Observations of Frontal Instabilities on an Upwelling Filament

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ABSTRACT

Frontal instabilities were observed along a density front on the cyclonic boundary of an upwelling filament that formed north of Pt. Arguello, California in October 1983. Observations of the instabilities were conducted using satellite sea surface temperature images and in situ sampling. The instabilities formed on the southern (cyclonic) boundary of the filament at a wavelength of about 20 km and consisted of two lobes, one warm and one cool, each with a width of about 4 km. The time scale for formation of the instabilities is about 1 day. Near-surface distributions of temperature, salinity, and density within the cool lobes of the instabilities are consistent with local upwelling at the rate of about 30 m d^{-1}. A simple model based on conservation of potential vorticity is presented, which accounts for the observed upwelling. Based on isopycnal displacements and the distribution of salinity, the signature of the instabilities appears to be confined to the upper 50 m of the water column.

1. Introduction

Upwelling filaments are commonly observed features along the west coast of the North America and are important mechanisms for transporting coastally upwelled water far offshore beyond the continental shelf. The vertical structure of these filaments is that of a surface intensified, baroclinic jet (Flament et al. 1985; Reinecker et al. 1985) with maximum offshore surface velocities exceeding 0.5 m s^{-1} (Kosro 1987; Huyer and Kosro 1987). Velocity surveys employing surface drifters (Davis 1985) and Doppler acoustic current profilers (Kosro 1987) with contemporaneous satellite sea surface temperature (SST) images indicate that these high offshore surface velocities often occur in association with cool filaments extending offshore up to 200 km. The time scale of these filaments, based mainly upon SST imagery, appears to be on the order of a few to several weeks, although the details of their growth and decay are largely unknown at present. Recent laboratory experiments in a rotating tank have shown that plumelike features resembling upwelling filaments can form in association with bottom topography such as a ridge (Narimousa and Maxworthy 1985). Other experiments (Narimousa and Maxworthy 1987) have shown that plumelike features can also result from baroclinic instability of an upwelling front in the absence of bottom topography. These features in the latter experiments were nonstationary and occurred randomly along the upwelling front.

To date, the occurrence of these filaments has been most completely documented along the California coast north of San Francisco, particularly in the CODE region. However, they are not limited to this portion of the coast. Temperature structures resembling upwelling filaments often form off central California. For example, an irregular region of relatively cool surface water (about 2°C cooler than surrounding water) extending about 80 km southward from Pt. Conception is present in the 12 October 1984 image reported by Chelton et al. (1987). Another SST image obtained by Sheres and Kenyon (1987) from 14 May 1984 shows a cool region resembling a filament that extends at least 100 km offshore just north of Pt. Arguello.

This paper presents observations of an upwelling filament that formed off the central California coast during October 1983. From the SST imagery, the filament appears to be rooted north of Pt. Arguello in roughly the same coastal region as the filament in the image of Sheres and Kenyon (1987). A sequence of SST images was obtained, which shows the growth and decay of the surface temperature field of the filament over a two week period. In situ measurements show that the offshore extent and volume transport of this filament are similar to those of filaments off the northern California coast.

A primary objective of the in situ sampling of this experiment was to observe the formation and evolution

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of transient, coherent structures that often form in association with upwelling filaments but are also observed in oceanic flows in general. Of particular interest are instabilities that are observed on frontal boundaries and that may be important in the dissipation of kinetic energy and in the horizontal exchange of heat, mass, and momentum in the upper ocean. During the experiment described in this paper, near-surface tow-yo transects through the upper 100 m of the water column were conducted across a pair of frontal instabilities, which were observed on the cyclonic boundary of the filament. The term instability is used here to characterize these features because their amplitudes grow in time and their evolution and development are at least superficially similar to shear instabilities observed in laboratory flows. Comparable images of shear instabilities were reported by Flament and Armi (1985). Frontal eddies or "shingles" have been observed on larger scales (100 to 200 km) along the cyclonic front of the Gulf Stream (Lee et al. 1981; Lee 1975; Lee and Mayer 1977; Lee and Atkinson 1983). Cyclonic eddies observed by Kennelly et al. (1985) on the boundary of a warm-core ring also appear to be similar to the instabilities observed on this filament, although the eddies which they observed were larger both horizontally and vertically and had longer lifetimes.

A major difficulty in sampling these transient features is that their appearance cannot be assured for a particular experiment nor can their location be predicted. Therefore, considerable flexibility must be incorporated into the experimental design if they are to be observed by methods other than satellite SST imagery. Real-time direction of shipboard sampling is also an essential element of the observing plan. The details of the experiment and instrumentation are given in section 2. A line of CTD stations across the filament is used to examine the water mass characteristics and to compute a section of the geostrophic velocity field across the filament. These observations of the large scale structure of the filament are discussed in section 3. Sections 4 and 5 deal with the SST imagery of the instability evolution and with the tow-yo transects across the instabilities. Simple conceptual models to explain some of the details of the near-surface finestructure and property distributions are also given.

2. Experimental description and data processing

In situ CTD data of this paper were collected northeast of Pt. Arguello, California on 22 and 23 October 1983. In addition, satellite SST images of this region were obtained over a two week period from 11 to 25 October 1983 and are used to examine the evolution and decay of the filament. During the sea-going portion of this experiment these images were received at the Scripps satellite facility and transmitted to the R/V *Ellen B. Scripps* via radiotelephone and telecopiers and used to direct the in situ sampling.

The instrument platform carries a Neil Brown Instrument Systems CTD and consists of a section of steel channel, which has a large fin protruding rearward to maintain the orientation of the CTD sensors into the flow as the platform is moved through the water. This platform has been used on several cruises and is stable at the low tow speeds of this experiment; a drawing of the platform is given by Washburn and Deaton (1986). Sections of temperature (T), salinity (S), and potential density anomaly (σ_θ) are obtained from "tow-yos" by winching the instrument package vertically at about 0.6 m s^{-1} with the ship underway at about 1.5 m s^{-1}. Continuous sea surface water properties are obtained by pumping seawater from a depth of about 2 m through a small chamber on deck which contains a Neil Brown conductivity cell and a platinum resistance thermometer. However, during this experiment, excessive bubble formation in the chamber resulted in unusable surface salinity and density data; the surface temperature data is unaffected by this problem. The transit time of the water through the pumping system is sufficiently rapid such that no measurable heating or cooling of the water occurred. During the in situ sampling the wind speed was typically about 10 m s^{-1} from 310° (along coast and upwelling favorable) as shown in the upper plot of Fig. 1. The surface mixed layer was consistently about 30 m deep, so both the 2 m pumped temperature data and the satellite SST images give representative temperatures for the surface layer. Wind data of Fig. 1 are from the ship's log in which wind speed and direction are recorded every 2 h.

Conductivity, temperature, and pressure are recorded at 32 Hz and logged on nine-track tape at sea. Absolute calibration of salinity is based upon in situ bottle sampling while temperature calibration is based upon measurements from reversing thermometers and pre-cruise/post-cruise laboratory calibration. The results of this paper are mainly dependent on relative changes in temperature and salinity; significant relative changes for temperature and salinity are conservatively estimated to be 0.01°C and 0.001, respectively. During post-cruise processing conductivity data are filtered with a 1-pole low-pass digital filter with an effective analog time constant of 0.1 s. This time constant was found to minimize salinity spiking in most sections of the data records, although some residual spikes remain. Because of the constantly varying platform speed, a single value of this time constant will not be ideal for all of the data. After filtering the conductivity data, one-second running means of temperature, conductivity, and pressure are formed and subsampled to a point about every 0.5 s. These running means are used to compute salinity and density. Further smoothing is obtained by averaging all variables into "pressure bins" of 2 db. Erroneously large and small values of salinity and density due to residual spiking are detected during processing and are not included in the 2 db averages.

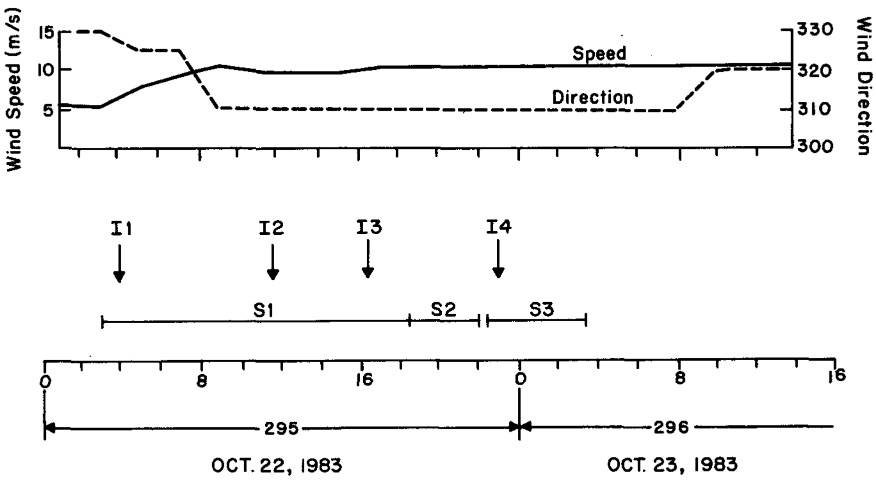


FIG. 1. Upper panel shows time series of wind speed (solid line) and direction (dashed line) from ship's log. Lower panel scale indicates time in Julian days with tick marks every 2 h. Bars above scale show time intervals of CTD survey (S1) and tow-yo surveys (S2 and S3). Arrows above survey bars labeled I1, I2, I3 and I4 give times of SST images shown in Figs. 5a–d.

Vertical gradients of temperature, salinity, and density are obtained from a centered, first difference between adjacent 2 db averages of these quantities.

Ship's position from a Loran receiver is logged continuously throughout the experiment every 20 s and is used to convert time to distance for the production of two dimensional maps of property sections. Relative distance along a particular ship track as a function of time is obtained from smoothed values of latitude and longitude and is interpolated from a piecewise linear fit to the position data. Occasional "dropouts" are found in the Loran data and during these times position is reconstructed by linear interpolation. Dropouts occur in less that 5% of the position data.

3. Large scale development and structure of filament

A sequence of four satellite infrared images of the distribution of SST covering the area northwest of Pt. Arguello over a two week period is shown in Figs. 2a–d. Dark areas indicate relatively warm waters and light areas are cooler waters. Maximum SST differences in the images are about 4°C. Spacing between tick marks in the images is one degree in both latitude and longitude. The first image in the sequence, Fig. 2a, is from 2330 UTC 11 October (JD 284) and shows cold water associated with the upwelling center located between Pt. Conception (identified as PC in Fig. 2c) and Pt. Arguello (PA). The upwelling center evident in this image is a recurring feature of this region of the California coast and is similar in extent to that observed by Atkinson et al. (1986) during the OPUS experiment. A broad region of relatively warmer surface water may be seen offshore from the upwelling center; this warm water extends northward past Pt. Buchon (PBu) and is suggestive of a northward coastal flow. Other observations discussed by Chelton et al. (1987) indicate that the warm, northward flowing "Davidson Current" is seasonally present along this section of the coast beginning in September or October. The broad light colored area along the western boundary of the image is a layer of clouds.

A very different pattern of SST is seen in the second image of the sequence, Fig. 2b, taken at 2217 UTC 17 October (JD 290). The upwelling center at Pt. Conception is absent in this image with warmer surface waters replacing the cooler upwelled waters at the coastline. The offshore region of warmer water, visible in Fig. 2a, is still present but is bisected by the developing filament of colder water that extends offshore along 34°50′. In the third image taken at 1137 UTC 22 October (JD 295) the SST expression of the filament extends about 120 km offshore with a width of about 50 km. The boundaries of the filament are most clearly defined in this image compared with any of the other cloud-free images obtained over the two-week period of these observations. The CTD tow-yo surveys of the filament discussed in section 4 were conducted over a 25-h period, which includes the time when this image was taken. Clouds are found in the image along the coast immediately north of Pt. Conception and around the Channel Islands in the lower right hand corner.

In the final image of the sequence taken at 1514 UTC 25 October the well-defined filament structure of the previous image has been replaced with irregular elongated patches of cool water. A broad region of cool surface water, present at the coast north of Pt. Buchon in Fig. 2c, has extended farther south in Fig. 2d. Based on the SST expression, the time scale for formation and decay of this filament is about two weeks, although the velocity field associated with the feature may have a different time scale.

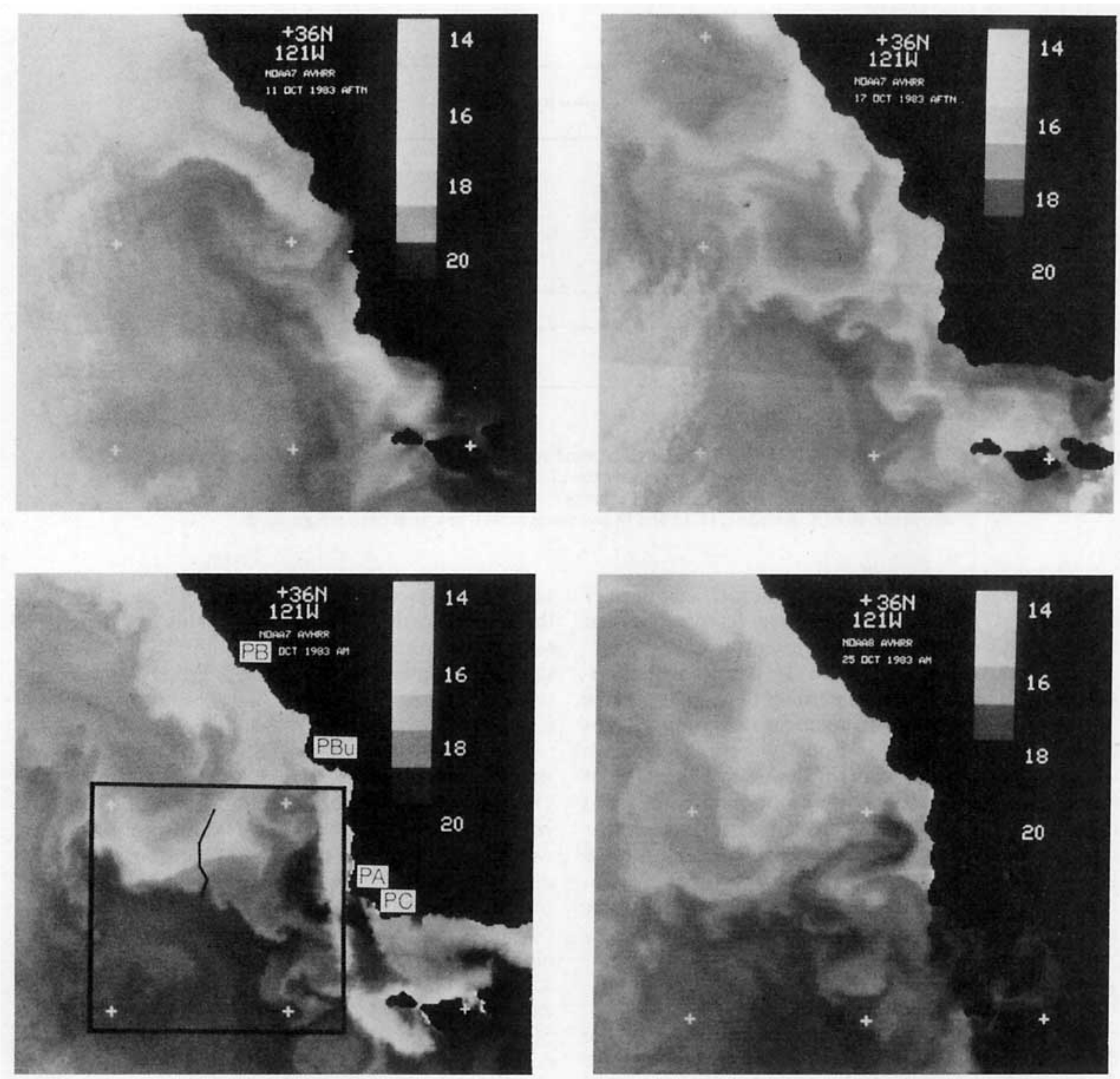
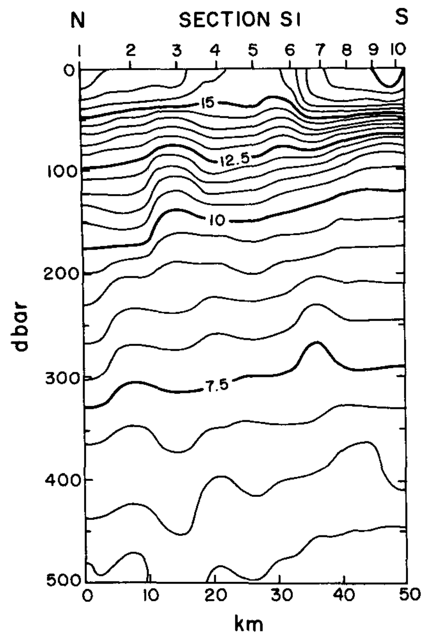
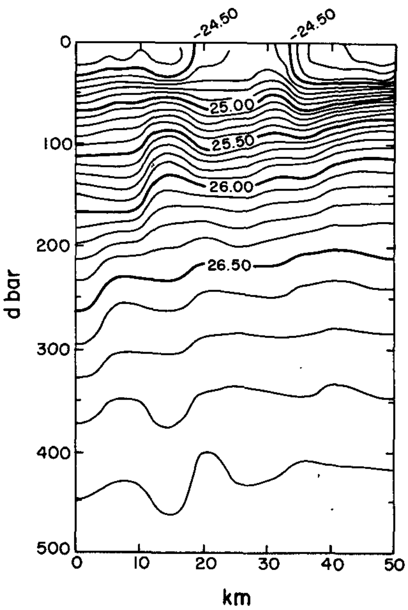


FIG. 2. (a: top left) Satellite SST image of area around Pt. Arguello taken at 2330 UTC 11 October 1983 (JD 284). Light areas indicate relatively cool water and dark areas indicate warmer water. The point 36°N, 121°W is identified at the top of the image. Spacing between tick marks is 1° in both latitude and longitude. (b: top right) As in (a) but image taken at 2217 UTC 17 October 1983 (JD 290). (c: lower left) As in (a) but image taken at 1137 UTC 22 October 1983 (JD 295). PC is Pt. Conception, PA is Pt. Arguello, PBu is Pt. Buchon, and PB is Piedras Blancas. Ship track across filament for Survey S1 is indicated with a black curve. Area covered by images in Fig. 5 is indicated with black rectangle. (d: lower right) As in (a) but image taken at 1514 UTC 25 October 1983 (JD 298).

In order to examine the subsurface distribution of temperature, salinity, and density across the filament, a section of CTD stations was obtained along the ship track shown superimposed on the SST image of Fig. 2c. A time line that summarizes the sequence of the CTD stations, tow-yo sections, satellite imagery, and wind conditions for the time period 0000 UTC 22 October to 1600 UTC 23 October 1983 is given in Fig. 1. The time interval during which the section of CTD stations was obtained is labeled S1 in Fig. 1 and extends from 0301 to 1832 UTC 22 October 1983 (Julian Day, JD 295). The arrow above the time line labeled I2 marks the time when the SST image of Fig. 2c was taken; this occurs about midway through the time interval S1. The entire sequence of images I1–I4 with an average spacing of about 6 h is discussed in section 4. The intended ship track was a straight line between the end points of the actual track shown in Fig. 2c;





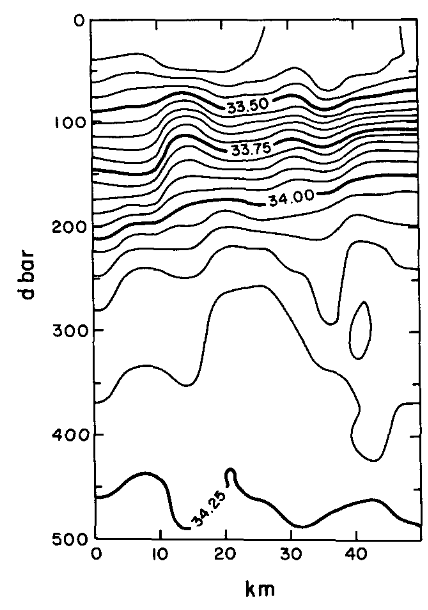


FIG. 3. (a: left) Map of isotherms to 500 db for 10 CTD stations of Survey S1. Contour interval is 0.5°C. Station locations are shown along top edge of the figure. North is to the left and the x-axis gives relative distance along the ship track. (b: middle) As in (a) but for salinity. Contour interval is 0.05 practical salinity units. (c: right) As in (a) but for σ_θ. Contour interval is 0.10 kg m^{-3}.

however, the ship was apparently set to the west by the near-surface velocity field of the filament.

The maximum sampling depth of each CTD station is about 500 m with an average spacing of 5.5 km. A contour section of temperature to 500 db is given in Fig. 3a with a contour interval of 0.5°C. Tick marks and numbers along the top of the figure identify the location of the 10 CTD stations. North is to the left and the x-axis gives relative distance along the ship track. Surface temperatures decrease over the first half of the section and reach a minimum of about 15.25°C at Station 5. Farther along the section, surface temperatures rise steadily reaching a maximum of 18.2°C in the warm water just south of the filament. A consistent tilting of the isotherms is evident below about 50 db with isothermal surfaces shoaling to the south. The 8°C isotherm, for example, rises by about 60 m over the 50 km width of the section giving an average isotherm slope of about 10^{-3}. Salinity contoured along the same section with an isohaline spacing of 0.05 is shown in Fig. 3b. At the surface, salinity increases monotonically from north to south reaching a maximum value of about 33.41 at Station 10. Below 250 db vertical salinity gradients are generally low within the filament, although a weak salinity maximum is present in Station 8 at about 300 db. Overall, the filament waters are free of obvious thermohaline intrusions above 500 db and are diffusively stable in the mean.

Potential density anomaly σ_θ is contoured in Fig. 3c for the section with a contour interval of 0.05 kg m^{-3}. The map is very similar to the temperature map of Fig. 3a because the filament waters are thermally stratified. As with temperature, a consistent tilt of the isopycnals across the section is evident below 50 db. An upward doming of isopycnal surfaces is coherent vertically over about 100 db and may be seen centered at 120 db at about 14 km into the section.

Density profiles along the line of CTD stations are used to construct contours of the geostrophic flow field referenced to 480 db across this part of the filament. To produce a smoothed estimate of geostrophic velocity relative to 480 db at a given station, profiles of geopotential anomaly from stations on either side are differenced. The spacing between stations used in computing geostrophic velocity is typically about 10 km. Examples of these profiles from three stations across the filament are shown in Fig. 4a. Positive velocity indicates onshore flow. A profile is identified in Fig. 4a by the stations that are used to produce it; profile 1–3, for example, is obtained from geopotential anomaly profiles at Stations 1 and 3. Maximum relative current velocities of -0.38 m s^{-1} are found for this profile at around 60 db and the vertical velocity structure of the filament is that of an offshore jet with velocity increasing in the upper water column. The velocity with respect to 480 db increases to zero at around 300 db and is positive everywhere below. The second profile 5–7 in Fig. 4a has roughly the same shape as profile 1–3 except that the velocity maximum is only about -0.16 m s^{-1} and the flow with respect to 480 db is offshore throughout the profile. The third profile 8–10 located

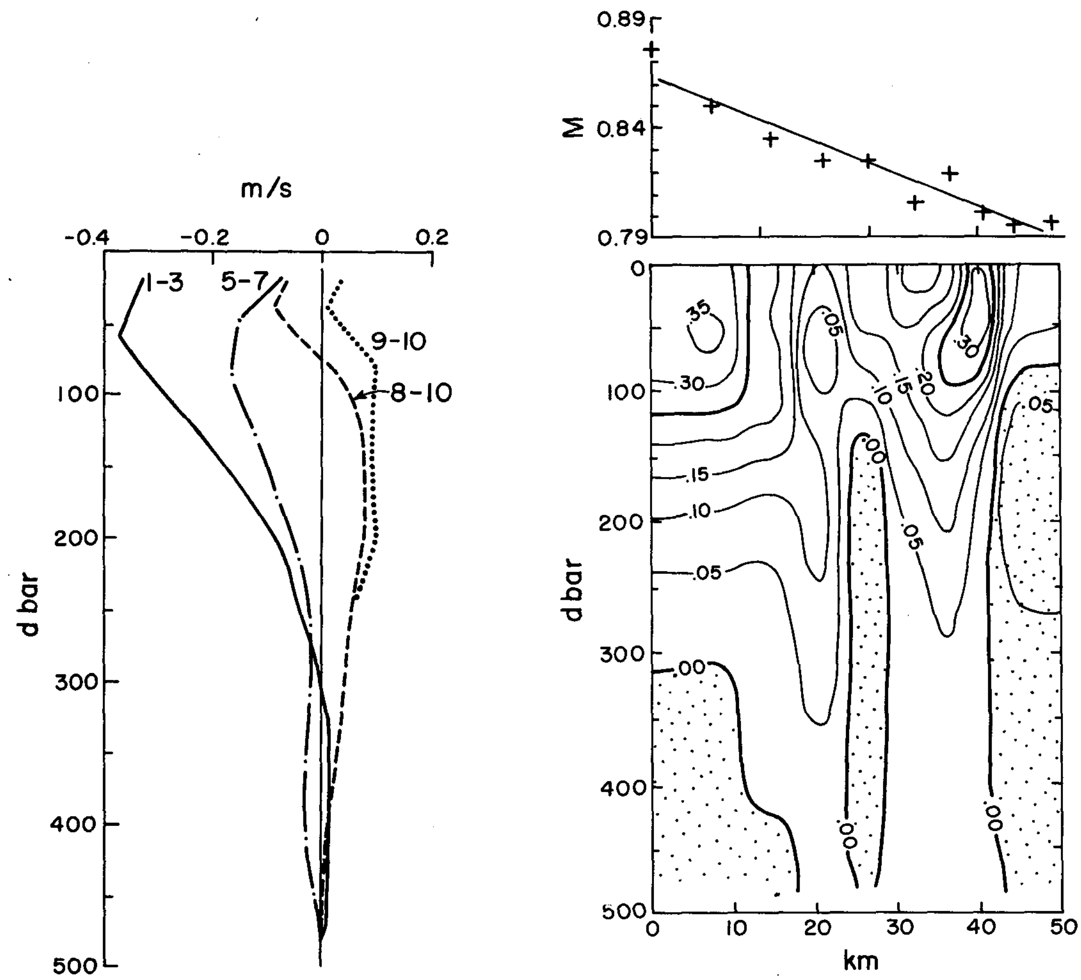


FIG. 4. (a) Profiles of geostrophic velocity computed with respect to 480 db between various stations of Survey S1. Labels identify stations used in computing profiles. Positive velocities indicate onshore flow. (b) Upper panel gives dynamic height in m for 0 db with respect to 480 db for the 10 stations of Survey S1. Individual values are indicated with plus (+) signs and the solid line is a least squares fit. Lower panel is a map of isotachs of geostrophic velocity for Survey S1 in m s^{-1} computed with respect to 480 db. Stippled areas indicate onshore flow.

at the SST front on the southern end of the line of CTD stations indicates an onshore return flow everywhere below 70 db with maximum velocities of about 0.05 m s^{-1}. Onshore flow probably extends all the way to the surface south of the SST front as suggested by a velocity profile computed between Stations 9 and 10, which are more representative of conditions south of the filament. A portion of this profile to 250 db is shown with a dotted line in Fig. 4a and indicates onshore flow everywhere above 480 db. In Figs. 3a and 3b it can be seen that the mean slope of isotherms and isopycnals is smaller between these two stations consistent with a weakening offshore geostrophic flow at the filament boundary. It should be pointed out, however, that the distance between Stations 9 and 10 is only about 4.5 km; at small scales internal waves can contribute as much or more to local isopycnal gradients as the larger scale geostrophic flow.

The geostrophic velocity contours of Fig. 4b, which are computed from profiles as described above, show a near surface maximum at the northern end of the section and indicate that the velocity field of the filament was not completely crossed in this section. Another more narrow surface velocity maximum is found near the southern end of the cross section at about 40 km. The maximum relative velocity here is about 0.30 m s^{-1} and is just north of the boundary between the offshore flow of the filament and the weak onshore return flow to the south. Onshore flow in Fig. 4 is indicated by stippling and maximum onshore current velocities which exceed 0.05 m s^{-1} are found at 200 db in the region of return flow south of the filament. This boundary between offshore and onshore flow also corresponds to the temperature front in the SST image of Fig. 2c near the southern end of the ship track.

The total geostrophic transport (relative to 480 db) through the section is offshore with a magnitude of about 1.7×10^6 m^3 s^{-1}. This is an underestimate of the total volume transport because the entire velocity field of the filament was not crossed in this survey and the actual depth of the velocity field probably exceeds the reference level of 480 db. A deeper penetration of the geostrophic velocity field is indicated in Fig. 3c by the slope of the deepest isopycnals: the $\sigma_\theta = 26.80$ surface, for example, has an average slope of about 0.6×10^{-3} across the section. A profile of dynamic height (in dynamic meters) across the section is given at the top of Fig. 4b with individual station values indicated by plus signs. The difference in elevation between the sea surface and the 480 db surface is about 0.08 m across this part of the filament and a least-square fit, indicated by the solid line, yields a sea surface slope with respect to 480 db of 1.5×10^{-6}.

Filaments such as the one observed here are likely to be important in exchanging water masses over the continental shelf along the central California coast. This filament is rooted in a section of the coast between Piedras Blancas to the north (PB in Fig. 2c) and Pt. Arguello to the south, a distance of 133 km. The 500 m isobath lies offshore at an average distance of approximately 26 km in this region and the volume shoreward of this line is roughly 9×10^{11} m^3. Based on the observed offshore geostrophic transport with respect to 480 db, this volume would be exchanged in about 6 days if the filament simply moved the coastal water off shore. This estimate of an exchange time scale is less than the lifetime of the filament derived from the SST imagery of Fig. 2. Therefore, the filament persists long enough for such an exchange to occur.

The order of magnitude estimate from the simple exchange model given above indicates that extensive volumes of coastal waters may be affected by the occurrence of these transient filaments. However, the nearshore velocity fields of the filaments may be very complex with the exchange being part of a larger scale meandering current system. Such a meandering current system with associated filaments was observed by Ikeda et al. (1984) off Vancouver Island.

4. Instability development and frontal structure

a. SST imagery of instabilities

Following the section of CTD stations across the filament, two high resolution tow-yo surveys were conducted through a pair of instabilities, which developed on the southern (cyclonic) front of the filament. During this portion of the experiment, a sequence of four relatively cloud free images was obtained over 19 h which covers the period of development of the instabilities. The times of these images relative to the time intervals of the line of CTD stations (S1) and the tow-yo surveys (S2 and S3) are indicated in Fig. 1 by arrows labeled I1 through I4. The first image I1 of the sequence is shown in Fig. 5a and was taken on 22 October (JD 295) at 0357 UTC. The scale of this image is expanded by a factor of four compared with Figs. 2a–d and the size of individual pixels (1 km) is evident. The instabilities evolve on the frontal feature which stretches from the southeast to the northwest in the image and the positions of the developing instabilities are indicated by arrows numbered 1 and 2.

The second image I2 of the sequence, Fig. 5b, shows the same region at 1137, some 7.5 h after Fig. 5a. This image is a subsection of the image of Fig. 2c and is indicated in that figure by a rectangle. The initial instability shown by arrow 2 in Fig. 5a has evolved into a cusplike feature in this image. The feature clearly shows two lobes, one warm and one cool, which appear to wrap around each other in a cyclonic rotation. A second feature that is similar may be seen on this same front farther to the southeast, although this latter feature was not examined in our surveys. Cool surface water may also be seen penetrating northward on the eastern edge of the initial instability indicated by arrow 1; this region of the front is found between arrows 1

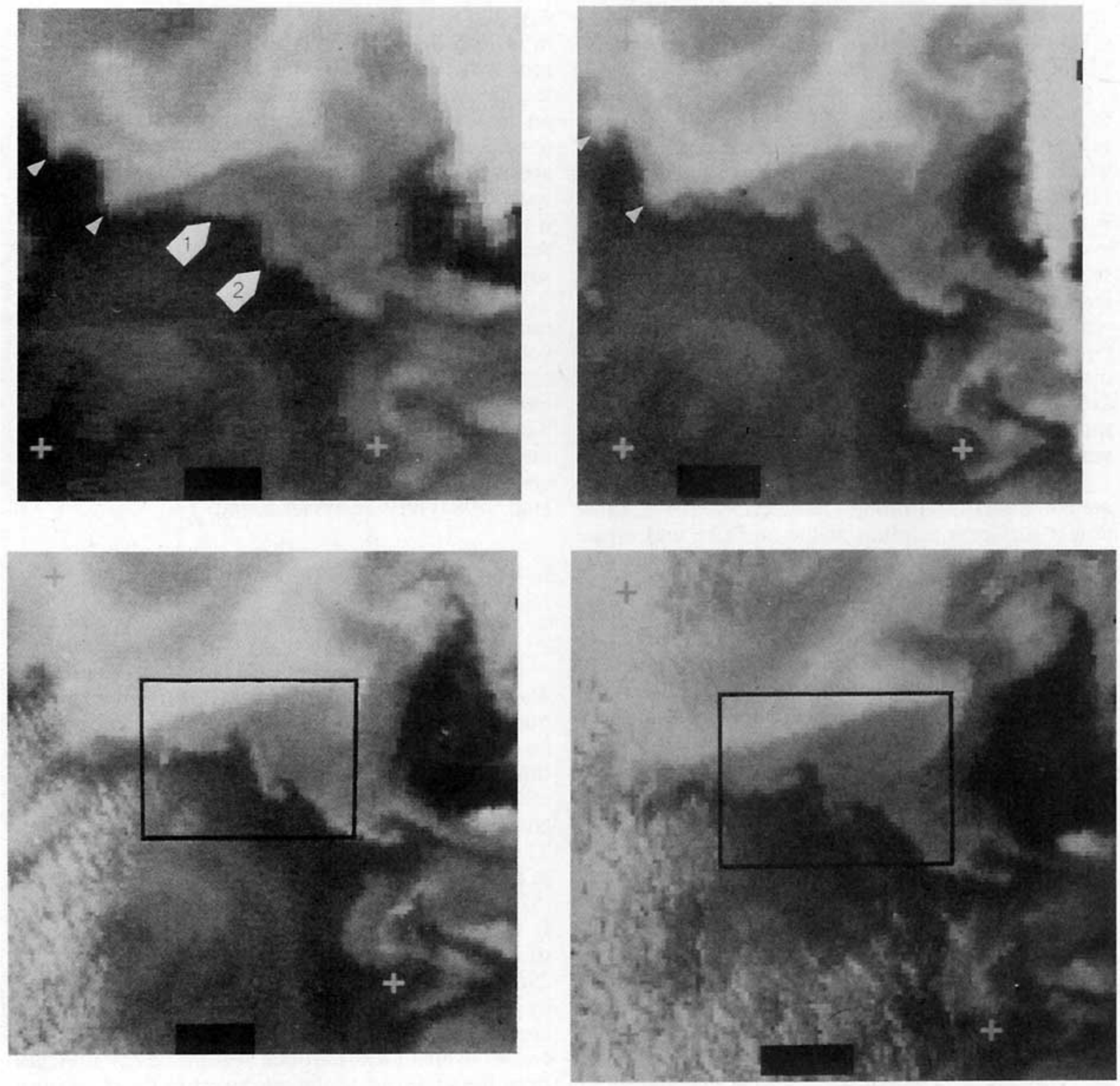


FIG. 5. (a: upper left) SST Image I1 of area enclosed by rectangle in Fig. 2c taken at 0357 UTC 22 October 1983 (JD 295). Shading as in Fig. 2a. Arrows 1 and 2 indicate developing frontal instabilities. (b: upper right) As in Fig. (a) but for Image I2 taken at 1137. (c: lower left) As in Fig. (b) but for Image I3 taken at 1619. Black rectangle encloses area shown in schematic of Fig. 6a. (d: lower right) As in Fig. (a) but for Image I4 taken at 2258. Black rectangle encloses area shown in schematic of Fig. 6b. Scale of this image is slightly larger than that of Figs. 5a–c.

and 2 in Fig. 5a. Northwestward propagation of some features of the filament may be seen through careful comparison of Figs. 5a and 5b. Two southward extensions of the filament, which are indicated with small arrows along the same front as the instabilities, clearly have a northwestward displacement in Fig. 5b compared to Fig. 5a. The magnitude of this displacement is about 7 to 9 km over the 7.5 h period. This implies surface velocities between 0.26 and 0.34 m s^{-1}, comparable to geostrophic current velocities relative to 480 db along the ship track of Fig. 2c.

The third image I3 of the sequence, shown in Fig. 5c, was taken at 1619, about 5 h after the image of Fig. 5b. At this point, the cool lobe of Instability 2 is more clearly defined while Instability 1 has evolved into a cusplike feature similar to Instability 2. The spacing of these two instabilities along the front is in the range 15 to 20 km and the cross-frontal extent of the warm and cool lobes of Instability 2 is about 8 km. A layer of clouds may be seen intruding into the image from the southwest.

The final image of the sequence I4 taken at 2258, some 6.5 h after the preceding image, is shown in Fig. 5d. Now the warm and cool lobes of Instability 1 have clearly evolved and the entire structure appears to have rotated in a cyclonic sense. An approximate estimate of the cyclonic rotation of Instability 2 is about 14° between images I3 and I4. This implies a rotation rate of about 1×10^{-5} s^{-1} or about $0.1f$ where f is the Coriolis parameter ($f = 8.2 \times 10^{-5}$ s^{-1} at this latitude). The cross-frontal extent of Instability 2 has increased somewhat at this point and is now in the range of 9 to 10 km. Clouds obscure much of the sea surface along the western and southern boundaries of the image. Subsequent images over the next several hours are entirely covered with clouds, which preclude any further examination of the evolution of the instabilities. The third feature, mentioned above, along this front, which resembles Instabilities 1 and 2 and which lies to the southeast, appears to be stretching in the alongfront direction with very little rotation.

The tracks of the research vessel followed during the two high resolution tow-yo surveys are shown in Figs. 6a and 6b. The schematic charts of Fig. 6 also show the approximate locations of the 17.6°C isotherm at the sea surface obtained from the satellite images of Figs. 5c and 5d. This isotherm is chosen because it marks the boundary between the filament and the warmer waters to the south and outlines the instabilities. The latitudes and longitudes which form the rectangles are labeled in Fig. 6a and a scale is included for reference. The boundaries of these diagrams are indicated by rectangles in Figs. 5c and 5d. Note that the scale of Fig. 5d is slightly larger than the scale of the other images in Fig. 5. The first track (S2), indicated with a solid line in Fig. 6a, began about two hours after the image I3 was taken at 1619 as shown in Fig. 1. This track is 16.5 km long and took about 3.5 h to complete. A turn to the northeast was executed at the beginning of this track in order to intercept the warm lobe and to position the ship for a southward track through the second instability. The second tow-yo survey (S3), indicated by a solid line in Fig. 6b, was underway when the image I4 was taken at 2258; the ship's position at this time is indicated by an arrow to the right of the ship track. The length of this track is about 28 km and took 5.2 h to complete.

b. Property distributions along ship track S2

A horizontal profile of sea surface temperature along track S2 is shown in Fig. 7a and is obtained from the CTD, which is "tow-yoed" between about 5 and 110 db, typically. The right-hand side of the figure is to the northeast. A schematic diagram of a typical section of the "saw tooth" path followed by the CTD through

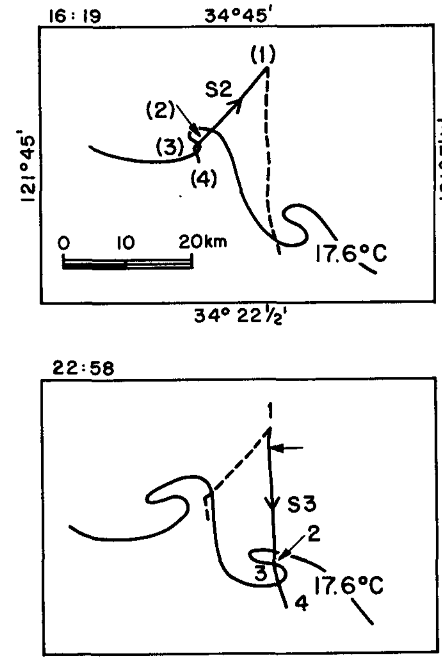


FIG. 6. (a) Schematic diagram showing ship tracks S2 (solid line) and S3 (dashed line) in relation to instabilities at 1619 UTC 22 October 1983. Instabilities are outlined by the 17.6°C isotherm which is shown with a solid line. Points (1) through (4) are also indicated in the T, S, and σ_θ profiles of Figs. 7a and 9a. Position of area shown here is indicated by rectangle in satellite SST image of Fig. 5c. (b) As in (a) but at 2258 UTC 22 October 1983. Points 1 through 4 are also indicated in the T, S, σ_θ profiles of Figs. 10a and 11a. Position of area shown here is indicated by rectangle in SST image of Fig. 5d. Arrow to right of ship track S3 indicates position of research vessel when image shown in Fig. 5d was taken.

the water column is shown in Fig. 8. Points which make up the SST curve (solid line) of Fig. 7a are obtained by averaging together all temperatures that lie above 15 db for each near-surface approach of the CTD; this yields a sea surface temperature about every 500 m along the ship track as indicated in Fig. 8. The mixed layer depth is typically 20 to 30 m along these sections so the average temperature above 15 db is representative of the mixed layer. Continuous temperature from 2 m depth measured on deck as described in section 2 is shown with a dashed line in Fig. 7a. The agreement between these horizontal temperature profiles indicates that sea-surface water properties may be obtained from the CTD used in tow-yo profiling. The difficulties with bubble formation in the continuous 2 m pumped data discussed in section 2 required this method of extracting sea surface salinity and density. Sea surface salinity along the track is also shown in Fig. 7a with a dot-dash line; the salinity scale is given at the right of the figure.

The pattern of warm–cool–warm–cool sea surface temperature in Fig. 7a verifies that both lobes of Instability 1 are crossed by the ship track, although just barely. Locating this instability at sea proved difficult because the satellite image used for finding it was a few

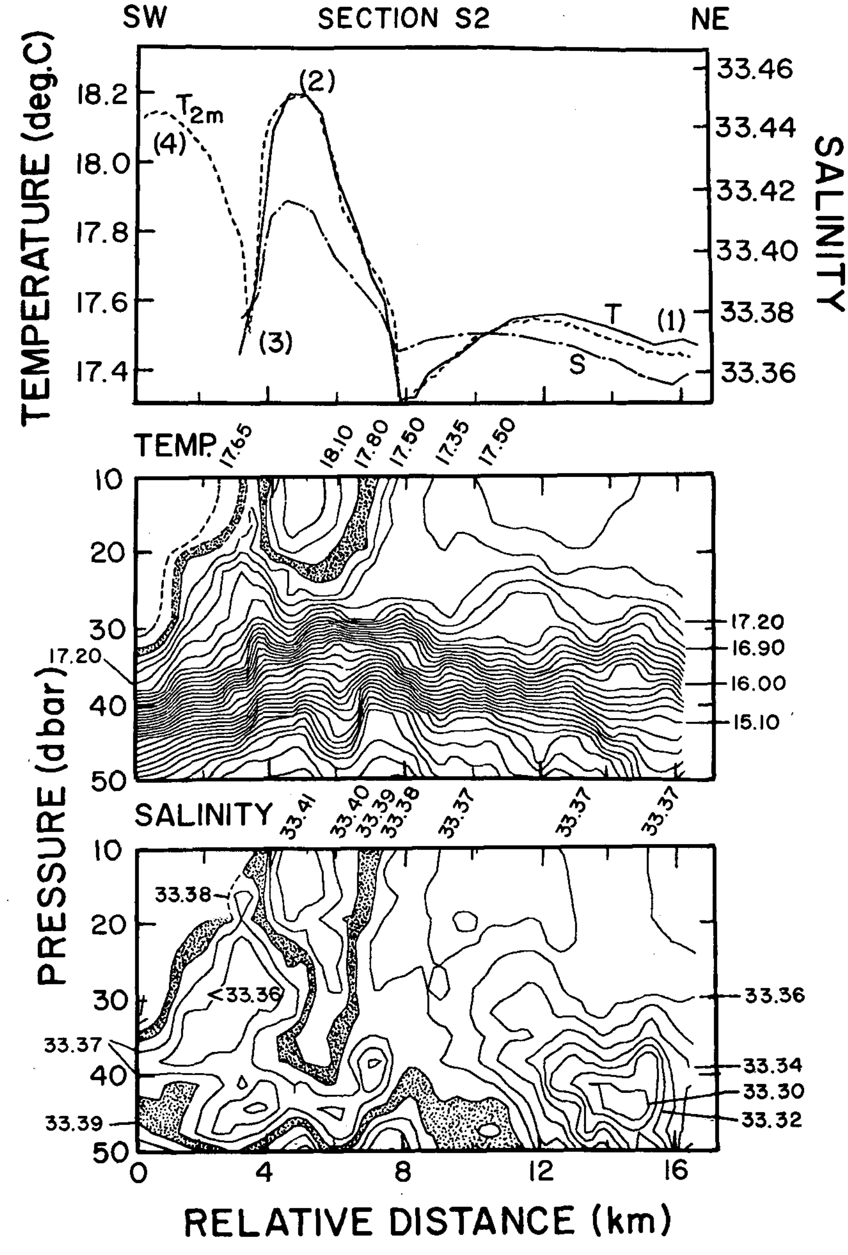


FIG. 7. (a) Horizontal profiles of temperature (solid line) and salinity (dash-dot line) averaged over upper 15 db obtained from profiling CTD along ship track S2. Continuous sea surface temperature (dashed line) at 2 m depth is labeled as $T_{2\,m}$. Points (1) through (4) are indicated in relation to instabilities in Fig. 6a. Direction of Section S2 indicated by SW (southwest) and NE (northeast) at top of figure. (b) Map of isotherms between 10 and 50 db along ship track S2. Contour interval is 0.15°C. Temperatures between 17.65 and 17.80°C are stippled. (c) As in (b) but for salinity. Contour interval is 0.01. Salinities between 33.38 and 33.39 are stippled.

hours old and the instability itself was moving westward and evolving over the period of these observations. At the beginning of the track in the warm water south of the filament, the sea surface temperature at point (4) is 18.15°C. (Only surface temperatures from the 2 m pumping system are available here because the CTD was being raised from the final 500 m cast at Station 10 along S1.) By 3.4 km at point (3) the surface temperature has dropped to 17.5°C and is similar to filament temperatures. Farther on at point (2) at 5 km, the surface temperature is back up to 18.19°C and is close again to temperatures south of the filament. The minimum temperature of 17.35°C is encountered at about 8 km and may be part of a band of cool water

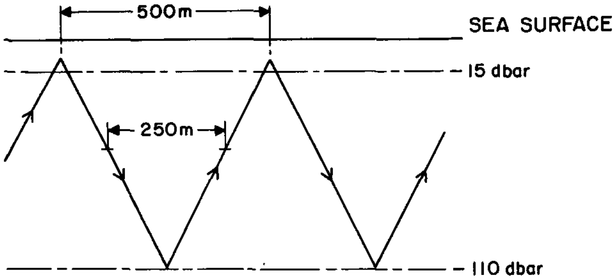


FIG. 8. Diagram of "saw tooth" tow-yo path followed by profiling CTD.

found elsewhere along the front in the satellite images. The temperature and salinity of point (1), which lies at the north end of the track, is representative of water properties in the filament. For reference, points (1), (2), (3), and (4) are indicated in the schematic diagram of Fig. 6a.

A cross section of isotherms between 10 and 50 db along S2 is given in Fig. 7b with a contour interval of 0.15°C. A total of 60 profiles is used in constructing this isotherm map and the spacing between midpoints of adjacent profiles is about 250 m as indicated in Fig. 8. The warm waters to the south of the filament are only partially mapped above 30 db over the first 3 km, although the 2 m pumped temperature data of Fig. 7a indicates that S2 begins in the warm water to the south. The warm lobe of the instability is clearly indicated between 3.7 and 7.3 km by the presence of water above 24 db that is warmer than 17.65°C. To better define the limits of the water to the south of the filament and the position of the warm lobe of the instability, the region between the 17.65° and 17.80°C isotherms is stippled in Fig. 7a. Dashed sections of the 17.65°, 17.80° and 17.95°C isotherms are extrapolated to the surface based on the 2 m pumped temperatures. A narrow ribbon of water cooler than 17.65°C, which is less than 500 m wide (poorly resolved given the profile spacing) and which extends to the surface, separates the warm southern waters from the warm lobe. This ribbon of cool water is at the edge of the cold lobe of the instability. A distinct doming of isotherms is evident above 30 db between 1 and 5 km. The minimum pressure, 21 db, of the 17.20°C isotherm occurs in this region; elsewhere throughout this entire section the 17.20°C isotherm is found only at greater pressures.

The corresponding map of salinity, contoured with an isohaline spacing of 0.01 practical salinity units, is shown in Fig. 7c. Regions containing salinities between 33.38 and 33.39 have been stippled as before to show where waters from south of the filament and from the warm lobe of the instability are found. A much deeper extent of the warm lobe is indicated by the salinity field compared with the temperature field. The stippled region indicating high salinity extends to about 40 db between 4 and 6 km while a distinct thermal signature of the warm lobe is confined to the upper 24 db. The only source of salinities of 33.38 and higher above 40 db is the frontal region south of the filament. A region of uplifted isohalines occurs in the same region as the doming of isotherms discussed above. A salinity minimum centered at 45 db with minimum salinities of 33.30 is found between 12 and 16 km and may be the signature of intrusive activity occurring in the filament. A broad region of nearly isohaline water is found above 30 db north of the high salinity, warm lobe.

A horizontal profile of σ_θ along track S2 is given in Fig. 9a; the dashed section of the profile in the first 3.5 km has been extrapolated based on the 2 m pumped temperature shown in Fig. 7a and the surface T–S relation in the warm lobe of the instability. From Fig. 9a it is clear that surface temperature gradients also correspond to surface density gradients. The highest surface densities are found in the cool lobe and just northeast of the warm lobe at 8 km. The surface density fronts found here stand in contrast to other observations in which surface thermohaline fronts are encountered on filament boundaries which are nearly compensating in density (cf. Flament et al. 1985). Typically, density compensated fronts are found farther offshore than the frontal regions discussed here.

A map of isopycnals down to 100 db is shown in Fig. 9b with a contour interval in σ_θ of 0.05 kg m^{-3}. The similarity of the σ_θ and temperature maps indicates that the water is thermally stratified. The extent of the instability is shown by stippling of regions in the map above 50 db for which salinity lies between 33.38 and 33.39. An uplifting of isopycnals that extends down to about 35 db is evident under the cool lobe of the instability between 1 and 5 km. A weakening of the seasonal pycnocline is found farther into filament waters beginning at about 13 km. Here the isopycnal spacing increases at around 40 db compared to the spacing nearer the beginning of the section.

This high resolution map of the isopycnal field suggests that the slope of the deeper isopycnals increases to the northeast (right-hand side of figure) in the section. Consider the σ_θ = 23.30 surface, indicated with a bold line in Fig. 9b. While small scale variability in the slope is present, the average slope of this density surface is relatively small over the first 4 to 8 km and then gradually increases over the rest of the section. This change in slope is consistent with lateral shearing of the geostrophic velocity field and in this case the thermal boundary of the filament nearly corresponds to the boundary of the geostrophic flow.

c. Property distributions along ship track S3

Horizontal profiles of surface temperature (solid line) and salinity (dash-dot line) are shown in Fig. 10a for segment S3. The ship track is oriented roughly north to south, and north is to the left in the figure. Note that the final points at the right of Figs. 7 and 9 nearly correspond to the beginning points of track S3. The

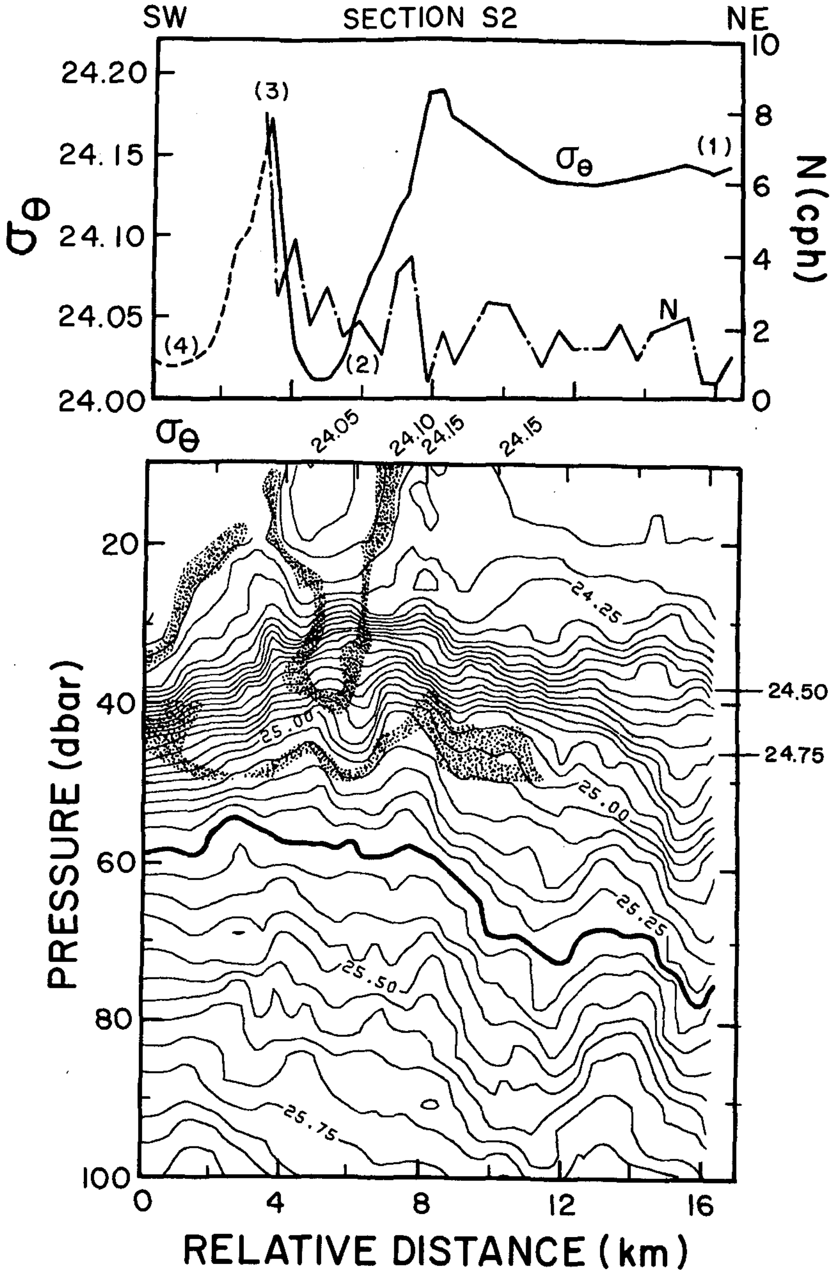


FIG. 9. (a) As in Fig. 7a, but for σ_θ (solid line) and buoyancy frequency N (dash-dot line). (b) As in Fig. 7b, but for σ_θ between 10 and 100 db. Contour interval is 0.05 kg m^{-3}. Stippling as in Fig. 7c. Bold line indicates σ_θ = 25.30 surface.

sea surface temperature pattern of cool–warm–cool–warm indicates that both lobes of the instability are crossed along this ship track. Cooler, low salinity filament water is found at point 1; warmer higher salinity water at point 2 in the warm lobe; cooler, low salinity water at point 3 in the cool lobe; and finally warm, higher salinity water at point 4, which lies to the south of the filament. The four points are labeled 1, 2, 3 and 4 and correspond to points (1), (2), (3), and (4) in the sea surface temperature and salinity plots of Fig. 7A.

A map of isotherms between 10 and 50 db along this section is shown in Fig. 10b with a contour interval of 0.15°C. The map is formed from 114 profiles. The warm lobe of the instability is clearly evident at 18 km in the map where water warmer than 17.50°C penetrates down to about 20 db. Regions for which the temperature lies between 17.65 and 17.80°C are indicated with stippling as before. Waters in the surface layer above 30 db are relatively well mixed both in the filament between points 1 and 2 and to the south around

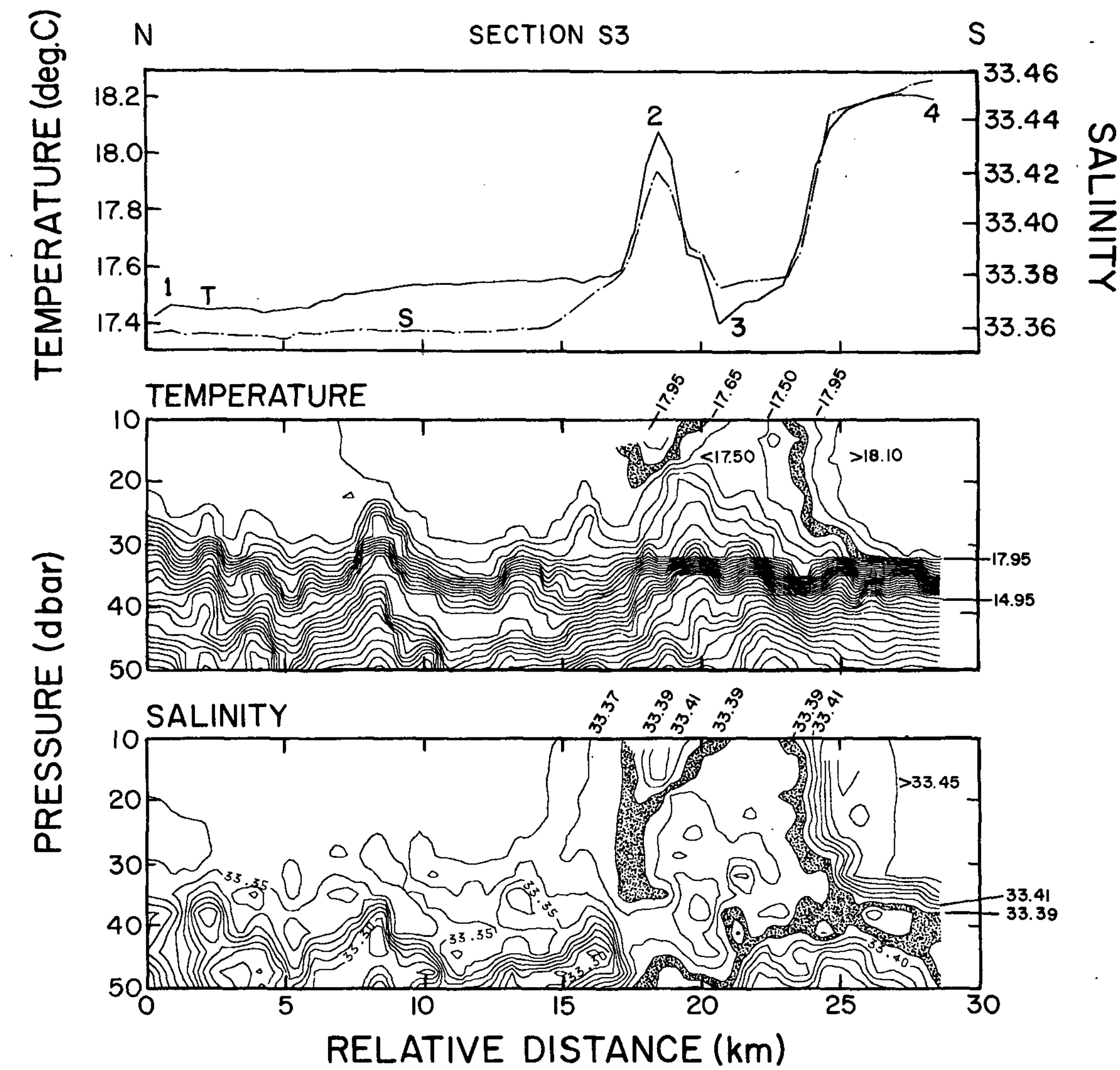


FIG. 10. As in Fig. 7, but for ship track S3. Points 1 through 4 are indicated in relation to instabilities in Fig. 6b. Direction of Section S3 indicated by N (north) and S (south) at top of figure.

point 4. Doming of isothermal surfaces is apparent around 19 km above 30 db. The 17.35°C isotherm is generally found below 25 db throughout the section except for this region of uplifting where it is found at 18 db. Water cooler than 17.50°C breaks out at the sea surface between about 21 and 22.5 km; these temperatures are encountered elsewhere at the surface only over the first 7 km of the section. Comparison of surface salinities in Fig. 10a reveals that, while the temperatures at points 1 and 3 are similar, the salinity at point 3 (33.375) is slightly higher than at point 1, (33.357) indicating that the water masses are different. While this is not a large salinity difference, it is significant and repeatable [compare the salinities along track S3 between points 1 and 3 with those along track S2 between (1) and (3)] and is not an artifact of the two-yo measurement technique.

A deeper penetration of the warm lobe water is again indicated in the contour map of salinity of Fig. 10c where salinities exceeding 33.38 are found down to 38 db at around 18 km. To better illustrate this, regions of Fig. 10c for salinity between 33.38 and 33.39 have again been indicated with stippling. As discussed above, the only source of this relatively saline water is in the frontal region and seasonal thermocline south of the filament. The salinity contour interval in Fig. 9c is 0.01. Salinities in the mixed layer and seasonal thermocline within the filament north of the warm lobe are generally less than 33.36. A pool of nearly isohaline water is found above 30 db over the first 15 km of the section.

A profile of σ_θ along segment S3 is shown in Fig. 11A with a solid line; it is again clear from this profile that surface temperature gradients also correspond to surface density gradients. Points 1, 2, 3 and 4 are num-

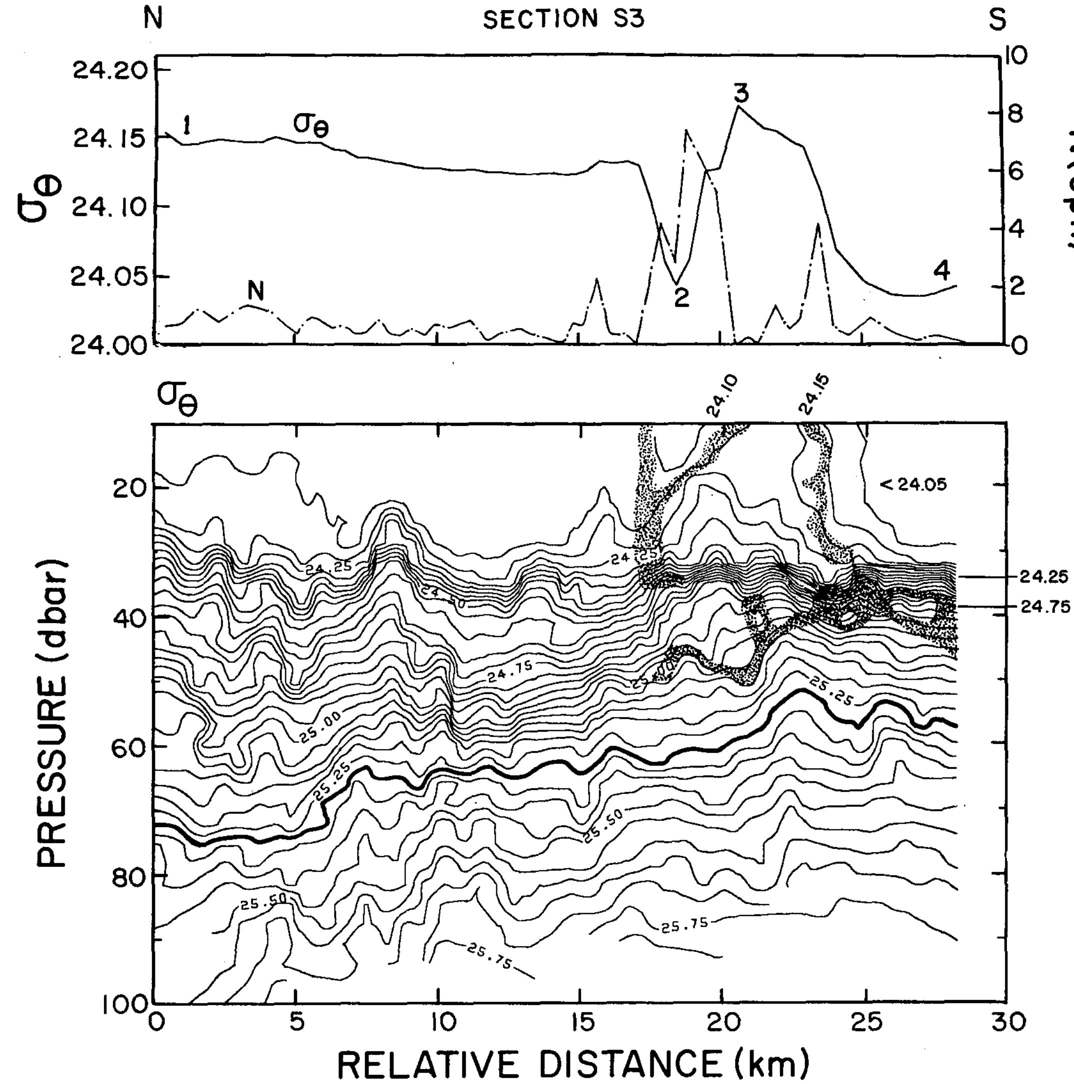


FIG. 11. As in Fig. 9, but for ship track S3.

bered for reference. The isopycnal (σ_θ) field above 100 db is given in Fig. 11b with an isopycnal contour interval of 0.05 kg m^{-3}. The extent of the instability is indicated by stippling ($33.38 < S < 33.39$) above 50 db. The only points at the surface where σ_θ exceeds 24.15 are in the cold lobe of the instability between 21 and 23.5 km. An uplifting of isopycnals is evident in this same region above 30 db. Over the first 23 km the general slope of σ_θ surfaces below 50 db is upward to the right in the figure (consistent with offshore geostrophic flow with respect to 480 db). This trend is present for the $\sigma_\theta = 25.30$ surface, shown with a bold line, which shoals by 20 db over the first 23 km of the section. Beyond 23 km, the slope of the 25.30 surface changes sign and produces a slight deepening of this surface in the water south of the filament. A general tendency for isopycnal slopes to become more level below the filament boundary may also be seen for other isopycnal surfaces and again indicates that the water property boundary coincides with the boundary of the geostrophic velocity field of the filament.

d. Near-surface stratification

An interesting feature of the near-surface layer (pressures less than 15 db) along ship track S3 is that water within the warm lobe of this instability is highly stratified compared with surrounding waters in the surface layer. This stratification is found despite 10 m s^{-1} winds and 30-m deep mixed layers elsewhere in the section. A plot of the buoyancy frequency N near the surface is shown in Fig. 11a with a dash-dot line; the scale for N is on the right-hand side of the figure. Individual values of N are computed from 2 db averages of density and sound velocity. Points shown in Fig. 11a are averages of all of these individual values above

15 db for each near-surface approach across the section; this procedure is the same as that used in computing surface water properties discussed previously. A maximum in N is found at Point 2 in Fig. 11A between 17 and 20.5 km with a peak value of 7.3 cycles per hour (cph). Typical maximum values of N in the seasonal thermocline are in the range 15 to 21 cph while mixed layer values are generally less than 1 cph. The secondary peak at 23.3 km is an artifact of the tow-yo sampling scheme and is actually due to a high horizontal gradient of density. A detailed examination of individual profiles throughout both regions of maximum N in Fig. 11a reveals that relatively large vertical density gradients are found near the surface at point 2 while at the smaller maximum, water is well mixed to the surface. A somewhat similar near surface increase in N within the cool lobe of Instability 1 may be seen at about 3 km in Fig. 9a. Possible mechanisms for producing the observed increases in near surface stratification are given in section 5.

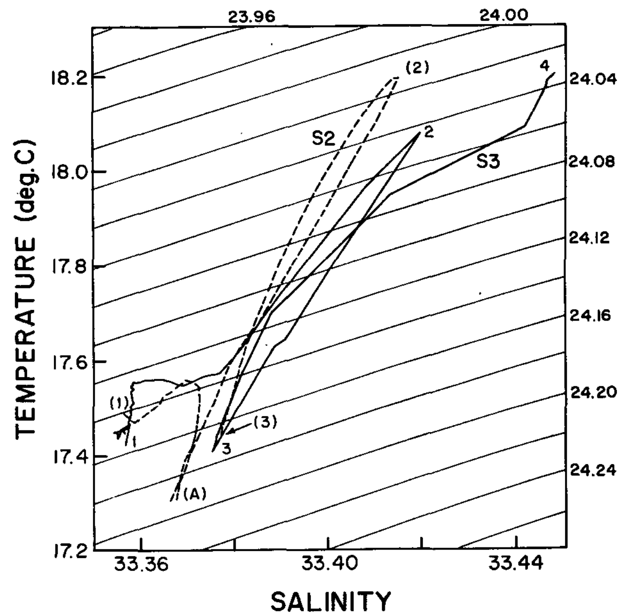


FIG. 12. Surface T–S diagram for points in Figs. 7a (dashed line) and 10a (solid line). Points (1) through (4) and 1 through 4 are shown in relation to instabilities in Figs. 6a and 6b.

e. Surface T–S relation

In order to sort out the various surface water mass types, a T–S diagram is constructed along Tracks S2 and S3 and is shown in Fig. 12. Points in Fig. 12 represent average properties in the upper 15 db of the water column. The T–S points along Track S2 are shown with a dashed line and points along S3 are shown with a solid line; numbered reference points in this figure are the same as the numbered points in Figs. 7a and 10a. Track S2 begins in the cool lobe of Instability 1 (see Fig. 6a) at point (3) and has T–S values that are very similar to those in the cool lobe of Instability 2 along track S3 at the corresponding point 3. Apart from water at point (A), the water at 3 and (3) has the highest density along either track and cannot be formed by simple horizontal mixing of water masses 1 and 4. A straightforward way of accounting for the appearance of "type 3" water at the surface is by local upwelling from a depth of about 30 m. Because temperatures at 1 and 3 are nearly equal, it might be concluded from the SST imagery alone that water at 3 is surface water from the filament which has simply been advected due to the instability. Pure advection, however, would not explain the higher salinity observed at 3. The locus of points on the T–S curve between points 2 and 3 is almost a straight line and is consistent with strong horizontal mixing across the boundary between the two lobes of Instability 2.

The highest density surface water found is at point (A) which lies north of the warm lobe along track S2. Based solely on the shapes of the T–S curves at points 3, (3) and (A), it might be concluded that similar processes result in the appearance of these water masses at the surface. However, no uplifting of isotherms which might indicate active upwelling is found in association with water mass (A) as may be seen in Fig. 7b. Furthermore, the temperature field is more well mixed above 25 db at (A) compared with 3 or (3). Water mass (A) is not found along S3.

5. Discussion

Horizontal shear in the near-surface velocity field at the cyclonic boundary of the filament is a likely driving mechanism for these instabilities. Although we do not have direct velocity measurements (e.g., using an acoustic Doppler velocity profiler), we can estimate the shear in the geostrophic velocity field at the filament boundary from our observations. Below, we present horizontal shear estimates based upon the observed velocity difference across the cyclonic front of the filament and upon two length scales: the distance between stations which span the front and an estimate of the width of the shear layer.

The difference in geostrophic velocity averaged between 20 and 40 db (within the depth range of the property distributions of the instabilities) relative to 480 db computed between Stations 8 and 9 is -0.32 m s^{-1}. The distance between these two stations Δy is about 4 km which yields a horizontal shear estimate $\Delta U/\Delta y \sim -8 \times 10^{-5}$ s^{-1} $\sim -f$. This probably represents a lower bound on the actual shear because only the geostrophic component of velocity is used in this estimate and the actual shear scale may be smaller than the station spacing.

An estimate of the shear layer width may be obtained if it is assumed that the alongfront spacing of the instabilities equals that of the fastest growing mode of a barotropic disturbance in a uniform shear flow between two regions of constant velocity. The alongfront wavenumber k of this mode is specified by $kL \approx 0.4$ where L is half of the shear layer width h (cf. Gill 1982). From the observed alongfront spacing of the instabilities λ of 15–20 km (section 4 and Fig. 5), an estimate of the width of the shear layer $h = (0.8/2\pi)\lambda$ is in the range 1.9 to 2.5 km. Therefore, taking h to be of order 2 km along with the velocity difference given above yields a horizontal shear $\Delta U/h$ approximately -2×10^{-4} s^{-1} $\approx -2f$. It is interesting that this estimate of h is nearly equal to the width of the surface T–S front on the southern boundary of the filament in Fig. 10a.

These order of magnitude shear estimates are subject to error since geostrophic velocities are computed from closely spaced CTD casts (about 8 km). However, as discussed in section 5, a consistent change in isopycnal slope at the filament boundary is evident over these scales, which indicates strong horizontal shear. Moreover, the order of magnitude estimates given above are comparable to several observations of strong near-surface horizontal shear in the range 10^{-4} to 10^{-3} s^{-1} as discussed by Sheres et al. (1985). Kosro (1987) reports observations of the velocity field in the upper 200 m of the water column during the Coastal Ocean Dynamics Experiment (CODE) from which horizontal shear may also be estimated. Horizontal shear at the cyclonic boundary of an upwelling filament off Pt. Arena (see his Fig. 4c) and in the alongshore current (see his Fig. 7b) are both of order 10^{-4} s^{-1}.

The growth rate of these instabilities is consistent with the fastest growing mode of a barotropic disturbance, which was assumed above, in determining the width of the shear layer. The e-folding time for the growth of a purely barotropic disturbance is about $(0.2dU/dy)^{-1}$, cf. Gill (1982), and when evaluated for these observations is about 8 h using the horizontal shear estimate of $-2f$. While a precise e-folding time is difficult to define from a sequence of satellite images, this is of the same order as the time between images I1 and I3 (12.5 h) in which Instability 1 evolves from a small perturbation on the front into a feature which has two fairly well defined lobes.

The distributions of isopleths of temperature, salinity, and density indicate that the flow field of the instabilities is three-dimensional with local upwelling occurring in the cool lobes of the instabilities. Compensating downwelling presumably occurs on the instability boundaries. Assuming that the $\sigma_\theta = 24.15$ surface, found at about 30 m (see Fig. 11) on either side of the instability, migrates to the surface over the time of formation of the instability (1 day), an upwelling rate of about 30 m d^{-1} is implied. Following Lee and Atkinson (1983), upwelling within the instabilities can be accounted for based on conservation of potential vorticity,

$$\frac{d}{dt}\left(\frac{\zeta + f}{D}\right) = 0 \tag{1}$$

where

$$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$$

is the relative vorticity, f the planetary vorticity, and D the height of a fluid column. Eastward and northward are the positive x and y directions.

We hypothesize that as fluid columns move from the narrow region of horizontal shear on the cyclonic boundary of the filament into the frontal instabilities, they undergo a decrease in relative vorticity. To compensate for this decrease in ζ, the height of the fluid columns must also decrease according to (1) so that angular momentum is conserved. The rate at which this decrease in D occurs may be found by differentiating (1) and solving for dD/dt,

$$\frac{dD}{dt} = \left(\frac{D}{\zeta + f}\right)\frac{d\zeta}{dt} \tag{2}$$

where the change in f with displacement in the y-direction can be neglected.

While insufficient measurements are available to accurately determine the various terms in (2), order of magnitude estimates suggest that changes in ζ might account for the observed upwelling. The initial relative vorticity ζ_1 is taken to be $\zeta_1 \approx -\Delta u/\Delta y \sim -\Delta U/h \sim 2f$ as given above. Idealizing the flow field of the instabilities as vorticies rotating at angular velocity $\omega_2 \sim 0.1f$ as discussed in section 4, the relative vorticity within the instabilities is of order $\zeta_2 \equiv 2\omega_2 \sim 0.2f$. Further, assuming that the time of formation of the instabilities τ is of order one day leads to an estimate of $d\zeta/dt \sim (\zeta_2 - \zeta_1)/\tau \sim -2 \times 10^{-9}$ s^{-2}. With a mixed layer depth of 30 m, (2) yields an estimate for $dD/dt \sim -18$ m d^{-1} which is of the same order as the upwelling rate given above of 30 m d^{-1}.

Additional evidence that the velocity fields of the instabilities are three dimensional is the high near surface stratification found within regions of the instabilities, such as in the warm lobe of Instability 2 (see Fig. 11a between 17 and 20.5 km). One way an increase in stratification could occur is by near-surface vertical shear of the horizontal velocity field across the high gradient frontal region between the instability lobes. This would result in a tilting of isopycnal surfaces from a near vertical orientation initially to an inclined orientation under the influence of the shear; after tilting both vertical and horizontal density gradients would be present. The increase in N to about 8 cph observed in the cool lobe of Instability 1 in Fig. 9a at 2.6 km occurs at the edge of a local upwelling region where sloping isopycnals extending to the surface also produce both vertical and horizontal density gradients. Here the upwelling causes isopycnals to rotate from an initial horizontal orientation to a more vertical one.

Frontal instabilities such as these are probably not important mechanisms for the dissipation of the total kinetic energy of upwelling filaments due to their rather small size and limited depth. Their overall role in filament dynamics may be as local vertical and horizontal transport mechanisms in horizontal shear layers acting over scales of a few kilometers. They extract kinetic energy from the shear flow at the filament boundary as eddy kinetic energy associated with their rotation and by converting kinetic energy to potential energy through upwelling. They may also be sites of increased dissipation of kinetic energy into heat through the action of small scale turbulent processes. Inasmuch as they are not purely barotropic disturbances, they may also extract available potential energy from the density field associated with the larger scale geostrophic flow. A crude estimate of the eddy kinetic energy is about 6 $\times 10^8$ J (joules) assuming an eddy to consist of a disc of seawater 4 km in radius rotating at $0.1f$ and having a vertical extent of about one mixed-layer depth (30 m). This is probably an underestimate since the initial rotation rate of the instability may have been larger than $0.1f$. An estimate of the potential energy gain through upwelling is larger. Idealizing the cold core of the instabilities as a volume of water 4 km by 4 km by 30 m deep and assuming this volume is raised 30 m into the mixed layer and displaces water that is 0.10 kg m^{-3} less dense, the potential energy gain is about 1 $\times 10^{10}$ J. The kinetic energy of the geostrophic flow field of the filament with respect to 480 db in Fig. 4b is about 3×10^8 J m^{-1} along the axis and for a 100-km long filament yields a total kinetic energy of about 3×10^{13} J. The kinetic energy lost to a single instability, therefore, is less than 0.04% of the total.

6. Summary and conclusions

Frontal instabilities were observed on the cyclonic boundary of an upwelling filament, which appeared north of Pt. Arguello along the central California coast. The instabilities observed form at an alongfrontal wavelength of 15 to 20 km and have a cross-frontal extent of about 8 km. These instabilities are composed of two lobes, one warm and one cool, which rotate about each other in a cyclonic sense. Based on a sequence of satellite SST images, the time scale for instability formation is about 1 day and the relative vorticity within the instabilities is about $0.2f$. The growth rate and alongfront wavenumber of these instabilities are consistent with those of the fastest growing mode of a barotropic disturbance in a uniform shear between two regions of constant velocity.

The satellite images were used to direct in situ sampling such that the subsurface structure of the instabilities could be observed during their evolution. Continuous tow-yo CTD sections through the upper 100 m of the water column were obtained across two instabilities and maps of the temperature, salinity, and density fields are presented. Comparison of the observed in situ SST profiles along the ship tracks with the satellite SST imagery verifies that the instabilities were crossed during shipboard profiling. The depth of penetration of the instabilities appears to be about 40 m based on the salinity field and isopycnal displacements. Salinity is a better indicator of the vertical extent of the instabilities than either temperature or density. The frontal boundary that separates cool fresher waters of the filament from warmer more saline waters to the south is composed of strong horizontal gradients of temperature, salinity and density. Waters are thermally stratified throughout the regions observed.

The near surface distribution of water properties and finestructure indicates that the velocity field of the instabilities is three dimensional. In particular, a distinct doming of isotherms and isopycnals is found within the cool lobes of the instabilities and is consistent with local upwelling at a rate of about 30 m d^{-1}. Surface water masses observed within the cool lobes of the instabilities cannot be formed from advection and simple horizontal mixing of surface waters found inside and outside of the filament. However, a nearly linear T–S relationship is observed for the surface waters between the warm and cool lobes suggesting that strong horizontal mixing processes occur within the instabilities over scales of about 3 km. High stratification ($N > 7$ cph) is observed within the surface layer in the upper 15 m of the water column and may result from vertical shear in the near surface velocity field.

Upwelling within frontal instabilities over horizontal scales of order 5 km may represent an important vertical transport mechanism that acts at filament boundaries. One effect of this upwelling would be to inject scalars such as nutrients and nearly neutrally buoyant particles, which are initially contained in the seasonal thermocline, into the surface layer in a time period of about 1 day. Due to the sloping isopycnal surfaces within the instabilities, strong isopycnal mixing processes could also transport scalars vertically from the seasonal thermocline into the surface layer. This isopycnal transport could possibly persist even after the larger scale forcing that produced the instabilities has subsided. Clearly, high resolution mapping of the near-surface velocity field and mixing processes are needed for a better understanding of the ageostrophic dynamics and transport processes of the instabilities.

In addition to the high resolution tow-yo observations, a section of CTD stations across the filament was obtained and is used to derive a section of the geostrophic velocity field of the filament. The geostrophic velocity field with respect to 480 db is that of a baroclinic jet with velocity increasing upwards in the water column. Maximum offshore current velocities are about 0.4 m s^{-1} at 60 db and the net transport of the filament is at least 1.7 Sv (1 Sv $\equiv 10^{-6}$ m^3 s^{-1}). This transport is large enough to exchange water masses over large sections of the continental shelf (order 130 km) in a time period that is less than the lifetime of the filament. This indicates that upwelling filaments are important transient circulation mechanisms along the central California coast.

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