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AN UPWELLING FILAMENT NORTH-WEST OF CAPE TOWN, SOUTH AFRICA

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Satellite images frequently show a thin filament of water stretching from the Cape Peninsula upwelling cell to beyond the shelf edge north-west of Cape Town. The filament carries nutrients and weakly motile biological organisms from the shelf zone towards the coastal transition zone, where eddies are observed. In order to probe the dynamics of the filament and its generating mechanism, a cruise was undertaken from 7 to10 February 1996. At that time, there appeared to be a filament in retreat, following an upwelling episode. There were two apparent eddies beyond the shelf edge observed in the satellite-derived SST. The southern feature did not have a clear hydrographic structure, but the northern one did, and it appears to be a semi-permanent feature. The region is forced by a number of oceanic and meteorological boundary conditions, none of which is entirely predictable. One is the sporadic advection of warm water from the Agulhas Bank onto the southern shelf. The cruise took place following such an event. The anticipated shelf-edge jet was greatly diminished and forced inshore. The possible effect of barotropic shelf waves on the configuration of the upwelling tongue and the formation of filaments is discussed.

In the past two decades, research interest has focused on the eddy fields of continental shelves, and filaments in the Coastal Transition Zones (CTZs) of eastern boundary regions. Brink and Cowles (1991) here given an overview of the extensively studied California-Oregon area, pointing out the salient features and dynamics of filaments and other CTZ processes. Similar studies have been pursued since then, withpapers relevant to the present one by Barth and Smith (1998), who studied the stability of a coastal upwelling front, and Strub *et al.* (1998), who compared circulation in the Benguela and California systems using altimetry and surface temperature. A few studies have related the CTZs in other geographical areas, notably on frontal mesoscale eddies off Portugal (Haynes and Barton 1991), the kinematics of the South-West African upwelling and frontal systems (Lutjeharms and Stockton 1987) and filaments and cross-frontal water exchange off North-West Africa (Kostianoy and Zatsepin 1996). Hutchings *et al.* (1986) discussed the biological characteristics of frontal structures in the Benguela system. Filaments in the Benguela system were first studied by Shillington *et al*. (1992).

The summer California-Oregon filaments are features rich in chlorophyll *a*- and nutrients up to 100 km wide. They have their origin in the coastal upwelling process, and stretch hundreds of kilometres offshore. A typical core velocity is 50 $cm \cdot s^{-1}$. Filaments appear to sink as they move offshore, but mixing with surrounding water could also account for their apparent disappearance. The instability of the summer California Current results in meandering, with the formation of eddies that can be instrumental in drawing water offshore as filaments. In the Benguela system, filaments appear to correspond in length roughly to the shelf width. South of 29°S, the shelf has a nominal width of 170 km, narrowing quite abruptly at 33°S to 60–80 km. Bathymetry appears to play an important role in the Benguela system.

A feature often observed on NOAA thermal satellite images in the narrow southern zone between Cape Town and Cape Columbine (Fig. 1) is a prominent westward excursion of water beyond the shelf edge. It is seen in both summer and winter. Its latitudinal position varies between 33 and 34°S, and it may have an associated eddy structure on its north-western extremity. Occasionally, it appears to develop a mushroom head with counter-rotating eddies.

Early examples of this formation appear in a synopsis of images from the first Coastal Zone Colour Scanner survey of the southern Benguela, the *Nimbus-7 CZCS* (Shannon *et al.* 1985). Of particular note are Figures 8 (Orbit 1634, 19 February 1979) and 13 (Orbit 2726 9 May 1979) of that study, both of which show excursions of water extending more than 100 km from the coast beyond the shelf break. The configuration is sketched in Figure 1. In the latter case, there is an impression of a thin stream, or “squirt”, of water moving south-eastwards from the shelf edge north of Cape Columbine, then mushrooming to form a pair of eddies, the southern cyclonic one entraining the filament emanating from the Cape Peninsula.

Evidence will be presented of a semi-permanent eddy west of Saldanha Bay over the Cape Canyon

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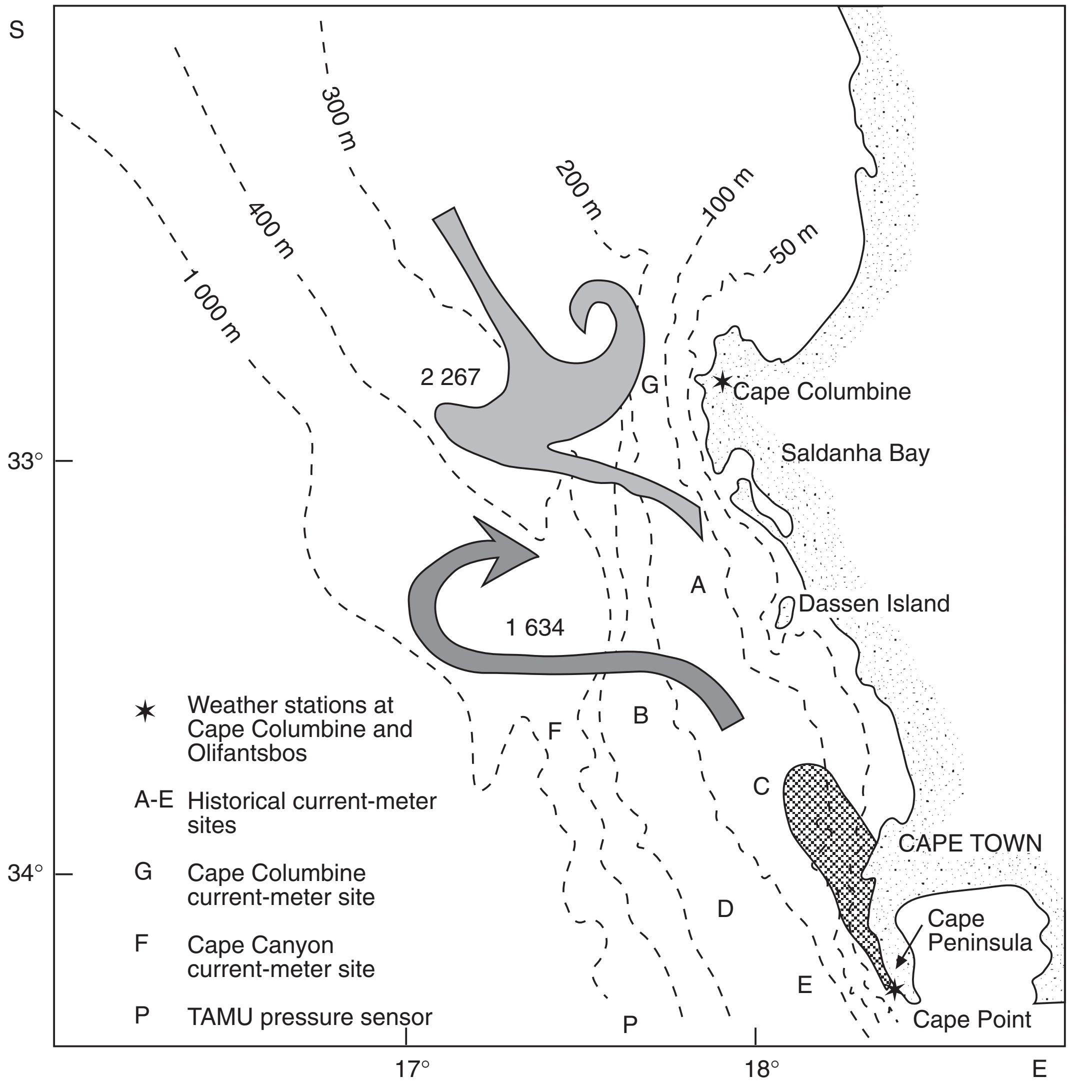

Fig. 1: Map of the Cape Canyon area of South Africa's west coast, positions of historical current-meter moorings and two coastal weather stations. The shaded areas represent eddy and filament formations, extracted from *Nimbus-G* orbit numbers 2267 and 1634 (see Shannon *et al.* 1985). Point P is the site of the Texas A&M (TAMU) University pressure sensor. The hatched area relates to the area of coldest upwelled water on 7 February 1996 (see text)

(Fig. 1). A pre-existing eddy would draw in shelf water, which in turn would entrain water from an upwelling tongue. Lutjeharms and Matthysen (1995) identified 21 eddies from NOAA images between 17 October 1985 and 10 December 1993. Four of these, measuring 60, 37, 72 and 44 km in diameter, were in the vicinity of the Cape Canyon. The others, mostly of smaller average dimension, were observed farther south between the 200- and 400-m contours, possibly generated by shear between equatorward flow on the outer shelf and poleward flow where it breaks the surface. The current structure in this area has been discussed by Nelson (1985).

It is not possible to draw dynamical conclusions

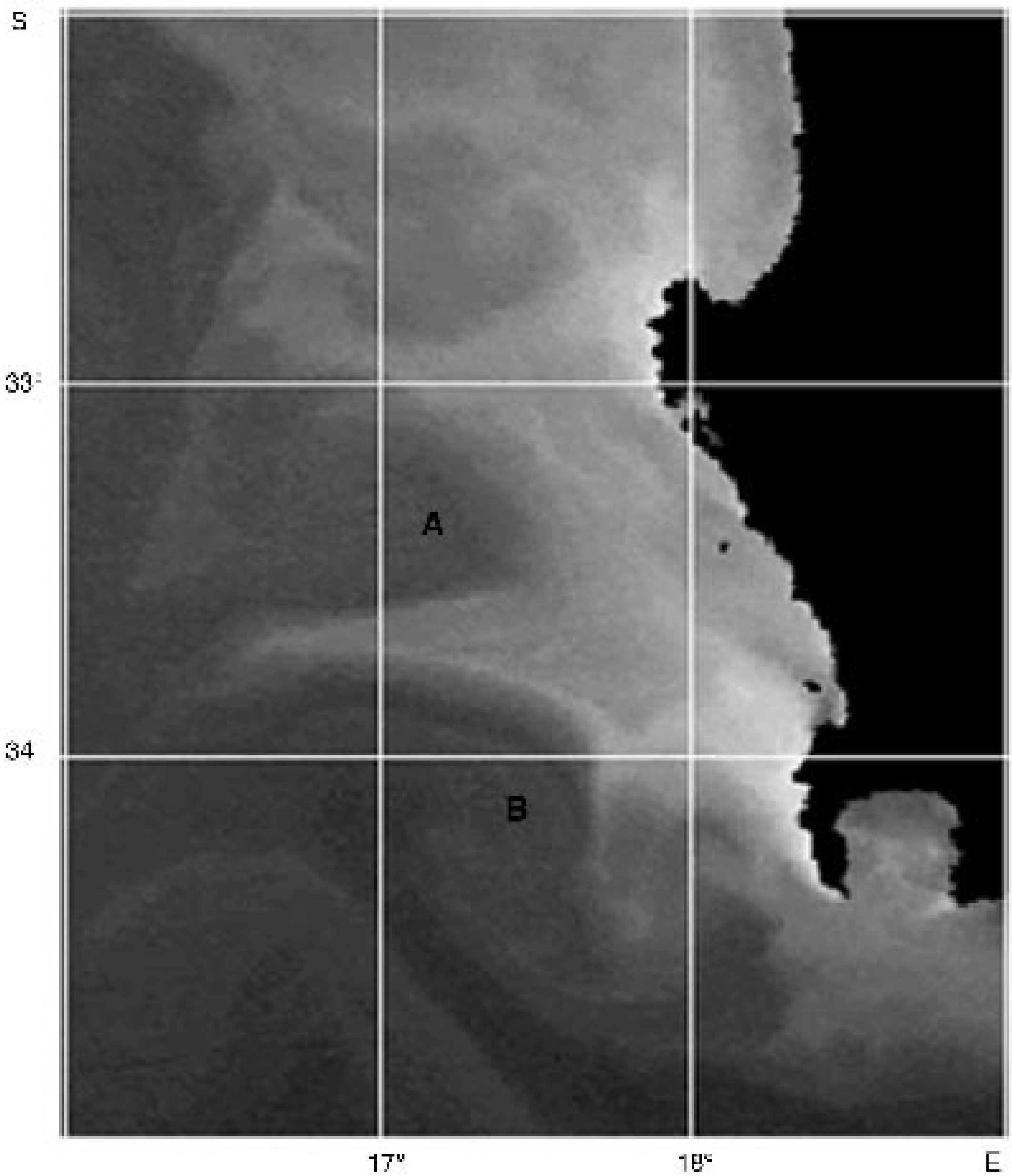

Fig. 2: NOAA satellite thermal image, 7 February 1996, with eddy-like features at A and B

from satellite images, but the large number of instances of the eddy and filament feature seen on NOAA images prompted a hydrographic investigation of the dynamics. With a view to studying these phenomena, a cruise was planned between 7 and 11 February 1996 in the hope of finding active filament formation. Indeed, a filament of upwelled water originating from the coastline of the Cape Peninsula had developed and appeared to be retracting at the time of the cruise. The near-surface movement of water there is of critical impor-

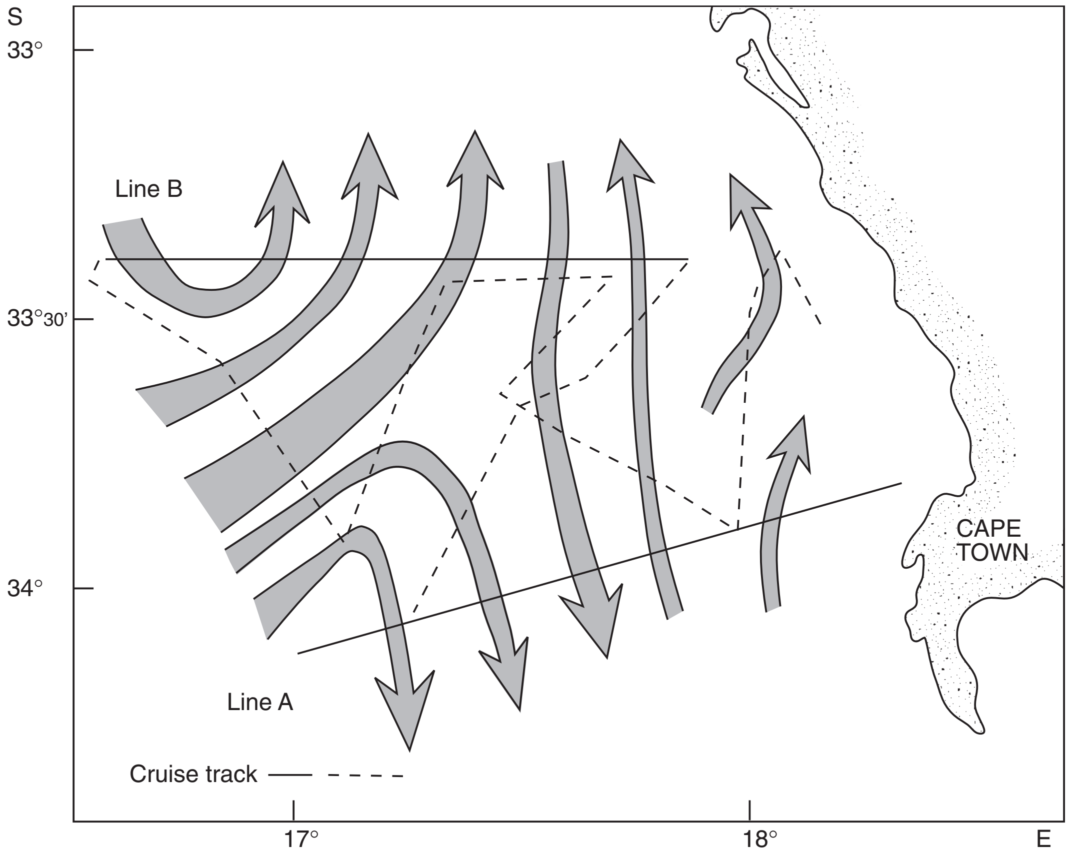

Fig. 3: Schematic flow pattern based on integrated ADCP measurements (8–72 m), the cruise track and Lines A and B referred to in the text

tance to the transport of eggs and larvae of pelagic fish from the western Agulhas Bank to nursery grounds farther north.

The text that follows details the currents, hydrography, surface features and meterological conditions prevailing during the cruise. Satellite-derived SST of the area on 7 February 1996 (Fig. 2) aided planning the ship's track.

RESULTS OF THE FIELD MEASUREMENTS

Currents obtained from the ADCP

A schematic of the 8–72 m depth-averaged current patterns derived by means of an RDI 150 kHz hull-mounted ADCP profiler, with the cruise track overlaid, shows the upper-layer circulation (Fig. 3). The station separation was 5 miles and currents were derived from both underway snapshots at typical speeds of 8 knots and on-station profiles, where hydrographic work was done. The latter measurements carry more weight in the synthesis of the flow pattern (Fig. 3). The instrument was run continuously with a 300-second integration time and bottom-fixed motion correction in depths shallower than 350 m. In deeper water, non-differential Global Positioning System fixing was used to obtain the ship vector. At each of the hydrographic stations, 3–6 profiles were averaged, omitting those affected by manœuvering of the ship while on station.

The persistence of thermal features over several days, as seen in the satellite images, supports the synopticity of the ADCP flow-field, completed over four days. The impression obtained from the satellite image is that of two eddies, one either side of a filament (A

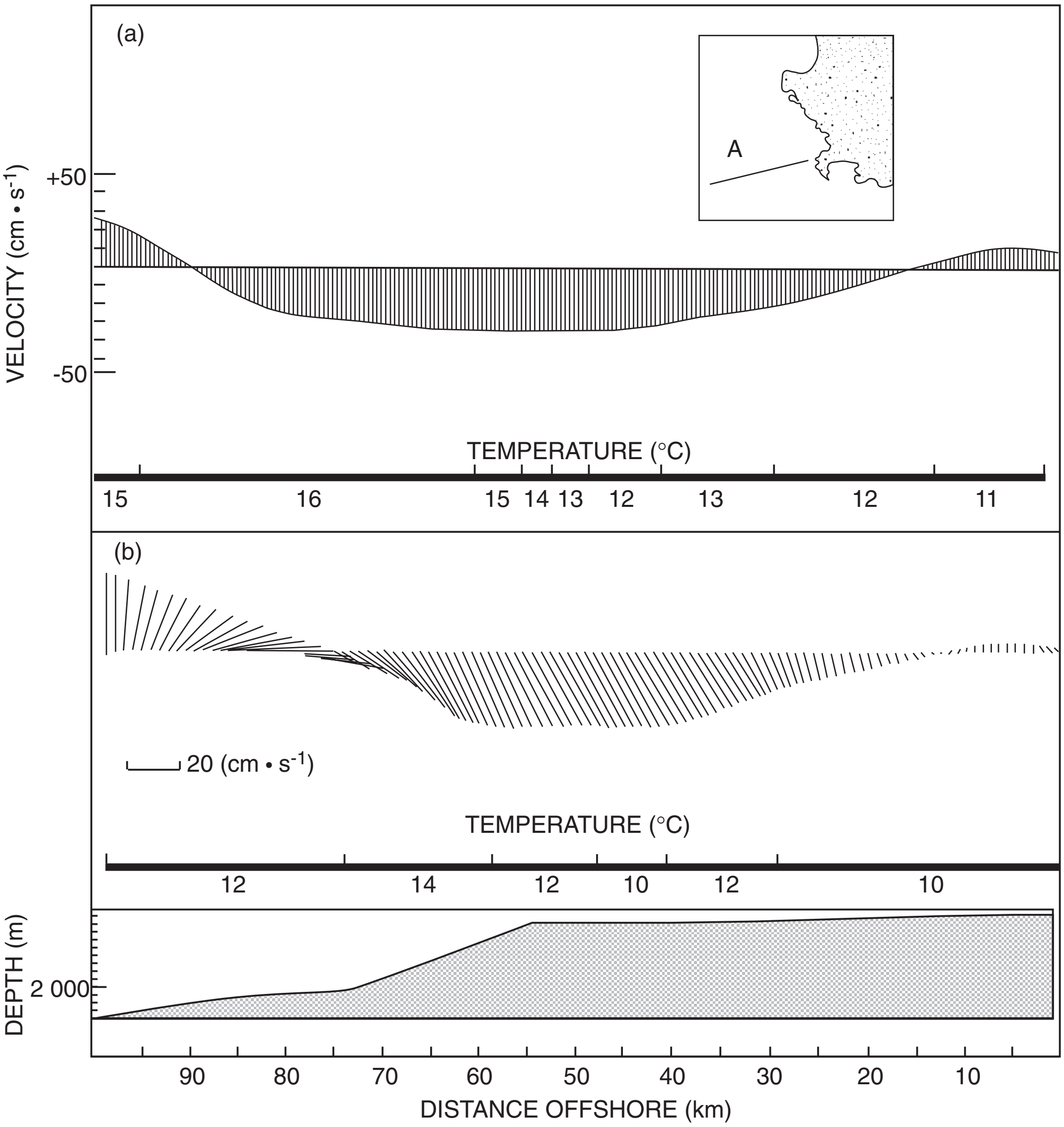

Fig. 4: (a) Depth-averaged (8–72 m) North-South (North is +) current profile on Line A with surface temperatures derived from CTD measurements and (b) entrainment of water beyond the shelf break by poleward-flowing water 64 m deep (vector referenced to true North), with temperature at that depth derived from CTD measurements, and bathymetry

and B in Fig. 2). The ADCP current field strengthens this impression, revealing anticyclonic motion centred about 33.5°S, 16.7°E and cyclonic motion centred about 34°S, 17.2°E.

East of these features, there was a narrow ribbon of poleward flow, adjacent to what appeared to be a weak baroclinic equatorward jet. Only in this region and east of it was there strong vertical current shear. Elsewhere, the average current from 8 to 72 m deep was barotropic.

A criticism of a synthesis such as this is that an ADCP snapshot of what is assumed to be the ambient flow can be significantly distorted by transient tidal

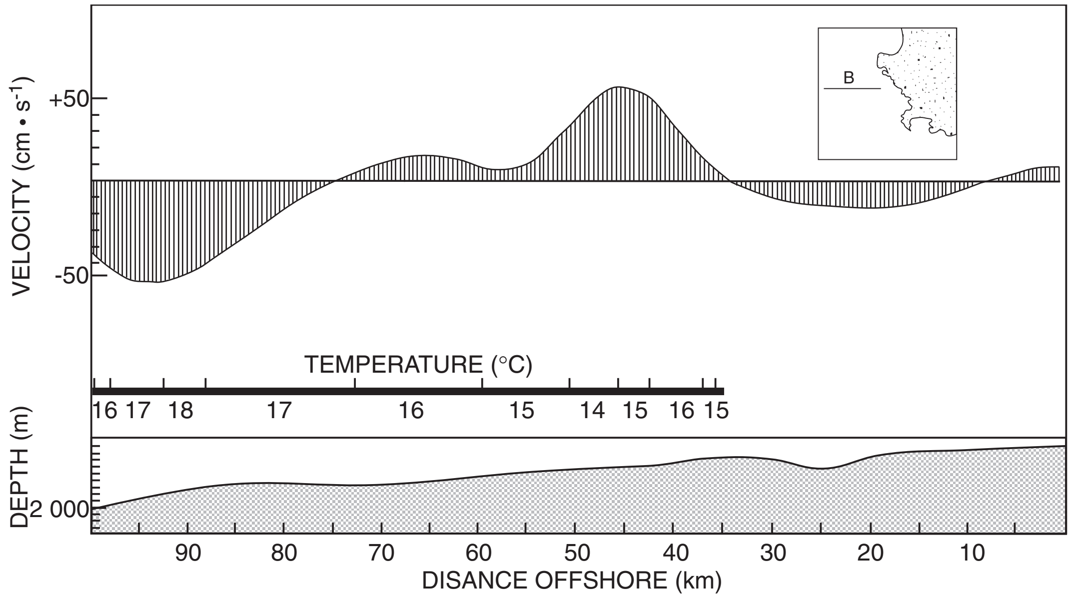

Fig. 5: Depth-averaged (8–72 m) flux normal to Line B (North is +) with surface temperatures (CTD) on western part of the line, and bathymetry

and inertial motions. Using historical data from current-meter moorings (labelled A – E in Fig. 1) in the vicinity of the Cape Peninsula (Nelson and Polito 1987), the transients were found to be smaller than the contributions from the ambient poleward flow or the inner-shelf equatorward flow. These records were low-pass filtered and then subtracted from the original series. The residual hourly vectors were sorted into 16 sectors of 22.5°, and the average speed in each sector was obtained. The supratidal frequencies had amplitudes <4 cm·s^{-1}. The root mean square values were typically between 2 and 4 cm·s^{-1}. Maxima up to 22 cm·s^{-1} occurred, but they were sporadic and sparse in the time-series, all of which were in excess of three spring cycles in length.

Figure 4a shows the depth-averaged (8–72 m) north-south current profile on Line A (see Fig. 3). Vectors have been interpolated between 12 stations using a cubic spline fit. The flow is polewards over most of the line. The stick vectors at 64 m on the line (Fig. 4b) indicate a strong entrainment of oceanic water into the poleward flow at the shelf break. This raises questions concerning the integrity of the southern eddy feature, the hydrography shedding more light on the matter. Temperatures at the surface (Fig. 4a) and 64 m deep (Fig. 4b) were obtained from CTD data, with cubic spline interpolated between the 12 measured values.

Figure 5 shows the flux normal to Line B (Fig. 3), integrated between 8 and 64 m, together with the surface temperature for the outer part of the line where CTD casts were done. On both Lines A and B, the speeds are considerably in excess of the transient supratidal frequency values.

Hydrography

The ship was equipped with a General Oceanics Mark 3b CTD, calibrated to a known accuracy of 10^{-2} in both temperature and conductivity and an accuracy better than 1 dB over a full depth range of 0–2 000 m. The acquisition, processing and filtering software was developed locally, following international standards.

Temperature at 6 m from an underway thermosalinograph did not give convincing evidence of a filament at the time of the cruise. However, the filament signal appeared clearly in the vertical structure of salinities from the CTD. During this apparent decay phase of the filament, salinity proved to be a more reliable signal than temperature, because of its conservative property. The hydrography during the cruise is detailed in Duncombe-Rae *et al.* (in press).

On Line A (Fig. 3), upwelled water of lower salinity was present at the surface, and a core of relatively high salinity water (35.00×10^{-3}) appeared over the 200-m

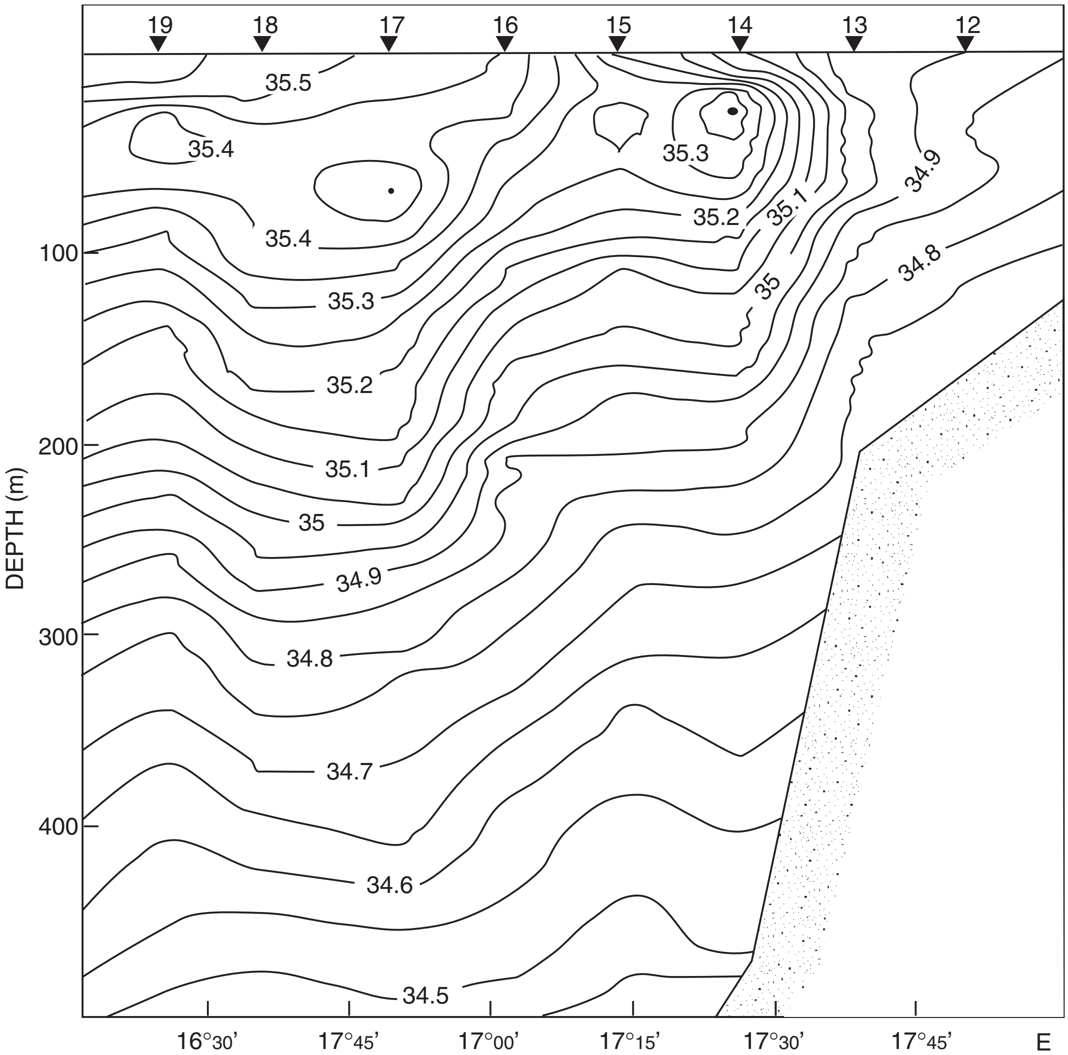

Fig. 6: Isohalines on Line B showing evidence of an eddy centred on 16.75°E. The station numbers are at the top

isobath at 18°E and at 50 m deep, east of a haline front. This may have been the remnant of the anticipated shelf-edge jet, normally situated much farther west (see review by Shannon and Nelson 1996). The section does not support the hypothesis that the southern feature in the satellite image is an eddy. From combined ADCP and temperature data, it appears that warm water was being entrained by the relatively strong poleward flow, causing it to rotate cyclonically at the southern extremity of the survey.

On Line B (Fig. 3), an eddy structure is clearly evident in the depression of isolines from 40–50 m down to 350–400 m in both temperature and salinity. The eddy had its clearest definition in salinity between 100 and 200 m (Fig. 6), indicating a diameter of some 30–40 km. Also on this transect is a core of high salinity water (35.3×10^{-3}) at 40 m over the 450-m isobath, between the eddy and a haline front over the shelf edge. This is in the region of equatorward-flowing water at 17.4°E on what appears to be the periphery of the eddy.

Inshore, a 20–30 m layer of upwelled water overlies

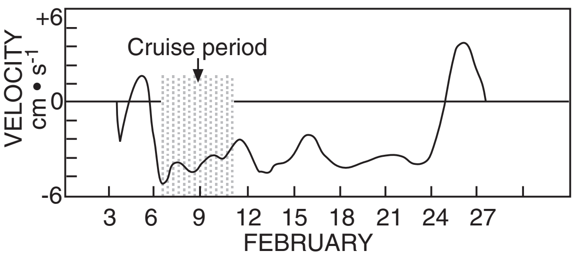

Fig. 7: Cape Columbine longshore current component (North is +) showing poleward flow from 3 to 27 February (speed inaccurate)

a 50–60 m thick layer of South Atlantic Surface Water. Together with the ADCP data, the salinity data show that the frontal structures were decaying at the time of the cruise, with the filament in retreat.

Current-meter records

The only current meter in the area at the time of the cruise was an Aanderaa RCM4 near Cape Columbine, situated 111 m deep in 174 m of water. Unfortunately, it appears that the rotor counter became defective some 40 days after deployment, and so the recorded speeds are slow. Nevertheless, the directional information is reliable and shows poleward flow over the period 6–24 February (Fig. 7) for the derived north-south component of flow, OSU low-pass filtered (Beardsley *et al.* 1985) to remove tides. The more energetic start to the series is typical of longshore motion propagating along the coastline from Cape Columbine to Cape Point, with periods of 3–6 days. Longshore currents may attain amplitudes up to 55 cm·s^{-1}. At Cape Columbine, they correlate strongly with local wind, but the correlation diminishes southwards to Cape Point.

From a 13-year time-series of currents at the Cape Columbine current-meter site (Fig. 1), the deep, poleward undercurrent is found to be strongest between mid-February and mid-April, the autumn period. Averaged over this 90-day period for the 13 years, the autumn speed is 11.0 cm·s^{-1}, compared to 6.8, 7.4 and 8.3 cm·s^{-1} for selected 90-day windows in winter, spring and summer respectively.

The unfiltered temperature time-series also correlates well with the longshore current, with cooling on the poleward phase, indicating entrainment of upwelled water, and warming in the equatorward phase. High-frequency spikes are caused by tidal mixing and contribute to the slow seasonal increase in temperature, from 8.8 to 9.6°C, at the current-meter depth.

Winds and atmospheric pressure

Air pressure at Cape Columbine and Olifantsbos correlate strongly, indicating that the distance between the two sites is smaller than the synoptic pressure length scale generating the marine geostrophic wind. Because of thermal boundary effects and the travelling coastal lows, longshore winds do not correlate as well, either in phase or amplitude. However, the waves in synoptic pressure, typically 10–15 hPa in amplitude with periods of 3–6 days, are coincident at both sites, and the geostrophic wind then dominates other thermal effects close to the coast (see Nelson 1992 for a theoretical discussion of synopticity and periodicity). Historical currentmeter and wind data show that upwelling events and the generation of longshore

Table I: Upwelling events, durations, total Ekman volumes per metre of coastline (*Ev*), based on square of longshore wind component (*Vn*), and estimate of average upwelling rate, assuming a 10-km wide upwelling strip, with reference to the time-series in Figure 8. Ekman drift calculated from $Ev = C_d \rho_{Air} V_N^2 / f$ at 34°S, where $C_d = 0.0013$, $\rho = 1.22$ kg·m^{-3} and f is the Coriolis parameter

Upwelling event	Duration (h)	Total Ekman volume ($m^3 \times 10^6$)	Average upwelling rate ($m \cdot day^{-1}$)
2	130	0.4195	8.1
3	545	3.7658	16.7
4	179	2.2171	29.7
5	87	1.0746	29.6
6	96	1.0341	25.8
7	125	0.3141	6.0
8	70	0.5119	17.5
9	42	0.1285	7.3
10	276	3.6976	32.1
11	117	0.5872	12.0
12	63	0.2647	10.1
13	120	0.5489	11.0

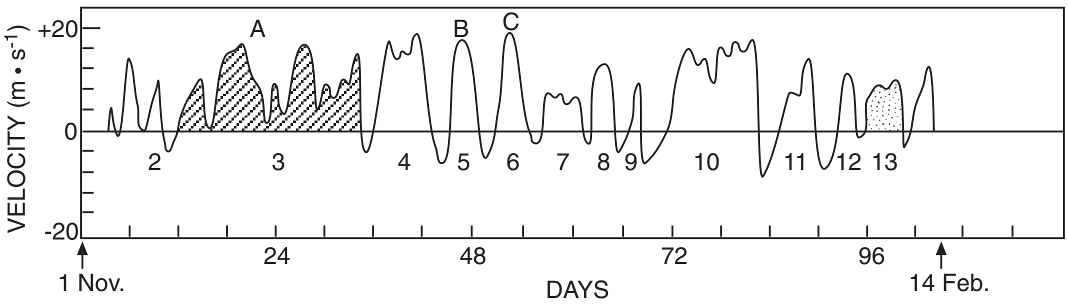

Fig. 8: Filtered longshore wind component (North is +) at Olifantsbos, 1 November 1995–14 February 1996. The hatched area represents a period of no wind reversals and the stippled area denoted Episode 13 of the series (see text). The numbers refer to the upwelling events listed in Table I

currents associated with coastal trapped waves are coincident at both sites, but that the correlation between wind and longshore current is not always in phase. Coastal-trapped waves could propagate freely with their own period from the northern site into the southern one, where locally forced and free waves would mix to introduce complexity into the longshore current. There is a secular change to lower air pressure towards summer with, in this case, a 9-hPa drop in average pressure between mid-October 1995 and early February 1996 at both sites.

During the filament cruise, there were two south-easterly wind events reaching speeds of up to 8 $m \cdot s^{-1}$, with the wind moderating but not reversing between them. The OSU-filtered longshore (north-south) wind component from 1 November 1995 to 14 February 1996 at Olifantsbos with half power at 2.5 days (Fig. 8) shows the synoptic scale variability with local thermal boundary effects removed.

DISTINGUISHING FEATURES OF THE STUDY AREA

What is the nature of this extreme southern Benguela system and how does it differ from other South-West African upwelling systems and from other eastern boundary systems? Nelson and Hutchings (1983) pointed to the significance of the southern extremity of the African continent lying in the belt of westerly winds in winter, and north of it in most summer seasons. Consequently, there is a powerful modulation of upwelling wind in summer as cyclones pass the southern land boundary, with typical periods of 3–6 days. Upwelling is intense but ephemeral. By contrast, upwelling at Lüderitz, the most productive area in the Benguela system, as in the California-Oregon system, appears to be more continuous within the upwelling season (Boyd 1987).

Table I gives estimates of the volume of upwelled water and upwelling rates over a nominal 10-km wide coastal strip for each upwelling-favourable event shown in Figure 8. These are the positive regions between zero crossings. The hatched area A in the figure is 545 h long and represents a merging of several 3–6-day episodes during which there was no wind reversal. This is clearly quite a different situation from the sharp peaks of B and C (87 and 96 h long respectively) or even smaller ones to the right, representing short, sharp injections of upwelled water favourable for filament generation. The biological consequences of the two types of event must differ widely. In principle, the greater the upwelling rate, the farther offshore the upwelling tongue will move. However, cross-shelf movement of water associated with shelf waves modifies the configuration. Events 1 and 14 are truncated by the filter and have been excluded. Peaks 2 and 3 have no zero crossing between them and represent one event.

A comparison of the estimated Ekman transport from longshore wind can be made to the upwelling observed in the satellite image. Taking Episode 13 in Figure 8 and in Table I, and an estimate of the coastline divergence length of 27.8 km, a total volume of upwelled water of 1.53×10^{10} m^3 is obtained for the six days of this active upwelling episode, ending between 7 and 8 February. A pixel count of the coldest water on the satellite image for 7 February (Fig. 2) gives an area of 5.79×10^8 m^2. An Ekman depth of 26 m would then make the observed tongue area compatible with the Ekman estimate. The uncertainties here are in the length of coastline and the fairly arbitrary temperature demarcation of the primary upwelling water, not calibrated in this image. Other unknowns are the heating and mixing of upwelled water around the perimeter of

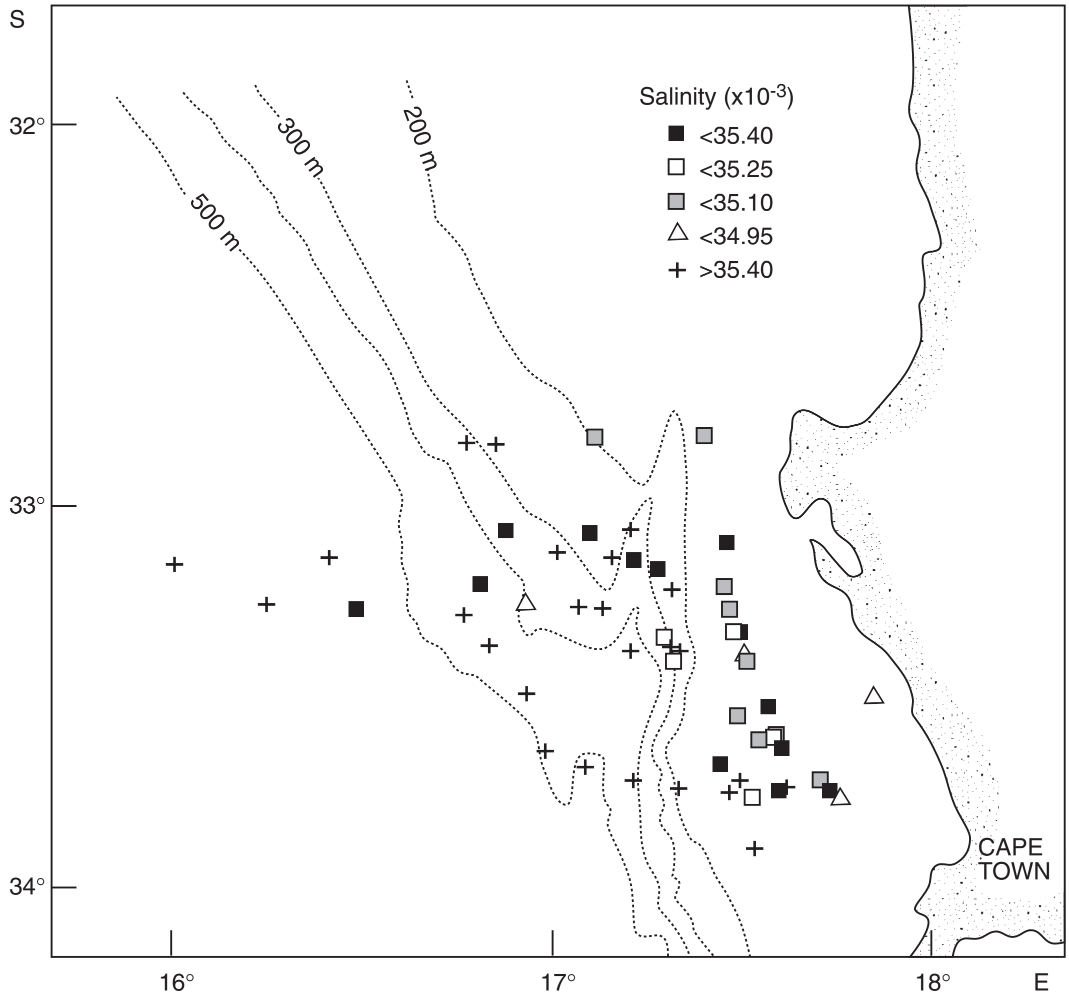

Fig. 9: Low salinity 100 m deep from historical CTD data in the vicinity of Eddy B (see Fig. 2)

the tongue. Nevertheless, the volumes are in reasonable agreement.

A second feature of the southern Benguela lies adjacent to different water masses. The advection of water onto the shelf can affect the SST signal and the interior dynamics markedly. Of particular significance here is water leaving the steep western side of the Agulhas Bank, a shallow plateau some 10^5 km^2 inshore of the 300-m isobath, and wedged between the Indian and Atlantic oceans. Although the TS characteristics of this water mass, perhaps more appropriately called the Agulhas Sea, does not differ significantly from Indian and Atlantic water on either side, it is generally warmer than the Benguela west coast water at any given depth, and must alter its potential vorticity when barotropic flow occurs over the western edge, appearing as a warm tongue curling onto the shelf of the West Coast. Recent examples of water crossing the western boundary of the Agulhas Bank into the Benguela system, as evidenced by thermal satellite imagery, were seen in mid-December 1996 and early January 1997. These episodes appeared to be more extensive than the February 1996 event, under discussion here. (J. Mantel, Sea Fisheries [SF], pers. comm.). The early 1996 event has received attention in the periodic egg and larval sampling surveys reported in Boyd and Nelson (1998).

A third feature is the advection of large-scale Agulhas rings northwards along the eastern deep boundary and their interaction with shelf waters (Duncombe Rae *et al.* 1992). These rings are estimated to average six per year (Gordon and Haxby 1990), usually moving north-west into the Cape Basin, but on occasion moving adjacent to the South-West African shelf.

A fourth feature is the secular variation in bottom pressure at Site P (see Fig. 1). A change of some 0.08 dB over a period of 100 days was observed at 980 m. At the time of the cruise, the pressure was 0.04 dB below average (T. Whitworth, Texas A&M University, pers. comm.). The variation was not compensated by atmospheric pressure, which dropped by 4 hPa, or a 0.04 dbar equivalent, in mean water pressure from winter to summer. Using satellite altimetry, Fu (1996) identified an annual 6-cm amplitude variation in sea-level change close to the tip of Africa and a semi-annual amplitude of 1–2 cm. Approximate 100-day mesoscale variability is confined to the Agulhas retroflection area near 40°S.

A reduction in bottom pressure in this region could cause water to flow from the Agulhas Bank, or increase the flux of the poleward undercurrent, so retarding the surface Benguela Current. The Agulhas Bank water with different density, displacing water west of Cape Town, would account for a part of the sea level change, but not the bottom pressure.

FACTORS AFFECTING THE FORMATION OF FILAMENTS NEAR CAPE TOWN

There is some evidence that the northern eddy (A, Fig. 2) is a semi-permanent feature. This would play a role in returning filamental water to the shelf, as appeared to be the case during the February 1996 cruise. What could be the reasons for its existence? The salinity depression measured at its centre was the strongest signal, showing a depression of the 35.40×10^{-3} isohaline at 100 m (Fig. 6). Historical CTD data were examined to establish the frequency of such an event. Salinity at 100 m was examined in an area bounded by 32.25–34°S and 1618°E. Excluding those points lying beyond the 500-m contour, where salinities in excess of 35.40×10^{-3} at 100 m may be expected, there was a tendency towards high salinity to occur in the vicinity of the canyon, and southwards onto the shelf (Fig. 9).

Historical current-meter data at 897 m (Nelson and Polito 1987) showed that water flows down the Cape Canyon (Site F in Fig. 1). A vector average of 2.9 km·day^{-1}, with a mean speed of 11.3 cm·s^{-1} and a maximum of 64 cm·s^{-1}, was attained. If this deep, poleward flow entrains water at the base of the eddy at 400–500 m, it will create a surface convergence with resulting anticyclonic flow. At 500 m in the canyon, and for some distance along its axis, the slope is 1:100. At a flow rate of 10 cm·s^{-1}, water would be removed from the base of the eddy at the interface with the poleward flow at a rate of 10^{-1} cm·s^{-1} over the canyon width of some 20 km, close to the diameter of the eddy. The process is equivalent to vortex stretching.

An alternative explanation is that there is an unstable shelf-edge jet and that, in its meandering, counter-rotating eddies are formed on either side of it. Shannon and Nelson (1996) proposed mechanisms for the generation of the observed jet, alternative to geostrophic flow at an upwelling front. Not all SF ADCP data obtained from survey cruises reveal continuity between the Cape Peninsula jet and the Cape Columbine jet. This was indeed one of the motivations for the cruise. Water seems to have a north-west orientation west of Dassen Island (Boyd and Oberholster 1994). However, several drogue studies have shown this continuity (Shannon and Nelson 1996).

Can shelf waves in any way influence the formation of filaments? An accelerating wind causes both upwelling and coastal-trapped waves. The longshore and cross-shelf motions in these waves can be distinguished in low-pass filtered data. Historical current-meter data, particularly at Site C (Fig. 1), show a westward movement of water, which would play an important role in carrying upwelled water offshore, assisting the surface Ekman drift. This record runs from 1 July to 14 September 1981. Filtering the series (OSU filter, Beardsley *et al.* 1985) and then integrating the west-flowing parts, 14 episodes are obtained, which divide into nine of small displacements <8 km and five significant events of 27.7, 25.3, 17.6, 26.6 and 18.0 km. The last two corresponded to early spring south-easterly wind events. These distances are large enough to carry water into the vicinity of the shelf edge, the Ekman transport in the top friction layer riding on the barotropic cross-shelf movement.

The effect of a wave propagating into the Cape Peninsula area from its generation site at Cape Columbine with active upwelling taking place would be to move the upwelling filament offshore. As the wave passes and its phase reverses, the effect would be to move the filament back onshore. However, if upwelling along the Cape Peninsula was initiated on the reverse phase of a coastal-trapped wave generated remotely as it moved through the area, the filament would be carried onshore (Fig. 10).

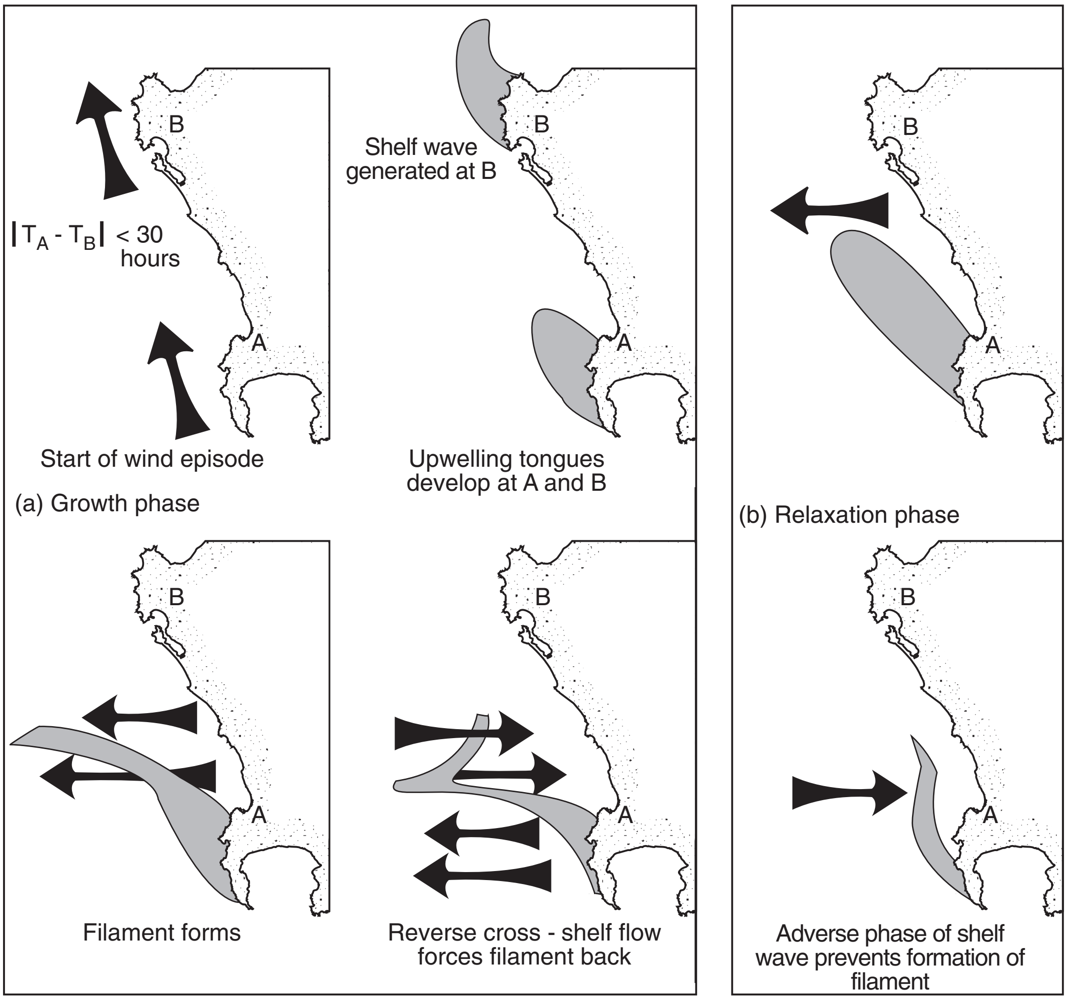

Fig. 10: Schematic diagram showing the effect of shelf waves on an upwelling tongue during (a) the growth and (b) the relaxation phase

CONCLUSION

One of the unexpected results of the cruise measurements was the absence of a strong shelfedge jet off the Cape Peninsula. The area over the 300–400-m isobaths is often occupied by a strong, equatorward-flowing baroclinic jet, first identified by Bang and Andrews (1974), who assumed it to be an upwelling frontal formation. Only a weak, equatorward baroclinic structure was found, much farther inshore than usual (Fig. 3). A search through unpublished ADCP data, together with the results of Boyd *et al.* (1992), has shown that the jet is variable and can occur outside the upwelling season. For example, a jet attaining a flow of 35 $cm \cdot s^{-1}$ was found at a depth of between 20 and 60 m in May 1992, during a period of northwesterly winds, conditions unfavourable to upwelling. Shillington *et. al*. (1992) observed the jet flowing strongly in March 1990.

The results of this study have raised more questions than were answered, but these questions provide pointers to further research. What is a normal situation for the flow-field on the southern Benguela shelf? Was this an anomalous period because a pool of warm water intruded on the shelf zone, and if so, how frequently does it occur? Was it influenced by a drop in the pressure field and did this also cause an acceleration in poleward flow? Further studies of the shelf-edge jet are required to understand its generating mechanism and whether it meanders and generates counter-rotating eddies, as in the California current.

From a biological point of view, the results and consequent questions raised by this investigation point to the need to break away from simple conceptual models and to accept that a high level of complexity exists in fish recruitment, not only in the biological processes themselves, but also in the influence of the physical environment. Passive transport by currents is one such example of the complex processes influencing survival. Fowler and Boyd (1998) and Huggett *et al*. (1998) demonstrate some of this complexity.

It is not enough simply to relate "wind run" to recruitment in a phenomenological and linear manner. There are myriad possibilities for water movement. A particular configuration of tongue, filament and eddies has little chance of ever being repeated in fine detail, implying that there is much diversity in the way that the environment can affect survival and recruitment of fish.

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