Shipboard Acoustic Doppler Profiler Velocity Observations near Point Conception: Spring 1983

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During April 1983, shipboard Doppler acoustic log current profiles were collected in an effort to characterize the flow field near points Conception and Arguello, California. Subsurface velocity maps derived from these profiles have been used to describe spatial flow structures both on and off the shelf and to investigate flow variability as a function of time and of wind stress. Persistent westward flow out of the northern half of the Santa Barbara Channel and eastward flow into its southern half were observed regardless of the direction of the local wind stress. During one well-documented upwelling-favorable wind event, currents responded in the form of an energetic (maximum 21 m/s speeds of >60 cm s⁻¹) offshore squirl of cold water. During weak or downwelling-favorable winds, currents continuous with the Santa Barbara Channel outflow were observed flowing to the northwest following the local isobaths before turning offshore west of Point Arguello. Evidence for wind forcing of current fluctuations nearshore between the points and north of Point Arguello was found. Lack of a thermal wind balance between directly measured velocity shear and horizontal density gradients was explained by the presence of large accelerations in the momentum equations. Lack of a consistent relation between velocity and temperature gradient illustrates the difficulty in estimating velocity from temperature information alone in this area.

1. INTRODUCTION

The interdisciplinary 1983 Organisation of Persistent Upwelling Structures (OPUS) field program had the objective of mapping the near-surface physical and biological fields within and around a localized upwelling center as a function of time. The goal of these observations was to provide a synthesis of how the physical structures and variability worked to determine the observed biological (e.g., phytoplankton) fields. A major part of this mapping effort was underway acoustic Doppler log (DL) current profiling, which is reported on here. Other physical measurements included current meter data [Brink and Muench, 1986], drifters, hydrography, and aircraft wind and temperature measurements [Atkinson et al., 1986].

The region around Point Conception, California (34°31’N), was chosen for this study because a considerable body of literature [e.g., Sverdrup, 1938; and satellite observations have suggested the presence of an intense coastal upwelling center in the area. In situ measurements during the 1981 OPUS pilot study tended to confirm earlier impressions, although no direct velocity measurements were made at that time [Brink et al., 1984].

In the following, we will attempt to exploit the DL current data to provide descriptions of spatial flow structures and their variability as a function of time and of the wind stress. The approach will be first to describe the data and its accuracy, then to discuss descriptive aspects of the flow field. Third, time series of spatially averaged currents will be treated, followed by a quantitative discussion of the relation between directly measured velocity and the hydrographic fields. Finally, some conclusions will be drawn, and the results summarized.

2. METHODS

2.1. Sampling Plan

The governing philosophy of the sampling scheme was to make systematic, repeated measurements that would define spatial structures of the upwelling feature and allow resolution of their temporal evolution. Toward this end, three principal hydrographic lines were defined [Atkinson et al., 1986]. The east-west A line and the north-south C line (Figure 1) were set in order to define the boundary conditions on the central OPUS region. The central G line (Figure 1) was meant to sample the core of the upwelling center. These lines were not viewed as sufficient to map anticipated structures, so two intermediate lines (GC and AG) were also sampled during mapping exercises. The overall sampling scheme was based on a 6-day cycle as follows: (1) a G line conductivity-temperature-depth (CTD)/bottle sample section; (2) a map covering, at least, the A, AG, G, GC, and C lines with expendable bathythermograph (XBT) and surface underway sampling; (3) an A line CTD/bottle sample section; (4) a G line CTD/bottle sample section; (5) a C line CTD/bottle sample section; and (6) a map as in (2).

One day was reserved between 6-day cycles for drogue releases, zooplankton net tows, or other sampling. The DL profiling system was run continuously through all phases of the measurements. Individual CTD/bottle lines took an average of 15 hours, and individual maps an average of 24 hours. The overall scheme allowed very good sampling of the main hydrographic lines during the time that the Doppler log was available (April 7–22): 9 A lines, 18 G lines, and 10 C lines.

2.2. Acoustic Doppler Log

Acoustic DL measurements were made from R/V New Horizon using a hull-mounted transducer operating at 300 kHz. Acoustic pulses (pings) of 20-msec duration were transmitted every 1.2 s in four beams directed obliquely through the water column. These acoustic signals were scattered by small particles (including phytoplankton and zooplankton), turbulence, or temperature microstructure. The Doppler frequency shift of the returned signal was measured in each of 31 time intervals. Each time interval represented a measurement over a 6.4-m column of water, yielding a total measurement range from the bottom of the ship to 200 m. By combining the measured Doppler frequency shift from each of the four beams, a measurement of horizontal velocity rel-
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![Map showing the locations of hydrographic lines, current meter moorings (solid circles) and NDBC buoys (open circles). Depths are in meters.](image)

Fig. 1. Map showing the locations of hydrographic lines, current meter moorings (solid circles) and NDBC buoys (open circles). Depths are in meters.

ative to the ship as well as two independent measurements of vertical velocity were obtained. Heading from the ship's gyrocompass and position from LORAN-C were used to calculate the ship's velocity, which was then combined with the acoustically measured relative velocity to yield absolute horizontal velocity. A more detailed description of the system can be found in the work of Joyce et al. [1982].

To decrease inherent system noise and the effect of surface wave motion on the measured velocities, a 10-min average was made from the 12-2 s pings. Thus each measurement record was constructed from up to 500 independent acoustic pulses. The percent of pulses returned varied with depth. The shallowest reliable data were obtained from a bin centered on 21 m. At depth, the percent of good returns fell below a threshold value (set at 27% corresponding approximately to the point where the two independent vertical velocity measurements deviated from each other by 1 cm s⁻¹), limiting the effective depth of the instrument to around 160 m. Data were not collected between 45 and 60 m because of instrumental problems.

Data were collected continuously from April 7, 2000 (all times are UT) until April 15, 2000 when the R/V New Horizon went into port and then again from April 18, 0800 until April 29, 0700. During this time the ship conducted hydrographic surveys in a 6-day cycle as described earlier. This data set, consisting of 10-min records, was subjected to the following editing procedures. Obviously spurious data associated with abrupt changes in the ship's velocity during a 10-min record were deleted. Data from the deeper bins contaminated by bottom reflections of the acoustic signal were also deleted. Records taken while the ship was in one location (e.g., conducting a CTD cast) were averaged to produce one record for that location. Finally, to reduce small record-to-record discrepancies in absolute velocity due to errors in LORAN-C navigation, the velocities over the earth were smoothed by a three-record or 30-min running mean filter. Note that no smoothing of the relative velocities was performed, and the measured vertical shear was preserved. Comparisons (presented in detail below) between absolute DL and current meter velocities with and without applying the smoothed ship velocity indicated an approximately 23% decrease in the standard deviation of the difference of the two measures of velocity using the smoothed DL data. This three-record running average was applied over the entire measurement period to produce the final data set. (An informal data report can be obtained by request from the authors.)

To assess the effect of tides on the DL measurement of velocity, the following analysis was performed. Hourly time series of depth-averaged velocity were calculated using all the current meter data at moorings C1 and C2 (Figure 1). A least squares fit using the two major tidal constituents (M₂ semidiurnal, period = 12.42 hours; K₁ diurnal, period = 23.93 hours) was applied to the data. Amplitudes and phases derived from the fit were used to calculate the tidal velocity during each 10-min DL record for a representative 24-hour time period. Tidal velocities were not expected to be uniform in space, and this analysis was intended only to provide an estimate of the percent of the total velocity signal attributable to the tides. The rms speed of the tide was 5 cm s⁻¹ while the total velocity signal measured by the DL was typically much greater. Because of this small tidal signal the original DL data set was left unaltered.

Earlier studies of the acoustic DL measurement system yield the following estimates of the accuracy in measuring current velocity [Joyce et al., 1982]. The accuracy in a 10-min average record of relative velocity estimated from the Doppler frequency shift is 1–2 cm s⁻¹. Errors in absolute velocity, due for the most part to insufficient LORAN-C resolution, are 5–10 cm s⁻¹ in a 10-min average record. To test the accuracy of the system used during OPUS, two comparisons were made between the acoustic DL velocity and velocity measured by conventional current meters on fixed moorings. The locations of current meters used for comparison are shown in Figure 1. Two moorings between points Arguello and Conception (C1 and C2) were deployed as part of the OPUS observational program [Brink et al., 1985]. Each mooring consisted of six vector-measuring current meters distributed throughout the 70-m-deep water column. In addition, five subsurface moorings (P1–P5) equipped with Aanderaa current meters were placed across the mouth of the Santa Barbara Channel (SBC) as part of a Mineral Management Service program conducted by Science Applications International Corporation (Bellevue, WA) and Dynalysis of Princeton [Brink and Muench, 1986]. Finally, a subsurface mooring (CHA) was deployed south of Point Arguello near the 200-m isobath for Chevron U. S. A., Western Region (La Habra, CA) [EG&G, 1985]. The mooring had either acoustic or vector measuring current meters at depths of 9, 11, 15, and 20 m, but only the 9-m instrument's data are used here.

In the first comparison, absolute velocity measured by the acoustic DL was compared to velocity measured by the appropriate current meter whenever the ship was within 1.3 km of any of the above moorings. The results, for a sample of 24 10-min records, showed a small mean difference between the DL velocity and the current meter velocity (−0.8 cm s⁻¹).
for the east component and -0.2 cm s$^{-1}$ for the north component), and the standard deviation of the difference was 8.1 cm s$^{-1}$ for the east component and 12.3 cm s$^{-1}$ for the north component. These results are in line with Joyce et al's [1982] estimate of 5-10 cm s$^{-1}$ error due mostly to LORAN-C navigation errors. The small mean difference indicates that the error in measuring velocity with the DL is essentially random. This fact will be exploited later when averages of the 10-min records are presented in section 4. Koero [1988], in an analysis of a much larger data set collected as part of the Coastal Ocean Dynamics Experiment (CODE) off northern California, found differences in the means of 0.6 cm s$^{-1}$ or less for cross-shelf velocity and 1.6 cm s$^{-1}$ or less for alongshelf velocity. Standard deviation of the differences were found to be 4.1-5.4 cm s$^{-1}$ in cross-shelf velocity and 3.6-4.4 cm s$^{-1}$ in alongshelf velocity. The smaller standard deviations in the CODE comparison may be attributable to increased LORAN-C accuracy in that area of the California coast (P. M. Koero, personal communication, 1985).

The second comparison involved the measurement of current shear over the largest common depth of the two measurement systems. Inaccurate navigation introduces no error in the DL measurement of the vertical shear of current. To obtain an adequate number of data points and in anticipation of the analysis presented in section 5 where section-averaged data are used to compare shear measured directly by both DL and current meters to geostrophically calculated shears, east-west DL velocities were averaged over seven crossings of the C line across the mouth of the SBC (Figure 1). Time averages of hourly current meter data from moorings P1-P4 were computed for the same time intervals as the C line crossings. East-west velocity differences were calculated between shallow (30 m for P1-P4) and deep (65 m for P1, 135 m for P2-P4) current meters and between the corresponding depth bins for the DL. The mean of the difference between these two measures of velocity difference ($\Delta u_{OM} - \Delta u_{DL}$) was 0.2 cm s$^{-1}$, and the standard deviation about the mean was 1.0 cm s$^{-1}$. These values indicate that the DL does indeed measure current shear accurately. (Note that since errors in velocity difference were essentially independent of depth separation, absolute error in velocity shear decreases as vertical separation increases.)

2.3. Drifters

In addition to the DL measurements described above, approximately 90 current-following drifters were deployed in the OPUS study region during April and May 1988. The construction, testing and performance of the drifters are described by Davis et al. [1982]. Surface drifters and drifters with subsurface drogues were deployed by both the R/V Velero IV (1 and 25 m) and R/V New Horizon (1 and 50 m). Deployments, consisting of mostly surface drifters, took place in cross-shelf transects along lines C, G, and A (Figure 1) and along two additional cross-shelf transects north of Point Arguello. In addition, 25 drifters were released close inshore between points Arguello and Conception in order to track the course of waters upwelled there. Drifters were tracked by radio direction finding techniques from aircraft, from a mobile shore station and from the R/V Velero IV. A sum-
Fig. 3. As in Figure 2 but for the G line with alongshelf velocity (\(v\)) positive to the northwest and cross-shelf velocity (\(u\)) positive to the northeast or toward the coast.

Fig. 4. As in Figure 2 but for the A line with alongshelf velocity (\(v\)) positive to the north and cross-shelf velocity (\(u\)) positive to the east or toward the coast.
Fig. 5. Time series of wind stress, where north is up, obtained from NDBC buoy 46023 (NC). Light and dark bars indicate time-averaging intervals.

mary of the drifter data obtained during the entire spring 1983 OPUS field program can be found in the work of Davis and Roper [1984].

3. A Descriptive View

3.1. Section Statistics

Mean and standard deviation sections of DL velocity components have been computed for the principal hydrographic lines. The approach was first to sort all 10-min averages into ~3.5-km bins along each line. For each bin, a mean and standard deviation were then computed for the local alongshelf (v) and cross-shelf (u) velocity components. Local coordinate systems were chosen as east-west along the A and C lines and rotated by 45° for the G line. The sign convention is such that positive alongshelf velocities correspond to flow with the mainland to the right and positive cross-shelf velocities correspond to flow toward the main-

April 9-10, 1983

21m Doppler Log Velocities

SST(°C) Surface Drifter Velocities

Fig. 6. (a) Doppler log velocities obtained at 21-m depth during April 9–10, 1983. The base of the velocity vector obtained at the beginning of the measurement period is circled. (b) Sea surface temperature (SST) in degrees Celsius and 12-hour averaged surface drifter velocities for the same time period as in Figure 6a. Wind stress at NC is displayed at the upper right, where the radius of the circle is 1 dyn cm⁻².
land. Following the computation, the results were contoured using computer algorithms described by Spyrou and Keffer [1983]. We estimate measurement errors in these mean velocities to be less than about 3 cm s\(^{-1}\). Some caution is required in interpreting plots of standard deviation, because the value expected from purely instrumental error is itself of O(10 cm s\(^{-1}\)).

The C line sections (Figure 2, an average of 11 individual sections) are very similar to those obtained from current meter data [Brink and Muench, 1986]. Along the northern side of the SBC, a westward mean flow with peak velocities greater than 15 cm s\(^{-1}\) was observed. Slightly weaker (~10 cm s\(^{-1}\)) eastward mean flow predominated south of about 20 km from the coast and was accompanied by a slight northward flow component. Strongest east-west mean flow occurred at depths shallower than about 75 m. Standard deviations of alongshelf and cross-shelf flow were in the range of 10–16 cm s\(^{-1}\) for both velocity components, decreasing only slightly with depth and varying little in the horizontal.

At the G line (Figure 3, an average of 13 individual sections), mean flow was dominated by a jet centered at a depth of about 40 m, 20 km from shore. The peak speed of the jet was greater than 20 cm s\(^{-1}\), and the feature was apparently a continuation of the westward flow on the northern side of the SBC. The jet existed just beyond the shelf break and tended to follow the approximate local isobaths, especially below about 60 m. Flow became weaker and more depth-independent away from the core of the jet. Mean DL velocities agreed within ±2 cm s\(^{-1}\) with averages computed over the same time period from current meters located on mooring CHA (Figure 1). Curiously, even though the G line crossed through the core of a persistent upwelling center, there was only a faint sign (at the shelf break) of a shoreward mean flow to supply the upwelled water. This may be due to lack of sufficient observations inshore of the shelf break or, as will be discussed later, to the presence of an alongshelf source for the upwelled waters rather than a directly offshore source typical of two-dimensional upwelling circulation models. The standard deviations of both velocity components ranged from about 20 cm s\(^{-1}\) in the upper 50 m down to about 12 cm s\(^{-1}\) at 120 m. There was a relative maximum in cross-shelf flow variability near the jet core but
a relative minimum in alongshelf variability near the same location. This apparently means that the jet was a persistent feature but that its location and direction tended to vary slightly. The maximum in the standard deviation of alongshelf velocity occurred seaward of 30 km offshore in the upper part of the water column. This region often contained strong south-southeastward flows during upwelling-favorable (southeastward) winds.

At the A line (Figure 4, an average of 9 individual sections) west of Point Arguello, the alongshelf flow was southward at depths less than 90 m and relatively uniform across the shelf. This shallow southward alongshelf flow occasionally appeared to continue across the seaward end of the G line, contributing to the high standard deviation observed there. In contrast to the G and C lines, the flow was so strongly sheared in the vertical that it was nearly zero at depths greater than 90 m. The sense of the alongshelf vertical shear was the same as that found by Strub et al. [1987] during April 1982 in a study of current meter velocities located 20 km offshore of Point Puraism. The cross-shelf velocity was directed offshore at all depths and locations, with speeds as great as 20 cm s⁻¹. Thus there was apparently a net mean convergence of alongshelf flow between the A and G lines, balanced by a strongly offshore directed cross-shelf flow along and, presumably, between the two lines. As on the other two principal lines, the A line standard deviations decreased with depth. The maximum of alongshelf standard deviation (>26 cm s⁻¹) was the largest observed along any of the hydrographic lines.

3.2. An Upwelling Cycle

While the statistics of the observed currents may be interesting, it is also useful to show the data in its basic form.

We will do this by describing the course of a significant upwelling event during April, from initiation to recovery. This will allow an appreciation for the richness of the data set and of the importance of relatively small scale features. In the following, 10-min-averaged acoustic DL data will be used, plotted so that the base of the arrows lies at the point of measurement. Further, concurrent 12-hour-averaged surface drifter velocities obtained from objectively interpolated drifter tracks [Davis and Reiger, 1984] are superimposed on the sea surface temperature (SST) (measurements from 2-m-deep ship intake) charts for comparison. The wind record is summarised in Figure 5.

The map of April 9–10 was obtained during a brief period of weak winds between two substantial upwelling events. The westward flowing jet from the SBC was clearly evident, although it appeared to meander as it crossed the OC line (Figure 6a). At the same time, a southwestward flow occurred in the northern region about 20 km from the coast. The two major flow structures appeared to converge between the AG and G lines and presumably acted to transport water farther offshore. One feature in common with all other maps was the generally weaker flow over the shelf. Surface drifter velocities (Figure 6b) seem to bear a qualitative resemblance to the 21-m DL velocities, especially when the 10 cm s⁻¹ inaccuracy of the DL measurements is accounted for.

Perhaps one of the more striking aspects of the April 9–10 map is the fronts evident in 2-m temperature (Figure 6b). The northern front (running north-south about 10 km from shore) had a sea surface temperature jump of about 1°C, with warmer water offshore. This feature, apparently a remnant upwelling front from an event peaking around April 2, had little apparent signature in the 21-m currents. One
possible exception is the southward jet on the northernmost transect. The surface drifters, on the other hand, hinted at some interesting structure. The two drifters directly west of Point Arguello both appeared to be converging on the front from the inshore (cold) side. In contrast, two drifters just offshore of the front appeared to be in an equatorward surface jet associated with the front.

A second front was evident at the extreme offshore end of the G line. This feature was perhaps more evident in the 21-m currents than in the 2-m temperatures, as a pronounced jet and velocity shear ($\sim 10^{-4} \text{ s}^{-1}$, comparable to the Coriolis parameter) occurred at the front. The only nearby surface drifter again appeared to be converging on the front from its cold water (shoreward) side. This particular front can be investigated in more detail by studying the concurrent temperature (XBT) and DL alongshelf current sections.

The surface temperature front 35 km offshore was associated with the doming of isotherms (Figure 7a) in the upper 100 m of the water column. Subsurface isotherms (e.g., $12^\circ \text{C}$) reached their shallowest depths about 25–30 km offshore. Maximum horizontal velocities to the southeast, seaward of this point, and to the northwest, inshore of this point (Figure 7a), were associated with strong horizontal temperature gradients. The boundary between these two oppositely directed flows extended to at least 100 m. If temperature can be used as a proxy for density, then the velocity shear in the vertical of these two horizontal flows is in qualitative agreement via the thermal wind with the isotherms sloping down and away from their shallowest levels $\sim 30$ km offshore. A quantitative test of thermal wind balance is presented in section 5.

The next complete map of the region (Figure 8) was made on April 12, when the upwelling-favorable (southeastward) wind stress was approaching a maximum (Figure 5). Cold surface temperatures were already appearing close to shore, particularly along the G line, at the estimated location of the local upwelling center [e.g., Atkinson et al., 1986]. Water cooler than $13^\circ \text{C}$ was found along the entire 50-km length of the G line, suggesting an incipient upwelling filament. The westward jet from the SBC appeared to deflect southeastward along the G and GC lines in a manner roughly consistent with offshore advection of the cold water. At the southeastern corner of the map, there was a suggestion of water warmer than $13.5^\circ \text{C}$ being advected northeastward into the area of eastward flow into the SBC. Interestingly, even though upwelling was apparently active, there were no obvious fronts observed.

The April 14–15 map (Figure 9) was made immediately after the peak of the upwelling event, but while southeastward wind stress was still near 1 dyn cm$^{-2}$ (Figure 5). The cold filament was strongly developed by this time, represented by 2-m temperatures less than $11.5^\circ \text{C}$ (Figure 9). Again, the coldest water was found near the coast on the G line. The 21-m currents and surface drifter velocities appeared to show a pattern of convergence onto the nearshore end of the G line and then an offshore jet with peak speeds greater than 60 cm s$^{-1}$. A front ran nearly east-west south of this feature, with some evidence for an eastward jet on its offshore side. Surface drifter velocities south of this front in conjunction with DL velocities along the offshore ends of the C and GC lines suggest that this mild ($\sim 30$ cm s$^{-1}$) jet appeared to be part of a larger clockwise flow pattern carrying water from the south into the SBC. Fronts in the nutrient and biological fields were observed to be associated with this temperature front as is discussed in detail by B.
H. Jones et al. (manuscript in preparation, 1987). On the northern side of the filament there was a less abrupt (non-frontal) but strong onshore-offshore temperature gradient. Throughout this gradient region, there was an energetic (up to 75 cm s$^{-1}$) southward flow which deflected westward to parallel the southwestward path of the upwelling filament. Thus the only sharp front on the side of the filament was along the equatorward side, just as was found by Flament et al. [1985] at a similar feature farther to the north.

At greater depths (Figure 10), the flow pattern remained similar, except that velocities were weaker and the G line flow was oriented more toward the northwest than the southwest. Flow along the AG line at 99 m was north-northeast, in opposition to the 21-m currents. A crude lower bound estimate of the transport down the axis of the filament is about $0.5 \times 10^6$ m$^3$ s$^{-1}$, a good deal smaller than down-filament transports estimated by Koosro and Huyer [1986] and Koosro [1987] for the region near Point Arena, California, during midsummer 1981 and 1982. The transport will be addressed again in more detail below.

Following the April 14–15 map, the wind stress weakened and then turned strongly poleward for a brief period during April 17–20 (Figure 5). Aircraft sea surface temperature charts (not shown) suggest that the upwelling filament of April 14–15 had largely disappeared by April 16, breaking up into a few small (~10-km diameter) patches [see Atkinson et al., 1986]. The next well-resolved survey was conducted on April 20–22, after the end of the downwelling event. The sea surface temperature pattern at this time was relatively featureless and disorganized (Figure 11b). In contrast to the temperature pattern, the April 20–22 21-m current structure was fairly well defined (Figure 11a), with a westward jet proceeding out from the SBC, along the shelf break and then due west and northwest from Point Arguello. Surface drifter velocities were generally weaker than the 21-m velocities. This appeared to be a fairly representative situation during weak or downwelling-favorable wind stress.

A few conclusions can be drawn from this case study. First, the westward jet extending from the SBC appeared to be a very persistent feature, although it did meander somewhat, especially during the peak of the upwelling event. Second, the flow over the shelf tended to be a good deal weaker than that farther offshore. The shelf region, interestingly, was where concrete evidence for traditional wind-driven upwelling was obtained [Brink and Muench, 1986]. In water depths greater than about 100 m, flow patterns tended to be more complex and less obviously related to the winds in any simple sense. This contrast between shelf and broader scale flow patterns off the U.S. west coast has been noted by other authors as well [e.g., Dye, 1986].

This study of the current’s response to the upwelling-favorable wind event of April 10–15 raises the question of the mass balance for the offshore squirl of cold water. While a detailed balance proved elusive with the present sampling scheme, a conceptual picture can be presented based on the available data (Figure 12). A T-S analysis of the surface upwelled water found along the axis of the squirt establishes the source of the shoreward 2/3 of the cold squirl to be from a depth of approximately 50 m in the SBC. Farther out the squirt, T-S properties of the upwelled water become like those of the southward flowing coastal current north of Point Arguello. Identification of water masses based on T-S properties is not expected to be influenced by local heating, since salinity provides the main distinctions. The boundary between these two water masses is shown by the dotted line in Figure 12. The coldest water anywhere in the study area
Fig. 11. Same as in Figure 6 but for data collected during April 20–22, 1983. SST is from an aircraft overflight which took place on April 21, 1983 (courtesy of D. Stuart).

was observed between the points, lending further proof to the existence of intensified local upwelling. A simple transport estimate due to offshore Ekman flow along the coast between the points yields a value of $0.06 \times 10^6$ m$^3$ s$^{-1}$. Earlier, a lower bound estimate of $0.5 \times 10^6$ m$^3$ s$^{-1}$ was made for the total flow out the axis of the cold squint between points C and D indicated on Figure 12. Clearly, another source of water is required to balance mass.

An estimate of the transport in the southward flowing coastal current west of Point Arguello between points A and B is $2 \times 10^6$ m$^3$ s$^{-1}$. This southward coastal current carries water upwelled farther north along the coast and appears as a cold band next to the coast north of Point Arguello in satellite infrared SST advanced very high resolution (AVHRR) images and SST maps derived from aircraft overflights of the area [Atkinson et al., 1988]. This flow was not observed to cross the G line (Figure 9(a)) so it must either downwell (highly unlikely) or deflect to the southwest. Our sampling scheme crossed part of this deflected coastal current along the central G line, as is indicated by the water mass boundary discussed above. Thus our lower bound estimate of $0.5 \times 10^6$ m$^3$ s$^{-1}$ flowing out the axis of the squint includes part of the transport attributable to the deflected coastal current. We hypothesize that much of the coastal current is deflected seaward between points A and C and that inclusion of this transport would make the total transport offshore in this cold feature on the order of $2 \times 10^6$ m$^3$ s$^{-1}$. This would bring the estimate of offshore transport in this seaward jet in line with Kosro and Huyer's [1986] and Kosro's [1987] values for similar features off the
central Californian coast. They suggested the source of mass for the seaward jets be deflected coastal currents. Our measurements here seem to confirm this scenario. Apparently, then, while most of the transport in this offshore jet is attributable to the deflected coastal current, the signal in the hydrographic and biological (B. H. Jones et al., manuscript in preparation, 1987) fields is dominated by local upwelling on the shoreward end of the G line.

Examination of DL velocities at 67- and 99-m depth (Figures 10a and 10f) shows flow to the northwest (orthogonal to the axis of the cold feature) along the G line and the very inshore end of the GC line. This northwestward flow, which is likely a continuation of the westward flow out of the northern half of the SBC, may be supplying the upwelling water inshore between the points. This is consistent with the T-S properties of the upwelled water mentioned before. In conclusion, this three dimensional conceptual model of the mass balance in the cold squirl includes contributions from locally upwelled water supplied from depth in the SBC and from the deflected southward flowing coastal current, which is itself made up, at least partially, of upwelled water.

4. Time Series

To reduce the random error in the individual 10-min DL measurements of velocity, to obtain a large-scale realization of the flow field, and to create time series of currents which can be correlated with other physical parameters (e.g., the wind), the DL data were averaged as follows. The OPUS study area was divided into five regions (Figure 13): block 1, the area north of Point Arguello along the essentially north-south coastline, where an alongshelf southward flow was often observed (Figure 4a); block 2, the area inshore of the 500-m isobath between points Conception and Arguello, where evidence for wind-driven, localized upwelling was obtained (Brink and Muench, 1986); block 3, the area south of Point Arguello, west of Point Conception, and offshore of block 2, representing the flow region in the deep waters off the shelf; block 4, the northern half of the SBC, where westward flow was generally present; and block 5, the southern half of the SBC, where eastward flow was generally observed (Figure 2a and Brink and Muench, [1986]). Because of the sampling program detailed in section 2, the OPUS study area was traversed fully or partially during 16 independent time periods (Table 1). The spatial coverage during these sampling periods ranged from complete mapping (north of Point Arguello to well into the SBC) to measurements only on the G line (entirely contained by blocks 2 and 3). The length of these 16 time periods ranged from 8 to 48 hours, with the average being around 25 hours (Table 1).

During each of these time periods, all the 10-min DL measurements of velocity occurring in a particular block were

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<td>18:58</td>
</tr>
<tr>
<td>4</td>
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<td>April 12, 0124</td>
<td>April 11, 1157</td>
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<td>5</td>
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<td>April 12, 1710</td>
<td>April 12, 0917</td>
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<td>April 15, 1515</td>
<td>April 15, 0505</td>
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<td>April 18, 1520</td>
<td>April 19, 1048</td>
<td>April 19, 0004</td>
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<td>April 20, 1610</td>
<td>April 20, 0141</td>
<td>28:56</td>
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<td>April 22, 1600</td>
<td>April 21, 1605</td>
<td>47:50</td>
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<tr>
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<td>April 24, 0215</td>
<td>April 23, 0908</td>
<td>34:15</td>
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<tr>
<td>12</td>
<td>April 24, 0215</td>
<td>April 24, 2230</td>
<td>April 24, 1223</td>
<td>20:15</td>
</tr>
<tr>
<td>13</td>
<td>April 24, 2230</td>
<td>April 25, 2010</td>
<td>April 25, 0920</td>
<td>21:40</td>
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<tr>
<td>14</td>
<td>April 25, 2010</td>
<td>April 27, 1830</td>
<td>April 26, 1920</td>
<td>46:30</td>
</tr>
<tr>
<td>15</td>
<td>April 27, 1830</td>
<td>April 28, 0240</td>
<td>April 27, 2235</td>
<td>8:10</td>
</tr>
<tr>
<td>16</td>
<td>April 28, 0240</td>
<td>April 29, 0140</td>
<td>April 28, 1410</td>
<td>23:00</td>
</tr>
</tbody>
</table>

All dates are in 1983.
Figure 13 shows mean 21-m DL velocities computed over all 16 time intervals from April 7-29, 1983. Mean winds for this period were upwelling-favorable (southeastward), but only weakly so. In contrast, mean currents flowed westward from the northern half of the SBC and continued westward in blocks 2 and 3 against the east-west component of wind stress. Just as in the mean vertical sections along the A and G lines (Figures 3 and 4), there seemed to be a convergence of 21-m alongshelf flow southwestward of Point Arguello. A strong mean eastward flow into the SBC was observed in block 5.

Time variations of 21-m velocities are illustrated by block averages computed for individual sampling periods. For example, block-averaged velocities computed during April 14-15 at the height of the upwelling-favorable wind event discussed in section 3 are displayed in Figure 14 which can be compared to Figure 9. Strong (>35 cm s⁻¹) southward flow existed north of Point Arguello as indicated by both the DL and drifter measurements of velocity. This flow was primarily along-isotherm (Figure 9c). Southwestward flow existed in blocks 2 and 3 as indicated by the DL measurements, CHA currents, and the surface drifters. This offshore flow was directed out the axis of the cold front feature discussed in section 3 (Figure 9b). Convergence of upper level waters from the north and from the SBC into the cold southwestward squirt is apparent in Figure 14. The SBC outflow-inflow was present as observed in the mean (Figure 13). Note the weak (<10 cm s⁻¹) northwestward flow on the shelf next to the coast between the points as measured by current meters at P1, C1, and C2. This contrast between local patterns offshore and onshore of the shelf break occurred in several of the 16 time periods.

Typical block-averaged currents during downwelling-favorable (northwestward) or weak winds are represented by spatial averages computed during April 22-24 (Figure 15). Inflow in the southern half of the SBC and strong outflow in the northern half were again observed. This channel flow pattern seemed to be present regardless of the direction of the wind stress, suggesting that it is maintained by processes other than local wind driving. Now, however, instead of southward or southwestward flow between the points there existed strong (>20 cm s⁻¹) northwestward flow, continuous with the SBC outflow and apparently following the local isobaths. Note that flow measured at C1 again did not compare well with either the DL measurement of velocity or the two current meter velocities at CHA and C2.

To quantify the relationship between the different measurements of velocity, the time series were interpolated using the complex inner (rotary) correlation [Mooers, 1973]. The nature of the method used to create the time series did not allow computations of lagged correlations between the block averages. As a result, independence time scales of the correlations were estimated using current meter data. Correlations between the block-averaged DL data and the appropriate current meters indicate good agreement between the two different measures of velocity, suggesting a successful reduction of error in the DL measurement. DL currents in block 2 between the points are well correlated with current meter velocities at CHA (correlation coefficient r = 0.73, DL velocities 5° to the left of current meter velocities) at zero lag (Figure 16). (They are well correlated at greater than 95% confidence. The number of degrees of freedom was estimated by using the technique of Davis [1976].

Fig. 14. Average currents during April 14-15, 1983. Base of block-averaged 21-m Doppler log velocity vectors appears at average position of all the samples occurring within a particular block. The number at the base of a velocity vector denotes the number of 10-min records making up the average. Block-averaged surface drifter velocities are plotted with the number of velocity estimates from adjacent fixes included in the average appearing at their bases. Average subsurface current meter velocities are also plotted. Wind stress at NC appears in the upper right.
on the major axis components of the two time series. This method is ad hoc but appears conservative because time scales associated with major axis components were generally longer than for minor axis components.) The block 2 DL current velocities are not significantly correlated with the current meters at C1 \( (r = 0.40, \text{DL velocities }^9 \text{ to the left of current meter velocities}) \) and C2 \( (r = 0.50, \text{DL velocities }^8 \text{ to the left of current meter velocities}) \) (Figure 16). Lack of significant correlation between the block-averaged DL velocities and the inshore current meters is probably related to the fact that the DL measurements making up the block average occurred mostly offshore of the shelf break. Moorings C1 and C2 were inshore of the shelf break and often measured flow quite different from that observed farther offshore, as was noted above. Block 4 DL currents are correlated with current meters at P2 \( (r = 0.71, \text{DL velocities }^9 \text{ to the right of current meter velocities}) \) and P3 \( (r = 0.62, \text{DL velocities }^8 \text{ to the left of current meter velocities}) \). Block 4 DL currents are not significantly correlated with P1 \( (r = 0.51, \text{DL velocities }^10 \text{ to the left of current meter velocities}) \), presumably because this was again a shelf mooring. Block 5 DL currents do not correlate significantly with P4 or P6, perhaps because of the infrequent and limited (few 10-min records in each average) DL observations in this area.

Block averages of velocity obtained from the surface drifters can be correlated with the appropriate current meters and with the DL measurements of velocity for the same spatial block. Only in blocks 1, 2, and 3 were there an adequate number of drifter observations to create a meaningful time series. Surface drifter velocity averages in block 2 (Figure 16) are correlated with subsurface velocities at CHA (9 m; \( r = 0.61 \), drifter velocities \( 4^\circ \) to the left of current meter velocities), C1 (20 m; \( r = 0.61 \), drifter velocities \( 33^\circ \) to the right of current meter velocities), and C2 (20 m; \( r = 0.59 \), drifter velocities \( 16^\circ \) to the right of current meter velocities). Surface drifter velocity averages are correlated with subsurface DL velocity averages for blocks 1 \( (r = 0.65 \), drifter velocities \( 29^\circ \) to the left of DL velocities) and 3 \( (r = 0.74 \), drifter velocities parallel to DL velocities). The drifter-DL correlation for block 2 is lower but still marginally significant \( (r = 0.41 \), drifter velocity \( 15^\circ \) to the right of DL velocities). Complex correlations between block-averaged DL 21-m velocities and wind stress are displayed in Table 2. Block 1 and 2 currents correlate with the wind such that the wind led by 2\( \frac{1}{2} \) and 1\( \frac{1}{2} \) days, respectively. The angular offset between the two vector series is less than 5° in each case. Block 3 and 4 currents do not correlate significantly with the local winds. Driving of the currents in the northern half of the western entrance to the SBC by remote winds via

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**Fig. 16.** Time series of average velocities computed for the time intervals indicated by the dark and light bars. (a) Wind stress at NC. (b) CHA 9-m currents. (c) Block 2 21-m block-averaged Doppler log velocities. (d) C1 20-m currents. (e) C2 20-m currents. (f) Surface drifter velocities averaged for block 2.
free poleward propagating waves is a possibility, but Brink and Muench [1986] found no evidence for this mechanism. In fact, their study of the large-scale space-time correlations between currents and winds was unable to resolve the nature of the wind driving (if any) in the SBC. Finally, currents in block 5 also correlate with the wind such that wind led by 2 days and currents were 94° to the right of the winds. Similar results for the relationship between the block-averaged currents and the local winds can be obtained by performing scalar correlations between single components of both the currents and wind. Maximum lagged correlation magnitudes are slightly higher at lags and angular offsets nearly identical to those discussed above.

Complex empirical orthogonal functions (EOFs) [Kundu and Allen, 1976] can be used to describe the spatial correlations between the vector time series of block-averaged DL velocity. These complex modes provide information about relative flow directions but do not indicate absolute flow directions. Before computing the complex EOFs for the DL data, gaps in the five time series were filled using drifter or current meter velocities, relying on the high correlations between the three measures of velocity. Gaps filled in this manner represent 15% of the total data set. Remaining gaps, representing 7% of the total data set, were filled by linear interpolation. The first empirical mode, representing 60% of the variance, is displayed in Figure 17. This mode is significantly correlated with the wind, \( r = 0.56 \), where the currents lag the wind by 1.2 days. The velocity mode is scaled by the velocity mode–wind stress regression coefficient where a 1 dyn cm\(^{-2}\) wind stress drives a current on the order of 15 cm s\(^{-1}\) depending on location. The orientation is such that the wind stress is aligned in its principal axis orientation. The mode shows alongshelf flow in blocks 1, 2, and 4 in response to wind deviations aligned northwest-southeast. The SBC inflow-outflow resembles the flow pattern observed in the individual mean current realizations. The flow in block 3 does not significantly contribute to the structure of this mode. The second empirical mode represents less than 25% of the variance and does not correlate significantly with the wind. Alternatively, real EOFs can be computed using the block-averaged east-west and north-south velocity components together. The structure and scalar correlation with the major axis component of the wind nearly duplicates the results for the complex modes discussed above. Multiple regression analysis shows that the major axis component of the wind maximizes the regression of winds and currents. Most importantly, information on absolute flow direction available from the real EOFs confirms the orientation displayed in Figure 17. We choose to present the complex mode because it accounts for a larger part of the observed variance. The EOF analysis has served to extract a wind-driven mode of variability with flow parallel to the coast and limited to the nearshore region. This more traditional along-isobath shelf flow contrasts the flow pattern present during a strong upwelling-favorable wind event as described in section 3. There is a strong cross-isobath flow extending far offshore (squirt) accompanied along-isobath flow closer inshore.

Spatial averaging of the 10-min DL records has reduced the random error in the measurement technique and yielded useful information about the flow field. Large-scale flow patterns have emerged characterizing the response to strong, upwelling favorable wind events and to weak or downwelling-favorable winds. Block averaging creates time series which, when correlated with the wind, showed evidence for local wind driving in an area between the points and north of Point Arguello. A complex EOF analysis shows the dominant mode of variability to consist of a coherent alongshelf flow in regions bordering the points with opposite flow in the southern half of the SBC. This mode is correlated with the wind such that a northwest-southeast wind deviation drives locally alongshelf flow.

5. DISCUSSION

The DL system yields an independent measure of velocity which can be used to test relations between velocity and the hydrographic fields. Specifically, a test of the thermal wind relation, a common method used to estimate velocity from

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**Table 2. Wind Versus Block Averages of 21-m Doppler Log Velocities**

<table>
<thead>
<tr>
<th>Block</th>
<th>( r )</th>
<th>( \theta ), deg</th>
<th>( r )</th>
<th>( \theta ), deg</th>
<th>Lag, days</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.18</td>
<td>-78</td>
<td>0.63</td>
<td>4</td>
<td>2.50</td>
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<td>-82</td>
<td>0.64</td>
<td>-6</td>
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</tr>
<tr>
<td>3</td>
<td>0.50</td>
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<td>0.56</td>
<td>-44</td>
<td>-1.00</td>
</tr>
<tr>
<td>4</td>
<td>0.10</td>
<td>-42</td>
<td>0.61</td>
<td>-94</td>
<td>-2.76</td>
</tr>
<tr>
<td>5</td>
<td>0.62</td>
<td>-94</td>
<td>0.68</td>
<td>-94</td>
<td>2.00</td>
</tr>
</tbody>
</table>

*Significant at 95% confidence level.

Lag is positive for wind leading; angle is positive for wind to the right of u.

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**Fig. 17. First complex empirical orthogonal mode of spatially averaged 21-m Doppler log velocity. The mode is scaled by the velocity mode–wind stress [at NC] regression coefficient. The orientation is such that the wind stress \( r \) is aligned in its principal axis orientation.**
the density field, will be performed. Second, a relationship between velocity and the gradient of temperature on a level surface will be investigated. This latter test is interesting because a consistent relation between $u$ and $\nabla T$ would be helpful for estimating velocity directly from readily available satellite sea surface temperature maps.

5.1. Thermal Wind Comparisons

Density sections along lines A, G, and C (Figure 1) were used to calculate geostrophic velocity between station pairs typically separated by 3–4 km. A reference level of 350 m was chosen to allow dynamic height calculations in the SBC. A deeper reference level makes a negligible difference, less than 2 cm s$^{-1}$. Dynamic heights for stations shallower than 350 m were calculated using the Montgomery extrapolation technique described by Reid and Mantyla, [1976]. Comparisons of the DL and geostrophic velocities for individual CTD sections (not shown) are noisy (velocity differences of up to 45 cm s$^{-1}$), but often yield qualitative agreement. Major differences between the two velocity fields may be due to the presence of ageostrophic internal waves, inertial motions, and other small-scale, time-dependent motions and/or the choice of reference level and/or the extrapolation technique. Averages over many transects were computed for both velocity fields along each line in an effort to reduce the ageostrophic noise and velocity differences (in the vertical) were calculated to eliminate any offset due to the choice of reference level or use of the extrapolation technique. Specifically, the DL velocity at the bottom of each average profile was subtracted from the DL measurements of velocity above. The geostrophic velocity at the same depth as the deepest DL observation was subtracted from the rest of the geostrophic velocity profile above. Velocity shear is easily obtained from these fields by dividing the velocity difference between two levels by the vertical distance between them. For the remainder of this section the term velocity "shear", however, will be used to mean vertical velocity "difference".

Figures 18 and 19 show average DL and geostrophic velocity shears for the G (an average of 7 individual sections) and C (an average of 4 individual sections) lines, respect-
TABLE 3. Ratios of Accelerations to Coriolis Forces

<table>
<thead>
<tr>
<th>Current Meter Location</th>
<th>Hourly</th>
<th>Low-Passed*</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>s.d. ( \text{u}_h )</td>
<td>s.d. ( \text{u}_l )</td>
</tr>
<tr>
<td>P2 (30 m)</td>
<td>0.72</td>
<td>2.18</td>
</tr>
<tr>
<td>P2 (135 m)</td>
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<td>1.57</td>
</tr>
<tr>
<td>Average</td>
<td>1.06</td>
<td>1.87</td>
</tr>
</tbody>
</table>

Standard deviation in an alongshelf \( (\text{e}) \)-crossshelf \( (\text{u}) \) coordinate system calculated for the time period April 7-29, 1988.

*Half power point of 1.96 days.

tively. Coordinate systems have been rotated to local alongshelf orientations (see insets, Figures 18 and 19). Results from line A are not presented, since it was only occupied twice while DL and CTD data were simultaneously being collected. Geostrophic and DL velocity shears along the G line show good qualitative agreement, with positive values inshore of 25 km associated with mean poleward flow (Figure 3a). Maximum positive velocity shears of between 4 and 12 cm s\(^{-1}\) were located 17 km offshore between depths of 20 and 60 m. Negative velocity shears existed offshore of 25 km with maximum values of \( | -14 | \) cm s\(^{-1}\). The geostrophic velocity shear field showed strong positive values near the offshore end of the G line which were absent in the DL measurements. The C line velocity shears again showed good qualitative agreement, with a region of positive velocity shear associated with westward flow out of the SBC north of 15 km offshore and a corresponding region of negative velocity shear, associated with eastward flow, on the opposite side of the SBC (Figure 2a).

A quantitative comparison over a 104-m depth interval between the directly measured velocity shear and velocity shear calculated from the thermal wind relation was performed. The comparison was done for 18 mean profiles along the G and C lines. The mean of the difference between the two measures of shear \( \Delta \text{u}_{DL} - \Delta \text{u}_{geostrophic} \) was 0.1 cm s\(^{-1}\) with a standard deviation of 0.7 cm s\(^{-1}\). Correlation between the two shears was 0.77. Geostrophic velocity shears can also be compared to velocity shears measured by moored current meters. The comparison was done for the mean computed over three C lines only, using moorings P1-P5 (Figure 1). Velocity shears between current meters at a depth of 30 m and the deepest instrument on each mooring (P1 and P5, 65 m; P2 and P4, 135 m; and P3, 350 m) were compared to geostrophic velocity shears over the same depth intervals. The mean of the difference between the current meter observations and the geostrophic calculations \( \Delta \text{u}_{CM} - \Delta \text{u}_{geostrophic} \) was \(-2.5\) cm s\(^{-1}\), with a standard deviation of 6.5 cm s\(^{-1}\). Correlation between these two measures of velocity shear was 0.70. Comparison between the two direct measures of velocity shear, DL and current meter, presented in section 2 showed excellent agreement over the average of seven C line crossings (0.2 cm s\(^{-1}\) mean and 1.9 cm s\(^{-1}\) standard deviation).

Krosz and Huyer [1986] performed a similar comparison of velocity shear for a larger survey near Point Arena farther north along the California coast. They smoothed both the geostrophic and DL velocity fields with a Gaussian filter with a smoothing length of 20 km and obtained vector correlations in the range 0.8-0.9. The lower correlation and high standard deviation (when compared with the standard deviation of the DL-current meter shear comparison) found during OPUS are probably due to the inability to filter out submesoscale, ageostrophic motions when sampling on such short space scales.

If geostrophy is not a good description of the momentum balance for observations collected in the OPUS study area, then other terms in the momentum equations must be important. Time series of velocity from current meters on moorings P1-P3 were used to compare the size of the acceleration terms in the momentum equations to the size of the Coriolis forces. Representative ratios computed at mooring P2 for the time interval during which the DL-geostrophic comparison was made are presented in Table 3. For hourly data, \( | \text{u}_h | \) and \( | \text{f}_h | \) (where \( \text{u}_h \) is the acceleration of northward flow and \( \text{f} \) is the Coriolis parameter) are of comparable size, suggesting a lack of geostrophy. For low-passed (half power point of 1.96 days) data, \( | \text{u}_l | \) \( \ll \) \( | \text{f}_l | \), meaning the alongshelf component of flow can be in geostrophic balance. For both hourly and low-passed data, \( | \text{v}_l | \approx | \text{f}_l | \), so the cross-shelf component of velocity will not be geostrophic in either case. Since the DL sampling frequency (one record every 10 min) is comparable to the hourly current meter data, the alongshelf observed momentum balance cannot be expected to be geostrophic.

5.2. Comparison of Currents With Temperature Gradients

The purpose of relating the velocity field to temperature gradients is not to establish any particular dynamics, but to investigate the possibility of an operational scheme for estimating the velocity field from temperature information alone. For example, Kelly [1983] has explored estimation of near-surface currents from satellite surface temperature observations. Temperature gradient fields at 21 m were calculated from XBT measurements by interpolating temperature onto a regular grid, smoothing and using finite differences to approximate the derivatives. To relate velocity and VT, the complex inner correlation [Mooers, 1973; Kundu, 1976] was employed. The resulting correlation is a complex number with modulus representing the degree of correlation and phase representing the angular offset between the two vector quantities. The important point in choosing this measure of mean veering is that each observation point is weighted by the magnitude of the individual vectors. That is, velocities which are strong (well above the noise level) dominate in the estimated mean veering. Likewise, data points in regions of weak velocities and VT with supposedly random orientations between the two vectors contribute little to the overall measure of veering.

Correlations were performed for each of seven maps which included observations on all five sampling lines. In addition, for each map, correlations were calculated on each line separately to identify regions with different \( \text{u}, \text{VT} \) relationships. Velocity and temperature at 21 m obtained at the end of a strong upwelling-favorable wind event described in detail in section 3 are displayed in Figure 20. Correlations and veering angles computed using the bold DL vectors only are displayed at the end of each line and for the map as a whole. An angle near 0° (flow cold to warm) or 180° (flow warm to cold) indicates cross-isotherm flow, while an angle near ±90° indicates along-isotherm flow. Note that if temperature was an exact representation of density and isotherms remained parallel with depth then a "geostrophic" relationship between
Fig. 20. Complex correlations between velocity and temperature gradient for data collected during April 14–15, 1983. Doppler log velocities at 21 m (same as in Figure 6a) are plotted on top of 21-m temperature contours obtained from an XBT survey. Correlation coefficient and angular offset (r, θ) appear at the end of each sampling line where only the bold velocity vectors have been used in the computation. Correlation parameters for the map as a whole appear in the lower left corner. Wind stress at NO appears at the upper right.

Fig. 21. Same as in Figure 20 but for data collected during April 24–25, 1983.

### Table 4: Complex Correlation Between u and VT

<table>
<thead>
<tr>
<th>Sampling Interval*</th>
<th>Number of Points</th>
<th>r</th>
<th>θ, deg</th>
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</tr>
<tr>
<td>5</td>
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</tr>
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<td>7</td>
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<td>0.50</td>
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</tr>
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<td>9</td>
<td>83</td>
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</tr>
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<tr>
<td>15</td>
<td>32</td>
<td>0.42</td>
<td>-169</td>
</tr>
</tbody>
</table>

*See Table 1 for UT times.

u and VT would yield an angle of −90° for the definitions adopted here. High correlations exist for the northernmost three lines in Figure 20. Flow off Point Arguello had equal cross- and along-isotherm components, while the flow on the AG line was mostly along-isotherm in a “geostrophic” sense. Along the center of the upwelling squirl, flow was almost completely cross-isotherm, and the offshore jet apparently flowed out the axis of the cold thermal feature. The low correlations on the GC and C lines makes interpretation of the mean veering angle suspect. For the whole map, a substantial correlation is found with a mean veering angle almost completely along isotherm in a “geostrophic” sense.

Velocity and temperature at 21 m obtained during a period of northwesterly winds are displayed in Figure 21. There was little structure to the temperature field except that in contrast to the temperature field during upwelling-favorable winds, the warmest water was found between the points. Consistent relations are found on four of the five lines. Lines A, AG, and G all had flow with comparable cross- and along-isotherm components. Line GC also showed flow with cross- and along-isotherm components but in the opposite sense to the above three lines: cross-isotherm flow from cold to warm. The correlation for the whole map is low but again shows a mean veering angle indicating flow almost along isotherm in a “geostrophic” sense.

Results for all seven maps (Table 4) indicate that there was no consistent mean veering angle between u and VT in this region. One can, however, identify local regions and times where there was a consistent veering angle between u and VT. It must be kept in mind, though, that no filtering of the velocity field has been done to remove signals due to small-scale velocity features which may have no direct relation to the smoothed, large-scale temperature field. Correlations may be higher and veering angles more consistent with smoothed fields.

In conclusion, lack of a consistent relation between velocity and temperature gradient may make it difficult to estimate flow fields from thermal imagery alone in the complicated Point Conception area. Calculation of along-isotherm flow through the assumption of a thermal wind balance using temperature as a proxy for density can, at best, recover a part of the velocity field. Further, estimation of horizontal cross-isotherm flow by assuming conservation of temperature will certainly be inaccurate in an area of active upwelling.

### 6. Conclusions

Shipboard acoustic DL velocity measurements have been used to characterize the flow near Point Conception during April of 1983. The data provided realizations of the velocity...
field over the large spatial scales essential for describing the contrast between flow on and off the shelf. Current meter measurements on the shelf [Brink and Muench, 1986] showed a classic wind-driven upwelling response, but offshore flow, as revealed by acoustic DL velocity measurements, consisted of eddies, jets, and fronts which are not easily described by traditional linear coastal circulation theory. Any successful characterization of circulation in this area must include both of these flow regimes.

Some conclusions can be made from this analysis. Persistent eastward flow into the southern half of the SBC and westward flow out in the northern half were observed. Currents responded during one particular upwelling-favorable wind event in the form of an energetic offshore squall (>60 cm s⁻¹) of cold water. Evidence for convergence of northern coastal current water and westward flowing water from the northern half of the SBC onto the squall was found. In situ velocity and hydrographic data confirmed the existence of this squall of cold water to at least 50 km offshore, while satellite infrared SST maps [Atkinson et al., 1986] suggested its continuation to at least 100 km offshore.

A conceptual model for the mass balance in this particular offshore jet of cold water included a contribution due to local upwelling, but most of the transport out the axis of the squall was apparently from the deflected equatorward flowing coastal current. A transport estimate for the coastal current west of Point Arguello between points A and B (Figure 12) is 2 × 10⁶ m³ s⁻¹. Measurements out the axis of the cold jet included only part of the deflected coastal current, with the remainder inferred to turn offshore. Inclusion of the total deflected coastal current will bring the estimate of offshore transport of cold water to > 2 × 10⁶ m³ s⁻¹. This is in line with estimates made by other investigators for similar seaward jets off the west coast of the United States [Koero and Huyer, 1986].

In contrast, during periods of weak or downwelling-favorable winds, continuous flow was observed emanating from the northern half of the SBC and flowing to the northwest following the local isobaths until the latitude of Point Arguello. On average, this flow proceeded directly offshore west of Point Arguello, but several individual realizations of the velocity field suggested continuation of this flow poleward of Point Arguello.

Correlations between spatially averaged DL velocities and the local winds confirmed the existence of wind-driven flow north of Point Arguello and inshore between the points. Farther offshore and in the SBC, no meaningful correlation between local winds and currents was found. Brink and Muench [1986], after obtaining a similar result between local winds and currents in the SBC, were unsuccessful in an attempt to find evidence for remote wind driving of the currents. They concluded that fluctuating currents in the SBC may be associated with large-scale meteorological patterns, but the actual driving mechanisms remain unknown.

A comparison of directly measured DL velocity shears with velocity shears estimated from the density field in this area showed a lack of thermal wind balance. High-frequency motions generated acceleration terms in the momentum equations large enough to negate a geostrophic balance. No consistent relation between DL velocity and VT was found, exemplifying the difficulties of estimating velocity from temperature information alone in this area.

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REFERENCES
Koero, P. M., Shipboard acoustic current profiling during the Coastal Ocean Dynamic Experiment, SIO Rep. 85-8, 119 pp., Univ. of Calif., San Diego, La Jolla, 1985.
Kundu, P. K., Ekman veering observed near the ocean bottom, J. Phys. Oceanogr., 6, 238–242, 1976.


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