized set-up is that it allows exact calculations of both the EFS and the Reynolds stress, and we can thereby attribute dynamical relevance between the two without ambiguity (noting that these quantities are not usually diagnosed from global climate models or observations). However, the simplified vertical structure in this two-layer system leads to an exaggerated impact of topography, as small values of topography are more dynamically relevant to the ACC. Idealized simulations may also overstimulate barotropic instability near

Figure 4. Parameterized cumulative southward transport across the jet axis plotted against the resolved cumulative transport, at each zonal grid point along the time-mean jet axis, for (a) flat bottom, (b) 150 m ridge, and (c) 300 m seamount. Data points are colored from light to dark as we move downstream. In each case, the parameterization is the best linear fit (by least-square method) that minimizes the total error between local values of transport and local 1) time-mean cross-jet isopycnal slope $\bar{S}$, 2) EKE times $\bar{S}$, 3) barotropic EPE ($EPE_{bt}$), 4) zonal growth of total EE ($dx_{EE}$), and 5) EFS, where each variable was smoothed and sub-sampled to a 80 km resolution.
topography (e.g., Barthel et al., 2017; Youngs et al., 2017). Despite this caveat, we argue that the two-layer set up provides useful dynamical insight, given that evidence of mixed barotropic-baroclinic instability is also observed in the Drake Passage (Foppert, 2019) and may be important for the momentum balance in the ACC (Constantinou & Hogg, 2019).

In this study, we focused only on eddy-induced isopycnal thickness fluxes and showed that eddy-driven upwelling does not occur in regions of high EE, but rather in regions of along-stream EE growth by baroclinic instability. However, the presence of high EE, and potentially high EKE in particular, may contribute to enhancing other types of transport, such as the upwelling of tracers through increased isopycnal stirring (Abernathey & Ferreira, 2015; Dufour et al., 2015). In addition, the net meridional transport in the Southern Ocean is forced by a combination of factors, including wind stress, surface buoyancy fluxes, and diabatic processes in the surface mixed layer; these factors are dominant where layers outcrop at the surface and emphasize the role of the vertical structure of eddy processes in the ACC that are omitted from this study. Results from the adiabatic simulations considered in this study best inform on interior upwelling processes, away from frictional boundaries such as the surface and bottom Ekman layers, and away from locations where diabatic mixing dominates (e.g., close to rough topography).

Keeping in mind the above caveats, the detailed dynamical analysis of these idealized simulations provides an important insight: assuming that high values of EKE and/or EPE indicate regions of strong eddy-driven transport is a misconception. In the Southern Ocean there is increasing evidence of mixed instability near topography, where both barotropic and baroclinic instability mechanisms contribute to the dynamics (Barthel et al., 2017; Foppert, 2019; Youngs et al., 2017). The distinct role of each instability mechanism, and their interaction, need to be considered when developing eddy transport parameterizations that will respond physically to changes in ocean dynamics.

Data Availability Statement
The simulation data and scripts used in the study are freely available on the Zenodo repository at DOI:10.5281/zenodo.2542957.

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