Energetic Submesoscale Dynamics in the Ocean Interior

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ABSTRACT

Submesoscale ocean processes, characterized by order-1 Rossby and Richardson numbers, are currently thought to be mainly confined to the ocean surface mixed layer, whereas the ocean interior is commonly assumed to be in quasigeostrophic equilibrium. Here, a realistic numerical simulation in the Antarctic Circumpolar Current, with a $1/48^\circ$ horizontal resolution and tidal forcing, is used to demonstrate that the ocean interior departs from the quasigeostrophic regime down to depths of 900 m, that is, well below the mixed layer. Results highlight that, contrary to the classical paradigm, the ocean interior is strongly ageostrophic, with a pronounced cyclone–anticyclone asymmetry and a dominance of frontogenesis over frontolysis. Numerous vortices and filaments, from the surface down to 900 m, are characterized by large Rossby and low Richardson numbers, strong lateral gradients of buoyancy, and vigorous ageostrophic frontogenesis. These deep submesoscale fronts are only weakly affected by internal gravity waves and drive intense upward vertical heat fluxes, consistent with recent observations in the Antarctic Circumpolar Current and the Gulf Stream. As such, deep submesoscale fronts are an efficient pathway for the transport of heat from the ocean interior to the surface, suggesting the presence of an intensified oceanic restratification at depth.

1. Introduction

Oceanic mesoscale and submesoscale turbulence has been extensively studied in the past decade (Klein and Lapeyre 2009; Mahadevan 2016; McWilliams 2016). Results emphasize the existence of submesoscale fronts (≤ 50 -km width), predominantly confined to the surface mixed layer (ML) and particularly energetic in winter when ML instabilities are active (Fox-Kemper et al. 2008; Callies et al. 2015). These fronts, mostly produced by co-interacting mesoscale eddies (50–300 km size), are associated with important positive vertical heat fluxes (Su et al. 2018). In contrast, submesoscale vertical heat fluxes in the ocean interior are thought to be small. This is because, in the classical paradigm, motions below the ML are broadly assumed to be in quasigeostrophic

(QG) balance, preventing the formation of strong density gradients at depth.

However, growing evidence suggests that interior ocean dynamics significantly depart from quasigeostrophy and may be strongly ageostrophic, as proposed by Molemaker et al. (2010). These authors show that relaxing the QG assumptions in an idealized model of the ocean interior leads to the emergence of large Rossby number and energetic frontogenesis driven by mesoscale eddies. Seismic imaging has also long revealed the existence of ageostrophic mesoscale eddies (50–100 km) in the ocean interior (Biescas et al. 2008; Menesguen et al. 2009; Barbosa Aguiar et al. 2015), such as the subsurface anticyclones of the North Atlantic Ocean known as “meddies” (Armi et al. 1988) or the coherent eddies of the Gulf Stream (Gula et al. 2019). In addition, two recent in situ studies diagnosed strong upward vertical heat fluxes in the ocean interior (Siegelman et al. 2020; Yu et al. 2019), believed to be produced by ageostrophic dynamics. In particular, Siegelman et al. (2020) reported an enhanced vertical heat flux at deep submesoscale ocean fronts in the Kerguelen region in spring and summer, seasons traditionally associated with weak submesoscales (Sasaki et al. 2014; Callies et al. 2015).

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FIG. 1. Map of buoyancy anomaly (shading; the “e” in the values indicates that the numeral preceding the e should be multiplied by 10 raised to the sign and numerals following it), defined as the anomaly with respect to the domain-averaged value, at 99 m superimposed with contours of SSH ranging from -0.9 to 1.15 m at 0.03 -m interval. The map is a randomly selected snapshot taken at 0200 UTC 28 Oct 2012, which is representative of the time period considered in this study.

Here, ageostrophic dynamics of deep submesoscale ocean fronts are studied in the Antarctic Circumpolar Current (ACC), offering a dynamical explanation for the observational results of Siegelman et al. (2020). The numerical simulation is described in section 2. The Ertel potential vorticity (PV), used to characterize meso–submesoscale turbulence and therefore ocean-scale interactions, is briefly introduced in section 3 in terms of nondimensional numbers. The ageostrophic character of this turbulence, along with frontal dynamics and vertical heat fluxes are analyzed in section 4. Some conclusions and perspectives are provided in section 5.

2. Realistic numerical simulation

A primitive equation global ocean simulation with a horizontal resolution of $1/48^\circ$, 90 vertical levels and internal tides (appendix A) is used to study ocean-scale interactions over a broad range of scales, from 10 km to basin scales. A subdomain of the ACC, just north of the Kerguelen Islands, that spans 55° – 73° E, 40° – 46° S (~ 1300 km $\times$ 700 km) is analyzed. The domain size is sufficiently large to capture part of the large-scale Subantarctic Front (SAF; Fig. 1) (Kim and Orsi 2014), and the model resolution is sufficiently high to resolve multiple mesoscale eddies and to permit the emergence of submesoscale features such as elongated fronts and submesoscale vortices (Fig. 2).

The time period ranges from 15 October to 15 November 2012, that is, late spring in the Southern Hemisphere. This season lies intermediate to the submesoscale-energetic winter and summertime, when submesoscales are thought to be inhibited by shallow ML (Sasaki et al. 2014; Callies et al. 2015). Here, the mixed layer depth

(MLD) is relatively shallow with an average value of 50 m but having local maxima of 100 m within cyclonic eddies.

3. Ertel potential vorticity as a tracer of ocean-scale interactions

The Ertel PV is a key dynamical quantity for the study of a stratified fluid in rotation. It is conserved along a Lagrangian trajectory and is only modified by sources, sinks, and friction. As such, the Ertel PV experiences a direct cascade in which numerous submesoscale filaments are generated by co-interacting mesoscale eddies, the latter emerging from baroclinic instabilities of large-scale flows (Pedlosky 2013a). The Ertel PV is used here to characterize the flow field in terms of mesoscale and submesoscale turbulence.

The Ertel PV can be expressed as

$$q = (f + \zeta)b_z + (\mathbf{k} \times \mathbf{u}_z) \cdot \nabla_H b, \quad (1)$$

where $b = g(1 - \rho/\rho_0)$ is buoyancy, with g being Earth’s gravitational acceleration, ρ being potential density, and ρ_0 being a reference density of 1027.5 kg m^{-3} , f is the Coriolis parameter, $\zeta = v_x - u_y$ is the vertical component of the 3D vorticity vector (i.e., the relative vorticity), $\mathbf{k} \times \mathbf{u}_z = (-v_z, u_z)$ is its horizontal component, and the spatial derivatives of w are neglected (Holton 1973). Subscripts denote partial derivatives.

The Ertel PV can thus be decomposed into two main components: $(f + \zeta)b_z$ and $-v_z b_x + u_z b_y$. It can also be expressed in terms of the nondimensional Rossby number Ro ($\equiv \zeta/f$) and Richardson number Ri [$\equiv f^2 N^2 / (\nabla b)^2$] (assuming thermal wind balance) as

FIG. 2. Maps of (a)–(c) Ertel PV, superimposed with SSH contours (in black) ranging from -0.9 to 1.15 m at 0.1 -m interval, and Ertel PV's (d)–(f) first component $(f + \zeta)N^2$ and (g)–(i) second component $-v_2b_x + u_2b_y$, at (left) 99, (center) 299, and (right) 506 m. All maps are for the same randomly selected snapshot taken at 0200 UTC 28 Oct 2012. The white dashed lines in (d) correspond to the vertical sections presented in Figs. 3, 4, 7, and 14, below. The red lines in (d) correspond to the two submesoscale fronts presented in Fig. 13, below.

$$q \approx fN^2(1 + \text{Ro} - \text{Ri}^{-1}), \quad (2)$$

where $N^2 \equiv b_z$ is the Brunt–Väisälä frequency squared (Thomas et al. 2008).

For QG flows, $\text{Ro} \ll 1$ and $\text{Ri} \gg 1$, which corresponds to a balance between pressure and Coriolis forces at leading order and a small isopycnal slope, respectively. For ageostrophic flows, $\text{Ro} = O(1)$ and $\text{Ri} \leq 1$, indicating a break of geostrophic balance and a large isopycnal slope, respectively. Commonly accepted values for Ro in an ageostrophic regime start at ~ 0.3 – 0.5 , corresponding to Ri^{-1} of ~ 0.1 – 0.2 . This is because $\text{Ro}^2 = \text{Ri}^{-1}$ when the horizontal length scale of the flow is close to the first Rossby radius of deformation (Molemaker et al. 2005). In the next sections, ageostrophy is explored in terms of the Ertel PV, Ro , and Ri .

4. Results

a. Ocean-scale interactions in the Antarctic Circumpolar Current

1) LARGE-SCALE BACKGROUND FLOW

The large-scale meandering SAF has a width of $O(100\text{ km})$ and strong currents reaching 1.5 m s^{-1} and separates dense waters in the south from light waters in the north (Fig. 1). This geostrophic jet, is, to leading order, in thermal wind balance within the permanent thermocline, which exhibits a sharp stratification gradient; N^2 increases from 3×10^{-5} to $5.5 \times 10^{-5}\text{ s}^{-2}$ over just 50 km from south to north and the permanent thermocline's depth deepens from 200 to 700 m (see at 43.5°S in Fig. 3a). This is reflected in the large-scale features of Ertel PV that are essentially governed by fN^2 [Eq. (2)] and follow contours of sea surface height (SSH; Fig. 2). North of the SAF and down to 506 m, the Ertel PV has a low magnitude of $\sim 3 \times 10^{-9}\text{ s}^{-3}$, indicating the presence of moderately stratified fluid ($N/f \sim 160$) sitting above the permanent thermocline. South of the SAF and starting from 99 m, the Ertel PV has a high magnitude of $\sim 5 \times 10^{-9}\text{ s}^{-3}$, indicating the presence of strongly stratified fluid within the permanent thermocline ($N/f \sim 215$). Large-scale patterns of Ertel PV become more prominent with depth (Fig. 2), covering the quasi totality of the domain at 900 m (not shown).

Numerous mesoscale cyclones and anticyclones are released through baroclinic instabilities of the SAF. The eddies have a size of ~ 150 and 50 km south and north, respectively, of the SAF (Fig. 1). This striking size difference is a consequence of the stratification difference that is reflected in the first Rossby radius of deformation, $L_d = NH/f$, with H being the depth scale of the flow, corresponding to the depth of the main thermocline, and

f being the Coriolis frequency. Mesoscale buoyancy anomalies (in color in Fig. 1) are mostly present in the vicinity of the SAF and are consistent with the thermal wind balance, that is, positive within anticyclones and negative within cyclones. However, there is a slight phase shift between these mesoscale buoyancy anomalies and large-scale SSH contours, as can be seen at 43°S , 62°E (Fig. 1). This offset is key to the generation of sub-mesoscale buoyancy fronts and filaments, because the strain field will then be able to stretch and compress these background anomalies (Klein et al. 2019). These large- and mesoscale characteristics are common to the two other most energetic currents: the Kuroshio Extension (Sasaki et al. 2014) and the Gulf Stream (Chassignet and Xu 2017).

2) MESO- AND SUBMESOSCALE TURBULENCE

Meso- and submesoscale turbulence is characterized by the spontaneous emergence of numerous small-scale filaments and vortices with a horizontal size of tens of kilometers (McWilliams 2016), not directly identifiable in SSH (Fig. 1) but evident in tracer fields, such as the Ertel PV (Klein et al. 2019).

Meso- and submesoscale features are strongly heterogeneous throughout the domain, as can be seen in the maps of Ertel PV (Figs. 2a–c). North of the SAF, numerous small eddies ($\sim 50\text{ km}$ in size) are associated with negative Ertel PV anomalies extending down to at least 506 m. These eddies are surrounded by filaments of Ertel PV at their periphery and in between them, elongated over distances of tens of kilometers. Instances of depth-intensified turbulence also occurs, such as at 506 m around 41.5°S , 69°E (Fig. 2c). South of the SAF, that is, in the large-scale-dominated and high-Ertel-PV region, few large mesoscale eddies ($\sim 200\text{ km}$ size) are mostly confined to the first 200–300 m. They are associated with elongated filaments over distances greater than 100 km.

The Ertel PV is mostly explained by its first component, that is, $(f + \zeta)N^2 = fN^2(1 + \text{Ro})$, as can be seen on the Ertel PV decomposition shown in Fig. 2, highlighting the dominant contribution of the relative vorticity at meso and submesoscale. The second component, that is, $-v_z b_x + u_z b_y \approx -fN^2\text{Ri}^{-1}$, is generally an order of magnitude lower than the first one, except at the location of strong submesoscale fronts where they become comparable (see section 4b). This latter component is mostly dominant in the vicinity of the SAF and its width increases with depth (Figs. 2g–i).

The ageostrophic character of the ocean interior becomes obvious on vertical sections of Ertel PV (Figs. 3 and 4), especially north of the SAF where the permanent thermocline can be deeper than 500 m. Three classes of eddy emerge. First, surface-trapped mesoscale eddies

FIG. 3. (left) Meridional and (right) zonal vertical sections at 0200 UTC 28 Oct 2012 of (a), (b) vertical stratification N^2 , (c), (d) Ertel PV, (e), (f) Ertel PV's first component $(f + \zeta)N^2$, (g), (h) Ertel PV's second component $-v_z b_x + u_z b_y$, and (i), (j) relative vorticity normalized by ζ/f . The two sections are highlighted by the white dashed lines in Fig. 2d. The red arrows in (c) and (d) correspond to the two submesoscale fronts presented in Fig. 13, below.

FIG. 4. As in Fig. 3, but for the other zonal and meridional transects highlighted by the white dashed lines in Fig. 2d.

are conspicuous in the domain, see at 62.5°E in Fig. 4a for instance. They extend down to 100–200 m, have a horizontal size of about 50 km, and are associated with low Ertel PV (Figs. 4c,d). They are mostly anticyclones ($\zeta/f < 0$), characterized by $|\text{Ro}| \geq 0.5$ and surrounded by positive rings of ζ/f (Figs. 4i,j), which explains the high-Ertel-PV rings located at their periphery. Strikingly, these eddies are encircled by strong vertical and horizontal gradients of buoyancy, apparent in both N^2 (Figs. 4a,b) and the second component of the Ertel PV (Figs. 4g,h). Since the Ertel PV and buoyancy fields are conserved along a Lagrangian trajectory, these submesoscale fronts should act as dynamical barriers that prevent mesoscale eddies from getting destroyed by their interaction with neighboring ones. As a consequence, mesoscale eddies become more coherent and energetic and their lifetime increases (Mariotti et al. 1994).

Second, subsurface mesoscale eddies are ubiquitous throughout the domain—see at 57°E in Fig. 4a or at 41.8° and 42.5°S in Fig. 4b for instance. Their vertical extent is of 400–600 m, and their horizontal size is 50–100 km. They have a less pronounced surface signature than surface-trapped ones (Fig. 2). Once again, these eddies are mostly anticyclones associated with low Ertel PV anomalies and surrounded by rings of ζ of opposite sign as well as enhanced buoyancy gradients, explaining the contours of high Ertel PV located at their edges. High $|\text{Ro}|$ (≥ 0.5) is found both in their core and at their periphery. These subsurface eddies may result from the subduction of low buoyancy waters located south of the SAF below higher buoyancy waters north of the SAF (Fig. 1). These eddies are also reminiscent of those recently observed in the Gulf Stream (Gula et al. 2019) as well as of the well-known meddies encountered in the North Atlantic (Menesguen et al. 2009; Barbosa Aguiar et al. 2015).

Third, deep mesoscale eddies with almost no surface signature are present, see for example at 41°S in Fig. 3a and 69°E in Fig. 3b. These eddies appear to emerge from the instabilities of the permanent thermocline north of the SAF, contributing to inject deep and strongly stratified fluid in upper oceanic layers (see at 69°E in Fig. 3b where waters at 600 m are injected up to 200 m). Contrary to the previous two types, they are mostly cyclones ($\zeta/f > 0$) associated with anomalies of high Ertel PV that is explained by its first component (Fig. 3h). Yet, similar to surface-trapped and subsurface eddies, they are surrounded by intense submesoscale buoyancy gradients and rings of relative vorticity of opposite sign (Figs. 3i,j).

The ageostrophic character of the flow field down to 506 m, that is, well below the mixed layer, also occur away from the SAF, in the northern part of the domain (e.g., at 60°–66°E in Fig. 3b); high $|\text{Ro}| \geq 0.5$ is associated

with numerous submesoscale structures above the permanent thermocline (600–800 m; Figs. 3b,j). These submesoscale features exhibit alternating low and high Ertel PV anomalies (Fig. 3d), mostly explained by the relative vorticity (Fig. 3j). Indeed, they have a very weak signature on buoyancy gradients (Fig. 3h).

South of the SAF, shallow eddies above the permanent thermocline (~ 200 m, Fig. 3a) have similar characteristics as surface-trapped ones north of the SAF (Fig. 3b); they are surrounded by filaments of ζ of opposite sign and enhanced buoyancy gradients. Within the permanent thermocline, an unexpected signature of internal gravity waves (IGWs) is present on the Ertel PV and relative vorticity fields. IGWs exhibit patterns of radial and crisscross beams, as can be seen at 43.5°–45.5°S in Figs. 4d, 4f, and 4j. However, they should not impact the spatial distribution of Ertel PV because linear waves do not transport material or tracer (Kundu and Cohen 2004). Nevertheless, these IGWs can have a local and transient signature as their high vertical velocity (Pedlosky 2013b; Kundu and Cohen 2004) may affect the buoyancy field. This interesting feature is further developed below.

The simulation includes energetic IGWs, comprising internal tides, near-inertial motions and a large IGWs continuum at higher frequencies. As such, it is essential to be able to disentangle them from balanced motions, that is, flows in thermal or gradient wind balance that encompass meso and submesoscales (Klein et al. 2019). To do so, we use frequency–wavenumber (ω – k) spectra following the methodology described in Torres et al. (2018). The dispersion relation curve associated with the highest baroclinic mode resolved by the model (10th baroclinic mode, curved dashed line in Fig. 5) partitions IGWs, located above the curve, and balanced motions, located below the curve [see appendix B and Torres et al. (2018) for more details]. The ω – k spectra of KE, Ertel PV, ζ , and ∇b are shown at 39, 299, and 900 m (Fig. 5). However they have a similar shape from the surface down to 506 m and decrease linearly with depth (partially shown for 39 and 299 m). As expected, most of the KE is contained at mesoscale (>50 km) and low frequencies at both 39, 299, and 900 m (Figs. 5a,e,i). While the impact of IGWs is visible, especially for M2 tidal motions and near-inertial waves, their relative contribution is weak compared to that of balanced motions. The Ertel PV exhibits a different distribution. At 39 m, most of the variance is contained at submesoscales (<50 km, Fig. 5b). At 900 m, the Ertel PV is less energetic and the variance is contained at both meso and submesoscale (Fig. 5j). Once again, the impact of IGWs on the Ertel PV is substantially weaker than that of balanced motions. The relative vorticity field is even less affected by IGWs. At 39 m (Fig. 5c), most of the variance is

FIG. 5. Frequency-wavenumber spectra computed from 15 Oct to 15 Nov 2012 at (top) 39, (middle) 299, and (bottom) 900 m of (a),(e),(i) kinetic energy (KE), (b),(f),(j) EPV (EPV), (c),(g),(k) relative vorticity ζ , and (d),(h),(l) lateral gradients of buoyancy ∇b . These spectra are presented in a variance-preserving form, which allows one to directly compare the relative contribution of different time and spatial scales with the total variance (appendix B).

FIG. 6. Maps of (left) lateral gradients of buoyancy $|\nabla b|$ superimposed with SSH contours ranging from -0.9 to 1.15 m at 0.1 -m interval and (right) the frontogenesis function F_s at (a),(b) 99, (c),(d) 299, and (e),(f) 506 m. The maps are a randomly selected snapshot taken at 0200 UTC 28 Oct 2012. The dashed rectangle in (b) corresponds to an active submesoscale area used in Fig. 15 (below) to compute vertical heat fluxes.

contained at high wavenumbers, whereas at 900 m (Fig. 5k) the variance is distributed between meso and submesoscales, similar to the Ertel PV. Consistent with the features observed in physical space, ω - k spectra of ζ are similar to those of Ertel PV's first component $(f + \zeta)N^2$ (Figs. S1a-c in the online supplemental material). Lateral gradients of buoyancy $|\nabla b|$ are not affected by IGWs down to 506 m (Figs. 5d,h) but they are at 900 m (Fig. 5l). Down to 506 m, the variance is principally captured by submesoscales. At 900 m, the variance distribution is similar to the Ertel PV. Consistent with the features observed in physical space, ω - k spectra of $|\nabla b|$ are remarkably similar to those of Ertel PV's second component $-v_z b_x + u_z b_y$ (Figs. S1b-d in the online supplemental material). The variance of ∇b is mostly explained by submesoscales down to 506 m. At 900 m, the variance of ∇b is smaller. Overall, ω - k spectra

highlight that the Ertel PV and its components are principally explained by scales ≤ 50 km and that IGWs have only a weak impact. As such, these results emphasize the existence of energetic submesoscales in the ocean interior, and in particular of energetic frontal dynamics over the domain and time period considered in this study.

b. Ageostrophic dynamics in terms of lateral gradients of buoyancy, Ro, and Ri

1) LATERAL GRADIENT OF BUOYANCY

Lateral gradients of buoyancy $|\nabla b|$ are of particular interest because of their link with frontogenesis processes associated with large vertical buoyancy and heat fluxes (Hoskins and Bretherton 1972). Strong $|\nabla b|$ are mostly at submesoscale (Figs. 6a,c,e). They are found from the surface down to 900 m along the SAF as well as

FIG. 7. Four vertical sections of buoyancy anomaly b' (shading; the “e” in the values indicates that the numeral preceding the e should be multiplied by 10 raised to the sign and numerals following it), superimposed with $|\nabla b| > 5 \times 10^{-8} \text{ s}^{-2}$ (black contours) corresponding to the dashed white lines in Fig. 2d. The mixed layer depth is shown in gray and corresponds to a density increase of 0.03 kg m^{-3} from the density at 10 m. Shown is a randomly selected snapshot taken at 0200 UTC 28 Oct 2012.

at the periphery and in between mesoscale eddies, that is, in strain dominated regions, as inferred from SSH contours (Figs. 6a,c,e). They have a width of $\sim 10 \text{ km}$ and are meandering over length scales ranging from 50 km , in the intense submesoscale area located at the center of the domain (43°S , 63.5°E), to several hundreds of kilometers along the SAF. The typology of $|\nabla b|$ is very rich; whereas $|\nabla b|$ in the intense submesoscale area tend to be concentrated in the upper hundred meters of the water column, other $|\nabla b|$ have an enhanced subsurface signature, which is especially the case for subsurface eddies located north of the SAF (Figs. 6a,c,e and 7). Overall, $|\nabla b|$ are similarly distributed as Ertel PV's

second component $-v_z b_x + u_z b_y$ (Figs. 2g–i, 3g–h and 4g–h) as most of $|\nabla b|$ are in thermal wind balance in the alongfront direction (in red in Figs. 2g–i).

Furthermore, the 3D distribution of $|\nabla b|$ is remarkably consistent with that of mesoscale buoyancy anomalies. A striking relation between mesoscale buoyancy anomalies and submesoscale $|\nabla b|$ emerges (Fig. 7). One can observe the classical bowl shape of the buoyancy field within anticyclone, such as between 57° and 59°E from 0 to 350 m , and reverse bowl shape within cyclone, such as at 69° – 70°E between 300 and 800 m (Fig. 7b). Intense $|\nabla b|$, highlighted by the black contours in Fig. 7, are located at the periphery of mesoscale eddies, both

FIG. 8. Cumulative distribution of (a) $|\nabla b|$, (b) $|\zeta/f|$, (c) $F_s/(\nabla b)^2$ and (d) strain rate at different depths over the entire domain for a randomly selected snapshot taken at 0200 UTC 28 Oct 2012. The horizontal dashed line corresponds to $y = 0.25$.

within and below the MLD, consistent with the maps of $|\nabla b|$ (Figs. 6a,c,e). While some $|\nabla b|$ extend from the surface down to depths of 700 m, following deep reaching buoyancy anomalies (at 66.5°E in Fig. 7c for instance), other $|\nabla b|$ are only present at depth with a weak—or no—surface signature. This is the case for deep eddies (Figs. 7b–d). $|\nabla b|$ are slanted and follow the bowl-shaped buoyancy anomalies. Such slanted $|\nabla b|$ are known to result from the competition between horizontal strain and vertical shear (Haynes and Anglade 1997; Klein et al. 1998; Meunier et al. 2015).

Statistics over the domain indicate that $|\nabla b|$ at 39 m, that is, mostly within the ML, reaches values of up to $2.6 \times 10^{-7} \text{ s}^{-2}$ (Fig. 8a). These values are close to, although smaller than, what is found in the literature in both observational and numerical studies. Indeed, Siegelman et al. (2019, 2020) reported values of up to $4 \times 10^{-7} \text{ s}^{-2}$ using in situ observations collected by southern elephant seals in this region in spring and summer, whereas Rosso et al. (2014) documented $|\nabla b|$ of up to $5 \times 10^{-7} \text{ s}^{-2}$ at 50 m in a high-resolution model at 1/80° also in this region. Note that the lower $|\nabla b|$ obtained here compared to

Rosso et al. (2014) is likely due to the lower resolution of our model. However, a key and surprising result concerns the quasi-constant magnitude of $|\nabla b|$ with depth. Indeed, $|\nabla b|$ still reach values greater than $2 \times 10^{-7} \text{ s}^{-2}$ below the ML and cumulative distributions of $|\nabla b|$ at different depths are broadly similar down to 506 m: $|\nabla b| > 5 \times 10^{-8} \text{ s}^{-2}$ account for 25% of the domain at 39 and 99 m, 20% at 299 and 506 m, and 15% at 900 m (Fig. 8a). This weak depth dependence of $|\nabla b|$, once again, strongly contrasts with QG dynamics. Indeed, buoyancy anomalies in the QG regime have a spectral slope in k^{-2} near the surface and k^{-5} in the ocean interior, that is, a k^{-3} difference between surface and depth (Hua and Haidvogel 1986; Smith and Ferrari 2009; Molemaker et al. 2010). Here, the wavenumber spectrum of buoyancy displays a slope in k^{-2} down to 99 m and in $k^{-2.5}$ below (Fig. 9), which highlights the ageostrophic character of the dynamics at depth. The k^{-3} difference in spectral slope between the QG and ageostrophic regimes also applies to $|\nabla b|$. This implies that, if a mesoscale buoyancy gradient with a size of 100 km is similar at surface and depth, $|\nabla b|$ with a size of 10 km should be smaller by a

FIG. 9. Wavenumber spectrum of buoyancy at different depths over the entire domain for a randomly selected snapshot taken at 0200 UTC 28 Oct 2012.

factor of 30 at depth than at the surface, or smaller by a factor of 1000 in terms of variance. This is clearly not the case here (Figs. 5d,h,i), highlighting the striking departure from QG dynamics in the ocean interior. The ω - k spectra of $|\nabla b|$ also emphasize the importance of scales < 50 km and the weak impact of IGWs (Figs. 5d,h,i), consistent with the features in physical space.

2) ROSSBY NUMBER

The signature of many submesoscale eddies and filaments is apparent in ζ/f , at 39 (not shown), 99 (Fig. 10a), 299 (Fig. 10c), and 506 m (Fig. 10e). Similar to the maps of Ertel PV (Figs. 2a–c), larger eddies are found south of the SAF, whereas deeper subsurface eddies are mostly found north of the SAF, starting at 299 m. Filaments of ζ/f have a deep vertical extent reaching 500 m, as can be seen in Figs. 3i,j and 4i,j, consistent with Ertel PV's vertical structure (Figs. 3c,d and 4c,d). Interestingly, deep-reaching filaments of ζ/f are collocated with weak vertical stratification (Figs. 3a,b,i,j and 4a,b,i,j), highlighting the almost two-dimensional character of oceanic turbulence in this region and season (McWilliams 1984).

Large $|\text{Ro}|$ are found not only near the surface but also in the ocean interior, down to 900 m. Ro ranges from -1.8 to 3.5 at 39 m (not shown), between -1.7 and 3.1 at 99 m (Fig. 11a), between -1.4 and 2.1 at 299 m (Fig. 11b), between -1.4 and 1.8 at 506 m (Fig. 11c), and between -0.7 and 0.7 at 900 m (not shown), highlighting the rare occurrence of inertial instability ($\zeta/f < -1$) down to 506 m. Cumulative histograms indicate that 20% of $|\text{Ro}|$ are larger than 0.45 at 39 m, 0.3 at 299 m and even

0.25 at 506 m (Fig. 8b). These values are remarkably large given the relatively moderate model resolution (~ 1.7 -km horizontal resolution). As suggested by the results of the previous section, the distribution of Ro is positively skewed down to 299 m, with skewness values of 0.66 at 39 m, 0.42 at 99 m, 0.19 at 209 m, 0.07 at 299 m, -0.05 at 506 m, and -0.53 at 900 m. These skewness values indicate a dominance of cyclones ($\zeta/f > 0$) down to 299 m and a dominance of anticyclones ($\zeta/f > 0$) from 506 to 900 m, consistent with the idealized results of Roulet and Klein (2010). These domain-wide statistics further confirm the strong departure from QG, not only within the mixed layer, but also in the ocean interior.

Filaments and vortices of Ro (Figs. 10a,c,e) can be separated with the Okubo–Weiss quantity (appendix C). Using this partition, Roulet and Klein (2010) showed that filaments of vorticity have a larger positive skewness than vortices, said to be due to ageostrophic frontogenesis. This is quantified over the entire domain using the same Okubo–Weiss partitioning as Roulet and Klein (2010), aimed at excluding vortices of Ro . Results show that the relationship between $|\nabla b|$ and filaments of Ro is asymmetric and positively correlated down to 506 m; negative or positive filaments of Ro are collocated with weak or strong $|\nabla b|$, respectively (Figs. 12a–c). At 900 m, the relation becomes symmetrical, suggesting a quasigeostrophic regime with the absence of cyclone–anticyclone asymmetry (Fig. 12d).

3) INVERSE RICHARDSON NUMBER

Similar to Ro , the signature of many submesoscale vortices and filaments is apparent in the inverse Richardson

FIG. 10. As in Fig. 6, but for the (left) Rossby number ζ/f and (right) inverse Richardson number Ri^{-1} .

number, defined by $Ri^{-1} \equiv (\nabla b/fN)^2$ assuming thermal wind balance, down to 506 m (Figs. 10b,d,f). The 3D distribution of Ri^{-1} closely resembles that of $|\nabla b|$ (Figs. 6a,c,e), and large Ri^{-1} are collocated with large $|\nabla b|$. The Ri^{-1} reaches 1.5 at 99 m, 0.5 at 299 m, 0.2 at 506 m, and 0.13 at 900 m. These large Ri^{-1} (≥ 0.1 –0.2) further highlight the ageostrophic nature of the ocean interior over the entire domain, known to be conducive to intense frontogenesis. Note that the thermal wind balance is a reasonable assumption that only breaks at a few locations south of the SAF and below the main thermocline. These locations can be identified by the negative values of the second component of the Ertel PV (Figs. 2g–i). A possible explanation for the thermal wind imbalance is the impact of IGWs, that are present below the main thermocline south of the SAF (Figs. 3c–d and 4c–d), and are associated with a temporal term that breaks the thermal wind balance, as suggested by Danioux et al. (2012). This can also be seen in the ω – k spectrum of

the Ertel PV (Fig. 5f), where a part of the variance is captured by IGWs (above the dashed line).

Overall, these results highlight the generation mechanism of ageostrophic submesoscale fronts by mesoscale turbulence (Klein and Lapeyre 2009), not only within the ML but also in the ocean interior, as further detailed in the next section for two submesoscale fronts.

c. Ageostrophic frontal dynamics

1) CASE STUDY OF AGEOSTROPHIC FRONTAL DYNAMICS

In this section, frontal dynamics associated with $|\nabla b|$ in the ocean interior are examined for two submesoscale fronts with a width of ~ 10 km. One front is located at the SAF boundary (43.5°S , 71.3°E) and the other is on the edge of an elliptic mesoscale eddy in the northwest of the domain (41.8°S , 59.1°E), as inferred from SSH contours in Fig. 1. Both fronts are identified by the red arrows

FIG. 11. Histograms of (a)–(c) $|\zeta/f|$ and (d)–(f) $F_s/(\nabla b)^2$ at (left) 99, (center) 299, and (right) 506 m over the entire domain for a randomly selected snapshot taken at 0200 UTC 28 Oct 2012.

on the vertical sections of Ertel PV in Fig. 3b and by the red lines on the horizontal map of Ertel PV in Fig. 2d. Their departure from QG is quantified in terms of the Rossby and Richardson numbers. The nature of the frontal dynamics is investigated with the frontogenesis function F_s defined as

$$F_s \equiv -\mathbf{Q} \cdot \nabla b, \quad (3)$$

where $\mathbf{Q}$ is the frontogenetic vector of Hoskins et al. (1978), which can be expressed as

$$\mathbf{Q} = \begin{pmatrix} u_x & v_x \\ u_y & v_y \end{pmatrix} \cdot \begin{pmatrix} b_x \\ b_y \end{pmatrix}. \quad (4)$$

Since the equation of the evolution of a buoyancy gradient is given by

$$\frac{1}{2} \frac{d|\nabla b|^2}{dt} = F_s + \nabla w \cdot \nabla b, \quad (5)$$

with w being the vertical velocity field (Hoskins 1982). A positive F_s indicates the presence of frontogenesis, and a negative F_s indicates the presence of frontolysis (i.e., frontal destruction).

The exact location of the submesoscale fronts is apparent in the buoyancy anomaly field (blue curve in Fig. 13), which exhibits a sharp jump down to 299 m in both fronts. The fronts are associated ζ/f (red curve) ranging from -1 to 2 at 39 m, from -1 to 1 at 99 m, from -0.75 to 0.75 at 209 m, and from -0.5 and 0.5 at

both 299 and 506 m. These high $|\text{Ro}|$ (>0.5) are the signature of an ageostrophic regime down to 506 m. In addition, both fronts are associated with high inverse Richardson number (Ri^{-1} ; green curve). Similar to Ro , Ri^{-1} decreases with depth: Ri^{-1} reaches 0.3 and 0.6 at 39 m, 0.3 and 0.4 at 99 m, 0.25 and 0.4 at 209 m, 0.2 and 0.23 at 299 m, and only 0.05 and 0 at 506 m, for the front in the SAF and the one at the eddy's periphery, respectively. These high Ri^{-1} (≥ 0.2) further confirm the ageostrophic character of deep-reaching submesoscale fronts. Last, the frontogenesis function F_s spikes at the location of these fronts, from the surface down to 299 m, highlighting their frontogenetic nature. F_s is on the order of 10^{-18} – 10^{-17} s^{-5} , which, when normalized by $|\nabla b|^2$, leads to rapid time scales from one to several hours. Interestingly, instance of frontolysis are often observed on one side of the front adjacent to frontogenesis (see at 99 m in Fig. 13c and at 299 m in Fig. 13h, for instance). Frontolysis alone also occurs (see at 506 m in Fig. 13i), indicating the front's total collapse. Furthermore, the slanted shape of submesoscale fronts discussed in the previous section is clearly visible in Fig. 13, as emphasized by the lateral shift of the fronts with depth.

A conspicuous asymmetry between the dense ($b' < 0$) and light ($b' > 0$) side of the fronts is present, not only at the surface, but also at depth. Cyclonic vorticities ($\zeta/f > 0$; dense side) are considerably stronger than anticyclonic vorticities ($\zeta/f < 0$; light side) and frontogenesis ($F_s > 0$) occurs on the dense side of the front. Asymmetrical

FIG. 12. Scatterplots of $|\nabla b|$ and ζ/f conditioned by Okubo–Weiss normalized by f^2 such that $\sim 20\%$ of the domain points are kept so as to capture the filaments and exclude the vortices, following the method described in Roulet and Klein (2010) (see main text and appendix C) at (a) 99, (b) 299, (c) 506, and (d) 900 m. The scatterplots are computed over the entire domain for a randomly selected snapshot taken at 0200 UTC 28 Oct 2012. Each point represents the average over each grid interval on the abscissa (that has a total of 200 grid intervals), and thin vertical lines show the std dev around the averages. A strong asymmetry between positive and negative ζ/f and ∇b is present down to 506 m, highlighting an ageostrophic regime. However, at 900 m the relation is symmetrical, suggesting a QG regime, as discussed in section 4.

frontogenesis is a distinctive characteristic of strongly ageostrophic flows (Hoskins 1982; McWilliams et al. 2009; Hakim and Keyser 2001; Capet et al. 2008) as it leads to a secondary circulation tilted along isopycnals, whereas the QG circulation cell is not tilted (Hakim and Keyser 2001). As a result, the vertical velocity field is negatively skewed, which further amplifies the skewness of Ro through the vorticity equation (Hoskins 1982; Hakim et al. 2002; Thomas et al. 2008).

The frontal dynamics considered here are complex because the fronts are not idealized nor isolated. Multiple submesoscale fronts of varying intensity are present in the vicinity of a stronger one, as can be seen in the successive occurrences of positive and negative Ro of different magnitudes (Fig. 13). Consequently, it is challenging to

derive a pointwise correspondence between the different quantities considered here, which was also the case for the nice results of Capet et al. (2008). However, we refer the reader to the idealized surface front presented in Fig. 4 of Thomas et al. (2008), where these relations are flabbergasting.

Overall, these results highlight the ageostrophic character of deep-reaching submesoscale fronts, characterized by large Ro and Ri^{-1} and positive F_s . These findings point to a positive skewness F_s associated with these fronts over the entire domain, as explored in the next section.

2) AGEOSTROPHIC FRONTOGENESIS

The 3D distribution of the frontogenesis function F_s closely resembles that of $|\nabla b|$. Strong F_s coincide with

FIG. 13. Horizontal profiles across two individual submesoscale fronts at different depths, described in the main text: (a),(b) 39, (c),(d) 99, (e),(f) 209, (g),(h) 299, and (i),(j) 506 m. In all panels, the buoyancy anomaly b' is in blue, the frontogenesis function F_s is in black, the Rossby number ζ/f is in red, and the inverse Richardson number Ri^{-1} is in green. These sections are shown by the red lines in Fig. 2d and the red arrows in Figs. 3c and 3d.

strong $|\nabla b|$ at all depths and F_s is characterized by filamentary structures with a width of ~ 10 km down to 299 m. At depth, filaments of F_s are concentrated along the SAF (Fig. 6). This spatial correspondence confirms that submesoscale fronts are generated by frontogenesis processes in the ocean interior.

The time scale associated to frontogenesis and frontolysis can be retrieved from $F_s/(\nabla b)^2$ [Eq. (5)]. RMS values of $F_s/(\nabla b)^2$ range from 1.7×10^{-5} to 10^{-5} s^{-1} between the surface and 500 m, corresponding to time scales from 10 h to a day. Twenty percent of frontogenetic processes [$F_s/(\nabla b)^2 > 0$] are larger than $2 \times 10^{-5} \text{ s}^{-1}$ —that is, 14 h—at 39 m and $1 \times 10^{-5} \text{ s}^{-1}$ —that is, a day—at 900 m (Fig. 8c). Similar results are obtained for frontolytic processes [$F_s/(\nabla b)^2 < 0$] (Fig. 8c). These time scales are consistent with the strain rate, for which 20% of the values are larger than $5 \times 10^{-5} \text{ s}^{-1}$ —that is, 12 h—at 39 m and $1.5 \times 10^{-5} \text{ s}^{-1}$ —that is, 19 h—at 900 m (Fig. 8d).

The rms of positive $F_s/(\nabla b)^2$ is greater than that of negative $F_s/(\nabla b)^2$ by a factor of at least 1.4, regardless of depth, indicating that frontal creation is faster than frontal collapse. Similar to Ro, $F_s/(\nabla b)^2$ is positively skewed, with skewness values of ~ 0.2 down to 506 m (Figs. 11d–f), highlighting the dominance of frontogenesis over frontolysis in the ocean interior. Overall, the strong asymmetry of F_s in the upper 500 m emphasizes once again the ageostrophic character of frontogenesis in the ocean interior, which is known to be associated with enhanced vertical velocities and vertical heat fluxes.

d. Vertical velocities and vertical heat flux

Vertical sections of daily-averaged vertical velocities w reveal positive and negative w of up to 500 m day^{-1} (Figs. 14a,c,e,g). Structures of w have a width of ~ 10 – 20 km. They are intensified in the ocean interior, below the mixed layer down to at least 900 m. Strong

FIG. 14. Vertical sections of (left) vertical velocities w superimposed with $|\nabla b| > 5 \times 10^{-8} \text{ s}^{-2}$ shown in black and (right) VHF for the zonal and meridional sections highlighted by the white dashed lines in Fig. 2b. The maps are for a randomly selected snapshot taken at 0200 UTC 28 Oct 2012. The mixed layer depth is shown in gray.

buoyancy gradients (in black in Figs. 14a,c,e,g) are collocated with large w . The signature of divergence is apparent as intensified and thin vertical features of w are collocated with ζ in areas of weak vertical stratification (at 41°–43°S in Fig. 14a or at 68°E in Fig. 14c for example). Instances of strong w inside subsurface eddies also occur, with ∇b acting as a barrier (at 42.5°S in Fig. 14g). w are impacted by IGWs, as can be seen at 44.5°S in Fig. 14g. The rms of w over the domain is maximal at 350 m with a value of 45 m day^{-1} and is 20 m day^{-1} at both 50 and 900 m (dashed blue curve in Fig. 15).

Vertical heat fluxes (VHF) are estimated from temperature and vertical velocity anomalies as $\text{VHF} = \rho_0 C_p w' T'$, where the prime refers to the anomaly with respect to the domain-averaged value and $C_p = 3985 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of seawater (Figs. 14b,d,f,h). A positive or negative value respectively indicates an upward or downward heat flux. Positive values result from frontogenesis, whereas negative values arise from frontolysis. VHF have a local amplitude reaching 2000 W m^{-2} down to at least 900 m. VHF are enhanced at the location of strong submesoscale fronts that border surface trapped, subsurface and deep mesoscale eddies. Time and domain-averaged VHF ($\langle \text{VHF} \rangle$) are positive and enhanced in the ocean interior relative to the first 50 m, that is, below the ML (Fig. 15), shedding new light on the diabatic nature of the ocean interior. In the active submesoscale area (dashed rectangle in Fig. 6b), $\langle \text{VHF} \rangle$ over 5 days reach values of up to 370 W m^{-2} at 150 m and remain surprisingly large at depth; 260 W m^{-2} at 500 m and 140 W m^{-2} at 900 m (orange curve in Fig. 15). Over the entire domain and one month, $\langle \text{VHF} \rangle$ reaches a maximal value of 260 W m^{-2} at 120 m, 140 W m^{-2} at 50 m, and 30 W m^{-2} at 900 m (green curve in Fig. 15). These findings are consistent with the in situ observations of Siegelman et al. (2020). The ω - k cospectra (appendix C) corroborate these results and further show that, down to 299 m, VHF are explained by scales $< 50 \text{ km}$ and frequencies corresponding to time scales from a few hours to a few days (Figs. 16a–c). At 506 m (Fig. 16d) and below (not shown), VHF are explained by scales of 30–150 km. A key and striking result is that linear IGWs do not impact VHF, which are predominantly explained by balanced motions. However, this does not fully exclude the impact of non-linear IGWs and their interactions with balanced motions (Thomas 2017). These results shed light on the efficient pathway for the transport of heat from the ocean interior to the surface enabled by deep submesoscale fronts.

5. Summary and perspectives

The region considered in this study is a subdomain of the ACC, which is sufficiently large to capture ocean-scale

FIG. 15. RMS of daily-averaged vertical velocities over the entire domain on 28 Oct 2012 (dashed blue curve), along with domain-averaged vertical heat fluxes (VHF) over the entire domain and one month (15 Oct–15 Nov 2012) (orange curve) and over the active submesoscale area (dashed black rectangle in Fig. 6d) and 5 days (26–31 Oct 2012) (green curve). The $\langle \text{VHF} \rangle$ are directed upward (positive values), and they are enhanced at depth, especially below the mixed layer.

interactions, from the large-scale meandering SAF, to multiple mesoscale eddies and numerous submesoscale fronts. Mesoscale eddies emanating from the SAF interact and coexist on both sides of the jet, down to a least 506 m. These surface-trapped, subsurface and deep mesoscale eddies are associated with submesoscale structures characterized by large $|\text{Ro}|$ (> 0.5) dominated by cyclonic vorticity ($\zeta/f > 0$). Intense gradients of Ertel PV, Ro, and buoyancy are located at their periphery, acting as ageostrophic dynamical barriers, which increase the eddies' coherence and lifetime (Mariotti et al. 1994). Note that the generation mechanism of these subsurface and deep eddies remains an open question that is out of the scope of this paper and will be the object of a future study.

Lateral gradients of buoyancy, resulting from the straining generated by mesoscale eddies, have a remarkable weak depth dependence, in stark contrast with QG dynamics but in close agreement with ageostrophic Boussinesq flows (Molemaker et al. 2010). Resulting frontogenesis is associated with rapid time scales from a few hours to a day, comparable to the background strain

FIG. 16. Frequency–wavenumber cospectra of vertical velocities and temperature computed from 15 Oct to 15 Nov 2012 at (a) 39, (b) 99, (c) 299, and (d) 506 m. These cospectra are presented in a variance-preserving form, which allows one to directly compare the relative contribution of different time and spatial scales with the total variance ([appendix B](#)).

field. Frontogenesis is asymmetric, amplifying the positive skewness of Ro , which is the signature of ageostrophic dynamics. Consequently, frontogenesis statistically dominates frontolysis, consistent with the theoretical and idealized studies of [Klein and Lapeyre \(2009\)](#), [Roulet and Klein \(2010\)](#), and [Molemaker et al. \(2015\)](#). As a result, there is a net upward VHF at deep-reaching submesoscale fronts that is induced by the asymmetrical character of the ageostrophic frontogenesis ([Molemaker et al. 2015](#)). These VHF are larger in the ocean interior than within the ML, concordant with the in situ findings of [Siegelman et al. \(2020\)](#) and [Yu et al. \(2019\)](#), and suggesting the presence of an intensified oceanic restratification at depth.

Several caveats pertaining to the model resolution need to be mentioned. First, even though $|\nabla b|$ are strong and associated with large $|Ro|$ (>0.5), they remain weaker than in the observations, especially in terms of Ri^{-1} . This is likely due to the fact that buoyancy gradients are

partly captured by higher vertical normal modes than those resolved by the model because of its coarse vertical resolution at depth (5 m at a depth of 49 m vs 45 m at 900 m). In comparison, the in situ data in [Siegelman et al. \(2020\)](#) have a constant vertical resolution of 1 m. This points to the need of choosing a vertical resolution able to adequately resolve higher vertical normal modes. In addition, the horizontal resolution of the model seems to lie at the edge of being submesoscale resolving, as can be seen in [Fig. 13](#) where the sharp fronts are only captured by a single model grid point. As such, a higher 3D resolution is needed to fully represent the ageostrophic dynamics of the interior ocean, which are likely to be even stronger than what is reported here.

This study solely considers a region of the ACC in springtime. However, the vertical heat fluxes diagnosed here in the ocean interior are comparable to those obtained in the ML in winter on a global scale ([Su et al. 2018](#)).

As such, these results call for extended analyses of deep ageostrophic frontal dynamics at different seasons and throughout the World Ocean, in particular in other parts of the ACC, as well as in the Gulf Stream and Kuroshio Extension, in order to confirm the ageostrophic character of the ocean interior.

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APPENDIX A

Details of the LLC4320 Simulation

The outputs of an ocean general circulation model, enabled by NASA Advanced Supercomputing Division, are used to investigate ocean dynamics down to the submesoscale. The model is based on a global, full-depth ocean and sea ice configuration of the Massachusetts Institute of Technology general circulation model (MITgcm) (Marshall et al. 1997; Hill et al. 2007) and uses a latitude–longitude–polar cap (LLC) grid (Forget et al. 2015). The MITgcm was spun up in a hierarchy of numerical simulations with increasing horizontal resolutions with 90 vertical levels. The simulation analyzed here is the highest resolution, the LLC4320 at $1/48^\circ$, with a time step of 25 s. The prognostic variables are saved as instantaneous snapshots at hourly intervals. Control and forcing files as well as details of the high-resolution LLC model setups are available online (http://mitgcm.org/viewvc/MITgcm/MITgcm_contrib/llc_hires).

Surface fluxes are from the 0.14° European Centre for Medium-Range Weather Forecasting (ECMWF) atmospheric operational model analysis, starting in 2011. The model also includes tidal forcing for the 16 most significant components that are applied as additional atmospheric pressure forcing (Chaudhuri et al. 2013). Vertical mixing is parameterized based on the critical value of Richardson number and is implemented using

the K -profile parameterization (KPP) scheme (Large et al. 1994) that has been extensively used and evaluated in ocean modeling studies (Large et al. 1997; Fernández-Castro et al. 2014). More details on the LLC4320 simulation, in particular on its validation with observations, can be found in Torres et al. (2018).

APPENDIX B

Frequency–Wavenumber Spectrum and Cospectrum

The w – k spectrum of a given variable $\phi(x, y, t)$ is computed in a box of $700 \text{ km} \times 700 \text{ km}$ and over 1 month. We refer the reader to Torres et al. (2018) for the full method. However, briefly, before computing the w – k spectrum of a $\phi(x, y, t)$, its linear trend is removed and a 3D Hanning window is subsequently applied to the detrended $\phi(x, y, t)$ (Qiu et al. 2018). A discrete 3D Fourier transform is then computed to retrieve $\hat{\phi}(k, l, \omega)$, where the caret indicates the Fourier transform, k is the zonal wavenumber, l is the meridional wavenumber, and ω is the frequency. The 3D Fourier transform is used to compute a 2D spectral density $|\hat{\phi}|^2(\kappa, \omega)$, where κ is the isotropic wavenumber, defined as $\kappa = (k^2 + l^2)^{1/2}$. The transformation from an anisotropic spectrum to an isotropic spectrum is performed following the method described by Savage et al. (2017).

The dispersion relation curves for IGWs (Gill 1982) have been also estimated for each vertical normal mode (Torres et al. 2018). The curve related to the tenth baroclinic vertical mode, that is, the highest baroclinic mode resolved by the simulation, was found to be the most relevant to partition balanced motions (below the curve) and IGWs (above the curve). To better compare the variance explained by IGWs and BMs in different areas of the ω – κ space, the spectra are presented in a variance-preserving form, which is achieved by multiplying the w – k spectra by ω and κ (Torres et al. 2018).

The ω – k cospectra of vertical heat fluxes are computed similar to the ω – k spectrum, following the method described in Flexas et al. (2019). First, the Fourier transforms of vertical velocity $\hat{W}(k, l, \omega)$ and temperature $\hat{T}(k, l, \omega)$ are calculated. The cospectrum of vertical heat fluxes is then given by

$$\widehat{W \cdot T}(k, l, \omega) = \text{Re}[\hat{W} \cdot \hat{T}^*(k, l, \omega) + \hat{W}^* \cdot \hat{T}(k, l, \omega)],$$

where Re is the real part of the complex quantity and the asterisk indicates the complex conjugate. The 2D cospectrum $\widehat{W \cdot T}(\kappa, \omega)$ is retrieved using the same method as before. The ω – k spectrum and cospectrum are presented in a variance-preserving form for easier comparison across the frequency–wavenumber domain.

FIG. C1. Map of the Okubo-Weiss quantity λ normalized by f^2 at 99 m. Vortex cores dominated by ζ are in blue ($\lambda < 0$), and filaments dominated by strain are in red ($\lambda > 0$). Strain-dominated regions are particularly prone to the formation of submesoscale fronts because of the exponential growth of tracer gradients in this region of the flow.

APPENDIX C

Okubo-Weiss Quantity

The Okubo-Weiss quantity is defined as

$$\lambda = \frac{1}{4}(\sigma_n^2 + \sigma_s^2 - \zeta^2),$$

where $\zeta = v_x - u_y$ is the relative vorticity, $\sigma_n = u_x - v_y$ is the normal strain rate, and $\sigma_s = v_x + u_y$ is the shear strain rate. The Okubo-Weiss quantity, derived by Okubo (1970) and Weiss (1991), is used to partition the fluid into regions with different dynamical properties, that is, elliptic regions dominated by ζ ($\lambda < 0$) from hyperbolic regions dominated by σ_n and σ_s ($\lambda > 0$) (Fig. C1 and Fig. S2 in the online supplemental material). Under the assumption that the velocity gradient is slowly varying along a Lagrangian trajectory, the behavior of a tracer gradient can be determined by the sign of λ (Hua and Klein 1998). Indeed, tracer gradients do not grow in vortex cores where $\lambda < 0$. In this case, the gradient vector experiences a simple rotation. On the other hand, in strain-dominated areas where $\lambda > 0$, tracer gradients exponentially grow.

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