Elevated mixing at a front
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Abstract. The meso-, submeso-, and microscale structure of a front in the California Current was observed using a towed vehicle outfitted with microconductivity sensors. Thirty-three >60-km cross-front sections from 0–350 m in depth were covered in 3.5 days. Objectively-mapped data are fit via the Omega (ω) equation to obtain vertical velocity. A composite cross-front section shows elevated mixing on the dense side within 10-20 km of the front. Water downwells and gradients are elevated there; i.e., Rossby number (Ro), horizontal strain (γ), and vertical shear-vorticity barriers (i.e., cyclonic side of the front) are similar on the dense side, suggesting an energy cascade from the mesoscale via the submesoscale to the microscale. However, it is unclear whether frontogenesis, internal wave breaking by elevated vorticity, or internal wave trapping by large α produces the elevated mixing. The mean turbulent heat flux opposes the mean restratifying, mesoscale heat flux of 10 W m⁻² and may allow the front to persist. Turbulent nitrate fluxes are 0.1–0.3 mmol m⁻² s⁻¹. Chlorophyll fluorescence and beam transmission reveal a <6-km wide, ~100-km long along-front streamer which is a deep biomass maximum. Time scales for mixing and nutrient fluxes are 0.3–3 days, which are similar to phytoplankton growth rates and the time scale for frontal evolution.

1. Introduction

1.1. From Frontogenesis to Mixing

The cascade of energy from mesoscales of O(10–100 km)⁷ to submesoscales of O(1–10 km) and further to turbulence at microscales of O(1–100 mm) affects vertical buoyancy, momentum and biogeochemical fluxes [Capet et al., 2008c; Thomas et al., 2008; Klein and Laptey, 2009; Nagai et al., 2009; D’Asaro et al., 2011]. Submesoscale structures are found on the flanks of mesoscale fronts in subsurface observations, satellite images, and numerical models [Rudnick, 1996; Shearman et al., 1999, 2000; Capet et al., 2008a, b; Thomas et al., 2008; Johnston et al., 2009; Klein and Laptey, 2009] as thin, elongated streamers of pycnocline tracers, strong vertical velocity, and O(1) Rossby (Ro) and Richardson (R) numbers (e.g., Figure 1) [Barth et al., 2001; Nagai et al., 2008; Capet et al., 2008a; Thomas et al., 2010; Pallás-Sanz et al., 2010a, b]. Fronts in the California Current System are ubiquitous in satellite sea surface temperature (SST) [Castellano et al., 2006] and in models [Capet et al., 2008a].

Quasi-geostrophic (QG) and semigeostrophic theory and numerical models indicate that gradients and downwelling are intensified on the cyclonic, dense side of fronts with more diffuse upwelling on the light side [e.g., Hoskins and Bretherton, 1972; Thompson, 2000; Rudnick, 2001; Mahadevan and Tandon, 2006; Capet et al., 2008a; Thomas et al., 2008]. Observations confirm this, but reveal complex three-dimensional submesoscale structure with Ro = ω/f of O(1) [Rudnick, 1996], where the Coriolis frequency is f and the relative vorticity is ω = vₓ – uᵧ. (x, y, and z are coordinates in nates and u, v, and w are velocities, which are positive eastward, northward, and upward; coordinate subscripts denote partial derivatives.) The coastal upwelling front in this paper displays a submesoscale cyclonic circulation with Ro = 0.3, which is not resolved by altimetry (Figure 1). Strong downwelling at such sites with enhanced cyclonic vorticity distorts the lateral buoyancy gradients and may facilitate vertical shear- or strain-driven diapycnal mixing. In other words, mesoscale potential energy is converted to kinetic energy at submesoscales, which cascades to microscales where it is dissipated [Mahadevan and Tandon, 2006; Capet et al., 2008c].

Despite the ubiquity of fronts and associated submesoscale structure, few observations have had both the spatial resolution and coverage to show elevated mixing at fronts [Garret, 1978; Marmion et al., 1985; Dewey et al., 1993; Lilibridge et al., 1990; Hitchcock et al., 1994; Hales et al., 2009; D’Asaro et al., 2011]. Elevated turbulence on the dense side of a front has been noted, but with limited resolution either in the horizontal [Nagai et al., 2009] or at the centimeter scales of scalar dissipation [Lilibridge et al., 1990; Hitchcock et al., 1994; Hales et al., 2009].

Our results show elevated meso- and submesoscale gradients on the dense side of the front along with elevated microscale gradients and vertical mixing within 20 km of the front on the dense side. Such conditions suggest some mixing mechanisms: double diffusion at intrusions [Garret, 1978; Schmitt, 1994]; internal wave trapping in regions of lower effective vorticity (i.e., anticyclonic side of the front) [Kunze, 1985] and/or internal wave reflection at high effective vorticity barriers (i.e., cyclonic side of the front) [Rainville and Pinkel, 2004]; mesoscale straining of internal waves, which leads to exponentially-growing wavenumbers and subsequent internal-wave breaking [Bühler and McIntyre, 2005]; and loss of balance in submesoscale flows converting potential energy to kinetic energy at smaller scales, which in turn leads to shear instability and mixing [Molmath et al., 2010].

1.2. Frontal Dynamics and Mixing at This Coastal Upwelling Front

To better understand frontal dynamics and mixing, we conducted two quasi-synoptic mesoscale surveys of a coastal
upwelling front in the California Current System from 3b
July–3 August 2006 (survey 1) and again from 3–7 August 2006 (survey 2). Instruments included a vessel-mounted acoustic Doppler current profiler (ADCP) and a towed vessel (SeaSoar) equipped with a conductivity-temperature-depth (CTD) instrument and microconductivity sensors. The front separated cold upwelled water from warmer offshore water (Figure 1a). A long, thin, along-front streamer of chlorophyll was seen (Figure 1b) extending along the dense, cyclonic side (Ro > 0, Figure 1d). Using objectively mapped currents (section 2.2), we find the along-front streamer is consistent with the lateral, mesoscale strain (Figure 1c):

\[ \alpha = \sqrt{(u_x - v_y)^2 + (v_x + u_y)^2} \]  

which is also large on the dense side. Dynamic topography shows a low around which the streamer is advected (Figure 1b) and which is consistent with the in situ density (Figure 1d). Near-surface currents are alongfront (roughly south-eastward) on the light side (Figure 1c).

In our previous work, these two surveys were used to understand three-dimensional frontal dynamics from tracer and potential vorticity perspectives [Pallàs-Sanz et al., 2010a] and if the observed currents and hydrography to dynamics via a generalized Omega (\( \omega \)) equation [Pallàs-Sanz et al., 2010b]. Downfront wind forcing and nonlinear Ekman dynamics were important during survey 1, while frontogenesis was driven by mesoscale strain in survey 2. While survey 1’s downfront winds can contribute to mixing [Pallàs-Sanz et al., 2010a, b], mixing on the dense side of fronts as found in survey 2 arises from intensified gradients due to mesoscale strain is, unrelated to local winds, and is likely ubiquitous [Capet et al., 2008c]. Pallàs-Sanz et al. [2010b] use mixing parameterizations in a generalized \( \omega \) equation to show elevated mixing intensifies \( w \) and concentrates \( w \) closer to the surface. However, these results depend sensitively on the details of the parameterization.

Here, we relate observed microscale dissipation to realistic frontal dynamics—intensified gradients and downwelling on the cyclonic side. A restratifying, cross-front ageostrophic secondary circulation (ASC) links weaker upwelling on the light side to the stronger downwelling. The downwelling of a chlorophyll streamer and potential vorticity anomaly between surveys verifies the inferred downward \( w \) on the dense side [Pallàs-Sanz et al., 2010a]. Microstructure instrument problems were encountered during survey 1, which is not considered further here. This paper covers survey 2, from which 13 cross-front sections are averaged to obtain a composite two-dimensional cross-front section with sufficient microstructure data.

1.3. Outline

This paper is structured as follows. SeaSoar and microstructure methods are given in section 2. The composite front is also described. One example of a cross-front section is shown in section 3, while the composite results are in section 4. In sections 5–6, turbulent heat and nutrient fluxes are calculated to demonstrate the relevance of mixing at the front. In section 7, a deep chlorophyll and biomass maximum is described and relevant time scales are estimated. A summary follows.

2. Data and Methods

2.1. Overview

From 3-7 August 2006, SeaSoar and vessel-mounted ADCP were used to obtain 13 quasi-synoptic (for the mesoscale), zonal, cross-front sections of hydrography, currents, and microstructure at 6° or 11-km meridional spacing from 35° 18’N to 36° 30’N (Figure 1). These 13 sections (one example section is shown in Figures 2 and 3) were used to produce a composite cross-front section (Figures 4 and 5). The processing of hydrographic, current, and microstructure data is summarized below. A full explanation is in Johnston et al. [2011].

2.2. Hydrography and Currents

SeaSoar profiles from the surface to a depth of ~350 m, while towed at 4 m s\(^{-1}\). Raw 24 Hz temperature (\( T \)) and salinity (\( S \)) from the SeaBird SBE 9plus CTD are averaged to 1 Hz, combined with Global Positioning System position and time, and then averaged in 12-minute (or ~3-km horizontal) bins and vertically in 8-m bins. The time bin is longer than the time for SeaSoar to complete a dive cycle from the surface to depth and back again.

A Wetlabs C-star transmissometer measures the turbidity of the water via optical beam attenuation at 660 nm over a 25-cm path length. The exponential decay scale of the transmitted light is the inverse of the beam attenuation coefficient (\( c \)). The SeaPoint chlorophyll fluorometer outputs a voltage which scales linearly with chlorophyll concentration. This voltage is scaled by a factor of 10 to produce numerical values similar to chlorophyll concentrations in mg m\(^{-3}\).

Current profiles were measured with an RDI Ocean Surveyor 75 kHz ADCP at 8-m resolution in depth from 16-100 m. Single ping amplitude and velocity were despiked and averaged into 12-minute time bins like the SeaSoar data. Error from transducer misalignment was minimized over a reference layer from 48-264 m using a complex correction factor to ship-relative velocity [Rudnick and Lutyten, 1996].

Since it is difficult to observe \( w \) directly due to its small magnitude, it is often inferred from density and current data via the \( \omega \) equation, which relies on eliminating time derivatives in the density and momentum equations [Hoskins et al., 1978]. Full details of our approach are in Pallàs-Sanz et al. [2010a] and Pallàs-Sanz et al. [2010b], but a summary follows here. Hydrographic and current data from the 13 sections are objectively mapped to remove aliased signals and noise (i.e., internal waves [e.g., Johnston et al., 2011]). For the objective map, decorrelation scales obtained from the data covariance are 15 and 43 km in the zonal and meridional directions. Mean fields of potential density (\( \sigma_\theta \)) are planar and currents are constant, which is consistent with thermal wind [Rudnick and Lutyten, 1996]. To include frontal curvature, we use the absolute geostrophic velocity as a first approximation, but ageostrophic forcing terms could be included for higher order approximations [Pallàs-Sanz et al., 2010b]. Objective mapping is an important step because otherwise noise would obscure the second order derivatives of currents and density that are needed to force the \( \omega \) equation via the divergence of the horizontal frontogenetic vector or \( \mathbf{Q} \) vector:

\[ N^2 \nabla^2 w + f^2 w_{zz} = 2 \nabla \cdot \mathbf{Q} \]  

where the buoyancy frequency is \( N^2 = -g/\sigma_\theta/\rho_0 \), \( \nabla \) is the horizontal gradient operator, \( \mathbf{Q} = \mathbf{u} \times \nabla \sigma_\theta/\rho_0 \), \( \mathbf{u} \) is the horizontal current vector, \( g \) is gravitational acceleration, and \( \alpha_\theta \) is the mean potential density. Divergent (convergent) \( \mathbf{Q} \) leads to downwelling (upwelling).

Vertical and along-isopycnal gradients of potential temperature (\( \theta \)) and \( S \) reveal regions of thermohaline variability. The spice gradient or along-isopycnal \( \theta_x \) reveals where water masses have been stirred but not yet mixed [Ferrari and Rudnick, 2000]. Similar results are obtained with \( S \) gradients. To highlight regions of double diffusion, we take vertical \( \theta - S \) gradients from the binned data, scale them by
the thermal expansion ($\alpha_T$) and saline contraction ($\beta_S$) coefficients, and produce the density ratio:

$$R_\rho = \frac{\alpha_T}{\beta_S}$$

(3)

A more practical indicator of double diffusion is the Turner angle, which has equally sized domains for salt fingering and diffusive convection, is well defined for $S_z = 0$, and has a finite scale [Ruddick, 1983]:

$$T_u = \tan^{-1}\left[\frac{(\alpha_T - \beta_S S_z)}{(\alpha_T + \beta_S S_z)}\right]$$

(4)

With the convention that stable profiles have positive $\alpha_T$, the region where $-45^\circ < T_u < 45^\circ$ is doubly stable, while more positive (negative) values favour salt fingering (diffusive convection).

Bottle data were obtained from upcasts at about 1835 UTC on 2 August 2006 by the Monterey Bay Aquarium Research Institute at CTD station 63.4-60 (37° 2.5’N, 123° 11.8’W) as part of their ongoing time series near Monterey Bay [e.g., Collins et al., 2003]. Nitrate and phosphate concentrations, [NO$_3^-$] and [PO$_4^{3-}$], were obtained following procedures in Pennington and Chavez [2000].

2.3. Microstructure

To infer microscale temperature variance dissipation rates as in Johnston et al. [2011] and similar to Dillon et al. [2003], the Transmitting Microstructure System (TMS) was mounted beneath the SeaSoar with dual Seabird SBE-7 mid-croconductivity probes and 3 accelerometers sampling at a frequency ($f$) of 2048 Hz. Shear probes cannot be used because of vibrational contamination from SeaSoar at the scales of interest. Scalar microstructure is at centimeter scales and is little affected by SeaSoar displacements, which are small ($\sim$ 1 mm) in comparison. Microconductivity data, coherent with the accelerations are removed [Goodman et al., 2006]. TMS data are converted to physical units in 4-s intervals (roughly 4 m in the vertical) using the CTD data.

Microconductivity data are converted to temperature only when CTD-measured $\partial T/\partial C$ > 5°C S$^{-1}$ m to avoid regions where $S$ dominates $C$ [Washburn et al., 1996; Nash and Moun, 1999, 2002; Dillon et al., 2003]. An anti-aliasing filter attenuates signals with $f$ > 680 Hz.

Following Osborn and Cox [1972], we use microscale temperature gradients ($T_z^2$) to estimate temperature variance dissipation rates:

$$\chi = 6D_T T_z^2$$

(5)

for fully-resolved gradients, where $D_T$ is the molecular thermal diffusivity. However, the TMS has a typical vertical resolution of 5-10 cm. Therefore, we fit the observed vertical wavenumber spectrum of $T_z^2$ to the Batchelor spectrum, which is a universal form for turbulent scalar fluctuations [Nash and Moun, 1999], and a noise spectrum. Spectra are calculated from half-overlapping 2-s windows and averaged over 5, 9, and 31 bands for $f < 10$, 100, and 680 Hz. Our detection threshold is $\chi \sim 10^{-10}$ oC$^2$ s$^{-1}$. Downgoing profiles are used, when the TMS is oriented into clean flow. Microstructure data are averaged into 8-m vertical bins and 12-minute time bins like the hydrographic and current data, are calculated for all the data and for a subset of the doubly stable data, which is conservatively given by $|T_u| < 35^\circ$ and $R_\rho < 0.7$. In doubly stable water, we use $\chi$ and the mean temperature gradient to estimate turbulent thermal diffusivity:

$$K_T = \frac{x}{(2T_z^2)}$$

(6)

which is low in regions of strong $T_z$ above 100 m even if it is large [Johnston et al., 2011].

2.4. Composite Front

To emphasize frontal processes and reduce noise (e.g., from either internal waves or instruments) we make a composite front from all 13 cross-front sections. Furthermore, compositing is required to produce microstructure estimates of turbulent mixing at the front. The TMS had a data return of 25% because only downward profiles are used and some data are lost due to occasional biofouling. Also data are processed and eliminated with steps described here and in Johnston et al. [2011]. While the three-dimensionality of fronts is important [Ruddick, 1996; Pallás-Sanz et al., 2010a, b], compositing is necessary to increase the data in each distance-bin because of the data return and the intermittency of turbulent mixing.

All data are gridded in 3 km bins based on their eastward distance from front, which is defined by the position of $\sigma_\theta = 24.55$ kg m$^{-3}$ at a depth of 24 m as the region of maximum gradient for the mapped $\sigma_\theta$. The mean location of the maximum gradient for the 13 sections is at $x = -0.4$ km, but can vary by about 2 bins (6 km) on an individual section. Positive values of $x$ denote the dense side. Vertical bins are 16 m. We find some differences in detail, but no differences in the conclusions for the binning at 3 or 10 km and 16 or 24 m. The number of data points in each bin for hydrographic data ($n_x$, section 2.2) and for microstructure data ($n_x$, section 2.3) differ because of quality control steps for the microstructure data (section 2.3 and following paragraph). Note that the front is oriented roughly southeastward and $x$ is at an angle to the front.

One additional data editing step is taken with the microstructure data that is not done by Johnston et al. [2011]. Where values of $K_T$ in the composite exceed $10^{-3}$ m$^2$ s$^{-1}$, each contributing spectrum from the 4-s data and resulting estimates of turbulent quantities are examined. Such 4-s data are excluded as unrealistic if $K_T > 10^{-2}$ m$^2$ s$^{-1}$ and then the binned values are recalculated. About 20 points of >4000 contributing to the composite were excluded.

Lastly, to form a composite for $R_\rho$ and $\alpha$, data are taken from the objective map, while composite $w$ is from the QG fit (section 2.2). The QG approximation may not be strictly valid because on the dense side $Ro = 0.2\pm 0.5$ and extends to depths of 350 m. Non-QG effects, such as frontal curvature and large $Ro$, are considered by Pallás-Sanz et al. [2010b], but even these methods involve other assumptions. Here, we only use the QG model, which is the simplest realistic model of frontal dynamics.

3. One Cross-Front Section

An example of a zonal section across the coastal upwelling front at 36° 18′N shows cold, salty water on the dense side ($x > 0$ km) flowing northwestward and warm, fresh water on the light side ($x < 0$ km) flowing southeastward (Figures 2a, b, d, and e). The $\omega$ equation yields a downward (upward) $w$ on the dense (light) side (Figure 2f).

As expected, gradients are enhanced on the dense side. Cycloonic vorticity (i.e., $Ro > 0$) is elevated (Figure 2h). $\alpha$ is often larger within 20 km of the front from 100-200 m (Figure 2g). Spice gradients are found on either side of a saline intrusion (Figure 2h-c). Interleaving and favourable conditions for double diffusion and salt fingering are indicated by $|T_u| > 45^\circ$, which appear more common below 200 m (Figure 2i). Mapped currents are first smoothed with a 13 point running mean (104 m) and then reduced shear is calculated as:

$$S_{2R}^2 = u'^2 + v'^2 - 4\nu^2$$

(7)

which is larger (less negative) on the light side only due to the uplift of isopycnals (Figure 2i). $S_{2R}^2$ reflects the mean shear even without the mapping and smoothing of currents.
Limited data return from the microconductivity sensors does not provide a complete view of the front from a single section, but this example section shows one region of elevated \(\chi\) below 75 m from \(x = 0 - 20\) km and another region at the mixed layer base on the light side (Figure 2j). In the interior, \(\chi\) appears lower for \(x < 0\) km and \(x > 20\) km. \(\chi\) is a measure of vertical microscale temperature gradients regardless of their origin—turbulence or double diffusion. Although \([Tu]\) exceeds 45\(^\circ\) in both sides of the front (Figure 2i), these regions are not correlated with elevated \(\chi\) (Figure 2j). Greater stratification near the surface leads to smaller \(K_T\) there, while larger \(K_T\) is found below 75 m from \(x = 0 - 20\) km with smaller values away from this region similar to \(\chi\) (Figures 2j-k).

To further illustrate the structure of the dense side, depth-mean \(\chi\) means are calculated from 56-200 m. The depth-mean spice gradient is large and positive (negative) from \(x = 0 - 20\) km (20-40 km) on either side of the salty intrusion (Figure 3a). Cyclonic vorticity is highest (Ro \(\sim 0.2\)) from \(x = 0 - 10\) km (Figure 3b) and is mainly due to \(w\) (Figures 2e and 3e) which was observed with higher resolution than \(u\) and \(v\) (Figures 2f and 3f), while the depth mean of \(u\) appears not to vary much within 20 km of the front (Figure 3d). \(S_T^2\) is larger (less negative) on the dense side, which is due to isopycnal uplift (Figure 3g). Depth-mean \(\chi\) is largest at the front and elevated from \(x = 0 - 15\) km by a factor of 3-10 compared to the light side (Figure 3g), while \(K_T\) appears about an order of magnitude larger from \(x = 0 - 20\) km (Figure 3h).

4. Composite Front

This section describes our main point-mixing is elevated on the dense side of the front. To emphasize frontal processes and reduce noise, we form a composite cross-front section by averaging all 13 cross-front sections based on the position relative to the front (section 2.4). Due to the intermittency of mixing, microstructure quality control procedures, and microstructure data return, no single cross-front section provides a complete picture of mixing.

The composite front shows many of the same features found in the example of a single cross-front section described in the previous section. Cold, salty water flows northwestward on the dense side and vice versa on the light side (Figures 4a, b, d, and e). \(w\) is downward (upward) on the dense (light) side (Figure 4f). Ro is cyclonic and relatively large as all measured depths on the dense side, while weaker anticyclonic values are found on the light side (Figure 4h). Larger space gradients are found on the dense side (Figure 4c). \(Tu\) is calculated from binned values of \(\alpha_0\beta\) and \(\beta^2S_T^2\). \(Tu\) is largest below 200 m on the light side and near the surface on the dense side (Figure 4l). \(S_T^2\) is largest (least negative) in the interior, smallest at the mixed layer base, and is elevated on the dense side only due to tilted isopycnals (Figure 4i).

Microstructure is elevated on the dense side (Figures 4m-n). \(\chi\) is largest in the two regions mainly from \(x = 0 - 20\) km and also at the mixed layer base on both sides of the front (Figure 4m). Below 100 m, \(\chi\) is about an order of magnitude larger on the dense side than on the light side. Double diffusion may contribute to \(\chi\) as indicated by the larger values of \(Tu\) which approach 45\(^\circ\) (Figure 4l). However, \(K_T\) is largest from \(x = 0 - 20\) km at depths below 100 m (Figure 4n) and its spatial structure resembles not \(Tu\) but Ro (Figure 4h). To avoid double diffusive regions, \(K_T\) is composited based on individual latitude-distance-depth bins with \([Tu] < 35\) and thus reflects turbulence rather than double diffusion. The number of microstructure data in each composite bin is a factor of 2-5 less than in the hydrographic and current data but there are 10-20 data points per bin and little difference in sampling density from the dense to light sides (Figures 4j-k).

Elevated gradients and microstructure on the dense side are further illustrated by their depth means, which are calculated from 56-280 m (Figure 5). As noted above, currents are northwestward (southeastward) and \(u\) is downward (upward) on the dense (light) side (Figure 5d-e). Cross-front gradients for \(v\) and \(w\) are strongest at the front or on the dense side. Depth-mean Ro is largest from \(x = 0 - 10\) km, while \(\alpha\) has a broad peak on the dense side (Figure 5a). Vertical and isopycnal temperature gradients are largest on the dense side (Figure 5b-e). \(S_T^2\) is larger (less negative) on the dense side due to uplifted isopycnals (Figures 5i).

Depth-mean \(K_T\) and \(\chi\) are 3-10x larger on the dense side from \(x = 0 - 20\) km (Figures 5g-h). To exclude doubly diffusive regions \(K_T\) and \(\chi\) are calculated only from bins on individual cross-front sections where \([Tu] < 35\) (e.g., Figure 2i) and plotted as dashed grey lines (Figures 5g-h). Double diffusion proceeds when \([Tu] > 45\) and so our estimate is conservative. Figures 5g-h show little difference for depth-mean values of \(K_T\) and \(\chi\) from doubly stable water (dashed grey lines) or all data (black lines). Their maximum values are still \(\sim 3x\) bigger than away from the front. The number of data points excluded from the doubly stable depth mean is not considerable (lower black line and dashed line, Figure 5i).

The ASC at the upwelling front leads to restratifying (upward and positive) heat fluxes almost everywhere (Figure 6b). The heat flux is:

\[ J_{qw} = \rho c_p (w' \theta') \]  

where \(\theta'\) and \(\theta''\) denote deviations from the mean density, the in situ density is \(\rho\), and the heat capacity is \(c_p = 4186\) J kg\(^{-1}\) K\(^{-1}\). Here and below, mean values in depth and eastward distance are denoted by \(\langle \rangle\) and \(\langle \cdot \rangle_x\). This restratifying flux reaches a maximum, \(J_{qw} = 40\) W m\(^{-2}\), at 50-m depth (Figure 6d). The composite front is used for this calculation and averaging reduces our peak flux from the value of 60 W m\(^{-2}\) at 50-m depth in Pallás-Sanz et al. [2010b], who use the area in Figure 1c for their calculation. Warming above and below peak arises from flux convergence:

\[ -J_z = \rho c_p D_z \theta \]  

where \(D_z\) denotes a material time derivative. The warming is \(>0.2^\circ\) C day\(^{-1}\), while cooling peaks at 0.5°C day\(^{-1}\) at 60 m and reduces to 0.05°C day\(^{-1}\) at 130 m. These estimates represent an average over \(\sim 80\) km in the cross- and alongfront directions and would be higher if restricted closer to the front.

5. Heat Flux

Turbulent and mesoscale heat fluxes are calculated below to demonstrate the relevance of elevated mixing on the dense side of the front. The mesoscale heat fluxes at fronts due to ASC are comparable to climatological surface flux values and can be more effective at restratification because they warm the upper water column while cooling below [Rudnick and Layton, 1996]. As density gradients increase during frontogenesis, so do vertical shear and strain. Eventually mixing and turbulent heat fluxes must become important as Hoskins and Bretherton [1972] suggested nearly 40 years ago. Since fronts persist despite the restratifying ASC and heat fluxes, destratifying turbulent fluxes must be of comparable magnitude. This balance is not necessarily steady. Here we calculate mesoscale and turbulent heat fluxes at the composite front using two independent methods: frontal dynamics from the QG \(\omega\) equation [Pallás-Sanz et al., 2010b] and our microstructure measurements.

The ASC at the upwelling front leads to restratifying (upward and positive) heat fluxes almost everywhere (Figure 6b).
Our microstructure measurements yield heat fluxes of

\[ J_{\beta} = -\rho c_p K_\beta T_z \]  

which are downward (negative) everywhere because of the positive temperature gradient (Figure 6a). The downward flux, \( J_{\beta} \), reaches 70 W m\(^{-2}\) on average, and is below 20–30 W m\(^{-2}\) from 100–200 m and within 20 km of the front. Turbulent flux divergence leads to cooling from 80–140 m with a maximum of 0.2°C day\(^{-1}\), which is about twice as large as \(\langle q_k \rangle_x \) to \(\langle -q_k \rangle_x \) (Figure 6c-d). Little turbulent heat transport takes place further away from the front. Turbulent flux divergence leads to cooling from 80–140 m with a maximum of 0.2°C day\(^{-1}\). Warming of \(\sim 0.1°C\) day\(^{-1}\) extends below that to 240 m. The cooling from flux divergence above the microstructure peak at 125 m is within a factor of 2–3 of the mesoscale cooling.

While the ASC restratifies the front, turbulent mixing works to homogenize the water column. The associated heat fluxes have different spatial distributions, but over all the mean heat fluxes are comparable: \(\langle -J_{\beta} \rangle_z \) is 10 W m\(^{-2}\) and \(\langle J_{\beta} \rangle_z \) is 8 W m\(^{-2}\). The turbulent flux is largest on the dense side, while the restratifying heat fluxes are largest on either side of the \(-J_{\beta}\) maximum (Figure 6c). \(J_{\beta}\) peaks at the mixed layer base, while \(-J_{\beta}\) is larger further below (Figure 6d).

With turbulent heat flux estimates in hand, the next obvious question is: what is its energy source? Possible sources for the turbulence include internal wave breaking, submesoscale instabilities, and baroclinic instability. We are unable to definitively identify a source, but some evidence suggests a relationship between restratifying fluxes and submesoscale processes. For example, mixing and turbulent heat flux is greatest on the dense side of the front where the isopycnals are steepest and gradients \(\langle \sigma \rangle\) are largest. However, criteria for submesoscale instabilities were inconclusive: we found neither low Ri indicative of shelf instability nor a sign change in the difference between absolute vorticity and strain rate indicative of an unbalanced ageostrophic instability [Molenkoper et al., 2005; Mahadevan and Tandon, 2006]. We leave a dynamical investigation of the source of turbulent buoyancy for future work.

6. Nutrient Flux

Mean turbulent nutrient fluxes at the front are calculated by combining our microstructure observations with nutrient data from nearby CTD bottle samples. To assess \(\text{in situ}\) nutrient concentrations, we look at water samples which were taken at a nearby station (section 2.2). Nutrient concentrations are correlated with density (Figure 7a), from which we estimate mean nutrient concentrations, gradients, and fluxes at the front. Correlation coefficients for \(x^2 = 0.906\) for \(\text{NO}_3\) and \(0.909\) below 25 m. For the mixed layer, constant concentrations are used (for \(\sigma < 24.772\) kg m\(^{-3}\); \[\text{NO}_3 = 5.46\) mmol m\(^{-3}\) and \[\text{PO}_4 = 0.94\) mmol m\(^{-3}\). The ratio of the fitted slopes is \(m(\text{NO}_3)/m(\text{PO}_4) = 14.5\), which is near the expected photo-trophic or Redfield nitrogen-phosphorus ratio of 16 N/15 P.

To calculate nutrient fluxes, we assume our mean measured diffusivity, \(K_T\), applies not just to temperature but other scalars, such as nutrients. Then nitrate flux is calculated as:

\[ J_{\text{NO}_3} = -(K_T)z[\text{NO}_3]_z \]  

and similarly for \(\text{PO}_4\). Both nutrient fluxes are upward (positive) and have similar vertical structure (Figure 7c). But \(J_{\text{NO}_3}\) is about twice as large as \(J_{\text{PO}_4}\). These values are 1–4x lower than found using direct nutrient and microstructure measurements at a shelf break front [Hales et al., 2009]. Of particular interest here are the smaller flux peaks near 50 m (due to the nutrient gradient rather than diffusivity), which are discussed in the next section.

7. Deep Chlorophyll and Biomass Maxima

Deep chlorophyll and biomass maxima (DCM and DBM) are found along the northernmost 9 (and possibly 11) cross-front sections as a streamer of subsurface chlorophyll and turbidity (Figures 8–9). These DCM and turbidity maxima are found from 50–100 m, are distinct from the near-surface maximum, and may be due to differing phytoplankton constituents [Li et al., 2010]. The DCM are strongest near 36° 24' N, where we see a chlorophyll filament is advected and strained at the front (Figures 1b-c). In all cases, the DCM are accompanied by turbidity maxima and therefore these are DBM. The DBM are found mostly on the dense side of the front from \(x = 0–20\) km. These limits are indicated by thick grey lines in (Figures 8–9).

Next we estimate time scales associated with mixing and nutrient fluxes. Using a median carbon to chlorophyll ratio of 100 [Behrenfeld et al., 2005; Li et al., 2010] and applying this to the observed chlorophyll distribution with a maximum of about 0.5 mg m\(^{-3}\), we obtain a maximum carbon concentration of roughly 50 mg m\(^{-3}\) in the DCM. If all the upwelled nutrients are used in new production at the Redfield carbon:nitrogen ratio of 106 C:16 N, then the new production near 50 m is 10 mmol C m\(^{-2}\) day\(^{-1}\) to the nearest order of magnitude or 12 mg C m\(^{-2}\) day\(^{-1}\). A crude estimate for a typical time scale of the carbon flux needed to maintain a 1–10 m thick DCM is: (1–10 m) \times 50 mg m\(^{-3}\) / 12 mg C m\(^{-2}\) day\(^{-1}\) = 0.5–5 days. Such a time scale is comparable to the diffusion time scale \((= l^2/(4K_T))\) from the solution to the one-dimensional diffusion equation, Fick’s second law of 0.3–3 days across a length \(l = 10\) m based on the observed \(K_T = 10^{-5}–10^{-3}\) m\(^2\) s\(^{-1}\) (Figure 7b). These time scales are similar to time scales from typical phytoplankton growth rates of \(\Omega_{0.1–1}\) day\(^{-1}\) and time scales of \(\Omega_{0.1}\) day for frontal evolution in the California Current System [Abbott and Letelier, 1998; Behrenfeld et al., 2005; Li et al., 2010; Pallás-Sanz et al., 2010a]. However, the various assumptions, constants, and ratios used here vary depending on numerous biological and physical processes. A further caveat is warranted for the fluorescence-chlorophyll conversion (section 2.2).

While mixing ultimately diffuses the DBM, other physical processes can contribute to its formation and decay: i.e., vertical motion, sheared flow, or straining by horizontal currents at the front. The 5 northern sections have the strongest DCM (Figures 8–9), which decrease as a increases (Figure 1c).

In summary, our observations of the DBM suggest an initially nonuniform distribution of phytoplankton has been advected, sheared, and strained at the front to produce a long, thin, alongfront streamer. The calculated time scales for mixing and nutrient fluxes have an order of magnitude similarity to typical phytoplankton growth time scales and the time scale for frontal evolution.

8. Summary

Using a towed vehicle we conducted a spatial survey of currents, hydrography, and scalar microstructure at a coastal upwelling front. Obtaining simultaneous observations of the meso-, submeso-, and microscale is challenging. Here we show evidence from a composite of 13 cross-front sections for enhanced microscale gradients and mixing on the dense side of a coastal upwelling front, as suggested by Hoskins and Bretherton [1972]’s QG model of frontal dynamics.
Elevated microstructure at the front is found from \( x = 0 \) to 20 km on the dense side and from depths of 48-168 m. One of the sections across the front, \( \chi \) resembles \( \theta_w \) in the individual section suggesting mixing on the edge of a salinity intrusion. In the composite front, enhanced submesoscale gradients are seen in \( R_\alpha, \tau_u \), and spice gradients on the dense side of a front. Microscale gradients are also enhanced on the dense side of the front. \( K_\tau \) is largest on the dense side within 20 km of the front and its spatial structure resembles downward \( w \), positive \( R_\alpha \), and larger \( \alpha \) (Figure 16). Regions with \( [\tau_u] > 35 \) are excluded in the calculation of \( K_\tau \) at the composite front and even when they are included there, there is little difference and our conclusion remains: turbulent mixing on the dense side of the front is elevated.

Mixing can strengthen frontal circulation. Increased parameterized mixing in the upper 100 m leads to intensified there [Pallás-Sanz et al., 2010b]. Our observations of deeper elevated mixing suggest \( w \) is also larger at depth. The turbulent heat flux is also largest from \( x = 0 \) to 20 km on the dense side of the front from depths of 50-200 m. The re-stratifying mesoscale flux is maximum on either side of the downward turbulent heat flux, which also has a deeper maximum than the re-stratifying flux. Do the suppressing turbulent and re-stratifying heat fluxes affect the front? We cannot definitively answer this question, but it appears possible the mean re-stratifying heat flux due to the ASC and the mean destratifying turbulent heat flux are comparable at \( \sim 10 \) W m\(^{-2} \). This compensation between the re-stratifying and turbulent heat fluxes is remarkable because the re-stratification occurs at submeso- and mesoscales, while the turbulence is at microscales.

Though we cannot definitively differentiate between possible energy sources for the turbulence with available data, we speculate that elevated turbulent mixing at the front may be due to some combination of the following: (1) frontogenesis which produces elevated gradients including current shear on the dense side, (2) near-inertial internal waves generated east of the front during a wind shift prior to the frontal survey are blocked and reflected by elevated vertical viscosity on the dense side of the front, or (3) these internal waves experience horizontal strain which leads to growing wavenumbers and shear. The measured shear was indicative of the mean shear and there was no sign that finescale or \( O(10) \) m shear was higher on the dense side as would be expected from either frontogenesis, blocking, or straining. However, elevated \( R_\alpha \), and other gradients on the dense side of the front suggest the submesoscale is involved in an energy cascade from mesoscale to microscale [Capet et al., 2008a,c].

At this front, we find a deep chlorophyll and biomass maximum which is located within 20 km of the front on the dense side. This deep chlorophyll maximum is separated from another near-surface maximum, which was described in Pallás-Sanz et al. [2010a]. This deep maximum is likely advected, sheared, and strained at the front to produce thin, narrow, alongfront streamer. The time scales for mixing, nutrient fluxes, frontal evolution, and phytoplankton growth from our data are about 0.5-5 days, which are similar to other estimates [Behrenfeld et al., 2005; Hales et al., 2009; Li et al., 2010]. This result suggests phytoplankton are adapted to take advantage of micro-, submeso-, and mesoscale physical processes at fronts [Abbott and Letelier, 1998].

The strength of this front and its vertical exchange are typical or even small compared to other fronts [Lillicbridge et al., 1990; Pollard and Regier, 1992; Hitchcock et al., 1993; Rudnick, 1996; Shearman et al., 1999; Barth et al., 2001; Hales et al., 2009; Thomas et al., 2010; Thomas and Joyce, 2010]. Therefore, it is likely that other fronts will exhibit stronger turbulence and mixing than found here on them.


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Hales, B. D., Hebert, and J. Marra, Turbulent supply of nutrients, J. Geophys. Res., 95(C8), 15,087, England shelf break front.


Figure 1.  Colours indicate a) sea surface temperature (SST) from the Advanced Very High Resolution Radiometer on 31 July 2006, b) mixed layer chlorophyll concentration from Sea-viewing Wide Field-of-view Sensor (e-folding scale for light transmission is ~5 m based on our optical measurements), c) strain ($\alpha$) scaled by $f$ from the objective map at 24 m, and d) Rossby number from the objective map at 24 m. The ship track from 3-7 August 2006 shows the 13 cross-front sections, which were covered from north to south (black lines, Figures 1a, b, and d). $\sigma_\theta$ from the objective map at 24 m is contoured at 0.1 kg m$^{-3}$ intervals (e.g., 24.2 kg m$^{-3}$, 24.3 kg m$^{-3}$, and so on) with the thick line indicating 24.55 kg m$^{-3}$ and defining the location of the front, $x = 0$ km (grey lines, Figure 1d). Every fourth current vector from the objective map is plotted on the cruise track (Figure 1c). AVISO merged absolute dynamic height from 2 August 2006 is contoured every 4 cm (thin grey lines) above the low of 48 cm (thick grey line, Figure 1b).
Figure 2. An example along 36° 18′ N of a cross-front section shows the front is located near $x = 0$ km, which is the cross-front distance from the region of maximum gradient at $\sigma_\theta = 24.55$ kg m$^{-3}$ (thick black line) at a depth of 24 m. Other isopycnals (25, 25.5, 26, and 26.5 kg m$^{-3}$) are contoured (thin black lines). Individual plots show: a) potential temperature, b) salinity, c) spice gradient (potential temperature gradient on isopycnals but plotted on the isopycnals’ mean depths), d) eastward velocity, e) northward velocity, f) upward velocity from the $\omega$ equation, g) horizontal strain from the objective map scaled by the Coriolis frequency, h) Rossby number, i) reduced shear from low passed, mapped currents, j) microscale dissipation of temperature variance, k) microscale thermal diffusivity, and l) Turner angle on isopycnals but plotted on the isopycnals’ mean depths. Longitude is shown in Figure 2b) and with similar tick marks along the top axes of the individual plots. Isopycnal contours from the objective map are used in Figures 2f, g, and i.
Figure 3. An example cross-front section at $36^\circ 18'\text{N}$ with the depth means from 56–200 m of: a) spice gradient (cross-front potential temperature gradient on isopycnals), b) Rossby number, c) horizontal strain from the objective map scaled by the Coriolis frequency, d) eastward velocity, e) northward velocity, f) upward velocity from the $\omega$ equation, g) dissipation of temperature variance, h) thermal diffusivity, and i) reduced shear from low passed, mapped currents. The front is located at $x = 0$ km. Longitude is shown in Figure 3b and with similar tick marks along the top axes of the individual plots.
Figure 4. The composite front is a bin mean made from individual cross-front sections, such as Figure 2. The front is located at $x = 0$ km. Individual plots show: a) potential temperature, b) salinity, c) absolute value of spice gradient (the cross-front potential temperature gradient on isopycnals but plotted on the isopycnals’ mean depths), d) eastward velocity, e) northward velocity, f) upward velocity from the $\omega$ equation, g) horizontal strain from the objective map scaled by the Coriolis frequency, h) Rossby number from the objective map, i) reduced shear from low passed, mapped currents, j) number of SeaSoar data points in each bin, k) number of microstructure data points in each bin, l) Turner angle on isopycnals but plotted on the isopycnals’ mean depths, m) dissipation of temperature variance from all data, and n) thermal diffusivity from data with $|Tu| < 35^\circ$. Thick line is $\sigma_\theta = 24.55$ kg m$^{-3}$ and thin lines are 25, 25.5, 26, and 26.5 kg m$^{-3}$ from the objective map.
Figure 5. The composite front depth mean from 56–280 m of: a) Rossby number (black) and strain scaled by the Coriolis frequency (grey) with both from the objective map, b) absolute value of scaled vertical potential temperature gradient, c) absolute value of spice gradient (isopycnal cross-front temperature gradient), d) eastward (black) and northward (grey) velocities, e) vertical velocity from the $\omega$ equation, f) reduced shear from low passed, mapped currents, g) thermal diffusivity for all data (black) and doubly stable data (grey dashed), and h) dissipation of temperature variance for all data (black) and doubly stable data (grey dashed). i) Upper black line shows the total number of hydrographic data points in each distance bin, while the lower lines are for all microstructure data (black) and doubly stable data (dashed grey).
Figure 6. a) The downward turbulent heat flux, $-J_{qk}$, is largest (20–70 W m$^{-2}$) on the dense side of the composite front to depths of 168 m. The front is located at $x = 0$ km. The thick line is $\sigma_\theta = 24.55$ kg m$^{-3}$ and thin lines are 25, 25.5, 26, and 26.5 kg m$^{-3}$ from the objective map. b) The heat flux, $J_{qw}$, from the ASC is upward and restratifies the front. c) The depth mean turbulent heat flux, $\langle -J_{qk} \rangle_z$, is maximum within 20 km of the front on the dense side (red line), while the restratifying heat flux, $\langle J_{qw} \rangle_z$, is maximum on either side (black line). d) The horizontal mean turbulent heat flux, $\langle -J_{qk} \rangle_x$, is largest from 100-200 m (red line), while the restratifying flux peaks at 50 m (black line).
Figure 7. a) Nitrate (red) and phosphate (blue) concentrations are obtained from water samples at a nearby CTD station with sampling depths indicated by x’s. Potential density is in black. b) The horizontal mean of turbulent diffusivity, \( \langle K_T \rangle_x \), is maximum near 120 m, c) while nitrate (red) and phosphate (blue) fluxes show two peaks near 50 m (due to the nutrient gradient) and 120 m (due to the diffusivity maximum).
Figure 8. Chlorophyll fluorescence shows a deep maximum generally at depths of 50–100 m and mostly on the dense side of the front \( (x = -5 \text{--} 20 \text{ km}) \) at most cross-front sections. Only the 9 northernmost sections are plotted. Black contours are \( \sigma_\theta = 24.55, 25, 25.5, \) and \( 26 \text{ kg m}^{-3} \). Thick grey lines indicate the dense side of the front: \( x = 0 \text{ and 20 km at a depth of 50 m} \). The cruise track is shown (thin grey line with dots denoting the \( \sim 3 \text{ km} \) binning interval).
Figure 9. As Figure 8, but for beam attenuation coefficient, a measure of turbidity.