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NEAR SURFACE DYNAMICS OF COASTAL UPWELLING

by

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ABSTRACT

Recent advances in the understanding of the dynamics of the Ekman layer in coastal upwelling regions are reviewed. In particular, the question of Ekman transport as a local indicator of the strength of upwelling is investigated. Results over the last decade have highlighted the inhomogeneous nature of the interaction between the coastal upwelling zone and the offshore oceanic waters. The role of upwelling filaments, their relation to eddy structures, and their importance in lateral exchanges of nutrients and biota are investigated.

INTRODUCTION

Understanding of the basic process of coastal upwelling is predicated on the work of Ekman (1905). There is a widespread, albeit implicit, acceptance of the Ekman model in conditions which do not comply with the simplified homogeneous, stationary, constant eddy viscosity situation considered by Ekman. Questions remain as to what is meant by the Ekman layer, how appropriate is the adoption of the Ekman transport, what the implications of time varying wind stress are. Moreover, recent observations show that upwelling regions are far from uniform alongshore though the predominant meteorological scales are considered to be long. What is the significance of these observations and what do they tell us about the nature of upwelling systems?

EKMAN TRANSPORT

The Ekman-Sverdrup model of coastal upwelling allows an estimate of upwelling intensity, i.e. offshore near-surface transport, from a knowledge of only the local wind stress Tau (Sverdrup, 1938). The volume transport is given as

$$T=\frac{\tau}{\rho f}$$

where τ is the surface wind stress ($\tau = C_d \rho_a w^2$), ρ is the water density, and f is the Coriolis parameter. The applicability of this index was demonstrated on the climatic time scale by Wooster and Reid (1963) in their review of the Eastern Boundary Currents. Since that time the use of the index has been extended by Bakun (1975) to weekly and daily mean conditions. Numerous studies in upwelling regions have demonstrated a close relation between local wind stress and upwelling activity indicated by sea surface temperature anomaly or other measure. However, the practical difficulties of making reliable current observations in the upper layers of the ocean have largely precluded testing in any systematic and quantitative way of the reality of this model.

It is often tacitly assumed that the surface mixed layer, which develops in response to the strong surface wind stress, is identifiable with the Ekman layer or layer of offshore transport (*e.g.* Brink, 1982), and is of constant depth. It has to be noted, however, that the surface mixed layer is not of fixed thickness, but varies significantly as the wind forcing changes. Secondly, there is little *a priori* justification for equating the mixed layer with the layer of Ekman transport. These issues have been explored by Lentz (1992) on the basis of observations carried out off Peru, Northwest Africa and Northwest America during the major upwelling experiments of the 1970s and early 1980s. Using data from current meters and temperature sensors covering the upper layer with at best 5 m resolution and meteorological information from adjacent surface buoys, he investigated the properties of the surface mixed layer and the related current structure.

The results show that the mixed layer depth is distributed approximately as chi-squared with, obviously, a predominance of near zero values. The depth of the mixed layer in the various regions varied between zero and almost 60 m, greater depths being associated with stronger wind stress. It was found that a good parameterisation of the depth was given by

$$h = A\, u^* / (N f)^{1/2}$$

where u^* is related to wind stress by $u^* = (\tau/\rho)^{1/2}$, N is the Brunt-Vaisala frequency at the base of the mixed layer, and f is the Coriolis parameter. This formulation arises from one dimensional mixed layer models which neglect the effect of both surface and lateral heat fluxes (e.g. Pollard et al., 1973). Since significant surface heat flux is normal in coastal upwelling regions, it seems to be the case that the lateral advection of heat is sufficient to compensate.

Given a time varying surface mixed layer depth, the question of how this relates to the Ekman transport arises. Attempts have been made to compare the theoretical Ekman transport to the observed offshore flow in the near surface layer assuming a constant mixed layer depth (e.g. Smith, 1981). In the mean, it was demonstrated that there was "fair agreement" between the calculated Ekman transport and the offshore transport estimated from the near surface observations. In the three cases considered, the Ekman transport was 26% less, 89% more and 72% more than the estimated transport. The magnitudes of the standard deviations differed by 36%, 8% and 72%. Correlation of the time varying Ekman transport with the offshore transport (both de-meaned) in the constant thickness near surface layer was significant in all cases at better than the 99.9% confidence level, while the corresponding regression did not differ from unity by more than the standard error. Thus the variability, at least, of the offshore transport was similar to that of the Ekman transport.

Lentz (1992) noted that since the mixed layer depth varied through the observation period, the above method often excluded observations actually in the layer or included observations below it. By calculating the theoretical Ekman transport from hourly wind values and observed transport from hourly current meter observations within the actual observed mixed layer and then low pass filtering the series, he obtained better correlations (in the range 0.8-0.9 compared to Smith's 0.5 to 0.7). However, the results showed that the observed transport was consistently smaller than that predicted. He was able to show from the current observations that the layer of offshore flow, the depth of which also varied with the wind stress, was actually thicker than the mixed layer depth, by a factor of roughly 1.5. Repeating the calculation over this expanded depth range, including the "transition layer", provided better agreement between the predicted and observed transports.

Although it was concluded that the transports "agreed well", the mean observed value amounted to only 59% of the predicted value, averaged over all the observation sites used (Fig. 1). Excluding comparisons believed less reliable increases the observed to 66% of the predicted value. Lentz (1992) points out that the transition layer thickness was not adjusted for optimal agreement at each study site, so that an improved result could be obtained. So, the best estimates of actual and predicted offshore transport agree to better than a factor of two. Reasons why agreement is not even better are numerous.

The problem of defining the "offshore" direction is not trivial. Any mis-orientation of the axes can severely affect the estimate of offshore flow because the alongshore flows are so much stronger. Local depth contours may not parallel the coast line or other contours and may change direction irregularly alongshore, compounding the difficulty of selecting the length scale over which the directions should be defined. Generally, the definitions are derived from the data in the sense that some measure of the direction of maximum current variance is used to define alongshore flow. Good vertical resolution of the currents throughout the water column are needed to provide a reliable estimate in this case. Smith

(1981) showed that defining offshore as the direction of zero net transport produced inconsistent results, probably a result of poor coverage of the entire water column or local lack of two dimensional mass balance.

Another difficulty is that upwelling is not only a surface boundary phenomenon but also a lateral boundary one. Upwelling is supposed to take place within a distance of the coast defined by the Rossby internal radius of deformation. Any surface transport measurements made within this distance will not include the total Ekman transport since some water is being upwelled further offshore. Lentz (1992) suggested some of his poorer agreements might be due to the proximity of the mooring to the coast. He also noted that the surface mixed layer (and presumably offshore flow) depth in general decreased shoreward as bottom depth decreased.

The practical difficulties of measuring currents in the near surface layer again impose a major constraint on the reliability of these estimates. The effect of strong wave currents and surface buoy motion can have significant effect on the current measurements. Moreover, in many cases the current meter spacing has been irregular, due to instrument failure, and coarse in relation to the thickness of the surface layer so that vertical resolution of the flow has been poor.

In view of these considerations and noting that Ekman's formulation was for stationary conditions, it may be remarkable that the theory has proven so applicable. Nevertheless, Acoustic Doppler Current Profilers now allow us to measure the velocities at high accuracy and good spatial and temporal resolution through the major portion of the water column, while modern current meters make possible measurements in the near surface layer. Perhaps, the time is now opportune for renewed attempts to refine our understanding of the Ekman response in the context of the several continental shelf experiments proposed for the near future.

FILAMENT STRUCTURES

Granted the validity of the Ekman transport mechanism and the extended scale of the predominantly trade wind systems which drive it, one might expect that coastal upwelling would be quite uniform over long stretches of the eastern boundary. Satellite imagery of any coastal upwelling region, showing AVHRR sea surface temperatures or chlorophyll-like pigment from CZCS, reveals that there is little evidence of uniform conditions along the shelf. Instead, the impression is one of a contorted boundary between upwelling and oceanic waters, localised areas of stronger upwelling and tongues of cool, high chlorophyll water extending hundreds of kilometers offshore from the coastal zone. The tongues of cooler water, which are typically narrow, defined by strong lateral temperature gradients, and often terminate in eddy-like structures, are termed filaments (Brink and Cowles, 1991). This region of interaction between the coastal upwelling and the offshore waters is known as the coastal transition zone. Fig. 2 shows a dipolar filament off southwest Portugal.

Upwelling is generally a seasonal phenomenon. For example, off the Iberian peninsula it develops in late spring as a response to the onset of the Portuguese Trade winds. Brief episodes of near-shore upwelling may develop in winter as a result of short lived favourable winds, but it is not until the summer when persistent upwelling is found. Typically, in May or June a narrow band of colder water, often of quite uniform width, is produced along much of the coast. This band often appears to be edged by many narrow "fingers" of cool water extending 20-30 km offshore. If the winds relax or become poleward, this narrow band may disappear temporarily. As the season progresses, major filament structures develop and extend offshore.

A statistical analysis of the archive of AVHRR scenes for the Iberian region for the years 1982-90 shows the seasonal development of filaments (Haynes *et al.*, 1993). There were fewer images available in winter than in summer because of persistent winter cloud cover, but the relative frequency of occurrence of filaments in winter was small compared to other times of year (Fig. 3). The filaments grow to identifiable lengths of 80 km by about the start of July (day 180). At this time the width of the band of upwelling is around 50 km, so the filaments extend some 30 km seaward of the upwelling boundary. They grow over the following weeks to reach a maximum mean length of around 130 km in late September (day 270). The maximum observed filament lengths show the same growth trend, increasing from about 120 km at the start of July to over 270 km in late September. After this, both mean and maximum lengths decrease until the filaments become less common in October (around day 300). Their ultimate fate is not well documented since cloud cover obscures the region as soon as winds become unfavourable. However, it is probable that once upwelling ceases to renew the filament waters near to shore, that surface warming due to insolation quickly destroys their surface manifestation.

A sequence of images off Iberia from 1982 shows the later stages of development. On 9 August, nascent filaments were evident as bulges in the upwelling front at 42 30'N, 41 10'N and 40 20'N. Traces of the small scale fingers remain at this time. By 20 August, the bulges had extended offshore by about 100 km to form identifiable filaments. The offshore progress of the filaments continued until 2 September, when the longest filament had reached about 200 km extent. The observed rate of advance of the filament offshore was about 10 cm/s. After 2 September, warming commenced in the southern part of the region as a result of weakening winds. Cloud cover then increased and the structures had disappeared by the time of later images in October.

What is the significance of these features in terms of onshore-offshore transport? Although there are few *in situ* observations of filaments off the Iberian peninsula, recent work carried out off NW Africa reveals something of the structure of this type of feature. Satellite sea surface temperature imagery show an extended cool feature reaching from between Cabos Juby and Bojador towards the island of Gran Canaria. The filament is over 120 km in length and appears to terminate in a cyclonic eddy. The sub-surface distributions associated with the feature were mapped by CTD survey data (Fig. 4), in which can be seen the partial separation of the eddy from the main structure. Detailed cross sections normal to the filament show that the isolines upwarp strongly near-surface but also that a signal is still evident as deep as 400 m (Fig. 5). The steep inclination of the isopycnal surfaces indicates intense offshore baroclinic flow associated with the cool feature. This is confirmed by Acoustic Doppler Current profiler data from the cruise. High chlorophyll (fluorescence) values occur in the feature. The chlorophyll structure in the surface layer could be compatible with subduction on the anticyclonic side of the filament, as has been observed in the California Current region (Kadko *et al.*; 1991).

Large offshore transports occur in relation to the filament, many times greater than the Ekman transport. The filament therefore represents a potential mechanism for removing cool nutrient rich water and its constituent biota from the coastal region, if mixing takes place across the filament boundaries. In the present case, a cyclonic eddy of significant dimension terminates the eddy. The higher chlorophyll fluoresence values in the eddy are indicative of the export from near shore. Since the eddy is essentially a dissipative feature, the filament is producing a one way transport from the coastal upwelling region into the oligotrophic offshore areas. This mechanism is thought to be partly responsible for enhanced primary production around the Canary Islands.

DISCUSSION

Considerable effort has been expended in testing the applicability of the Ekman transport to coastal upwelling, with encouraging, but not close quantitative, agreement between theoretical and observed offshore surface layer transports. Further investigation of the near surface transports with modern techniques are required for better understanding of the time and space varying response. Once upwelling is established, the development of a field of filaments and eddy structures occurs. The strong horizontal flows in filaments coupled with dissipative eddies may be the predominant mechanism for export of upwelled water and its properties out of the coastal region to enrich the offshore oligotrophic zone.

ACKNOWLEDGEMENTS

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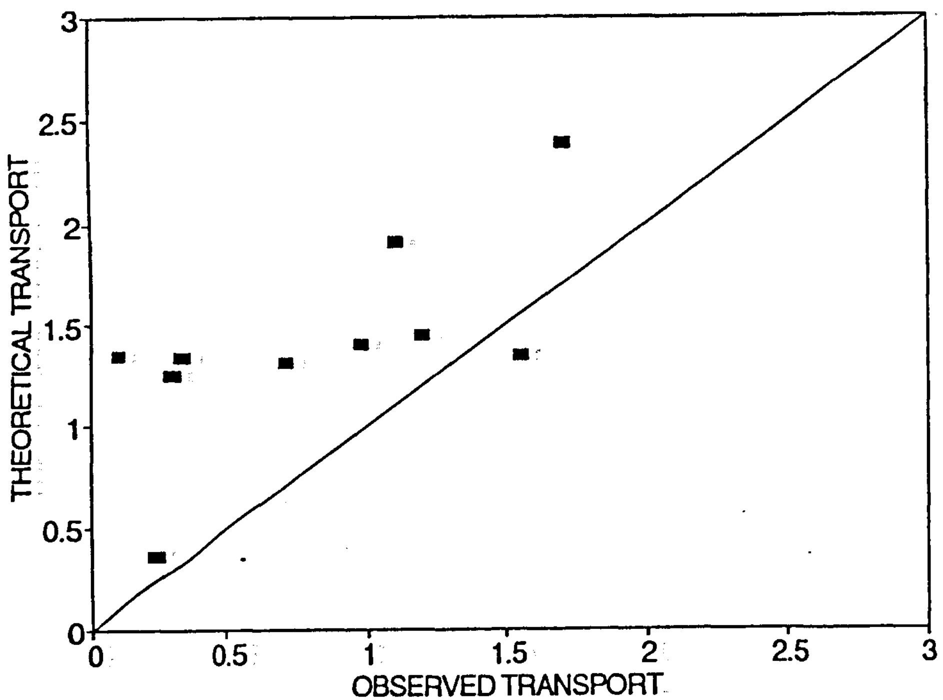

Fig.1 Mean theoretical Ekman transport vs. mean observed offshore transport in the near surface mixed layer and transition layer (Data from Lentz, 1992).

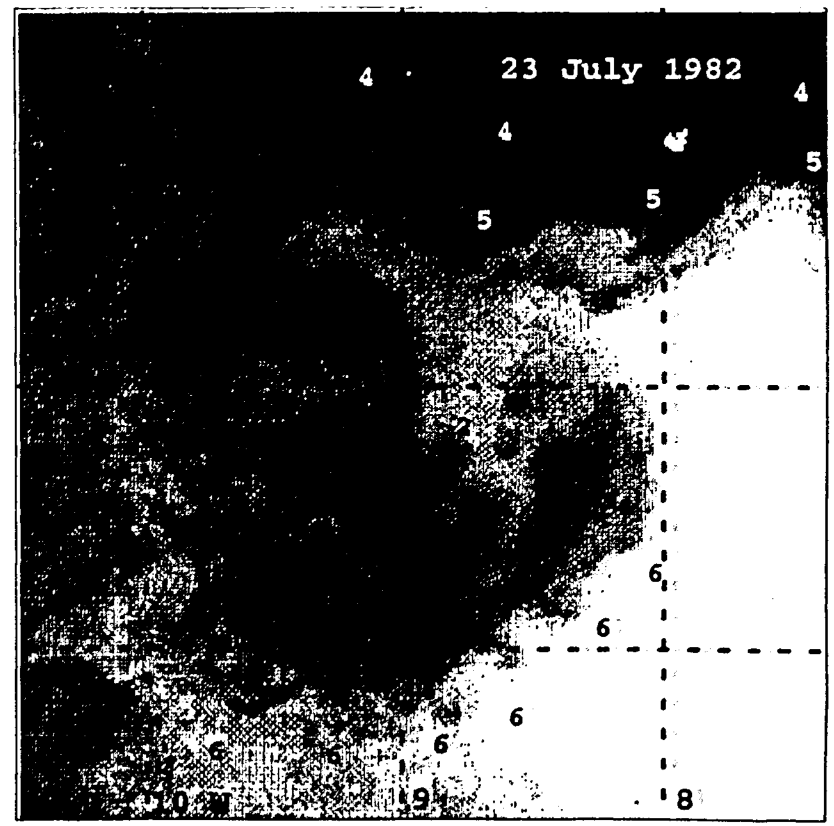

Fig.2 A filament of cooler (darker tones) water extending offshore of Cabo Sao Vicente, Portugal.

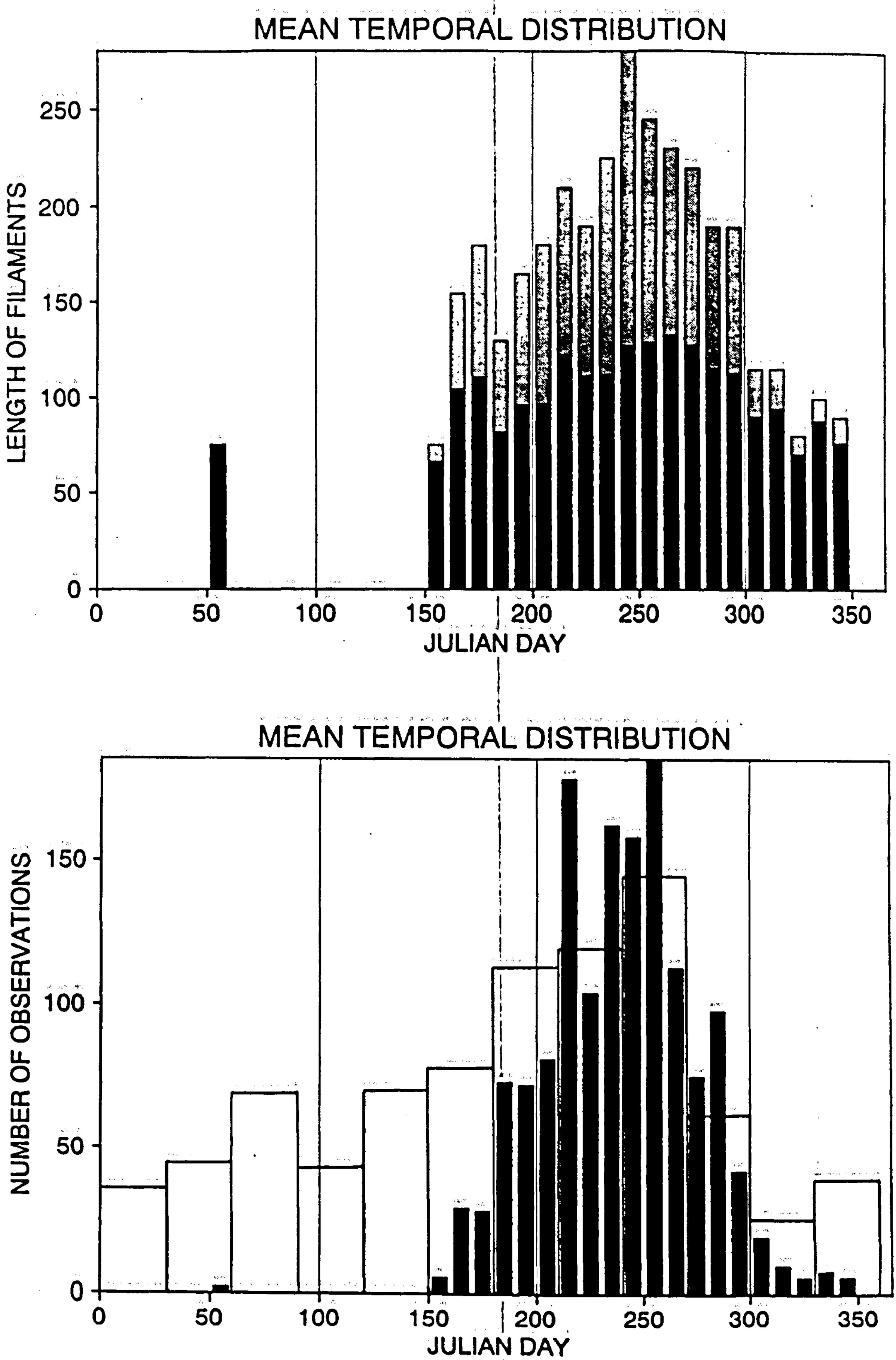

Fig.3 Seasonal variation of mean and maximum filament length (top) and number of available images and filaments observed (bottom) off Iberia for period 1982-90.

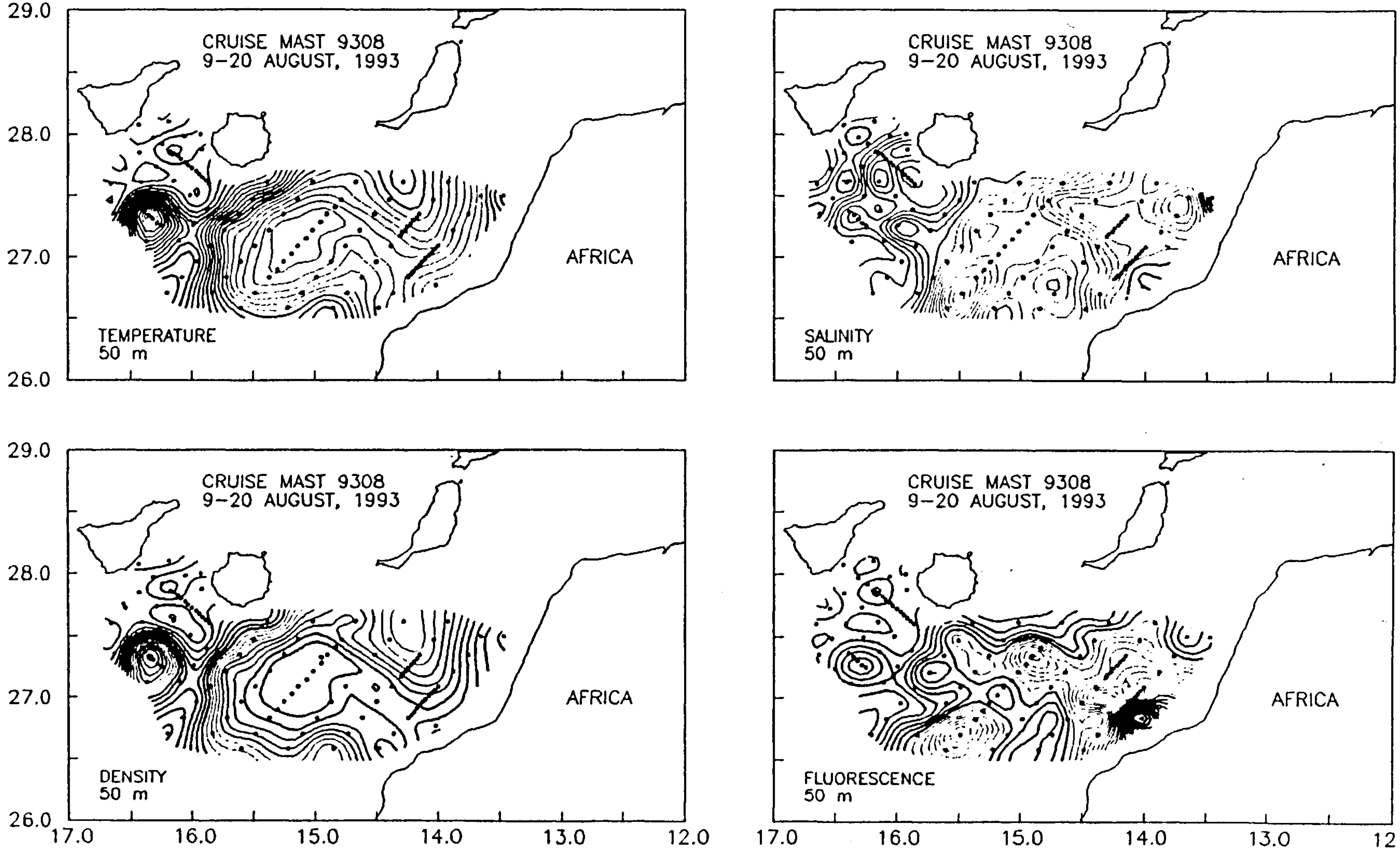

Fig.4 Maps of temperature, salinity, density and chlorophyll fluoresence off NW Africa in August 1993 showing filament and eddy structure.

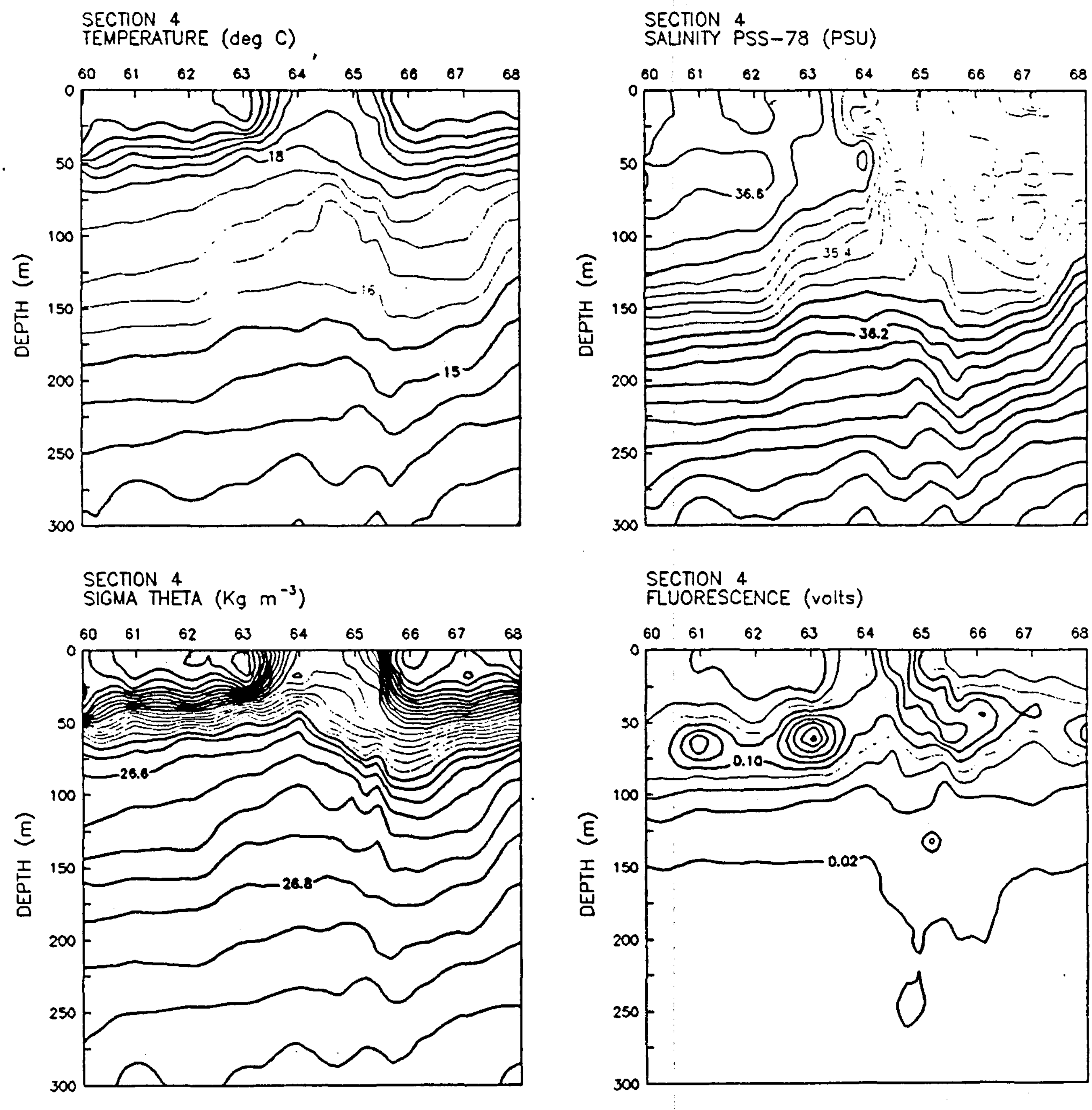

Fig.5 Vertical section of temperature, salinity, density and chlorophyll fluoresence across the filament of Fig 4.