Wind Stress Curl Forcing of the Coastal Ocean near Point Conception, California*

ANDREAS MÜNCHOW
Institute of Marine and Coastal Sciences, Rutgers, The State University of New Jersey, New Brunswick, New Jersey

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ABSTRACT

Near Point Conception, California, the atmospheric flow separates from the coast and a large wind stress curl results. Direct spatial wind field observations from 20 aircraft overflights in the spring of 1983 suggest that Ekman pumping of on average 4 m day$^{-1}$ contributes to the local dynamics. During strong and persistent upwelling events the curl-driven Ekman pumping reaches up to 20 m day$^{-1}$. A single complex empirical orthogonal function explains more than 72% of the spatial and temporal wind stress variance. It reveals large wind stress curl at the center of the western Santa Barbara Channel entrance. This dominant mode correlates strongly ($r^2 = 0.79$) with the wind stress observed at a moored buoy. The location of largest Ekman pumping in the ocean indicated by this mode coincides with the location of a laterally sheared, cyclonic flow. Cold upwelled waters enter the Santa Barbara Channel along its southern perimeter while warm, and thus buoyant, waters from the Southern California Bight exit the channel along its northern perimeter. Buoy wind stress and lateral current shear at 30-m depth correlate significantly at periods of about 3.5 and 6 days with phase lags of about 0.5 and 2 days, respectively. Hydrographic observations from 1983 do not, however, indicate effects of Ekman pumping on the internal mass field as winds vary in speed and direction at daily timescales. This contrasts with 1984 observations that do indicate strong doming of isopycnals over the center of the channel at its western entrance. Winds prior to and during hydrographic observations in the spring of 1984 were both stronger and more steady than they were in 1983. Cyclonic shears reach 0.4$f$ at the entrance of the Santa Barbara Channel near Point Conception; here $f$ is the planetary vorticity. The thermal wind balance explains the lateral shear of the alongchannel surface flow rather well.

1. Introduction

The study of wind-forced motions in the coastal ocean has a long history that probably began with Fridtjof Nansen who, during the 1893–1896 Norwegian expedition to drift with the ice across the Arctic Ocean, discovered peculiar ice motions to the right of the wind. He subsequently attributed this motion to the earth’s rotation (Nansen 1902). His qualitative conjecture motivated Walfried Ekman to study this effect using the laws of motion that Bjerknes (1901) had just formulated for the ocean and atmosphere. A landmark paper resulted (Ekman 1905), that, I believe, marked the beginning of coastal physical oceanography. Three decades later, another Arctic explorer, Harald Sverdrup, was the first to quantify wind-forced upwelling along the California coast. Using an inferred vertical upwell-
perturbations can then propagate upstream and their kinematic energy is often trapped near such regions. The wind field along the coast near Point Conception has been analyzed by Caldwell et al. (1986) with time series data. They emphasize temporal variability, but nevertheless find a wind field that varies along the coast. Their data, however, are inadequate to study the effect of the local wind stress curl since most of their wind stations are on land. The data thus do not resolve the wind field over the ocean where most of the spatial variability occurs. In a different study, Eddington et al. (1992) model the winds near Point Conception with a layer model that features an active lower marine layer and a stagnant atmosphere above. Their process study successfully reproduces both a wind maximum south of Point Conception and the large horizontal wind shears across the western entrance of the Santa Barbara Channel. Similar observational and theoretical studies are reported for the CODE region off Northern California by Enríquez and Frihe (1995) and Samelson (1992).

Here I investigate the spatially variable wind field at the western entrance of the Santa Barbara Channel near Point Conception. More specifically, I will discuss circumstantial evidence to present the hypothesis that the wind stress curl near Point Conception contributes to the oceanic circulation in the western Santa Barbara Channel. This study statistically analyzes wind data from aircraft overflights in the spring of 1983, demonstrates a strong correlation between wind and wind stress curl, shows that atmospheric conditions in the spring of 1983 are similar to those in 1984, and, finally, interprets oceanic conditions in the spring of 1984 with the wind patterns of 1983 in mind. The wind stress curl near Point Conception will emerge as one of several potentially important forcing mechanisms for the circulation in the western Santa Barbara Channel.

2. Data sources and study area

Aircraft wind observations constitute the main data source of this study. In April and May of 1983 the NCAR aircraft Queen Air surveyed the wind field near Point Conception. Friehe et al. (1984) describe the instrumentation aboard the aircraft and its capability to measure winds over the coastal ocean off northern California. The appendix presents details of the wind calibration, the estimation of wind stresses, and the errors in derived quantities. Figure 1 shows the track of each of 20 airborne wind field surveys that took about 3–4 hours to complete. The sea breeze in the study area can reach amplitudes of 1–3 m s$^{-1}$ (Brink et al. 1984; Caldwell et al. 1986), however, these winds diminish rapidly away from the coast. Nevertheless, some temporal aliasing is certainly present even though the synoptic winds are an order of magnitude larger than the sea breeze. Stuart and Linn (1983) describe the initial processing, which includes the interpolation of the data onto a regular grid with distances $\Delta x = 5.7$ km and $\Delta y = 3.7$ km where $x$ and $y$ represent longitude and latitude,
respectively. The gridded wind stress field is a function of position \((x_i, y_j)\) and time \(t_k\); that is, \(\mathbf{\tau} = \mathbf{\tau}(x_i, y_j, t_k)\) where \(i, j, k\) are integers from 1 to \(N, M,\) and \(K\). A bold variable, such as \(\mathbf{\tau}\), represents a vector while a nonbold variable denotes a scalar. An overbar indicates a temporal average; that is,

\[
\mathbf{\tau}(x, y) = \frac{1}{K} \sum_{k=1}^{K} \mathbf{\tau}(x, y, t_k)
\]  

(1)

and a vertical bar represents a spatial average; that is,

\[
|\mathbf{\tau}(t)| = \frac{1}{NM} \sum_{i=1}^{N} \sum_{j=1}^{M} \mathbf{\tau}(x_i, y_j, t).
\]  

(2)

The spatially variable part of a field can then be written as \(\mathbf{\tau}_c(x, y, t) = \mathbf{\tau}(x, y, t) - |\mathbf{\tau}(t)|\). The spatial average of this property is by definition zero; that is, \(|\mathbf{\tau}_c(x, y, t)| = 0\). The spatial gradients are identical to those of the original field, for example,

\[
\nabla \times \mathbf{\tau}_c(x, y, t) = \nabla \times \mathbf{\tau}(x, y, t).
\]  

(3)

Ekman pumping results from Ekman flux divergence and convergence, which in turn are caused by a spatially variable wind field. In order to conserve mass, a vertical velocity \(w\) results, which is (Gill 1982)

\[
w = k \cdot \left( \nabla \times \frac{\mathbf{\tau}}{\rho} \right).
\]  

(4)

where \(f, \rho, \mathbf{\tau},\) and \(k\) are Coriolis parameter, density of the water, wind stress, and vertical unit vector, respectively.

Current observations are available from instruments that in 1984 were moored inside the Santa Barbara Channel, which is approximately 150 km long, 10–30 km wide, and generally more than 200 m deep. A central basin is deeper than 600 m and two shelf regions are shallower than 200 m and narrower than 15 km. The present study uses current meter and CTD data from the 1984 experiment. Hydrography from 1983 is available, but it reflects conditions during weak and variable wind forcing only. Finally, wind and sea surface temperature data from NDBC meteorological buoy 46023 (shortened to EB23 in the following) complement the above data.

The present study synthesizes data from three observational experiments. Table 1 lists these and five more
TABLE 1. Experiments in the Point Conception area: DOE stands for Department of Energy, OPUS stands for Observations of Persistent Upwelling Systems, SBC stands for Santa Barbara Channel, and CCCCS stands for Central California Coastal Circulation Study. The data used in this study are indicated by an asterisk.

<table>
<thead>
<tr>
<th>Year</th>
<th>Experiment</th>
<th>Instrumentation</th>
<th>Peer-reviewed reference</th>
<th>Figure in text</th>
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<tr>
<td>1964</td>
<td>DOE</td>
<td>Bottles, drifters</td>
<td>Brink et al. (1984)</td>
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<td>AVHRR</td>
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<td>Barth and Brink (1987)</td>
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<td>Drifter</td>
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* Lagerlof and Bernstein (1988) use a map of SST from a CTD survey and record mean current vectors from moored current meters.

Conducted near Point Conception. Peer reviewed papers refer to publications that describe the data in detail where they are available. All except the 1983 OPUS studies were motivated by offshore oil drilling operations. Most of the data listed in Table 1 are archived at the Scripps Institution of Oceanography where it can be accessed through the World Wide Web (http://www-ccs.ucsd.edu/zoo).

3. Spring 1983: Wind observations

Dorman and Winant (1995) statistically analyze the 10-yr record of meteorological buoys deployed in the coastal ocean off the western United States. They describe the wind field at large spatial (>200 km) and temporal (>1 month) scales. They find the strongest winds off Point Conception in the spring. In contrast, this study discusses the local wind field near Point Conception at much smaller spatial (<50 km) and temporal (<1 month) scales in the spring of 1983. The following five subsections introduce the atmospheric conditions with data from rawindsondes (section 3a), discuss three cases of the observed wind fields \( \tau(x, y, t) \) (section 3b), describe the properties of the temporally averaged field \( \overline{\tau}(x, y) \) (section 3c), study all fields \( \tau(x, y, t) \) with a complex EOF (CEOF) analysis (section 3d), and close with a correlation analysis of wind stress and wind stress curl (section 3e).

a. Atmospheric stratification

Figures 2a and 2b show the wind vector and the sea surface temperature at buoy EB23 (see Fig. 1 for location) as a low-pass filtered time series for the time period when the aircraft surveyed the wind field in the lower atmosphere. The (Lanczos) filter has a half-power point at 17 h and passes 90% and 10% of the variances at periods of 27 and 12 h, respectively. It thus removes most of the diurnal variability. Weak (<10 m s\(^{-1}\)) and variable winds occur during April and early May of 1983 (Fig. 2a) while strong (>10 m s\(^{-1}\)) and upwelling favorable winds persist after day 126 (6 May). The sea surface temperature then drops by 3°C within 3 days near day 130 (Fig. 2b). The observational period of the aircraft data thus covers both persistent upwelling and intermittent up- and downwelling-favorable winds. The hydrographic data are more limited to periods of weak and variable winds (Fig. 2b).

At Vandenberg Air Force Base (see Fig. 1 for location) vertical profiles of atmospheric properties are measured twice daily with a rawindsonde. The temperature and pressure data from these profiles allow estimation of the stability or buoyancy frequency \( N \), which describes the vertical density stratification of the atmosphere for descending air (Gill 1982); that is,

\[
N^2 = \frac{g}{T} \left( \frac{\partial T}{\partial z} + \Gamma \right),
\]
Following Winant et al. (1988), I define a local Froude number \( F = U/(ND) \), which constitutes a useful parameter to qualitatively describe the atmospheric flow. Here \( U \) and \( (ND) \) are the particle and the internal phase velocity, respectively; \( D \) represents the vertical scale of motion. The Froude number becomes larger than 1 after day 126 when typical scales are \( U \sim 15 \) m s\(^{-1}\) for the wind velocity (Fig. 2a), \( D \sim 500 \) m for the vertical scale of the motion, and \( N \sim 0.04 \) s\(^{-1}\) for the stability frequency. The flow is then supercritical and no internal waves can propagate against the flow. The energy of internal perturbations is then either trapped near the cape or swept downstream by the flow. The upstream conditions remain unchanged.

### b. Case studies

The strongest upwelling favorable wind event lasting longer than 5 days occurred on day 128 (8 May, Fig. 3a). The spatial mean wind stress reached 0.31 Pa (Fig. 3b); the marine layer had a strong temperature inversion (not shown) and was about 500 m thick. The flow is supercritical (Fig. 2), separated from the coast, and Ekman pumping velocities reached 20 m day\(^{-1}\) off Point Conception (Fig. 3c). The spatial pattern of the large Ekman pumping extended into the western Santa Barbara Channel and was largely a result of the flow separation. Enriquez and Friehe (1995) find similar values for the wind stress curl near Point Arena off Northern California for both the 1982 CODE and 1989 SMILE data.

Four pulses of acceleration and deceleration (Fig. 3b) reveal a structure similar to that of an undular bore. These pulses do not, however, contribute strongly to the wind stress curl. They instead represent variations of the form \( \partial_{\phi} v_{r}/r \) where \( v_{r} \) and \( \phi \) are azimuthal velocity and distance in a radial geometry, respectively. A term \( v_{r} \partial_{\phi} v_{r}/r \), however, indicates the nonlinearity of the atmospheric flow suggestive of hydraulic dynamics (Winant et al. 1988; Samelson 1992).

A less extreme case of large Ekman pumping is depicted in Fig. 4 for day 122 (2 May). The winds are moderately upwelling favorable (Fig. 4a) and the spatial mean wind stress is about 0.14 Pa (Fig. 4b). The spatial pattern of the Ekman pumping (Fig. 4c), however, closely resembles that of Fig. 3c. The flow separation is less abrupt than in the previous case, but the cyclonic circulation in the lee of the cape corresponds to a wind stress curl that causes up to 10 m day\(^{-1}\) Ekman pumping at the entrance of the Santa Barbara Channel.

Finally, Fig. 5 depicts the atmospheric flow in the...
western part of the Santa Barbara Channel as a strong downwelling wind event passes the study area on day 109 (19 April) of 1983. The wind stress field appears trapped near the northern coast (Fig. 5a) where steep mountains rise to more than 1500-m height within 5 km of the coast. The atmosphere is continuously stratified during this event and no inversion is present. The laterally sheared flow, however, again corresponds to a cyclonic circulation to the southeast of Point Conception (Fig. 5b). Ekman pumping velocities at the entrance of the Santa Barbara Channel are thus again positive (Fig. 5c), that is, they indicate curl-induced upwelling by about 4 m day\(^{-1}\). Downwelling of similar magnitude, however, occurs farther to the south. The horizontal scale of the atmospheric flow is about 10 km in the across-stream direction.

c. The “mean” wind field

Figure 6 shows the temporal mean field \(\overline{\tau}(x, y)\) that is, the mean wind stress vector field subtracted from the data. The constant vector \(|\overline{\tau}(x, y)|\) is subtracted from the data. The constant vector \(|\overline{\tau}(x, y)|\) in Fig. 6b represents a scale while the mapped vectors represent the spatially variable part of the wind field; that is,

\[
\overline{\tau}(x, y) = \overline{\tau}(x, y, t) - |\overline{\tau}(x, y)|. \quad (6)
\]

Its dominant feature is the cyclonic circulation to the south of Point Conception.

The temporal mean wind stress causes about 4 m day\(^{-1}\) upwelling through Ekman pumping in an area 10 km wide and 80 km long (Fig. 6c). This area extends into the Santa Barbara Channel. Ekman pumping does not require a coastline and the vertical velocities are generally sheared flow, however, again corresponds to a cyclonic circulation to the southeast of Point Conception. The acceleration and separation of the flow results in a cyclonic circulation, which becomes particularly clear in Fig. 6b. Ekman pumping velocities at the entrance of the Santa Barbara Channel are thus again positive (Fig. 5c), that is, they indicate curl-induced upwelling by about 4 m day\(^{-1}\). Downwelling of similar magnitude, however, occurs farther to the south. The horizontal scale of the atmospheric flow is about 10 km in the across-stream direction.

d. The “variable” wind field

A complex empirical orthogonal function (CEOF) analysis systematically decomposes the spatial wind stress vector fields into a set of ordered statistical modes. Mathematically, these modes are the eigenvectors of the complex (Hermitian) covariance matrix of the observed vector field, which are written in the form \(\tau_t(x, y, t) = \tau_x + i \cdot \tau_y\), where the spatially variable vector \(\tau_t(x, y, t)\) is interpreted as a complex number and \(i^2 = -1\). The zonal and meridional wind stress components \(\tau_x, \tau_y\) represent the real and imaginary part of \(\tau_t\), respectively. The eigenvalue \(\lambda\) to each eigenvector represents the amount of variance explained by this mode. Kundu and Allen (1976) and Kelly (1988) describe the method in detail. The decomposition of the observed field is a statistical, not a physical, decomposition. A common problem in any EOF analysis of temporally and spatially variable fields is to remove the correct means. I could remove either the spatial or the temporal, or both, means in either order. Kelly (1988) suggests removing both means when analyzing spatial variability. Hence the fields analyzed are \(\tau_t(x, y, t) = \tau(x, y, t) - |\overline{\tau}(x, y)|\).

Figures 7 and 8 depict the spatial pattern and temporal amplitudes of the largest three modes, respectively. The temporal amplitudes of Fig. 8 represent the (complex) amplification factors for the spatial patterns of Figs. 7a, 7c, and 7d. A unit vector pointing upward in Fig. 8 represents the pattern shown in Fig. 7 while a unit vector pointing sideways represents the same pattern rotated by 90 degrees. The temporal amplitudes are scaled such that the variance of the amplitude time series for each mode has uniform variance. Each modal pattern is also rotated such that the arrows in Fig. 7 point in the direction of the principal axis of the amplitude time series; that is, principal axes of the time series in Fig. 8 are vertical.

The first three CEOF modes here explain almost 90% of the variance (Fig. 7); however, only the first mode (72.5%) is statistically distinct since, as Fig. 7b indicates, the uncertainty of its eigenvalue \(\Delta \lambda = \lambda(2/N)^{1/2}\) does not overlap with that of any other eigenvalue (North et al. 1982). Here \(N = 19\) is the degrees of freedom; the result does not change for \(N = 6\). In contrast, the eigenvalues of mode 2 (10.8%) and mode 3 (6.4%) do overlap within their error bars and thus are not truly orthogonal of each other. Both are distinct from mode 4, however, and thus will be discussed as a coupled mode.

Mode 1 indicates a clear separation of the flow field near Point Conception. This separation is the main feature of the spatial wind field near Point Conception. The CEOF mode indicates a spatially coherent region of large and spatially variable horizontal Ekman fluxes and thus Ekman pumping. A ridge of minimum flow, but maximum curl, extends into the Santa Barbara Channel from a location at the coast between Point Arguello in
Fig. 3. Maps of (a) surface wind stress vectors, (b) deviations of the surface wind stress vectors from the spatial mean, and (c) vertical Ekman pumping velocities on day 128 (8 May 1983). The spatial-mean vector is displayed in the upper-right corner of (b) as a scale. Note the four pulses of acceleration as the flow turns, emphasized by the four solid lines. The Froude number for this event is larger than 1.5; i.e., the flow is supercritical (see Fig. 2).

Fig. 4. As in Fig. 3 but for day 122 (2 May 1983).
Fig. 5. As in Fig. 3 but for day 109 (19 Apr 1983).

Fig. 6. As in Fig. 3 but for the temporal mean of all 20 maps.
the north and Point Conception 20 km to the southeast (Fig. 7a). This ridge represents the inshore boundary of the flow as it separates from the coast. Closer inspection of Fig. 7a also reveals that the flow accelerates inshore near Point Arguello only to decelerate and accelerate again toward Point Conception. Much weaker wind stress curl of the opposite sign occurs farther offshore as winds and their gradients diminish in that direction.

Figure 8a shows the amplitudes of the wind stress anomalies, demonstrating that all except two maps contribute to the first mode. The two exceptions are the maps from day 104 (14 April) and day 126 (6 May) when the wind stress field is very close to its temporal mean state during the observational period. Mode 1 amplitudes (Fig. 8a) suggest two distinct temporal domains. From day 102 to day 125 the vectors are largely negative; that is, they indicate a downwelling tendency that weakens the mean state (Fig. 6). It reaches its minimum on day 117 (27 April 1983) when the wind field is downwelling favorable for a few days and the wind field is both strong and anticyclonic (not shown). After day 126 the amplitude vectors become positive; that is, they indicate an upwelling tendency that strengthens the cyclonic mean state. Toward the end of the record, when the lower atmosphere becomes vertically stratified (Fig. 2c), mode 1 thus represents a cyclonic wind stress curl during the strong upwelling-favorable wind events. The flow at this time is supercritical; low-level temperature inversions are present at 500-m height. This inversion is absent during the first 20 days of the record.
Fig. 8. Amplitudes of the first three temporal CEOF modes whose pattern is shown in Fig. 7. Units of the amplitude functions are arbitrary. The time series are scaled and rotated such that the total variance of each modal time series is 1 and its principal axis is vertical. The 19 vectors are connected with a solid base.

Fig. 9. Magnitude of the complex vector correlation between CEOF mode 1 amplitude and the wind stress at EB23. (a) The data scatter (symbols) and linear regression (line). (b) EB23 wind stress (line with symbols) and CEOF mode 1 (line only) as a function of time. The latter is scaled to fit the same scale as the buoy stress.

Modes 2 and 3 combine to explain an additional 17.2% of the variance. Together they describe weak convergent/divergent atmospheric flow. Such a spatial feature corresponds to a spatial acceleration/deceleration of the flow. In addition, these two modes appear to contain a weak shear region extending into the Santa Barbara Channel more aligned with the orientation of the channel as compared to the more oblique angle with which the ridge of large curl of mode 1 extends into the channel. The temporal amplitudes (Figs. 8b and 8c) are somewhat scattered and disorganized; that is, these two modes provide both smaller time and space scale variability that modifies the dominant flow described by mode 1.

In summary, mode 1 is strongest and cyclonic following day 126 (6 May 1983). It is at this time that the Ekman pumping would be largest as both the temporally mean (Fig. 6a) and variable (Fig. 7a) winds combine to result in the largest wind stress curl in the lee of Point Conception. The question arises if such enhanced upwelling over the center of the channel can be seen in hydrographic sections across the channel entrance in the form of doming isopycnals. The answer for the spring of 1983 is a “no,” for all such sections were taken during weak, intermittent, and even downwelling-favorable conditions (Fig. 2). Furthermore, the last 1983 OPUS section across the western entrance of the channel was taken on 5 May 1983 (day 125) a few days prior to the onset of the strong and persistent upwelling. The next question then arises: are doming isopycnals over the center of the channel seen in different years and, if so, is it possible to predict the wind stress curl from single point wind measurements? The answer to both these questions will be a “yes.”

e. Wind stress curl prediction

The wind stress curl as represented by the first CEOF correlates strongly with the wind stress at EB23 (Fig. 9). Maximum vector correlation between the observed wind stress at buoy EB23 (Fig. 2a; see Fig. 1 for location) and the temporal amplitude of mode 1 (Fig. 8a) reveals $r_{12}^2 = 0.79$ that occurs at an angle of $-56^\circ$; that is, wind stress correlates best with the pattern shown in Fig. 7a after it is rotated $56^\circ$ clockwise from east. This orientation represents the direction of upwelling favorable alongshore winds. The correlation is significantly different from zero at the 95% confidence level for 18 ($r_{12}^2 > 0.2$) and 5 ($r_{12}^2 > 0.7$) degrees of freedom assuming wind stress decorrelation timescales of 2 and 6 days, respectively.

Figures 9a and 9b show the scatter and time series, respectively, of the vector components leading to this maximum correlation; in Fig. 9b the first CEOF mode $A_1(t)$ is scaled to fit the same graph as the wind stress. The wind stress at EB23 explains 79% of the variance of the first CEOF mode through the regression $A_1(t) = -0.9 + 6.8\tau$, where $\tau$ is the observed wind stress in Pa along $56^\circ$ clockwise from east. The constant offset $-0.9$ represents the temporal mean wind stress (0.13
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is still unknown. Harms and Winant (1998) and Dever
clonic circulation in its ®ve different ¯avors, however,
does this robust wind stress curl affect the internal mass
field at the entrance of the Santa Barbara Channel? Does
it contribute to the oceanic circulation in the western
approaches of this channel?
Harms and Winant (1998) identify six characteristic
patterns from EOF analyses of a year-long 1994 current
record of surface currents. All but one of these patterns
show cyclonic current shear at the western entrance of
the Santa Barbara Channel. The dynamics of the cy-
clonic circulation in its five different flavors, however,
is still unknown. Harms and Winant (1998) and Dever
e et al. (1998) comment on pressure gradient and local
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Wang 1997) adds that lateral friction and small-scale
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cumulistic evidence to justify the hypothesis that the
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Figures 10a and 10b compare time series of wind
observations from buoy EB23 for a 60-day period in
the spring of 1983 and 1984, respectively. Markers in
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servations are available for the period when the most
intense upwelling and Ekman pumping takes place. In
contrast, the 1983 aircraft surveys sampled all wind
conditions including the strong and persistent upwelling
following day 126 (Figs. 2, 3, and 9). It also contrasts
with the conditions during a single CTD survey con-
ducted in the spring of 1984, which, it appears in Fig.
10, sampled the density field during strong and persis-
tently upwelling favorable winds.
Figures 10c and 10d show a 60-day segment of a
6-month long record of 30-m currents on the southern
and northern shelf, respectively. From about day 128
through day 148 currents on the northern shelf \(u(y_2)\) are
out of the channel and reach 80 cm s\(^{-1}\) (Fig. 10d).
In contrast, currents on the southern shelf \(u(y_1)\) are into
the channel and reach 40 cm s\(^{-1}\) (Fig. 10c). The winds
at EB23 during this period of time (that includes a CTD
survey on day 139, 19 May) are upwelling favorable
and exceed 10 m s\(^{-1}\) for almost the entire 20-day long
period. Note that offshore wind, currents, and lateral
current shears fluctuate at a 5–10 day timescale. The
respond to the wind field is not always instantaneous
and forcing mechanisms other than the local winds con-
tribute to the flow field and its lateral shear.
Figure 11 shows the result of a frequency-domain
cross-correlation analysis of the full 6-month record
from January through July of 1984. It indicates signif-
icient coherence (variance explained) between the
across-channel shear \(d u = u(y_1) - u(y_2)\) and the wind
stress. Two distinct spectral peaks emerge at periods of
about 6 and 3.8 days that explain, respectively, 40% and
55% of the variance (Fig. 11). The correlated motion at

![Figure 10](image)

**Fig. 10.** Time series of (a) spring 1983 winds at EB23, (b) spring 1984 winds at EB23, (c) 30-m currents at the southern channel entrance, and (d) 30-m currents at the northern channel entrance. The symbols toward the top in panels (a) and (b) indicate the time hydrographic sections were collected across the western entrance of Santa Barbara Channel.
the short 3–4 day timescale lags the wind forcing by about 60° or 0.5 days. This short, almost instantaneous response is typical of a local, directly forced, wind-driven response. In contrast, the larger 120° phase lag at the longer 5–9 day timescale implicates a different forcing where the cyclonic shear lags the winds at EB23 by about 2–3 days. Auad et al. (1998) find similar correlations and phase lags in their more comprehensive analysis of all 1984 mooring data from the Santa Barbara Channel. While they comment on the poor correlation of surface currents with the winds, their EOF results for bandpassed data in a 4–6-day band indicates significant correlations with the wind. Their results and those presented here are consistent; that is, it is the cyclonic current shear [expressed in Auad et al. (1998) by the EOFs] not the surface currents that correlate with the winds at this frequency with a 2-day phase lag. Recall that the wind stress at EB23 correlates strongly with the wind stress curl near Point Conception (Fig. 9). It thus is conceivable that the wind stress curl results in the surface current shear at the 5–9 day timescale. A possible scenario on how the wind stress curl and Ekman pumping could set up such a circulation is presented next.

Figure 12a shows the density field at 30-m depth along with low-pass filtered current vectors at the same depth centered on the time of a 2-day CTD survey. The Lanczos filter has a half-power point at 84 h and passes 90% and 10% of the variance at 140 and 60 h, respectively. At the eastern entrance a 30 cm s⁻¹ strong inflow transports lighter waters into the channel; however, lateral density gradients here are small. In contrast, strong lateral density gradients prevail in the western part of the channel where dense waters enter along the southern island chain, while buoyant waters leave the channel near Point Conception to the north (Fig. 12a). These buoyant waters occupy a surface layer about 40 m deep on the northern shelf/slope region only (Fig. 13).

A different picture emerges from density observations below the surface layer. Figure 12b shows how the depth of a particular density surface varies horizontally. The 26.5-σt surface is about 100-m deep along the northern shore. In most of the channel it slopes toward the surface from the northern shelf to the southern shelf where it is only about 80 m deep. South of Point Conception, however, the isopycnal is only 60 m deep and it is shallowest over the center of the channel. It deepens toward the southern shelf. This apparent doming of the density field becomes particularly clear in Fig. 13, which shows the density distribution of transects A and B across the western channel (see Fig. 12b for locations). All iso-

**Fig. 11.** Frequency domain correlation (coherence squared) between lateral current shear and alongshore wind stress: (upper panel) the phase lag of the lateral shear with respect to wind stress and (lower panel) the fraction of the variance explained by a linear transfer function.

**Fig. 12.** Maps of the density and flow field of Santa Barbara Channel centered on day 140 (19 May 1984). (a) Horizontal distribution of density and currents at 30-m depth. Density data are from CTD casts while currents are from current meter moorings. (b) Depth of the 26.5-σt density surface. Labels “A” and “B” in (b) refer to the two density sections shown in Fig. 13. Filled circles indicate the location of CTD stations.
Pycnals below 50 m are raised toward the surface at the center of the channel. Curl-driven upwelling would raise the isopycnal surfaces preferentially near the center of the channel (Fig. 6). If these doming isopycnals are furthermore geostrophically balanced, then they would indicate a cyclonic surface circulation relative to a stagnant deeper layer.

The thermal wind relation

\[
\frac{\partial u}{\partial z} = g \frac{\partial \rho}{\partial y} - \frac{\rho_0 f}{\partial y}
\]

(7)
gives geostrophically balanced vertical current shears only. This shear is here arbitrarily referenced to zero flow at 250 m or, if the bottom is shallower than 250 m, to zero flow at the bottom. Figure 14 compares the geostrophic current profile across the channel at 30-m depth with observed low-pass filtered currents observed at the same depth during the CTD survey. The two curves generally agree to within 5 cm s\(^{-1}\) except at the northernmost mooring where the observed flow is about 20 cm s\(^{-1}\) stronger than the geostrophic flow. Hence, other mechanisms contribute to the flow here besides geostrophy if the reference level indeed represents a level of no motion. Nevertheless, the thermal wind relation reproduces the lateral current shear at 30 m rather well.

The relative vorticity \(\xi\) of the upper-layer fluid can be approximated by the lateral current shear; that is,

\[
\xi = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} = \frac{\Delta u}{\Delta y} = \frac{0.56 \text{ m s}\(^{-1}\)}{14 \text{ km}} \approx 0.36 f,
\]

(8)

where the values for \(\Delta u\) and \(\Delta y\) are from Fig. 12 and \(f = 1.1 \times 10^{-4} \text{ s}\(^{-1}\). The scale \(\Delta y\) closely corresponds to the internal deformation radius \(N D f\) where \(N^2 = (g/\rho_0)\partial \rho/\partial z\) is the square of the buoyancy frequency (about 6 cph) and \(D\) is the vertical scale of motion (about 50 m). These scales result in an internal deformation radius that is about 11 km. The ratio between the relative vorticity \(\xi\) and the planetary vorticity \(f\) corresponds to a Rossby number that measures the relative importance of nonlinear inertial to Coriolis forces (Münchow and Garvine 1993). Nonlinear inertial forces thus contribute...
to the dynamics of the flow near Point Conception. It is conceivable that the advection of cyclonic vorticity (a nonlinear process) generated near Point Conception by the wind stress curl contributes to the dynamics of the western Santa Barbara Channel. Since the across-channel balance appears to be in geostrophic thermal wind balance, the nonlinear advective forces must contribute to the along-channel balance. The dynamics at the western entrance of the channel thus appear to be semigeostrophic.

The link between the geostrophically balanced isopycnals and the wind stress curl is admittedly weak and circumstantial. The evidence appears sufficient, however, to justify the hypothesis that the wind stress curl at the western entrance of the Santa Barbara Channel affects the local circulation. The nonlinear advection of vorticity from this region into the channel is best investigated with surface drifter observations, such as described by Dever et al. (1998). In order to accurately predict the flow in the Santa Barbara Channel the wind stress curl thus must be included in numerical models of the channel circulation.

5. Conclusions

Near Point Conception the atmospheric flow separates from the mountainous coastline, which changes its orientation by about 90 degrees. A cyclonic circulation in the lower atmosphere emerges in the lee of the cape from both the temporal average of all wind data and from an analysis of the spatial variations from this average. More than 70% of the variance of the atmospheric flow is explained by a single statistical mode that describes cyclonic flow patterns during strong and persistent upwelling events. Supercritical flows exhibit a stronger than average cyclonic tendencies, but subcritical flows result in a cyclonic circulation as well.

The spatially variable atmospheric circulation corresponds to about 4–20 m day\(^{-1}\) vertical Ekman pumping velocities in the ocean. These inferred vertical velocities at the base of the surface Ekman layer could raise deep isopycnals and thus stretch vortex tubes in the ocean interior (Gill 1982). During strong upwelling events in the spring of 1984, isopycnals dome near the center of the Santa Barbara Channel where, in 1983, the core of the atmospheric cyclonic circulation was located. The persistent wind stress curl could thus have contributed to the doming of the isopycnals over the western center of the channel. In May 1984 moored current meters recorded a strong oceanic cyclonic circulation. During this event the thermal wind balance explains the across-channel structure of the along-channel flow; however, the across-channel current shear suggests that the along-channel momentum balance is not geostrophic. The dynamics is then semigeostrophic.

Near Point Conception many dynamically different regimes and processes overlap: eddies from the offshore California Current occasionally impinge upon the shelf (Atkinson et al. 1986); variable topography certainly scatters vorticity waves into smaller scale motions; and large temperature gradients between the two entrances of the Santa Barbara Channel force a buoyancy-driven circulation. All these processes interact with the wind-forced motions. It is hard observationally to separate these processes in order to understand the physics of their interaction. Idealized numerical models and laboratory studies, such as Wang (1997) and Klinger (1994), are needed in order to shed light on some of the physical processes that influence the flow near Point Conception and the western Santa Barbara Channel.

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APPENDIX

Calibration of Aircraft Winds

A careful calibration of the data is necessary in order to estimate the wind stress curl at the sea surface from aircraft measurements of winds aloft. During each survey the aircraft flew directly over meteorological buoys EB11 and EB23 (see Fig. 1 for locations). Figure A1 shows the scatterplot of 10-m buoy and 152-m aircraft winds along with linear regression curves for the east and north component of the wind vector. Table A1 summarizes the regression coefficients and their uncertainties at a 95% level to reject the null hypothesis. The correlation of buoy and aircraft winds is good; that is, more than 98% of the buoy variance is linearly related with the aircraft wind variance. The offset between the two measurements is zero with an uncertainty of 0.4 m s\(^{-1}\). The speed of the buoy winds at 10 m is about 72% of the aircraft winds at 152 m to within an error of about 0.5%. The results of the regression are similar to those of Brink et al. (1984) off Point Conception; however, the errors on the regressions are smaller here because of the larger sampling size. Nevertheless, instrument and calibration uncertainties lead to errors in the wind speed of about 1 m s\(^{-1}\) with about 40% of the uncertainty arising from the maximum offset and 60% of the uncertainty arising from the uncertainty of the slope of the regression.
The calibrated winds (Table A1) are converted to wind stress $\tau$ using the empirical bulk formulas of Wu (1980) $\tau = (\rho_a/\rho_w) C_{10} |U_{10}| U_{10}$ where $\rho_a/\rho_w$ is the density ratio between air and water and $C_{10}$ is an empirical drag coefficient for winds $U_{10}$ measured 10 m above the sea surface. The drag coefficient is $C_{10} = (0.8 + 0.0656 |U_{10}|)10^{-3}$.

The wind stress curl, finally, is computed by central differences of the two components of the wind stress vector. It is thus estimated from the difference of four imperfectly estimated quantities. Furthermore, the measured-quantity wind must be cubed in order to obtain the estimated-quantity stress. Empirical tests reveal that this uncertainty in the wind speed causes an uncertainty in the Ekman pumping velocity of between 30% and 80%. These errors are upper limits and they depend upon the wind speed. Enriquez and Friehe (1995) find from a more comprehensive error analysis that their wind stress curl estimates are accurate within 0.13 Pa/100 km, which corresponds to an Ekman pumping velocity of about 1.2 m day$^{-1}$.

### Table A1. Regressions between aircraft (152 m) and buoy (10 m) winds.

<table>
<thead>
<tr>
<th>Offset (m s$^{-1}$)</th>
<th>Error offset (m s$^{-1}$)</th>
<th>Slope</th>
<th>Error slope</th>
<th>$r^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>East</td>
<td>-0.22</td>
<td>0.411</td>
<td>0.73</td>
<td>0.051</td>
</tr>
<tr>
<td>North</td>
<td>0.00</td>
<td>0.408</td>
<td>0.71</td>
<td>0.042</td>
</tr>
</tbody>
</table>

### REFERENCES


