Frontal Instability and Energy Dissipation in a Submesoscale Upwelling Filament

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ABSTRACT

Based on high-resolution turbulence microstructure and near-surface velocity data, frontal instability and its relation to turbulence are investigated inside a transient upwelling filament in the Benguela upwelling system (southeast Atlantic). The focus of our study is a sharp submesoscale front located at the edge of the filament, characterized by persistent downfront winds, a strong frontal jet, and vigorous turbulence. Our analysis reveals three distinct frontal stability regimes. (i) On the light side of the front, a 30–40-m-deep turbulent surface layer with low potential vorticity (PV) was identified. This low-PV region exhibited a well-defined two-layer structure with a convective (Ekman-forced) upper layer and a stably stratified lower layer, where turbulence was driven by forced symmetric instability (FSI). Dissipation rates in this region scaled with the Ekman buoyancy flux, in excellent quantitative agreement with recent numerical simulations of FSI. (ii) Inside the cyclonic flank of the frontal jet, near the maximum of the cross-front density gradient, the cyclonic vorticity was sufficiently strong to suppress FSI. Turbulence in this region was driven by marginal shear instability. (iii) Inside the anticyclonic flank of the frontal jet, conditions for mixed inertial/symmetric instability were satisfied. Our data provide direct evidence for the relevance of FSI, inertial instability, and marginal shear instability for overall kinetic energy dissipation in submesoscale fronts and filaments.

1. Introduction

Submesoscale fronts and filaments, frequently observed at scales around 0.1–10 km in the surface boundary layer (SBL) of the ocean in the vicinity of frontal regions, have been extensively studied in the past due to their importance for SBL restratification, vertical transport of tracers, and energy dissipation (McWilliams 2016). These features are thought to be generated by mesoscale advective deformation of a preexisting horizontal buoyancy gradient, by baroclinic instability of the weakly stratified SBL (Fox-Kemper et al. 2008), or by other processes related to, e.g., upwelling, differential mixing, and river plumes. The classical view is that submesoscale fronts and filaments are quickly intensified by frontogenesis induced by a quasigeostrophic strain field (Hoskins 1982; Capet et al. 2008). There is, however, increasing evidence from recent numerical studies that late-stage submesoscale frontogenesis is affected, or even dominated, by an ageostrophic secondary circulation that is tightly connected to vertical turbulent mixing and atmospheric forcing (McWilliams et al. 2015; Gula et al. 2014; Thomas and Lee 2005). Submesoscale frontogenesis breaks the balance constraint of the large-scale quasigeostrophic dynamics, and makes submesoscale currents susceptible to various types of flow instability that are generally associated with a forward energy cascade (McWilliams 2016). Submesoscale instabilities therefore provide a potential conduit for the downscale transport of mesoscale energy, which may be especially relevant in the SBL, where other energy sinks (e.g., bottom friction, spontaneous emission of internal gravity waves) are likely less relevant.

It is unclear at the moment which processes ultimately determine frontal arrest and energy dissipation, but a number of recent studies point at the importance of a special class of instabilities that occur in stably stratified flows if the Ertel potential vorticity \( q \) and the Coriolis parameter \( f \) attain opposite signs: \( fq < 0 \). As summarized by Thomas et al. (2013), these include inertial/centrifugal instability and symmetric instability (SI). Both extract kinetic energy from the balanced background flow, inject it into the smaller scales of the fastest growing modes, and therefore induce a forward energy cascade. As the scales of these instabilities are still too large to be directly associated with viscous
dissipation, the current belief is, based on evidence from high-resolution simulations (Taylor and Ferrari 2009), that the final step toward the scales of three-dimensional turbulence and dissipation is bridged by secondary Kelvin–Helmholtz (KH) instabilities.

Thomas (2005) pointed out that conditions with \( f_q < 0 \) are likely to occur in fronts subject to atmospheric buoyancy loss (cooling) and/or downfront winds inducing a destabilizing cross-front Ekman transport (dense water advected on top of light water). Both processes extract potential vorticity (PV) from the SBL, thus maintaining conditions for inertial instability and SI. Large-eddy simulations (LES) of density fronts by Taylor and Ferrari (2010) suggest that the turbulent SBL exhibits, in this situation, a distinct two-layer structure with a gravitationally unstable upper layer (convection), and a stably stratified lower layer, in which turbulence is driven by “forced” SI (FSI in the following). Thomas and Taylor (2010) showed that the energy dissipation associated with FSI scales with the sum of the surface buoyancy flux and the Ekman buoyancy flux (the rate at which the destabilizing cross-front Ekman transport generates available potential energy).

A number of recent field studies including turbulence microstructure observations have confirmed that submesoscale fronts are regions of enhanced mixing (e.g., Nagai et al. 2009, 2012; Johnston et al. 2011). However, comprehensive investigations that combine turbulence and hydrographic data at resolutions sufficient to resolve the extreme gradients in narrow and quickly evolving submesoscale fronts, required to test the real-ocean relevance of the above concepts, are still largely lacking. A notable exception is the study of frontal instability and mixing in the Kuroshio by D’Asaro et al. (2011), who combined ship-based observations with turbulence data from a tracked Lagrangian float drifting along the frontal jet. They were the first to conclusively show that enhanced energy dissipation found in a narrow frontal region of the Kuroshio was generated by FSI, and scaled with the Ekman buoyancy flux. A follow-up study in a Gulf Stream front based on similar instrumentation (Thomas et al. 2016) arrived at the same conclusions, however, with the caveat that near-inertial waves may induce transient modifications.

Here, unlike previous studies focusing on fronts in large-scale current systems like the Kuroshio and the Gulf Stream, we concentrate on FSI and associated mixing in a narrow submesoscale front bounding a freshly forming dense upwelling filament in the Benguela upwelling system. Our data include turbulence microstructure data and near-surface velocity data from a towed research catamaran, allowing us to resolve the vertical and lateral structure of shallow submesoscale fronts in unprecedented detail. After a brief summary of the dynamics of frontal instability in the following section 2, we describe our study area and instrumentation in section 3, followed by a short overview of the filament structure and evolution in section 4. The main section 5 includes a detailed analysis of submesoscale frontal instabilities and associated turbulence. A brief discussion of our results and some conclusions are contained in section 6.

2. Frontal instability and dynamics

a. Classification of frontal instability

Summarizing the conditions for inertial and symmetric instabilities in stably stratified flows, Thomas et al. (2013) pointed out that these types of instability are expected to occur if the Ertel PV (\( q \)) and the Coriolis parameter (\( f \)) have opposite signs:

\[
f_q = f(fk + \nabla \times u) \cdot \nabla b < 0,
\]

where \( k \) is the upward unit vector, and \( b = -g \rho \varphi / \rho_0 \) the buoyancy, here defined based on the potential density \( \rho_0 \) (we use \( \rho_0 = 1026 \text{ kg m}^{-3} \) for the constant reference density).

The PV can be decomposed into vertical and baroclinic components, \( q_v \) and \( q_{bc} \), and (1) can thus be rewritten as

\[
f_q = f(q_v + q_{bc}) < 0,
\]

The first component is associated with vertical vorticity and stratification,

\[
q_v = (f + \zeta)N^2,
\]

where \( \zeta = k \cdot (\nabla \times u) \) is the vertical component of the relative vorticity, and \( N^2 = \partial b / \partial \zeta \) is the square of the buoyancy frequency. The baroclinic component can be written as

\[
q_{bc} = \left( \frac{\partial w}{\partial y} - \frac{\partial u}{\partial z} \right) \frac{\partial b}{\partial x} + \left( \frac{\partial u}{\partial x} - \frac{\partial w}{\partial y} \right) \frac{\partial b}{\partial z}.
\]

Two types of instability are generally distinguished in stably stratified flows with negative \( f_q \): SI is expected to occur for \( |f_{bc}| > f_{\zeta} \) with \( f_{\zeta} > 0 \) (absolute vorticity is cyclonic), whereas flows with \( f_{\zeta} < 0 \) are susceptible to inertial/centrifugal instability (Thomas et al. 2013).

Defining a coordinate system aligned with the front (with cross-front and downfront coordinates denoted as \( x \) and \( y \), respectively), and assuming that the front is straight and shows no variations in the \( y \) direction, these two components of the PV reduce to
where we introduced $M^2 = -\partial b/\partial x$ to denote the cross-front buoyancy gradient. Following our definition of the cross-front $x$ coordinate, we have $M^2 > 0$ inside the frontal region. From (2) and (5), it is clear that cyclonic shear has a stabilizing tendency, whereas, at least in geostrophically balanced fronts with $fq_{bc} = -M^4$, the baroclinic term is always destabilizing. As shown below, both effects are generally important in our data.

b. Forced symmetric instability

Thomas (2005) argued that SI can only be sustained by an upward PV flux through the sea surface, permanently draining PV from the SBL. He showed that conditions are favorable for such forced SI (FSI) if the sum of the atmospheric buoyancy flux $B_0$ and the Ekman buoyancy flux,

$$ B_E = \frac{\tau_w^* M^2}{\rho_0 f}, $$

is positive: $B_0 + B_E > 0$ (here, $\tau_w^*$ denotes the down-front component of the wind stress). FSI can thus only persist if the net surface forcing acts to gravitationally destabilize the SBL, either by a destabilizing atmospheric buoyancy flux (e.g., cooling) or by a cross-front Ekman transport that moves dense water on top of lighter water.

From idealized numerical simulations of FSI, Taylor and Ferrari (2010) identified a turbulent low-PV layer of thickness $H$ that is composed of two sublayers with distinct vertical structure and dynamics. They found that the upper part of the low-PV layer is characterized by a well-mixed region of thickness $h$, in which turbulence is directly driven by the surface forcing, as in classical convection. Underneath this convective layer, a turbulent forced SI layer was shown to develop, under the obvious geometrical constraint that $h/H < 1$. Different from the convective layer, the forced SI layer is characterized by significant stable stratification.

Taylor and Ferrari (2010) derived an implicit relation for the convective layer depth as a function of the buoyancy forcing $B_0 + B_E$, the cross-front buoyancy gradient $M^2$, and the thickness $H$ of the low-PV layer. Without any further assumptions, it is easy to show that their model equations (24) and (25) can be rewritten in the following form:

$$ \frac{(h/H)^4}{\kappa^2} \left( \frac{c}{\nu} \right)^6 \left( \frac{L_{MO}}{H} \right)^{-2} \left( 1 - \frac{h}{H} \right)^3, $$

where $c \approx 14$ is a model constant. Here, $\Delta u_g = HM^2/f$ denotes the (geostrophic) velocity difference across the low-PV layer, and

$$ L_{MO} = \frac{u_g}{\kappa (B_0 + B_E)}, $$

a modified version of the Monin–Obukhov (MO) length scale that also takes into account the effect of the Ekman buoyancy flux. The friction velocity, $u_g = \sqrt{\tau_w^*/\rho_0}$, is computed from the wind stress $\tau_w^*$, and $\kappa = 0.4$ denotes the von Kármán constant. Equation (7) shows that the relative thickness of the convective layer (or, equivalently, the thickness of the forced SI layer) is determined by two nondimensional key parameters: the velocity ratio $u_g/\Delta u_g$ and the ratio of the MO length scale to the total thickness of the low-PV layer.

Finally, it is important to note that Thomas and Taylor (2010) showed from their high-resolution simulations that the rate at which FSI extracts kinetic energy from the background flow scales with $B_0 + B_E$, and linearly decreases from $B_0 + B_E$ to the surface zero at the bottom of the low-PV layer. Assuming that this energy is locally dissipated (Thomas and Taylor 2010), the averages of the dissipation rate across the low-PV layer ($-H < z < 0$) and the forced SI layer ($-H < z < -h$) are $(B_0 + B_E)/2$ and $(B_0 + B_E)(1 - h/H)^2/2$, respectively. We will use this result below when comparing our turbulence observations to the model of Thomas and Taylor (2010).

3. Study area and methods

a. Study area

The Benguela upwelling system, located in the southeastern Atlantic Ocean off the Namibian coast, is one of the major upwelling systems of the global ocean. Upwelling in this region is linked to predominantly equatorward winds along the coast, and tends to be characterized by strong spatial and temporal variability of the wind field (Risien et al. 2004). Mesoscale and submesoscale eddies, filaments, and fronts are ubiquitous in the Benguela system, providing an ideal environment for the purpose of this study. Here, we focus on a filament in the vicinity of the Lüderitz upwelling cell, one of the most energetic upwelling hotspots of the entire Benguela system (Shannon and Nelson 1996). The experimental site is located at approximately 14°E, 27°S on the outer shelf at a water depth of 300–500 m. The local inertial period is $T_f = -2\pi/f = 26.5$ h, where $f = -6.6 \times 10^{-5}$ rad s$^{-1}$ is the Coriolis parameter. Figure 1a provides a regional overview of the sea surface structure on
1 December 2016, just before the start of the measurements, when a freshly forming upwelling filament evolved out of the main upwelling front in a region with strong mesoscale activity.

The key components of our ship-based observations were high-resolution turbulence microstructure and near-surface velocity measurements, obtained on 3–7 December 2016 during a cruise with R/V Meteor. These measurements were carried out on four repeated transects across the filament, either from the drifting ship near fixed stations distributed along the transect (transects T1 and T2), or from continuous “tow-yo” profiling from the slowly cruising ship (transects T3 and T4), as summarized in Table 1.

b. Ship-based measurements

Stratification and mixing data were obtained with the help of an MSS90-L turbulence microstructure profiler from ISW (Germany), equipped with two airfoil shear probes for turbulence measurement, a fast-response thermistor, a Seapoint turbidity sensor, and precision CTD sensors from Sea and Sun Technology (Germany). The free-falling profiler was balanced for a sinking speed of 0.6–0.7 m s$^{-1}$, yielding a 100-m profile every 5–6 min. Microstructure profiling was performed in a tow-yo mode from the stern of the ship while slowly [speed: 1–2 kt (1 kt = 0.51 m s$^{-1}$)] cruising against wind and waves. This resulted in an effective horizontal profile spacing of a few hundred meters, depending on ship speed.

Microstructure data, collected at 1024 Hz, were averaged down to 256 Hz for noise reduction before further processing. Data from the CTD sensors were then averaged into 0.1-m vertical bins, and converted into Conservative Temperature $\Theta$ and Absolute Salinity $S_A$ according to the international TEOS-10 standard for seawater (Millero et al. 2008; Feistel et al. 2010). The energy dissipation rate $\epsilon$ was estimated by integrating shear spectra across the dissipative subrange inside half-overlapping 256-point windows. The upper wavenumber

<table>
<thead>
<tr>
<th>Transect No.</th>
<th>Time</th>
<th>Type</th>
<th>CatADCP</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td>1207 UTC 4 Dec 2016–1732 UTC 4 Dec 2016</td>
<td>FS</td>
<td>No</td>
</tr>
<tr>
<td>T2</td>
<td>1615 UTC 5 Dec 2016–0428 UTC 6 Dec 2016</td>
<td>FS</td>
<td>No</td>
</tr>
<tr>
<td>T3</td>
<td>1522 UTC 6 Dec 2016–0138 UTC 7 Dec 2016</td>
<td>CS</td>
<td>Yes</td>
</tr>
<tr>
<td>T4</td>
<td>1447 UTC 7 Dec 2016–2048 UTC 7 Dec 2016</td>
<td>CS</td>
<td>No</td>
</tr>
</tbody>
</table>
for integration was found iteratively as a function of the Kolmogorov wavenumber with a correction for lost variance as described in Moum et al. (1995). These dissipation rate estimates were then averaged into 0.5-m vertical bins. As the upper meters of the water column are likely affected by the presence of the ship, data above 5 m depth are not shown in the following, and data above 10 m are excluded from the analysis.

Horizontal velocities were obtained from a vessel-mounted 75-kHz ADCP (RDI Workhorse, Ocean Surveyor), sampling the water column in 8-m vertical bins at 1 Hz. With the uppermost bin located at 17.5 m depth, this instrument was not able to provide velocity data in the important near-surface region, where the core of the frontal jet was located. Our ship-based velocity measurements were therefore complemented by data from a 300-kHz ADCP (RDI Workhorse), mounted on a 5.5-m research catamaran that was towed at a distance of 100–150 m behind the ship, in parallel with the microstructure measurements. This instrument sampled the upper part of the water column at 2-s intervals inside 2-m vertical bins with the uppermost bin located only 4.5 m below the surface. The catamaran was equipped with a remote-controlled rudder, allowing it to be laterally steered out of the ship’s wake in order to obtain near-surface velocity data that were undisturbed by ship effects. As acoustic bottom tracking data were not available due to the large water depth, the speed over the ground was estimated based on high-precision GPS data that were available independently on both platforms (R/V Meteor and the research catamaran). Velocity data from both ADCPs were averaged over intervals of 120 s and rotated into the alongfront and cross-front directions as described below. A comparison of velocity data from both ADCPs in the overlapping depth range between 17.5 and 73.5 m showed excellent agreement.

Alongtrack near-surface temperatures at 10-s intervals were obtained from an SBE38 temperature sensor (SeaBird, United States) mounted at 2 m depth inside the hull of the ship. An SBE21 thermosalinograph provided near-surface salinities.

A shipboard weather station, operated by the German Weather Service (DWD), provided continuous meteorological data during the entire cruise. These included wind speed and direction, air temperature and pressure, shortwave and longwave radiation, and humidity, all at approximately 35 m height. These data were used to estimate the cross-front and downfront components of the wind stress, \(\tau_x^c\) and \(\tau_y^c\), and the net upward heat flux (or net ocean heat loss) \(Q_{\text{net}} = Q_{\text{SW}} + Q_{\text{LW}} + Q_S + Q_L\), using the COARE 3.5 routines (Fairall et al. 2003; Edson et al. 2013). Here, \(-Q_{\text{SW}}\) is net downward shortwave radiation, and \(Q_{\text{LW}} = Q_{\text{LW}}^{\text{out}} - Q_{\text{LW}}^{\text{in}}\) the net upward longwave radiation, calculated from the difference between the outgoing longwave radiation \(Q_{\text{LW}}^{\text{out}}\) and the observed downward longwave radiation \(Q_{\text{LW}}^{\text{in}}\). The former was estimated using the Stefan–Boltzmann law (with an emissivity of 0.985), using ship-based SST measurements (Köhn et al. 2017). The sensible and latent heat losses, \(Q_S\) and \(Q_L\), were calculated from the COARE 3.5 bulk formulas. The contribution of precipitation to the heat flux was neglected as there was virtually no rainfall during our measurements.

c. Satellite data sources

Daily SST data at 1-km resolution, provided by the Ocean Biology Processing Group (OBPG, http://oceancolor.gsfc.nasa.gov), and sea surface height anomalies (SSHA) at 25-km resolution, taken from Aviso (http://www.aviso.altimetry.fr/), were used for the detection of fronts and filaments. SST data were strongly affected by clouds during large parts of our ship-based transect measurements (3–7 December 2016). However, a virtually cloud-free scene of the entire study area could be obtained just a few days before our measurements (1 December), as shown in Fig. 1a. For the wind field, daily ASCAT data with 0.25° grid resolution from the Asia–Pacific Data-Research Center (http://apdrc.soest.hawaii.edu/) were used.

d. Local coordinate system

Near-surface temperatures and velocities along the main transect T3 (Fig. 1b) show an approximately 15-km-wide cold filament, bounded by two sharp fronts in the northeast (NE) and southwest (SW), respectively. Remarkable is the strong, approximately 5–10-km-wide frontal jet associated with the southwestern front with near-surface velocities exceeding 1 m s\(^{-1}\).

For the subsequent analysis, we introduce a rotated coordinate system with the cross-front and alongfront directions denoted as \(x\) and \(y\), respectively, as indicated in Fig. 1b. This coordinate system was chosen such that the alongfront direction was aligned with the frontal jet, based on near-surface velocity data from our main transect T3 (see Table 1). This resulted in an angle of 61.3° between the \(x\) axis and the east direction. Vice versa, we determined \(\phi = 16.2°\) for the angle between the \(x\) axis and the alongtrack direction \(s\) (Fig. 1b). The origin of the cross-front coordinate, \(x = 0\), is collocated with the maximum cross-front density gradient at the surface. All data measured along the ship track were projected onto the cross-front coordinate \(x\), assuming homogeneity in the \(y\) direction. Similarly, gradients along the ship track were converted into cross-front gradients using the relation \(\partial/\partial s = \cos \phi \partial/\partial x\). Vectorial
quantities pointing into the cross- and alongfront directions are denoted by the subscripts \( x \) and \( y \), respectively, except for the cross-front and alongfront current speeds that are referred to as \( u \) and \( v \).

### 4. Filament evolution and structure

#### a. Filament evolution

Figure 1a suggests that the formation and evolution of the filament may be linked to the interaction of two mesoscale eddies (anticyclonic and cyclonic, marked as "A" and "C" in the figure) with the main upwelling front. This becomes clearer from the enlarged view in Fig. 2, showing the evolution of the filament for a 9-day period before and during the ship-based observations. During this period, the center of the cyclonic eddy remained at a stable position near the center of the domain shown in Fig. 2, and also the center of the anticyclonic eddy (only partly visible in the figure) showed little variability around its mean position at 13°20'E, 27°08'S. Winds were around 10 m s\(^{-1}\), directed equatorward along the coast with little spatial variability, thus providing persistent conditions for coastal upwelling.

These two eddies induced a confluent strain field, locally deforming the main upwelling front by entraining a filament of cold waters from the coastal upwelling cell. During the days from 28 November to 1 December (Figs. 2a–d), the filament intensified and was entrained farther westward into the warm waters outside the upwelling region. This process is consistent with the classical view of frontogenesis due to a mesoscale strain field (Hoskins 1982) that has been shown to dominate the initial phase of filament generation and intensification also in high-resolution simulations (e.g., Capet et al. 2008; Gula et al. 2014). From 2 December onward, cloud cover increased, and useful satellite images became rare. The last available snapshot before the start of the ship-based measurements is from 2 December, providing SST data only in the southern and eastern parts of the domain. Combined with our high-resolution thermosalinograph observations along the cruise track (3–5 December), these data nevertheless show that the filament remained a persistent feature until 5 December (Fig. 2e). Finally, on 6 December, SSHA data show first indications for a disintegration of the cyclonic eddy in the center of the domain, and an associated collapse of the eastern end of the filament, resulting from the advection of warm fluid patches into the transition region between the filament and the main upwelling cell (Fig. 2d). It is important to note, however, that the sharp front bounding the filament from the SW remained remarkably robust until 6 December, and did not show indications for meandering or other strong distortions in the vicinity of the main transect T3.
For this reason, most of the following analysis will focus on this frontal region.

**b. Vertical structure**

Figure 3 shows the evolution of the vertical filament structure along our four cross-filament transects (T1–T4, see Table 1) that were obtained on 4–7 December 2016. Transects T1 and T2 were coarse-resolution surveys with a limited number of microstructure profiles (typically 4–5) clustered around fixed stations along the transect (see black markers in the figure). Near the surface, these measurements are complemented by high-resolution temperature and salinity data from the ship’s thermosalinograph, which turned out to be helpful for the identification of sharp fronts not resolved by the coarse-resolution station data. During transects T3 and T4, microstructure measurements were obtained in a continuous tow-yo mode from the slowly cruising ship, yielding an order-of-magnitude improved horizontal resolution at the expense of a longer transect duration. All transects exactly followed the alongtrack coordinate $s$ shown in Fig. 1b, although with different start and end points, respectively. Data measured along the ship track were then projected onto the cross-front direction $x$ as described above in the context of Fig. 1b. Please note that our local coordinate system, based on transect T3, remains fixed for all four transects to more clearly illustrate the relative motion of the front.

The short transect T1 (Figs. 3a,b), observed on 4 December, only covered the frontal region bounding the filament in the SW. Outcropping isopycnals in this region bridge a lateral density contrast of approximately 0.6 kg m$^{-3}$, which is largely due to temperature differences that are only to a small extent compensated by salinity effects (the upwelled water inside the filament is slightly fresher). Near-surface thermosalinograph measurements show smooth temperature and salinity transitions across the entire transect with no indications for locally intensified, sharp fronts. This suggests that, in this case, lateral gradients in the near-surface region are well resolved even with our coarse-resolution station measurements. On the SW side of the front ($x < -5$ km, approximately),
a well-defined SBL, bounded by a sharp thermocline at 50 m depth, can be identified. This layer shows indications for a reduction of vertical stratification toward the surface but is nowhere “well mixed” except close to the surface at the SW end of the transect. Thermocline isopycnals predominantly outcrop within $-5 < x < 5$ km, which is reflected in an intensification of the cross-front buoyancy gradient at the surface.

The longer transect T2 (5 December) shows a similar picture on the SW side of the transect but now also reveals that the filament merges into a second, shallower front at the NE end (Figs. 3c,d). The high-resolution thermosalinograph data show that smooth cross-front density gradients found in transect T1 have now evolved into a sharp front at $x = -3$ km. A similar front is also observed at the NE end of the filament, near $x = 14$ km. Clearly, the horizontal structure of these sharp fronts is not resolved by our coarse-resolution microstructure stations.

The advantage of our high-resolution observations becomes evident for the main transect T3 (6 December, Figs. 3e,f), which is based on continuously repeated microstructure profiling with a horizontal spacing of a few hundred meters only. While the large-scale structure of the filament in T3 is comparable to T2, the high-resolution data for T3 reveal a wealth of additional small-scale features. Most importantly, these data resolve two sharp fronts on either side of the filament, and reveal a strong asymmetry in mixed layer depths, with a shallow (less than 20 m deep) mixed layer on the NE side, and a more than 35-m-deep mixed layer on the SW side.

Finally, data from transect T4, sampled on 7 December (Figs. 3g,h), show a collapse of the filament. This is consistent with Fig. 2f, in which indications for the beginning disintegration of the eastern part of the filament became evident already a day before.

c. Atmospheric forcing

Figure 4 shows the temporal variability of the atmospheric forcing shortly before and during the ship-based measurements along transects T1–T4. With an average around 0.4 Pa, the wind stress was strongest during transect T1, and decayed to approximately half this value during our main transect T3 (Fig. 4a). The downfront component of the wind stress $\tau_{w}$ was positive throughout the measurements, and generally dominated the total wind stress, except for transect T2, when winds were comparatively weak, however. This suggests that the SW front of the filament was characterized by destabilizing Ekman transport (dense water is moved on top of light water) while the front in the NE was permanently stabilized, as discussed in greater detail below.

The surface buoyancy flux $B_0$ (positive upward, i.e., positive for buoyancy loss) shows a clear diurnal signal peaking at approximately $B_0 = -4 \times 10^{-7} \text{ m}^2 \text{s}^{-3}$ around midday due to strong solar heating (Fig. 4b). During nighttime, atmospheric cooling induced a buoyancy loss that was typically in the range $B_0 = 4-6 \times 10^{-8} \text{ m}^2 \text{s}^{-3}$. This diurnal cycle was imprinted on the ship transects, depending on their relative timing. Transect T1 was exposed to strong daytime heating (buoyancy gain), while T2 was affected by only weak atmospheric heat fluxes with a slight cooling tendency. Our main transect T3 was characterized by a transition from late-afternoon weak diurnal warming to nighttime cooling just when R/V Meteor crossed the front in the SW on its way from...
the cold filament to the warm ambient waters farther offshore.

5. Frontal structure and stability

a. Frontal structure

The following analysis of frontal stability will focus on the sharp front bounding the filament in the SW (Fig. 5). This is motivated by the robustness of this feature during our measurements, the availability of both very high-resolution microstructure and undisturbed near-surface velocity data from the research catamaran, and conditions favorable for FSI due to downfront winds.

Figure 5 shows that the strongest horizontal buoyancy gradients (up to $M^2 = 2 \times 10^{-6} \text{s}^{-2}$) are found inside the cyclonic flank of the frontal jet ($-2 < x < 2 \text{ km}$), where downfront velocities of up to $1 \text{ m s}^{-1}$ are observed. Further SW ($x < -10 \text{ km}$), the main front merges into a 30–35-m-deep SBL with weak stratification and strong turbulence. Here, horizontal buoyancy gradients are substantially smaller compared to the main front but still on the order of few times $10^{-7} \text{s}^{-2}$. For comparison, Taylor and Ferrari (2010) used $M^2 = 2.1–8.5 \times 10^{-7} \text{s}^{-2}$ in their simulations of FSI in near-surface fronts, noting that their intermediate value ($M^2 = 4.2 \times 10^{-7} \text{s}^{-2}$) “represents a relatively strong front.” Also, their SBL depths were similar to our observations.

The downfront wind stress $\tau_w$ was relatively constant along the transect, and the Ekman buoyancy flux $B_E$ defined in (6) thus largely traces the variability of the horizontal buoyancy gradient (Fig. 5a). The total buoyancy flux $B_0 + B_E$ is positive throughout the
frontal region, providing favorable conditions for FSI. Due to the strength of the cross-front buoyancy gradients, and weak to negligible solar radiation during the transect (late afternoon and nighttime), we generally have $B_0/E \ll 1$ (red and black curves in Fig. 5a). On the SW side of the transect ($x < -10$ km), a weakly stratified SBL can be discerned with turbulence reaching down to the upper edge of the thermocline at 30–35 m depth (Fig. 5c). This is contrasted by the region with strongest cross-front gradients ($-2 < x < 2$ km), which is seen to be vigorously turbulent down to 25 m, however, without any indications for a well-mixed SBL. Both regions are separated by a transition zone ($-10 < x < -2$ km), where isopycnals from the deep thermocline in the SW part of the transect gradually rise toward the surface. For the subsequent analysis, we identified six subregions (see Fig. 5d), characterized by distinct dynamics and structure, respectively. As discussed in more detail below, each of these regions represents a different stability regime. The key parameters for each region are summarized in Table 2.

Regions I and II are both located in the SW part of the transect, characterized by deep mixed layers and moderate horizontal density gradients. While the well-mixed layer for region I reaches down to the thermocline, region II is vertically separated into a well-mixed surface region, and a deeper layer with significant stratification (see isopycnal structure in Figs. 5b,c) that is reminiscent of the structure in the numerical simulations described in Taylor and Ferrari (2010). Also the SBL depth, the horizontal buoyancy gradient $M^2$ and the buoyancy forcing $B_0 + B_E$ (see Table 2) are within the parameter range investigated by Taylor and Ferrari (2010), suggesting that region II provides a direct benchmark for their numerical study.

Regions III, IV, and V are located inside the main frontal zone where the horizontal density gradient and the vorticity on the cyclonic side of the frontal jet reach their maximum strength, respectively (Fig. 5d). The distinction between these regions becomes clearer from the enlarged view of the density and velocity structure inside the frontal region in Fig. 6. The data shown in this figure were horizontally filtered with a 2-km box filter, which was found to effectively remove random fluctuations in the raw velocity data without significantly damping the mean frontal velocity gradients (see Fig. 5d). This filter property should be kept in mind when interpreting cross-front gradients like the vorticity shown in Fig. 6c.

Figure 6 shows that most of the cross-front density gradient in the near-surface region ($z > 20$, approximately) is found in regions III and IV, while most of the cross-front (cyclonic) shear is confined to regions IV and V. Recalling the competing roles of cyclonic shear and

<table>
<thead>
<tr>
<th>Region</th>
<th>$M^2 (s^{-2})$</th>
<th>$B_0 + B_E (m^2 s^{-3})$</th>
<th>$\xi/f$</th>
<th>$H (m)$</th>
<th>$h (m)$</th>
<th>$L_{MO} (m)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>$1.19 \times 10^{-7}$</td>
<td>$2.05 \times 10^{-7}$</td>
<td>$-1.00$</td>
<td>27.5</td>
<td>18.5</td>
<td>15.8</td>
</tr>
<tr>
<td>II</td>
<td>$2.62 \times 10^{-7}$</td>
<td>$6.65 \times 10^{-7}$</td>
<td>$-1.36$</td>
<td>32.0</td>
<td>14.8</td>
<td>9.8</td>
</tr>
<tr>
<td>III</td>
<td>$6.51 \times 10^{-7}$</td>
<td>$2.17 \times 10^{-6}$</td>
<td>0.90</td>
<td>17.5</td>
<td>7.3</td>
<td>3.6</td>
</tr>
<tr>
<td>IV</td>
<td>$1.74 \times 10^{-6}$</td>
<td>$5.54 \times 10^{-6}$</td>
<td>4.80</td>
<td>—</td>
<td>—</td>
<td>1.6</td>
</tr>
<tr>
<td>V</td>
<td>$2.63 \times 10^{-7}$</td>
<td>$7.09 \times 10^{-7}$</td>
<td>2.27</td>
<td>—</td>
<td>—</td>
<td>12</td>
</tr>
<tr>
<td>ISI</td>
<td>$3.92 \times 10^{-8}$</td>
<td>$1.00 \times 10^{-7}$</td>
<td>$-0.67$</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
</tbody>
</table>

FIG. 6. Enlarged view of (a) density, (b) velocity, and (c) Rossby number at different depth levels for subregions III, IV, and V (see Fig. 5d). Black markers indicate locations of individual microstructure casts.
baroclinicity with regard to SI according to (2), our classification of regions III–V thus reflects the following physical regimes: strong baroclinicity and weak cyclonic vorticity (region III), strong baroclinicity and strong cyclonic vorticity (region IV), and weak baroclinicity and strong cyclonic vorticity (region V). We will see below that this distinction is also mirrored in the results of the stability analysis. Note that despite the small width of the frontal regions III–V [O(1) km], and different from most previous studies, horizontal gradients are well resolved by our velocity and microstructure measurements.

b. Analysis of frontal instability and turbulence

In the following, we apply the theory outlined in section 2 to investigate the stability properties of regions I–V. To this end, we first computed the region-averaged cross- and alongfront velocities \( \overline{\upsilon} \) and \( \overline{\upsilon} \) by averaging the raw (catamaran-based) velocity profiles at 2-m vertical resolution over each of the five regions, respectively. Similarly, region-averaged density profiles were computed at 0.1-m vertical resolution, based on the available microstructure casts inside these regions. These profiles were vertically filtered with a centered 2-m box filter to obtain representative density \( \overline{\rho} \) and buoyancy \( \overline{b} \) profiles at resolution comparable to the velocity data. This becomes important when vertical density and velocity gradients are quantitatively compared, e.g., in terms of nondimensional parameters like the gradient Richardson number (see below).

From these region-averaged profiles, we computed the (squares of) the total vertical shear,

\[
S^2 = \left( \frac{\partial \overline{\upsilon}}{\partial z} \right)^2 + \left( \frac{\partial \overline{\upsilon}}{\partial z} \right)^2,
\]

and the buoyancy frequency, \( N^2 = \overline{\partial b}/\partial z \), by vertical finite differencing. The region-averaged vertical vorticity, \( \xi = \overline{\partial \upsilon}/\partial x \), and cross-front buoyancy gradient, \( M^2 = -\overline{\partial b}/\partial x \), were computed from averaging the cross-front gradients over each of the five regions.

The “low-PV layer” that Taylor and Ferrari (2010) suggested as an extension of the classical concept of the “mixed layer” was identified here with the near-surface region with negative \( f_q \) (if existing). This definition of the low-PV layer thickness \( h \) has the advantages that it does not require any modeling assumptions, avoids the assignment of an (arbitrary) threshold for “low” PV, and is directly connected to the stability condition in (2). We generally find a sharp increase of \( f_q \) for \( z < -H \), such that practically little ambiguity arises regarding the thickness of the low-PV layer. The depth of the directly forced convective layer \( h \) was computed from (7), based on the region-averaged quantities compiled in Table 2.

In the following, we will also use to the Ozmidov scale, \( L_O = (\epsilon/\bar{N}^4)^{1/2} \) (with \( \bar{N}^2 \) based on Thorpe-sorted density profiles), and the wall-layer scale, \( l = \kappa |z| \) (with \( \kappa = 0.4 \)), to quantify the typical size of turbulent overturns limited by stratification and the presence of the free surface, respectively. For example, for \( L_O > \kappa |z| \), the overturning scales are likely controlled by the proximity of the surface rather than by stratification.

Finally, we will compare the observed alongfront velocity \( \overline{\upsilon} \) with the geostrophic velocity \( \overline{u}_g \) computed from integrating the observed thermal wind shear \( \partial \overline{\upsilon}/\partial z = -M^2/f \) from the bottom of the low-PV layer to the surface.

1) REGIONS I AND II: DEEP MIXING LAYERS

As discussed above, regions I and II are located SW of the main frontal region, on the warm side of the front, where mixed layers are deep and both atmospheric and Ekman buoyancy forcing are destabilizing during nighttime conditions \( (B_b + B_E > 0) \), see Table 2). Individual potential density profiles (not shown) in the SBL exhibit extended regions with unstable stratification, indicating convective mixing. These unstable regions are less clearly reflected in the region-averaged \( \bar{N}^2 \) (Figs. 7c,d), which is often dominated by a few stably stratified patches in the same depth range. Still, however, the averaged stratification in the near-surface region is very small \( [\bar{N}^2 = O(10^{-7}–10^{-6})s^{-2}] \), and depth intervals with \( \bar{N}^2 < 0 \) (circle markers) can be found in both regions even after the averaging procedure.

For region II, the thickness of this weakly/unstably stratified near-surface layer is in good agreement with the predicted convective layer depth \( h = 14.8 \) m (gray dashed line in Fig. 7d), computed from (7) according to the model by Taylor and Ferrari (2010). For \( z < -h \), gradually increasing stable stratification is observed, the Richardson number becomes too large for shear instability \( (Ri > 1/4) \), and the Ozmidov scale \( L_O \) starts to control the size of the turbulent overturns (red line in Fig. 8d). Most importantly, however, Fig. 8b shows that this stably stratified but vigorously turbulent layer below the unstratified near-surface region is characterized by \( f_q < 0 \) and \( f_q_c > 0 \), indicating conditions favorable for FSI down to a depth of \( z = -H = -32 \) m. The destabilizing baroclinic term \( q_{bc} \) dominates the total PV in the stability condition (2), and outweighs the slightly cyclonic, hence stabilizing, vertical component \( q_f \). Figure 7b shows that the velocity difference between the bottom of the low-PV layer and the convective layer is largely geostrophically balanced, although local deviations from a perfect geostrophic balance
can be discerned. The conditions for FSI (Taylor and Ferrari 2010) are thus satisfied in an approximately 17-m-thick forced SI layer in the range $-H < z < -h$. It is worth noting that the parameter space, the vertical scales, and the geometry with a forced SI layer underneath a directly forced convective layer are in close agreement with the numerical simulations of Taylor and Ferrari (2010).

Figure 8b shows that the dissipation rates inside the forced SI layer ($-H < z < -h$) slightly decrease with depth with an average value of $\varepsilon = 1.6 \times 10^{-7}$ W kg$^{-1}$. As discussed in section 2b, the model of Thomas and Taylor (2010) predicts $\varepsilon_{SI} = \left(B_0 + B_E\right)(1 - h/H)^{3/2}$ for the average dissipation rate inside the forced SI layer.

From the values in Table 2, we find $\varepsilon_{SI} = 1.0 \times 10^{-7}$ W kg$^{-1}$, which is within a factor of 2 of the observed value. This excellent agreement between model and data supports our interpretation of FSI as the main energy source for turbulence in this region. It is further tempting, although speculative in view of the limited vertical resolution of our shear measurements, to interpret $Ri' = 1/4$, found in the upper part of the forced SI layer (Fig. 7d) as evidence for the secondary shear instabilities observed by Taylor and Ferrari (2009) in their LES of FSI.

In region I, the well-mixed/unstably stratified surface layer is significantly deeper compared to region II, and persistent stable stratification is observed only below...
approximately 23 m depth (Fig. 7c). The model of Taylor and Ferrari (2010) in (7) yields $h = 18.5$ m from the values in Table 2, underpredicting the observed value by approximately 20%. Figure 8a shows that the PV in region I is generally much closer to zero compared to region II but nevertheless reveals a layer with negative $f q$ and positive $f q$ down a depth of approximately $H = 27.5$ m. This leaves room for a less than 5-m-thick forced SI layer between 23 and 27.5 m depth, where strong turbulence coincides with significant stable stratification. Overall, however, the evidence for FSI is less clear compared to region II.

2) REGIONS III–V: MAIN FRONTAL REGION

We conducted a similar analysis for regions III, IV, and V to determine the stability characteristics of the main front around $-2 < x < 2$ km. As discussed in the context of Fig. 6, region III is located in a zone with large $\mathcal{M}^2$ but relatively small cyclonic vorticity with an average Rossby number, $\xi/f$, close to 1 (Table 2). This is reflected in the vertical and baroclinic contributions to the total PV (Fig. 10a), showing a clear dominance of the destabilizing baroclinic component $q_{bc}$ over the stabilizing vertical contribution $q_z$. The resulting layer with negative $f q$ has a thickness of $H = 17.5$ m, substantially larger than the thickness $h = 7$ m of the directly forced near-surface layer computed from (7). In contrast to regions I and II, we cannot directly confirm the existence of unstable stratification here as measurements in the upper few meters are unreliable due to ship effects. Nevertheless, Fig. 9d does show a strong increase of $N^2$ below $z = -h$, and Table 2 suggests that the buoyancy forcing $B_0 + B_E$ is approximately twice the value of region II, such that

Fig. 8. (top) Vertical, baroclinic, and total PV as defined in (2) and (5) for (a) subregion I and (b) subregion II and (bottom) turbulence dissipation rate and Ozmidov length (both plotted on a logarithmic scale) for (c) subregion I and (d) subregion II. Dashed red lines indicate regions where the Ozmidov scale is larger than the wall length scale, $L_O > \kappa |z|$. Gray dashed and solid lines denote the convective layer depth $h$ and the low-PV layer depth $H$, respectively.
gravitationally unstable stratification near the surface should be expected.

The forced SI layer \((-H < z < -h)\) in region III is largely geostrophically balanced (Fig. 9a) and strongly turbulent with \(L_O = O(1) \text{ m}\), suggesting that turbulence is not directly affected by the presence of the free surface (Fig. 10d). The Thomas and Taylor (2010) estimate for the FSI-induced dissipation rate averaged across the forced SI layer, \(\varepsilon_{SI} = (B_0 + B_2)/(1 - h/H)^{3/2}\) (section 2b), yields \(\varepsilon_{SI} = 0.4 \times 10^{-6} \text{ W kg}^{-1}\) in this case. This value is of the same order of magnitude but slightly smaller than the observed value, \(\varepsilon = 1.5 \times 10^{-6} \text{ W kg}^{-1}\), pointing at an additional source of turbulence in the forced SI layer. Values of the gradient Richardson number are close to or below the threshold for shear instability throughout the forced SI layer (Fig. 9d) suggest that region III may be characterized by a mixture of shear instability and FSI.

The stability regime in region III is contrasted by region V, where cyclonic vorticity is relatively strong, while \(M^2\) is small, and may even become negative at greater depths (Fig. 6). In view of this, it is little surprising that the frontal jet in this region shows no indications for a geostrophic balance (Fig. 9c), and no layer with negative PV can be identified (Fig. 10c). SI can therefore not be expected, and other instabilities must be acting to maintain the observed high levels of turbulence (Fig. 10f) in the presence strong re-stratification (note that \(N^2\) increases toward the surface in this region, opposite to what would be expected in a classical oceanic SBL). With the Richardson number fluctuating around the critical value \(R_i = 1/4\) in the upper 20 m of the water column (Fig. 9f), shear instability appears to be the most likely source of turbulence in this region. In regions deeper than approximately 20 m, the shear of the frontal jet decays, and \(R_i\) becomes too large for shear instability. It may be speculated that the patchy turbulence in this region is driven by internal-wave breaking with vertical and temporal scales too small to be resolved by our measurements.

Especially interesting is the frontal core region IV, where both cyclonic vorticity and \(M^2\) are large (Figs. 6a,c), suggesting a competition between stabilizing and destabilizing effects according to (2). Figure 10b shows that the
extreme vorticity, with $\tilde{\zeta} f = 4.8$ on the average (Table 2) and local values nearly twice as large (Fig. 6c), is strong enough to fully compensate the destabilizing effect of baroclinicity throughout the SBL. SI therefore cannot play a role in this region.

Figure 9e shows that, similar to region V, the vertical shear associated with the frontal jet is sufficient to maintain a Richardson number close to the critical value for shear instability throughout the upper 25 m of the water column.

It is worth noting that the situation in regions IV and V is reminiscent of the “marginal stability” regime described by Smyth and Moum (2013) in the context of a stably stratified shear layer in the upper flank of the equatorial undercurrent, with the caveat that rotation effects play a more important role in our case.

6. Discussion and conclusions

The main results of our study are summarized in Fig. 11, distinguishing between processes on the warm side of the front, where mixed layers are deep (Fig. 11a), and the frontal core region, where the maximum horizontal density gradient is observed inside the cyclonic flank of the frontal jet (Fig. 11b). With reference to Fig. 5d, the former approximately corresponds to the range $-25 < x < -10$ km, which includes regions I and II, while the latter focuses on the core frontal region around $-2 < x < 2$ km (regions III–V).

As summarized in Fig. 11a, the warm side of the front is characterized by a deep SBL and a moderate horizontal buoyancy gradient, gradually fading away from the frontal region. This suggests a decreasing impact of the destabilizing Ekman buoyancy flux on SBL turbulence, which can be quantified by combining the MO length scale defined in (8) with the expression for the Ekman buoyancy flux in (6), assuming $B_\varepsilon \gg B_0$ (as in our data). This yields

$$\frac{L_{\text{MO}}}{H} = \frac{1}{\kappa \cos \alpha} \frac{u_\psi f}{HM^2} = \frac{1}{\kappa \cos \alpha} \frac{u_\psi}{\Delta u_g}$$ (10)

with $\alpha$ denoting the angle between the downfront direction and the wind stress (recall that $\Delta u_g = HM^2/f$ is the geostrophic velocity difference across the low-PV layer). Thus, while strong winds have a tendency to increase both Ekman-driven convective and directly
shows that for downfront component (cos of (7) for this special case (not shown) shows that vective layer thickness can be used to replace one of the two independent pa-

Kelvin–Helmholtz instability. Driven turbulence, respectively. KHI denotes regions prone to

Abbreviations “w” and “c” refer to wind-driven and convectively

cyclonic flank of the frontal jet (see frontal structure in Fig. 5).

The situation is somewhat different in the core frontal

region (regions III–V), where the strong cyclonic vor-
ticity inside the flank of frontal jet plays an essential role (Fig. 11b). More specifically, we find that the stability regime of regions III and IV is characterized by an intricate balance between strong destabilizing baroclinic effects ([f/\Omega] is more than an order of magnitude larger compared to region II) and a comparably strong stabilizing effect of the cyclonic vorticity (large [f/\Omega]) due to Rossby numbers up to [c/\Omega = O(10)] [see (2)]. The stability analysis in section 5b has shown that conditions for SI ([f/\Omega] > [f/\Omega]) are satisfied only in region III, while in region IV, SI is suppressed due to the considerably stronger cyclonic shear. For region III, we identified a two-layer structure analogous to region II, which is, however, more confined to the near-surface region due to the overall smaller thickness of the negative-PV layer (Fig. 11b).

It is noteworthy that the vertical structure and the governing processes in regions II and III are strikingly similar to those found in the idealized simulations of FSI by Taylor and Ferrari (2010). Moreover, the dissipation rates we observed in the forced SI layer in the lower part of the SBL are in good quantitative agreement with a simple model by Thomas and Taylor (2010), based on idealized simulations, showing that the rate at which FSI

wind-driven SBL turbulence, (10) shows that the latter effect dominates. This tendency may, however, be compensated by an increase in [M^2] if the wind stress contains a downfront component (cos \alpha > 0). Expression (10) also shows that for \beta_E \gg \beta_0, we have \LMO/H \approx u_g/\Delta \theta_g, which can be used to replace one of the two independent parameters in the implicit equation for the relative convective layer thickness [h/H] in (7). The numerical solution of (7) for this special case (not shown) shows that [h/H] is a monotonically increasing function of \LMO/H, which implies that the relative thickness of the forced SI layer (1 - [h/H]) decreases with increasing \LMO/H.

The behavior predicted by (10) is in excellent agreement with the variability of [h, \ H, \ and \ LMO] among

regions I–IV (region V is located outside the frontal density gradient, and therefore shows a different behavior). From the values in Table 2, it is easy to show that \LMO is (approximately) inversely proportional to \M^2, as predicted by (10), and therefore strongly increases with increasing distance from the main front (going from region IV to region I). For the symmetrically unstable regions I–III, we find that the relative convective layer thickness, [h/H], increases monotonically with \LMO/H, following the theoretical prediction from (7) for \beta_E \gg \beta_0 (see previous paragraph). Moving from region III in the core of the front to region I on the light side, the values in Table 2 yield: [h/H] = (0.41, 0.46, 0.67) and \LMO/H = (0.21, 0.31, 0.57). As summarized in Fig. 11, this suggests that the convective layer gradually expands at the expense of the forced SI layer at the bottom of the low-PV layer.

Finally, for the outermost region of the front (-25 < x < -20 km), where the horizontal buoyancy gradient has decayed to \M^2 = 7.8 \times 10^{-8} \text{s}^{-2}, we find \LMO = 35.3 m, comparable to the total SBL depth. SBL turbulence at the outer edge of the front is thus largely wind driven, and not significantly affected any more by the frontal buoyancy gradient (Fig. 11a). This suggests that region I approximately marks the transition point at which FSI starts to become a relevant process in the lower part of the SBL.

FIG. 11. Sketch of governing processes (a) for -25 < x < -10 km, corresponding to the region with a deep SBL on the warm side of the main front, and (b) for -2 < x < 2 km, corresponding to the main frontal region with outcropping isopycnals inside the cyclonic flank of the frontal jet (see frontal structure in Fig. 5). Abbreviations “w” and “c” refer to wind-driven and convectively driven turbulence, respectively. KHI denotes regions prone to Kelvin–Helmholtz instability.
extracts energy from the background flow scales with the Ekman buoyancy flux. Previous studies of energy dissipation associated with FSI in submesoscale fronts arrived at similar conclusions (D’Asaro et al. 2011; Thomas et al. 2016). These investigations were, however, based on SBL-averaged turbulence parameters from a Lagrangian float, which complicates a direct comparison to the idealized simulations of Thomas and Taylor (2010) because SBL turbulence production by processes other than FSI (direct wind forcing, wave breaking, Langmuir turbulence, etc.) is more difficult to quantify. Our microstructure-based measurements resolve the vertical structure of SBL turbulence, allowing us to explicitly focus on the forced SI layer that is less likely affected by these processes due to the protecting effect of stable stratification and its deeper location at the bottom of the SBL, remote from the wave-affected near-surface region.

In regions IV and V, located at the edge of the frontal jet, we did not find any evidence for FSI. As mentioned above, in region IV, FSI was suppressed by the strong cyclonic vorticity in the upper 20 m, underlining the importance of our catamaran-based near-surface velocity measurements to correctly determine the stability properties of this narrow frontal region. Instead of FSI, we found that in both region IV and region V, the vertical shear associated with the frontal jet brings the Richardson number close to the critical value for shear instability (\(R_i = 1/4\)), suggesting that the observed high mixing rates in this region are associated with marginal shear instability (Fig. 11b). The overall conclusion from this is that in general both FSI and shear instability contribute to energy dissipation in submesoscale upwelling filaments.

So far, we have not discussed region “ISI” located inside the anticyclonic flank of the frontal jet (Fig. 5d). As shown in the following, a large fraction of this region is characterized by \(f q < 0\) and \(\zeta f \approx -1\), suggesting that the flow is unstable with respect to a mixed type of inertial/symmetric instabilities (denoted as ISI in the following). This regime, rarely investigated in field observations so far, was recently studied by Grisouard (2018) in a series carefully designed numerical experiments. Focus of the study by Grisouard (2018) were SBL instabilities growing inside a region with \(f q < 0\) and \(\zeta f \approx -1\) on the anticyclonic side of an idealized, finite-width front, similar to the situation encountered in our study. The main findings of Grisouard (2018) were the following: (i) during the initial phase, the fastest growing perturbations were aligned with the tilted isopycnals, as theoretically predicted for SI in an unbounded domain; (ii) for finite-width fronts, however, the cross-front velocities associated with the growing instability eventually induced isopycnal displacements and even density overturns; (iii) this implies an exchange of potential energy between the instability and the background flow, which is different from the classical view that SI grows at the expense of the background kinetic energy.

These properties are largely consistent with our observations in region ISI. As shown in Fig. 12b, the central part of this region satisfies the condition for instability (\(f q < 0\), see pink contour lines) and contains extended areas with \(\zeta f < -1\) (blue shading), indicating that the flow is unstable with respect to “baroclinic” inertial instability (or inertial-like ISI). This is supported by Fig. 12c, showing a layered structure of the cross-front velocity that is in many respects similar to that found in the simulations (Grisouard 2018, see their Figs. 4 and 8). Consistent with the structure of the cross-front velocity field are pycnostads and local overturns at the edge of the unstable region (near \(x = -7\) km), similar to those visible in Fig. 5 of Grisouard (2018). Figure 12a shows that these overturns are associated with strongly enhanced energy dissipation. One important conclusion
from our study is therefore that inertial-like ISI provides a direct pathway for the downscale transport of large-scale potential energy toward small-scale turbulence and dissipation, as recently suggested by Grisouard (2018). Our data also suggest that in submesoscale fronts, inertial-like ISI prevents Rossby numbers smaller than $\xi f \approx -1$ on the anticyclonic side of the front, whereas the cycloidal side remains stable with respect to SI even for $\xi f \gg 1$ due to the damping effect of large cyclonic vorticity. This result may help to explain the striking diurnal cycle of FSI up to a complete shutdown of FSI around midday. Future work, focusing on a neighboring filament in our study area, will highlight the effect of diurnal variability on frontal instability.

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