Physical description of an upwelling filament west of Cape St. Vincent in late October 2004

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[1] The present paper describes the physical characteristics of an upwelling filament off Cape St. Vincent (southern Iberia) in October 2004, based on remote sensing, hydrographic, ADCP and wind data. It was associated with a meandering jet similar to eastern Pacific or Canary filaments, though its dynamical features were weaker as a result of the surface-trapped (top 100 m) and weak horizontal density gradients. The filament transported 0.9 Sv of coastal water offshore, of which 8% was due to Ekman transport; it formed following the destabilization of an alongshore equatorward upwelling jet close to the cape. The filament was structurally asymmetric, with anticyclonic geostrophic vorticity to the north and more strongly cyclonic to the south, as a result of the meanders and existing eddies. The filament comprised a sequence of cold, submesoscale (~30 km diameter) cyclones. Vertical velocities of ±15 m·d−1 were associated with meanders of the jet. The surface and subsurface circulation was coupled. At middepths (90–150 m) anticyclonic recirculation of northern water seemed associated with a separated, older upwelling jet. Mediterranean Water (MW) was present below 350 dbar as a topographically steered undercurrent and a meddy-cyclone dipole. These undercurrents contributed to the formation of the October filament through baroclinic instability associated with the enhanced vertical shear.


1. Introduction

[2] The ocean circulation off the Western Iberian Peninsula (IP) is dominated by seasonal reversal of wind driven mesoscale flows [Sánchez and Relvas, 2003; Peliz et al., 2005]. The temporally variable wind field [Sánchez et al., 2007] interacts with pre-existing alongshore pressure gradients [Relvas and Barton, 2002] and topography [Serra and Ambar, 2002] to produce complex circulation patterns.

[3] Upwelling-favorable winds occur from May through September [Sánchez and Relvas, 2003]. In response, a cold upwelling jet extends over the upper 100 m or more of the water column, advecting ~1 Sv (1 Sv = 1 × 106 m3·s−1) of low-salinity water equatorward.

[4] Upwelling filaments are cold-water tongues with their source in the upwelling zone that may reach some ~50 km of width and up to 300 km of length. They have been intensively studied in the west US coast since the early 1980s [Flament et al., 1985; Kosro and Huyer, 1986; Rienecker et al., 1985; Brink and Cowles, 1991; Ramp et al., 1991; Flament and Armi, 2000]. These studies show narrow baroclinic jets that extend offshore transporting a few Sv of cold surface water, bounded by warm and salty waters on the cyclonic (southeastern side) and warm and fresher water on the anticyclonic (northwestern side) side. A characteristic of the horizontal circulation is the shoreward, return flow on the cyclonic side. The recurrent activity of upwelling filaments is a major contributor to cross-shelf exchange [Sánchez and Relvas, 2003]. Although filaments appear in response to upwelling-favorable winds, instabilities of the coastal jet and interaction with the offshore field trigger their formation [Strub et al., 1991]. Under this assumption cold filaments form part of a continuous, meandering, southward jet.

[5] The scarcity of in situ measurements of Iberian upwelling filaments lead to controversy in explanations of their formation, maturation and decay. Satellite imagery studies [Haynes et al., 1993] observed that these preferentially root to coastline irregularities and abrupt bathymetric changes. The model of Batteen et al. [1992] attributed their development to wind stress curl effects, which are enhanced close to Cape St. Vincent (CSV). However, Roed and Shi [1999] suggest that they may form even along a straight coastline. It seems that a single mechanism cannot be held exclusively responsible for their formation.

[6] Satellite imagery reveals that filaments start to form in May–June and become common around July, when they have a mean length of 80 km [Haynes et al., 1993; Relvas, 1999]. As upwelling continues these may reach a maximum
length of 250 km in late August. Their evolution consists of periods of active growth interrupted by relaxation episodes.

[7] To date in situ observations of filaments in NE Atlantic are limited to NW Iberia [Barton et al., 2001] and a number of events in the Canary region [Barton et al., 2004]. In the first case the offshore transport was estimated at just 0.5 Sv in a thin surface layer, with a remnant subsurface return flow below [Barton et al., 2001]. The Canary observations [Barton et al., 2004] comprise a number of filaments with speeds exceeding 0.3 m s\(^{-1}\) over the upper 200 m close to Cape Juby. These appeared to be related to the presence of a quasi-permanent eddy located over the bathymetric trough downstream of the shallower Fuerteventura/Africa channel.

[8] Filaments cause a much larger horizontal cross-shelf mass transport than expected by the wind-driven Ekman forcing [Kosro and Huyer, 1986]. Additionally, vertical circulation has been inferred by observations and numerical modeling in the northeast Pacific [Brink and Cowles, 1991]. Subduction velocities of about 20 m d\(^{-1}\) along the offshore path of the upwelling filament are common [Washburn et al., 1991]. Dewey et al. [1991] reveals complex secondary circulation patterns related to convergences and divergences of the large Rossby number flow.

[9] This paper presents the physical description of an upwelling filament off Cape St. Vincent in late October 2004. These are the first in situ observations of such an structure in the southern Iberian region. Filament occurrence is rare at this time of the year [Haynes et al., 1993] but 2004 was characterized by persistent upwelling-favorable winds almost continuously since May, intensified during September and October.

[10] The objectives of this paper are to: (1) present an extensive physical description of the Cape St. Vincent filament, (2) evaluate its volume transport and the dynamical features, (3) evaluate its coupling with the eddy field and subsurface flows, and (4) investigate the mechanisms of vertical forcing within the filament. The paper is structured as follows: section 2 describes the data used in the study and its processing; section 3 contains a detailed analysis of in situ data acquired off Cape St. Vincent; section 4 discusses the results and compares the Cape St. Vincent filament with those on the east Pacific coast; finally, the conclusions are given in section 5.

2. Data and Methods: The October 2004 Observations

[11] Satellite data from the NOAA Advanced Very High Resolution Radiometer (AVHRR) were processed to derive sea-surface temperature (SST) for the Iberian Peninsula, using standard NOAA procedures except for a hybrid cloud-masking algorithm and low-noise SST equation [Miller et al., 1997]. NASA Sea-viewing Wide Field-of-view Sensor (SeaWiFS) data were processed into chlorophyll-a estimates using the OC4v4 algorithm and an automated system [Lavender and Groom, 1999]. SST maps showed a well-defined filament at Cape St. Vincent on 20 October 2004 (Figure 1). From 22 to 25 October 2004 (about 63 h) this filament was surveyed on board the R/V D Carlos I (ATOMS04 cruise, Figure 1D). The sampling was guided by satellite SST imagery transmitted to the ship in near-real time.

[12] A total of 42 CTD casts were distributed on an almost regular grid of 15 km mean spacing, with reduced spacing close to the fronts; this allowed for resolution of \(\sim 30\) km features (Figure 1D). We expected the baroclinic radius of deformation to be \(R_d \sim 30\) km, and full wavelength structures on the order of \(\pi R_d\). (\(R_d = f^{-1} \sqrt{g' h}\), where \(g' = g(\Delta \rho)\) the reduced gravity, \(h\) the column depth and \(f\) the Coriolis parameter.) This was inferred using climatological values from the surface to 400 m (\(\Delta \rho = 2.22\) kg m\(^{-3}\) [Sánchez and Relvas, 2003]). Considering the top 200 (100) m of the water column, expected \(R_d\) was \(\sim 22\) (20.5) km (\(\Delta \rho = 2.01\) (1.88) kg m\(^{-3}\)). Stations were sampled from the surface to a maximum depth of 400 dbar using an Idronaut OS316 CTD probe attached to a Rosette sampler. The CTD sampling rate was 20 Hz and the lowering speed was 1 m s\(^{-1}\).

[13] Although we consider here the data to be synoptic, the timescales associated with mesoscale features are not much longer than the sampling period. From 20 October the filament experienced first a northward push (Figures 1A–1D) after which an upwelling pulse occurred (Figures 1B–1C). This source of variability may handicap the retrieval of observed fields, and particularly compromise the determination of vertical velocities by the Omega Equation. Although precise determination of synopticity errors is largely case-to-case dependent, the experiment of Gomis et al. [2005] revealed that an along-front sampling results in errors of about 15% for dynamic height and 50% for relative vorticity and vertical velocity. Considering that our sampling strategy (preferentially across-front) produce smaller synopticity errors, and accounting for typical observation errors and sampling limitations (about 6% for dynamic height and between 15 and 30% for geostrophic vorticity and vertical velocity), the qualitative identification of the sign and magnitude of mesoscale features should be guaranteed [Gomis et al., 2005].

[14] Current speeds were measured along the ship’s track with a hull-mounted 38 kHz RDI acoustic Doppler current profiler (ADCP). The moderate steaming velocity (mean \(\sim 8\) knots) allowed a relatively good retrieval of returned echoes over the upper 700 m of the water column and permitted bottom tracking over depths shallower than 1500 m. Returned echoes were averaged in 5 min ensembles with a vertical bin size of 16 m. Data postprocessing was done with the Common Oceanographic Data Access System (CODAS) [Firing et al., 1995], then ADCP data were averaged over 2.5 km along track, and 20 m in depth, starting at 30 m down to 700 m. Contamination of the ADCP data by tidal currents was evaluated with local tidal solutions obtained with a regional configuration of the Egbert et al. [1994] OSU Tidal Inversion model (OTIS). Barotropic tidal currents in the ADCP locations at the acquisition times were about 0.12 m s\(^{-1}\) in a narrow band around Cape St. Vincent, remaining less than 0.04 m s\(^{-1}\) elsewhere. ADCP velocities were detided by simply removing the model currents.

[15] In order to calculate high-order spatial derivatives of the hydrographic fields, a 2D univariate spatial Optimal Interpolation scheme was used (OI, Bretherton et al. [1976]). In this analysis weight functions depend on the relative location of stations, and the output field is smoothed.
according to a given measure of observational errors (signal-to-noise ratio, SNR). The statistical mean was modeled by fitting a second-order polynomial function to the observations. Correlations between increment values were assumed to follow a Gaussian function of the type

\[ C_{ij} = \exp \left( -\frac{d_{ij}^2}{2L^2} \right) \]

where \( d_{ij} \) is the separation distance between locations and \( L \) the characteristic scale, related to the scale of dominant structures. \( L \) was set to 20 km based on sample lag-correlations using data from Sánchez and Relvas [2003] climatology. To allow for smoothing of the analyzed fields the SNR was conservatively set to 0.01 for CTD and 0.1 for ADCP data. The fields were further smoothed with an analytical filtering [Pedder, 1993] centered at a wavelength of 27 km. The result produced a number of 3D matrices for the CTD and ADCP data with 5 km of grid point spacing. Further gridding details are described by Sánchez [2005].

[16] Dynamic height was calculated relative to 400 dbar, while for shallower stations a linear extrapolation method was applied [Reid and Mantyla, 1976]. ADCP measurements revealed that 400 dbar was not a motionless reference level (e.g., Figure 2). A nondivergent stream function fitted to the gridded ADCP data at 400 m provided the reference field that was combined with CTD data to render the absolute dynamic height, as follows: If we assume Helmholtz theorem, we can decompose the horizontal flow into divergent and irrotational parts

\[ u_h = k \times \nabla \psi - \nabla \phi, \]

with vertical unit vector \( k \), nondivergent stream function \( \psi \) and potential \( \phi \). Taking the curl of this equation, we get the Laplace equation

\[ \nabla^2 \psi = k(\nabla \times u_h), \]

solved by applying an integral constraint for mass conservation at the boundaries defined by the 0.1 error covariance contour of the OI of ADCP data. \( \psi \) was scaled by \( f \) to make it equivalent to dynamic height (Figure 2). The root-mean-square (rms) difference between raw ADCP vectors and nondivergent velocity field was 0.074 m s\(^{-1}\) (0.058 m s\(^{-1}\) for the zonal...
and 0.045 m s\(^{-1}\) for the meridional component), whereas the RMS difference between the gridded ADCP and the nondivergent velocities was 0.049 m s\(^{-1}\) (0.041 m s\(^{-1}\) for the zonal and 0.031 m s\(^{-1}\) for the meridional component).

3. Results

3.1. Winds

[17] Northerlies were strong and persistent off the southwestern IP from September through October 2004, with two major periods of relaxation: one between 8–10 October and another between 17–27 October during which the cruise occurred (Figure 3, top). The ship’s meteorological station data revealed smaller-scale temporal variability (Figure 3, bottom). The first half of the cruise took place under weak southwesterly winds, while toward the end of 23 October the northerlies peaked again. The Ekman transport (\(T_E\)) across 9.50\(^\circ\)W (total length \(\approx 80\) km) using peak upwell-favorable wind velocities prior to the cruise (7.5 m s\(^{-1}\), 11–16 October) was about 0.07 Sv. (\(T_E = \frac{\rho_a}{\rho_w} \Delta \tau v_w\)) with distance \(\Delta\), mean density \(\rho_w\), and wind stress \(\tau\), given by \(\tau = \rho_a C_D |v_w| v_w\) with air density \(\rho_a\), drag coefficient \(C_D\) and wind speed \(v_w\), e.g., Ramp et al. [1991]).

3.2. Sea Level Anomaly Maps

[18] Sea level anomaly (SLA) contours for 12–22 October revealed a number of mesoscale structures (Figure 4). On 15 October a mesoscale anticyclone was situated at \(\approx 10.50^\circ\)W 38.50\(^\circ\)N (A1, SLA >9 cm in Figure 4C). On 19 October a smaller anticyclone was sitting south of Cape St. Vincent (A2, SLA >8 cm). In between, a cyclone occurred (C, SLA <3 cm).

[19] Figure 4D shows interaction of the coastal circulation with the offshore eddy field during the cruise (22 October). Cyclonic circulation occurred over the shelf around the cape, perhaps as a consequence of the pulse of northerlies in the second week of October (Figure 3). In addition C seems to have divided into two smaller cyclones C\(_1\) and C\(_2\) although it could be an artifact of the mapping. The cyclonic zone over the shelf seems to stretch from the coast toward C\(_2\), forming a joint structure similar to an upwelling filament.

3.3. Satellite SST Images

[20] SST imagery showed an interrupted coastal upwelling band, in response to the equatorward winds (Figure 1). Many of the features detected by altimetry left an imprint on SST imagery. A2 was seen in Figure 1C as a gentle bending of the thermal front south of the cape. A northern branch of A1 may have drawn upwelled water from Cape Espichel seawards to form an upwelling filament (Figure 1B). Another westward filament was noted at Cape St. Vincent on 20–22 October. Its length was about 167 km of which \(\approx 77\) km corresponded to a curling segment, and the width was \(\approx 45\) km at \(\approx 9.50^\circ\)W. Its SST signature was <18.5°C, contrasting with the ambient waters >20.5°C. Maximum temperature gradients were slightly stronger at the southern boundary (>0.25°C km\(^{-1}\)) than at the northern boundary.
The southern part of the filament appeared as a wavy feature, evidencing meanders of the front. The amplitude of these waves increased with time, which could be indicative of current shears. This filament grew during the intensification of northerlies between 13–16 October (Figure 3). Later on, its offshore segment curled anticyclonically, suggesting a northward push from the cyclone $C_2$ (Figure 4D).

### 3.4. Water Masses

Figure 5 presents the T-S plot of all CTD stations. Upper layer properties ranged 36.05–36.4 in salinity and 14–20°C in temperature, which were greatly influenced by coastal upwelling. Lower salinities (<36.25 at 10 m) corresponded to coastal stations and those situated below the influence of the coastal upwelling. Fewer stations were situated outside the upwelling influence (Figure 5), and exhibited relatively warm and saline types compatible with eastern North Atlantic Central Water of subtropical origin (eNACWt), typical of the Gulf of Cádiz [e.g., Relvas, 1999]. Below 14°C waters followed the linear T-S properties of the underlying eNACWt, although a slight departure from the line was observed below 13.5°C, attributed to the increase of subsurface salinities through turbulent detrainment of the Mediterranean Water (MW). A number of other stations showed MW influence as sharp subsurface salinity increases below 300 m (Figure 5 inset plotted as gray shades). If these are considered as MW tracers, then this reveals the shallow (<400 m) spreading of MW.

### 3.5. Thermohaline Fields

As indicated by satellite SST, the upper-layer plots show the influence of coastal upwelling (Figure 6, from 50–150 dbar). At 50 dbar coastal waters appeared separated from the open ocean by a relatively intense density front of ~1 $\sigma_T$ (Figure 6C, at 50 dbar). A patchy band extended from the coast into the open ocean forming a filament. The density anomaly front delineated this structure, with similar gradients at the north and at south boundaries. Salinity and temperature contours also illustrated the coastal filament penetrating the deep ocean west of 10°W with temperature <16.9°C and salinity <36.2 (Figures 6A–6B, at 50 dbar). The offshore extension of the filament curled poleward, suggesting anticyclonic bending around a warm, saline water pocket centered at 9.50°W 37.35°N. The sequence of patches forming the upwelling filament caused meanders along the northern and southern filament fronts. The southern front showed a wavy pattern that could also be observed in the satellite imagery (Figure 1D).

At 150 dbar the density front was 6 times weaker than at 50 db, with maximum differences of ~0.15 $\sigma_T$ between the shelf and offshore waters (Figure 6C, at 150 dbar). The southern and northern fronts were broken at about 9.50° and 10°W by southward extensions of the northern warm pool. Consequently the filament appeared partially divided, giving rise to a number of cells (Figure 6, at 150 dbar). As a result the isopycnals crossing the bathymetry bent sharply southward leaving a high-density (cool and fresh) structure on the coastward side, with temperature <14°C and salinity <36.05. Two offshore cells were less fresh and slightly warmer than the inshore one, and appeared as remnants of older upwelling pulses.

The chain of cold and fresh cells were still detectable at 250 db, though with weaker horizontal density gradients (<0.10 around the cape, Figure 6C, at 250 dbar). More saline (>36) and warmer (>14.15°C) waters occurred at the north and the south of the sampling area (Figures 6A–6B).

Although diffuse, the cold cells evident could still be perceived at 340 db, particularly north of St. Vincent Canyon (Figure 6, at 340 dbar). Offshore, weaker nuclei
indicated that this was the lower boundary of the coastal upwelling. At this depth the effect of MW was seen as a warm, saline wedge at the southwestern corner of the sampling area. The relatively high temperatures compensated for the large salinities, making this wedge more buoyant than the surroundings and hence formed a sharp density front. Immediately south of CSV a warm and particularly saline intrusion, best seen in the salinity plot, followed the shelf-break (Figure 6B, at 340 dbar). Finally, a well defined warm and saline spot was seen at 9.75°C/37.20°N. Individual T-S profiles suggested the presence of MW at these locations (Figure 5), and hence the coupling of Mediterranean outflow with the upwelling circulation. The spatial discontinuities between these saline structures, as well as their dissimilar density characteristics suggested that they had different dynamical origins. The inshore one was largely topographically steered, likely related to the upper core of the Mediterranean undercurrent around Cape St. Vincent. The offshore ones were probably associated with the eddy field generated by the Mediterranean undercurrent, as will be discussed below.

3.6. Geostrophic Transport and Flow in the CSV Region

[26] As inferred from the SLA map of 22 October (Figure 4D) the surface circulation was characterized by an upwelling flow interacting with the eddy field, although ADCP revealed a more complex circulatory pattern populated with small-scale meanders and recirculation cells (Figure 7A). Although the overall pattern is coherent, a mismatch between
the ADCP and the geostrophic flow at upper levels was noted at some locations, specially along the 37.30°N line. This part of the cruise was conducted after the upwelling-favorable wind event of 24 October, which may have added inertial noise and divergent flows that were filtered in the geostrophic field. A coherent flow characterized the northern half of the sampling area, whereas a more diffuse pattern occurred to the south. An equatorward jet experienced a sharp anticyclonic meander shortly before St. Vincent canyon north of 37.20°N, as inferred from the thermohaline fields: this meander could be related to the southernmost tip of A1 (Figure 4D). Part of the geostrophic transport deflected southwards at 9.40°W 37.20°N. As the density front strengthened across the northwestern front (Figure 6C, top) flow intensification occurred and a net speed gain within the jet was observed (>0.35 m s⁻¹). Around this site geopotential anomaly differences greater than 8 dyn cm were observed (not shown), corresponding to maximum offshore transport of 0.9 Sv (Figure 7A). This offshore branch experienced a sharp anticyclonic bending, that dissected two cyclonic cells. These were apparently related to the equatorward recirculation of the upwelling filament. The inshore one was confined east of 9.50°W and north of 37°N, while the offshore one occupied the southwestern corner of the sampling area, possibly linked to C2 (Figure 4C–4D). The geostrophic transport associated with this onshore recirculation amounted to 0.4 Sv across 9.75°W. This asymmetry between the offshore and onshore components of the transport is explained by the splitting of the jet west of 9.50°W. Part of the flow leaks anticyclonically, part recirculates cyclonically onshore. Part of this return flow experienced an anticyclonic turn close to CSV, with convergence of oceanic water toward the southern filament border. This feature spatially coincided with A2 (Figure 4C).
Figure 6. Thermohaline properties at selected pressure levels. (A) temperature (°C), (B) salinity, and (C) density $\sigma_o$. The 0.1 error covariance contour of the OI has been used to blank areas with too large interpolation errors. The 50, 100, 500, and 1000 m bathymetric contours are also shown.
At 120 dbar the upwelling jet meandered over a larger area (Figure 7B). Dynamic height gradients across the northeastern front weakened with increasing depth: not exceeding 5 dyn cm compared to >7 dyn cm at the western front. This revealed the baroclinic nature of the equatorward jet and the relative barotropicity of the poleward portion of the anticyclonic meander, suggesting that each was associated with a different process. The signatures of the recirculating cyclones disappeared at this level, believed to be the last of the return flow associated with the filament. In their place a cyclonic meander was seen flowing polewards west of 9.60°W and south of 37°N, which merged with the poleward deflection of the anticyclonic return at 9.90°W.

Figure 7. (A) Volume transport function at 5 dbar (units are Sv, or 10^6 m^3 s^-1) and along-track ADCP velocity vectors averaged over 0–40 m. (B) Absolute dynamic topography at 120 dbar (units dyn cm) and ADCP vectors averaged over 110–170 m; (C) same as Figure 7B but at 180 dbar (DH) and 160–220 m (ADCP); (D) same as B but at 340 dbar (DH) and 320–350 m (ADCP). The 0.1 error covariance contour of the OI has been used to blank areas with too large interpolation errors. The 50, 100, 500 and 1000 bathymetric contours are also shown.
37.10°N. The small anticyclone south of St. Vincent Canyon was present at this level.

[28] At 180 dbar both the hydrographic and ADCP data portrayed anticyclonic meandering of the coastal jet north of 37.50°N (Figure 7C). As observed in upper layers, part of this flow was seen to divert southwards in a sharp cyclonic shift west of 9.30°W, to later turn anticyclonically and join the larger meander. Compression of isolines occurred in the southwest, partly because of this feature, and partly because of the northward push from the southern cyclone. Accordingly, ADCP currents followed the dynamic height field and reached 0.20 m s⁻¹. Additionally, ADCP vectors south of the cape followed the bathymetry and suggested a smaller anticyclone anchored to the topography. Some velocity fine structure was seen over the canyon, not observed in the dynamic topography chart.

[29] At 340 dbar there was no evidence of the upwelling circulation and the flow was predominantly polewards. The anticyclone south of the canyon appeared to be replaced by the Mediterranean undercurrent following the shelf-break (Figure 6, at 340 dbar). Velocities at the southern cyclonic meander exceeded 0.20 m s⁻¹ at this level.

[30] In summary, the dynamic field showed a large anticyclonic meander and northward recirculation of 0.9 Sv corresponding to the upwelling jet. In upper layers this meander was populated with a number of smaller scale meanders associated with cyclonic cells on the southern side of the front. These cells were observed to convey approximately 0.4 Sv of the return flow related to the offshore segment of the upwelling filament. Of these cyclonic cells, only the inshore one showed consistency and weakly extended below 100 dbar. Underneath the offshore cyclone was a strong poleward (and also cyclonic) flow with velocities greater than 0.25 m s⁻¹. The Mediterranean undercurrent was observed to follow the bathymetry, though its signal (either hydrographical or kinetical) was lost close to St. Vincent Canyon, which would suggest local stretching of the flow. Local velocity fine structure occurred close to the canyon.

3.7. Vertical Sections

[31] Vertical cross-sections paint a more complex picture than the simple offshore projection of coastal upwelling (Figure 8 and Figure 9). Water with salinity <36 and temperature <15° was upwelled from below 120 dbar (offshore) onto the surface (near the shore, Figure 8). Isopycnals were elevated toward the coast and over the cyclonic cells spanning the upwelling filament. The elevation was noted from approximately 250 dbar, although with more intensity over the surface layer (0–100 dbar), which revealed the surface-trapped nature of the upwelling. The surface layer thickened to the north, where the warm pool was situated.

[32] Along the zonal section at 37.50°N, the N-S ADCP velocity revealed a thick surface equatorward flow (Figure 8A, top). The temperature and salinity plots discriminated two distinct cores within this flow. First, close to the coast the isotherms and isohalines intersected the surface at 9.20°W indicating the limit of the upwelling band (Figures 8B–8C, top). The maximum slope of the density anomaly field occurred above 70 dbar (Figure 8D, top), representing the upper core of the narrow equatorward jet that advected cold, low-salinity upwelled water with velocities greater than 0.20 m s⁻¹. The lower core was centered at 9.35°W, ~120 dbar, with maximum breadth at 70 dbar, progressively narrowing with depth until it disappeared below 250 dbar. This core conveyed warmer and more saline water than the upper jet and so may be associated with an upwelling jet formed before the downwelling-favorable episode of 17–23 October 2004 (Figure 3). The boundary between both flows was defined by the buoyancy frequency maximum at 70 dbar at 9.30° (Figure 8D, top).

[33] A weak poleward flow occurred close to the slope between 100–250 dbar (0.06 m s⁻¹ at 9.15°W, Figure 8A, top), comparable to other undercurrents observed in many coastal upwelling regions [e.g., Smith, 1995]. It was distinct from the topographically steered Mediterranean vein which featured a sharp bending of isolines and isotherms toward the bottom below 350 dbar (Figures 8B–8C, top), with poleward velocities of 0.04 m s⁻¹.

[34] A poleward undercurrent was also observed along the western border, between 100–350 dbar at 9.70°W. Instrument failure prevented acquisition of data out to 10°W, although the thermohaline pattern suggested that the poleward flow extended to the surface in the west (Figure 8A, top). As with the equatorward flow, the poleward transport appeared dual-cored, separated by a velocity trough. The doming of isotherms and isohalines close to 10°W could indicate a surface core (Figure 8C, top) featuring anticyclonic recirculation of the poleward underflow, a hypothesis that seems corroborated by the dynamic topography at 120 m (Figure 7B).

[35] At 37.20°N (Figure 8, bottom) the upwelling front was located at 9.40°W due to the offshore meandering of the upper jet (see Figure 7A). East of 9.50°W two features were observed. Inshore, alternating bands of negative-positive flow conveyed cold, low-salinity water (Figure 8, bottom). Isopycnals associated with these outcropped to the surface and buoyancy frequency maxima shoaled (Figure 8D, bottom). This flow resulted from the zonal alignment and small-scale meandering of the upwelling jet by the offshore shift across this section (Figure 7A). As observed in the previous transect, the upwelling jet appeared to ride a subsurface equatorward jet at 9.40°W. Sharp deepening of isopycnals produced a deeper buoyancy maximum at this spot (Figure 8D, bottom). The westward half of this transect featured an intense surface-trapped poleward current (0–110 dbar, >0.30 m s⁻¹) carrying cold, low salinity upwelled waters.

[36] At middepths (150–300 dbar) at 9.50°W a narrow, subsurface jet appeared to be related to the anticyclonic meander seen farther north (Figure 7C). West of 9.80°W under 200 dbar a rather barotropic flow with velocities >0.14 m s⁻¹ carried warmer and more saline water polewards, as inferred by the downward bending of isolines (Figures 8B–8C, bottom).

[37] Meridional sections showed a band of uplifted isotherms and isohalines intersecting the surface at 37°N (Figure 9). The zonal transport by the filament past the sharp meander was captured by the section at 9.55°W (Figure 9, top). The surface-trapped nature (0–100 dbar) of the cross-shelf flow was noted by the ADCP velocities
and in the density anomaly (Figure 9, top). The northern side of the transect was occupied by warm surface water that approached the coast with velocities exceeding 0.12 m·s⁻¹ (Figures 9B–9C, top).

[38] The vertical structure of the filament was discovered toward the south of the transect. A strong offshore flow carrying cold, low salinity water extended from the surface down to 80 dbar with velocities >0.36 m·s⁻¹. (Figure 9, top).
This was accompanied by an onshore return flow with velocities $>0.30 \text{ m s}^{-1}$ that closed a cyclonic loop west of $9.55^\circ \text{W}$. The same structure was seen across $9.75^\circ \text{W}$, some $20 \text{ km}$ farther offshore (Figure 9A, bottom). However, anticyclonic deflection of the westward flow caused the poleward curling of the filament, which resulted in strong deceleration of the zonal flow between the two transects.

Figure 9. Meridional sections of (A) zonal ADCP velocity (cm s$^{-1}$), (B) temperature ($^\circ \text{C}$), (C) salinity, (D) $\sigma_b$. Buoyancy frequency (cpd) shades are also overlaid, with colorbar included in Figure 9D. A chart with the transect location and the absolute dynamic topography contours at 10 dbar is included for visual reference for each plot row. Top: along $9.55^\circ \text{W}$; bottom: along $9.75^\circ \text{W}$. 
[39] A complex structure was noted below 100 dbar. Under the filament a trough in the thermohaline fields produced a saddle surface (Figure 9, top) from 110 to 350 dbar. As it was not observed shoreward (see the horizontal maps in Figure 6) or offshore (Figure 9, bottom) we suggest that this feature resulted after southward advection by small scale meandering at ~100 dbar of the northern anticyclone, as it was confirmed by the ADCP velocity around this warm trough.

[40] Under the filament northern flank (at 130 dbar), a vigorous flow >0.28 m s⁻¹ brought relatively saline, warm water westward (Figure 9A, top and bottom). The poleward recirculation indicated that this was the subsurface equatorward jet. Zonal velocities accompanying this feature suffered from strong deceleration between the two transects due to anticyclonic deflection of the flow. The deep flow south of 37.1°N was the westward component of the rather barotropic cyclone described before. It extended from 150 dbar down to at least 700 db (not shown) with velocities >0.28 m s⁻¹. Its thermohaline structure was associated with relatively warm and saline waters, as seen by the downward bending of isotherms and isolahnes south of 36.9°N.

[41] A schematic view of the circulation is presented in Figure 10. Over the top 100 m an energetic, baroclinic, meandering jet carried upwelled water equatorwards with velocities >0.30 m s⁻¹. Its thermohaline structure was associated with relatively warm and saline waters, as seen by the downward bending of isotherms and isolohnes south of 36.90°N.

[42] Geostrophic Vorticity

[43] Meandering of the surface flow was quantified by the relative vorticity calculated from the absolute geostrophic velocity (Figure 11):

\[ \zeta_g = \frac{\partial v_g}{\partial x} - \frac{\partial u_g}{\partial y}. \]

The location of the \( \zeta_g \) patches at 5 dbar placed along small-scale curvatures of the front, although shear vorticity occurred specially along the offshore branch of the jet (Figure 11A). Relative vorticity ranged from ~0.5f to +0.6f at sites well away from the boundaries. The filament featured a patchy cyclonic zone dissected by a number of anticyclonic intrusions. A strong cyclonic patch was situated at the root of the upwelling filament with values >+0.6f. Another cyclonic center with \( \zeta_g > +0.4f \) was observed at the offshore fragment of the filament. Anticyclonic intrusions exhibited smaller absolute values. Most of these were associated with the larger-scale anticyclonic circulation described in previous paragraphs.

[44] The weakening of horizontal density gradients with depth caused a reduction in geostrophic vorticity at 120 dbar (Figure 11B). Although the picture was similar to that at 5 dbar, the cyclones were barely visible and maximum \( \zeta_g \) values did not exceed +0.2f. The anticyclonic signals north of 37.20°N bore maximum \( \zeta_g < 0.3f \). This confirmed the highly baroclinic nature of the meandering flow associated with the filament.

3.9. Vertical Forcing

[45] The Rossby number given by the ratio of geostrophic relative vorticity to \( f \) was large near the surface, with maximum values > 0.6. The nonlinear part of the flow may be nonnegligible everywhere, and departures from geostrophy may exist. If we use quasi-geostrophic dynamics (QG), the 3D flow field can be approximately determined by the isobaric distribution of geopotential [Holton, 1992, p.150]. Under this theory, the omega equation supplies a valuable tool for the diagnosis of the vertical motion:

\[ N^2 \nabla^2 w + f^2 \frac{\partial^2 w}{\partial z^2} - f \frac{\partial}{\partial z} (U_g \cdot \zeta_g) + \frac{g}{\rho_0} \nabla^2 (U_g \cdot \nabla \rho) = 0 \]  

with buoyancy frequency \( N^2 \), Laplacian operator \( \nabla^2 \), vertical velocity \( w \), geostrophic velocity \( U_g = (u_g, v_g) \), reference density \( \rho_0 \), and density profile \( \rho \). The right-hand side (RHS) of equation (1) represents the vertical velocity forcing term. The vertical velocity \( w \), ageostrophic) is related to the horizontal velocity \( U_g \), geostrophic) and density fields. This vertical velocity is needed to maintain the thermal wind balance in the presence of variations in the horizontal advects of vorticity and density.

[46] Since equation (1) is a Poisson type equation, assuming that \( w \) is sinusoidal in the 3D space, if the sum of the RHS terms is negative, \( w \) is positive [Holton, 1992]. In mesoscale ocean motions \( \frac{\partial}{\partial z} (U_g \cdot \zeta_g) \) is dominant in the RHS of equation (1) [Tintore et al., 1991]. In these cases, if the geostrophic vorticity (and its advection) decreases with depth, the dominant term can be approximated by \( -U_g \cdot \nabla h \zeta_g \) (advection of geostrophic vorticity) [Tintore et al., 1991]. Figure 12A depicts this term at 5 dbar. (At 120 dbar \( -U_g \cdot \nabla h \zeta_g \) was barely 10% of that at 5 dbar, which allows this approximation). Positive and negative advection of geostrophic vorticity was noted along the meandering axis of the filament. As water parcels traveling with the geostrophic flow experienced the sequence of cyclonic and anticyclonic turns, they showed sharp advection of geostrophic vorticity. According to equation (1) vertical motion was inferred along a wavellike pattern of the surface flow.

[47] Mutual cancellation of both forcing terms on the \( f \)-plane hinder the quantitative diagnosis of \( w \) in equation (1). Rather we can use an alternative expression in terms of the spatial gradient of geostrophic velocities [Hoskins et al., 1978]

\[ \nabla^2 (N^2 w) + f^2 \frac{\partial^2 w}{\partial z^2} = 2 \nabla \cdot \mathbf{Q}. \]

In this equation the forcing terms in equation (1) are substituted by twice the divergence of \( \mathbf{Q} \), defined as:

\[ \mathbf{Q} = (Q_x, Q_y) = \frac{g}{\rho_0} \left( \frac{\partial U_g}{\partial x} \nabla \rho \cdot \frac{\partial U_g}{\partial y} \nabla \rho \right). \]
Convergence of $Q$ indicates regions where upward motion is taking place [e.g., Holton, 1992]. This equation is an elliptic partial differential equation that has a unique solution if suitable boundary conditions are given on all sides of the domain. We have solved the inversion of equation (2) using an iterative relaxation scheme as by Pinot et al. [1996]. The boundary condition is prescribed as $w = 0$ at the surface, bottom and lateral boundaries (Dirichlet). Changing the boundary from Dirichlet to Neumann (the gradient of $w$ is set to zero instead) implied variations in the solution less than 1%. The forcing term $2\nabla \cdot Q$ was assumed zero outside the domain. Results close to the boundaries are questionable, although the elliptic nature of the omega equation prevents the solution within the observed region from being affected by a lack of knowledge of the forcing further away.

The resulting vertical velocity field at 40 m is presented in Figure 12B. The spatial pattern showed length scales similar to the relative vorticity, suggesting the main role of the geostrophic vorticity advection in equation (1). At this level vertical velocities ranged ±15 m d$^{-1}$. Patches of alternating positive and negative $w$ of ~30 km in size were located in association with small-scale curvatures of the geostrophic flow.

The SeaWiFS chlorophyll-$a$ map on 23 October 2004 is plotted in Figure 12C. The filament could be observed as a discontinuous elongation of the coastal upwelling. It featured alternating patches of high and low chlorophyll-$a$, with a major discontinuity at 09.65$^\circ$N. Remarkably, patches of high chlorophyll-$a$ concentration approximately coincided with zones of upward vertical velocity as inferred from equation (2) (plotted as circles in Figure 10).
This improved our confidence in the estimation of the vertical forcing. The vertical velocity around the filament (along the 56.7 dyn. cm. streamline) is plotted in Figure 12D. If we assume QG dynamics, a water parcel following the line experiences a sequence of vertical displacements as it turns around the filament. The vertical distribution of $w$ appeared coherent from the surface to the bottom of the sampled layer, although maxima occurred between 30–70 dbar, fading to zero at 200 dbar and close to the surface. Maximum upwelling velocities are associated with regions of large vertical shear [see Figure 7 and Sánchez, 2005]. If not balanced by stratification, flows with strong velocity shear may be baroclinically unstable [e.g., Sánchez and Gil, 2004], which may be found in the northern meandering branch of the upwelling front.

4. Discussion

4.1. Cross-Shelf Transport

In section 3.1 it was shown that $T_E$ accompanying the filament was at most 0.07 Sv, which accounts for ~8% of the total transport (0.9 Sv) observed within the seaward jet. This equatorward transport was partitioned into two jets (Figure 8A, top). A narrow upper one (top 100 m) was related to a recently formed upwelling jet and advected approximately 0.4 Sv of cold, low-salinity upwelled water. A deeper one at ~120 dbar could be discriminated from the former by its warmer and more saline signature.

For the wind to be exclusively responsible for the generation of a 0.9 Sv current, a 8.5 m s$^{-1}$ wind over the whole western Iberian coast (800 km) would be required. Part of this $T_E$ may be lost by filaments to the north of Cape St. Vincent (e.g., Figure 1B). Considering the local coastal segment from Cape Espichel to Cape St. Vincent alone, a 12 m s$^{-1}$ wind (for example during the extreme upwelling-favorable event of 13–20 September 2004) could explain a $T_E$ of 0.4 Sv, equivalent to the transport in the upper jet. These results suggest that entrainment by local $T_E$ may account only for a marginal part of the equatorward transport. Rather, southward advection of northern waters and local eddy mixing seems to be essential to feed the filament in the CSV region.

Considering the present observation to be a typical CSV filament, and accounting for filament duration (present 100–200 d per year, with mean filament duration of 2–3 months [Haynes et al., 1993]) the associated annual mean cross-shelf export would be 0.25–0.50 Sv. The filamentation of the upwelling front in response to the meandering of the upwelling jet suggests a very effective mechanism for cross-shelf exchange close to CSV, as seen in the Spring-Summer climatology [Sánchez and Relvas, 2003].

4.2. Comparison With Other Filaments

Table 1 describes the characteristics of upwelling filaments in the eastern Pacific and eastern Atlantic, including the October 2004 filament. Common characteristics are the sharp SST contrast between the filament (12–13.5°C) and the ambient waters (16°C) and the presence of a seaward jet of about 10–70 km width, up to 300 km long over a relatively deep surface layer of about 300 m. Offshore velocity inside the jet may range 0.3–1.0 m s$^{-1}$,
decreasing in depth with a depth scale of 150 m. The seaward transport may reach up to 4 Sv. [55] The Cape St. Vincent filament was akin to observations in the NW Iberian peninsula in summer 1998 [Barton et al., 2001], rather than to the CCS examples (Table 1), as Iberian filaments were less energetic than the CCS counterparts. The thermohaline and surface structure was similar in all cases, with significant structure below 300 m, but the
Table 1. Comparative of Filaments Cited in Literature

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<th>Reference</th>
<th>Onshore vel., m·s⁻²</th>
<th>Offshore vel., m·s⁻²</th>
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*From left to right: offshore velocity (m·s⁻¹), onshore velocity (m·s⁻¹) (The method used to estimate velocities is annotated as: geostrophy, drifters, ADCP and sequence of satellite images), normalized relative vorticity, transport (Sv), filament dimensions (width/length in km, depth of the jet in m), maximum horizontal temperature gradient (°C·km⁻¹), maximum horizontal salinity gradient (°C·km⁻¹), vertical density gradient (kg·m⁻²), reduced gravity (m·s⁻²), Rossby baroclinic radius (km). Horizontal T/S gradients are based in hull-mounted probes or SST images.

These magnitudes are estimated over the top 200 m at the northern warm pool. For WOA01 the data used correspond to the 5° × 5° summer climatology at node locations: 37.5°N 127.5°W (CCS), 42.5°N 12.5°W (NW Iberia), 27.5°N 17.5°W (Canary), 37.5°N 12.5°W (SW Iberia). Sánchez and Relvas [2003] uses May–September data. The location is at 37°N 10°W. The literature review restricts to the northern Hemisphere EBCS.
Iberian filaments evidence smoother (horizontal and vertical) thermohaline gradients. The weaker velocities and shallower penetration of the jet resulted in a sharp reduction of the transport around the filament from 2–4 Sv to barely 1 Sv in the Iberian filaments.

[56] The CCS and Iberian system share many common features typical of EBCS: seasonality of coastal upwelling, squirts and jets deforming a contorted upwelling front, upwelling centers associated with capes and promontories, intense eddy field and poleward countercurrents. Regardless of local wind-forcing, a major difference between both systems is found in the vertical structure of the mass field. Table 1 also shows the vertical density gradients from the surface to 200 m at the time of the observations (most of them between June–August) as well as climatological summer averages. Vertical density differences (and consequently $g'$) were up to 35% greater in the CCS than in the Iberian or Canary regions. (These differences were about 30% over the top 500 m, not shown).

[57] To evaluate the consequences this may have on a equatorward jet, consider a 2-layer model in a semi-infinite domain, with an upper layer forming a surface front similar to that found in our observations [Killworth et al., 1984] (Figure 13). Assuming geostrophy, there is a meridional velocity dependent on the square root of the reduced gravity:

$$u(y) = \sqrt{g' H_0 \exp(y/R_d)}.$$  \hfill (4)

[58] As $g'$ is 50% higher for the CCS than the Iberian region, this model predicts a higher velocity for the CCS by

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**Figure 13.** The frontal model configuration used by Killworth et al. [1984].
a factor of \(\sqrt{1.5}\), which may partly explain the different velocities observed in these regions and could be crucial for their dissimilar formation and evolution of filaments.

[60] Killworth et al. [1984] argued that this configuration is unstable to small perturbations, being the wavelength of the fastest growing mode:

\[
\lambda_f = 5.5R_d \sqrt{|s|} r - 1,
\]

with a real component of the phase speed of

\[
c_f = \frac{0.113}{\sqrt{r-1}},
\]

with \(H_o\) the asymptotic thickness of the upper layer and \(H_o\) the total water depth.

[61] Let’s assume 500 dbar as the level of no motion for both systems [see, e.g., Relvas, 1999, and references therein]. Taking 200 m as the mean penetration of the upwelling jet in the CCS and 100 m for the Iberian system (Table 1), we assume \(r = 2.5\), \(R_d \approx 30\) km for the CCS (top 200 m), and \(r = 5\), \(g' = 0.012\) m s\(^{-2}\) and \(R_d \sim 15\) km (top 100 m) for the Iberian case. This would result in \(\lambda_f(\text{CCS}) = 165\) km, \(c_f(\text{CCS}) = 0.10\) m s\(^{-1}\), and \(\lambda_f(\text{Iberia}) = 115\) km, \(c_f(\text{Iberia}) = 0.06\) m s\(^{-1}\). This suggests that the dissimilar stratification is responsible for the larger wavelength of the fastest mode of instability (1.4 times greater in the CCS) and the faster \(c_f\) (1.7 times faster).

[62] The source of this dissimilar stratification is related to the salinity profiles. Low-salinity Pacific Subarctic Water and river inflows cause salinity to increase with depth in the CCS [e.g., Dewey et al., 1991]. Contrarily, the presence of MW underneath the eNACW cause a salinity minimum at approximately 400 m depth off Iberia [e.g., Sánchez and Relvas, 2003]. The result is that descending in the top layer, decreasing salinities partially compensate the temperature decrease with depth and the water column is less stratified off Iberia than in the eastern Pacific (see Table 1). As upwelling favorable winds bring subsurface water up to the surface, baroclinicity (understood as the isopycnal slope) is also weaker, contrasting with the stronger horizontal gradients in the CCS.

### 4.3. Upwelling Circulation: Structural Asymmetry and Vertical Motion

[63] The October filament was formed by mesoscale meandering of the upwelling jet. Between the seaward jet and the onshore return flow a cyclonic zone occurred, dissected by a number of anticyclonic turns (Figure 7A). These emerged in satellite (Figures 1 and 4) and the hydrographic observations (Figures 6 and 9). Structural asymmetry, typical of CCS filaments [Rienecker et al., 1985; Ramp et al., 1991; Dewey et al., 1991; Flament and Armi, 2000] was also observed here. The filament was bounded to the north by a large (~100 km) anticyclone, featuring relatively large transitions from cold to warm water (>20 km). Sharp cyclonic bending occurred to the south, what caused an onshore flow that gave continuity to the equatorward jet (Figure 7). Although velocities were similar for both the seaward and return jets (~0.3 m s\(^{-1}\)) horizontal shears greater than 0.4f occurred at the southern (cyclonic) boundary, coincident with a sharper SST front (Table 1). The sharp meanders were responsible for advecting of geostrophic vorticity and vertical velocities of ±15 m d\(^{-1}\) around the filament. Although more intense at the northern flank, upwelling spots occurred north and south of the main flow, separated by patches of downwelling (Figure 12B). These results showed a striking coincidence with SeaWiFS chlorophyll-a distribution (Figure 12C), which reinforced our confidence in these estimations. As well as the seaward export of upwelled, nutrient-rich water, the filament induced a patchy distribution of pelagic biota.

[64] Vertical motion has been inferred in filaments using different approaches, including primitive equation models, drifters and biochemical tracers. Subduction of cold water along the filament is related to the convergence with warm, oceanic water, being responsible for the sharp temperature gradient in the outer edge of the filament [Flament et al., 1985; Kosro and Huyer, 1986; Brink and Cowles, 1991; Brink et al., 1991; Washburn et al., 1991]. Flament and Armi [2000] reported cross-isotherm convergence from sequences of satellite images, surface drifters and accumulation of debris. The paper detailed the asymmetry between the anticyclonic and the cyclonic sides of the jet, and strong convergence at the cyclonic front. The authors also suggested frictionally driven ageostrophic secondary circulation superimposed on the geostrophic flow as well as isotropic divergence of 0.3f at the jet axis, which was indicative of local upwelling along the jet.

[65] These processes have not been explored in the present paper. Rather, vertical motion diagnosed here was based on the system response to the geostrophic advection on the geostrophic and hydrostatic balance (QG dynamics). QG approximation seems to be a reasonable choice to the dynamical study of the filaments, justified by the fact that filaments consist of relatively strong currents with narrow structure and relatively large Rossby number flows. Dewey et al. [1991] presented an extensive momentum analysis of a filament in Punta Arena and noted a maximum ageostrophic contribution of 6.5% of the total variance. The authors estimated vertical motions driven by the adjustments in relative vorticity [Dewey et al., 1991]. This was in agreement with our findings using a different approach and corroborates that vertical motion occurs after the meandering of the upwelling jet.

### 4.4. Coupling With Subsurface Flows

[66] A complex subsurface circulatory pattern interacted with the upwelling (surface) circulation (Figure 10). Between 90–150 dbar a coherent jet recirculated around a relatively saline and warm anticyclone, probably A1 of Figure 4D. The origin of this flow is unclear although we speculate that it could be part of a subducted, former upwelling jet (Figure 6) formed after the persistent northerlies of September 2004, that was overridden by the recent event of 8–18 October (Figure 3).

[67] The poleward flow below 200 dbar along the southwestern tip of the sampling area was rather barotropic (Figure 8) and exhibited a clear tendency for cyclonic motion. This circulation was coherent with \(C_1\) and \(C_2\) observed in the SLA maps at approximately that location (Figure 4D).
Portimão canyon, close to Cape St. Vincent branch could be part of a dipole formed by an anticyclonic possibility remains that this meddy feature and the cyclonic collect hydrographic data below 400 m but ADCP vectors MW influence (St. 11 in Figure 5). Unfortunately we did not cyclones are largely barotropic with velocities >0.20 m/C1 baroclinic and deep velocity structure, the accompanying 37.10.

5. Conclusions

Perhaps more interestingly, a well-defined and relatively isolated warm and saline spot occurred at 9.75°W 37.10°N (Figure 6, at 340 dbar). T-S properties revealed the MW influence (St. 11 in Figure 5). Unfortunately we did not collect hydrographic data below 400 m but ADCP vectors showed a coherent anticyclonic motion around this feature. This suggests the presence of a shallow meddy. The possibility remains that this meddy feature and the cyclonic branch could be part of a dipole formed by an anticyclonic meddy and an attached cyclone. Cyclones usually occur with MW salinity lenses detached by topographic forcing at Portimão canyon, close to Cape St. Vincent [Serra and Ambar, 2002]. While the meddies usually have a rather baroclinic and deep velocity structure, the accompanying cyclones are largely barotropic with velocities ~0.20 m s\(^{-1}\) at depth levels as shallow as 300 m. They have been traced by SLA anomalies and in clear SST images [Serra and Ambar, 2002; Serra et al., 2005].

A complex subsurface circulation underneath the upwelling flow was revealed using SLA maps, thermohaline properties and ADCP velocities. Observations of undercurrents, anticyclonic meanders and meddy dipolos in the area are not infrequent [Relvas, 1999; Serra et al., 2005]. Their role on the upper layer flows has not been assessed here, but we speculate that their presence is key. Strongly opposing subsurface flows increase vertical shear which is a source of baroclinic instability, contributing to the destabilization of the upwelling jet and generation of filaments, as suggested by Ikeda and Emery [1984]. This topic deserves further research.

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