Numerical model simulations of continental shelf flows off northern California

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Abstract

The three-dimensional circulation on the continental shelf off northern California in the wind events and shelf transport (WEST) experiment region during summer 2001 is studied using the primitive equation regional ocean modeling system (ROMS). The simulations are performed with realistic topography and initial stratification in a limited-area domain with a high-resolution grid. Forcing consists of measured wind-stress and heat flux values obtained from a WEST surface buoy. The general response shows a southward coastal upwelling jet of up to 1 m s\(^{-1}\) and a weakening or reversal of currents inshore of the jet when upwelling winds relax. Model results are compared to WEST moored velocity and temperature measurements at five locations, to CODAR surface current observations between Pt. Reyes and Bodega Bay, and to hydrographic measurements along shipboard survey lines. The model performs reasonably well, with the highest depth-averaged velocity correlation (0.81) at the inshore mooring (40 m water depth) and lowest correlation (0.68) at the mid-depth mooring (90 m depth). The model shows generally stronger velocities than those observed, especially at the inshore moorings, and a lack in complete reversal of southward velocities observed when upwelling winds relax. The comparison of surface velocities with CODAR measurements shows good agreement of the mean and the dominant mode of variability. The hydrography compares closely at the southern and northern edges of the survey region (correlation coefficients between 0.90 and 0.97), with weaker correlations at the three interior survey lines (correlation coefficients between 0.44 and 0.76). Mean model fields over the summer upwelling period show slight coastal jet separation off Pt. Arena and significant separation off Pt. Reyes. The cape regions also experience relatively strong bottom velocities and nonlinearity in the surface flow. Cross-shelf velocity sections examined along the shelf reveal a double jet structure that appears just north of Bodega Bay and shows the offshore jet strengthening to the south. We examine the dynamics during an upwelling and subsequent relaxation event in May 2001 in which the WEST measurements show evidence of a strong flow response. The alongshelf variability in the upwelling and relaxation response introduced by Pt. Reyes is evident. Analysis of term balances from the depth-averaged momentum equations helps to clarify the event dynamics in different regions over the shelf. A clear pattern in the nonlinear advection term is due to the spatial acceleration of the southward jet around the capes of Pt. Arena and Pt. Reyes during upwelling. Results from a three-dimensional Lagrangian analysis of water parcel displacement show significant southward displacement in the coastal jet region, including a strong signal from the double jet. Alongshelf variability in parcel displacements and upwelling source waters due to the presence of Pt. Arena and Pt. Reyes is also apparent from the Lagrangian fields. A cyclonic eddy-like recirculation feature offshore of Pt. Arena prior to the upwelling event causes large patches of onshore-displaced parcels. Additionally, across-shelf variability in the response

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of water parcels along the D line includes decreased vertical displacement and increased alongshelf displacement in the offshore direction.

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1. Introduction

As part of the coastal ocean processes (CoOP) program, the wind events and shelf transport (WEST) project has the overall goal of understanding the role that wind-driven transport plays in biological productivity over the northern California shelf. The region of interest for WEST extends from just south of Pt. Reyes to about 50 km north of Bodega Bay (Fig. 1). This is an area that has been studied previously by large observational programs, such as the coastal ocean dynamics experiment (CODE) (Beardsley and Lentz, 1987) and the shelf mixed layer experiment (SMILE) (Alessi et al., 1991), and results of those experiments have been used for guidance in design of the WEST field measurements. Moorings deployed at five locations from May 2001 to 2003 by the WEST project collected measurements of current velocity, temperature, conductivity, pressure, wind, chlorophyll fluorescence, and optical transmissivity (Dever et al., 2006). Additionally, surface heat flux parameters were measured at the central D090 mooring and fluxes were calculated using the algorithms specified in Beardsley et al. (1998) (E. Dever, personal communication). Land-based coastal radar (CODAR) systems were in operation during May 2001–September 2003 to measure surface currents over the region from Pt. Reyes to Bodega Bay (Kaplan et al., 2005). Ship surveys also were conducted during three summer upwelling periods and two winter downwelling periods between June 2000 and November 2002. These surveys were comprised of both vertical profiles and underway tow-yo sampling of water properties and current velocity (Largier, 2004).

In this study, we utilize the regional ocean modeling system (ROMS) to model the circulation over the continental shelf and upper slope in the WEST region during May–June 2001. During this period, the winds are generally upwelling favorable at the central D090 mooring with a mean wind-stress magnitude of 0.074 N m⁻² directed toward 120°T. The magnitude and direction of maximum standard deviation in wind-stress are similar to the mean for this period, i.e. 0.052 N m⁻² and 116°T. This wind-stress forces an upwelling circulation that persists throughout the summer with the exception of intermittent 1–3 day periods during which the alongshelf winds relax or reverse. The modeled velocity and hydrographic fields are first compared quantitatively with the WEST data sets. Following the model-data comparisons, we discuss the time mean response, which is dominated over the shelf by an upwelling circulation. We then proceed to focus on the dynamics of a specific upwelling and subsequent relaxation wind event that occurs from May 17 to 23, 2001. This event, which is representative in magnitude and duration of upwelling/relaxation events during the summer season, was selected because WEST shipboard observations are available. This allows for model-data comparisons with shipboard measurements as well as a resulting model analysis that corresponds to a period of focus for WEST investigators. We present a detailed

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Fig. 1. Observational layout for wind events and shelf transport (WEST) project.
description of the Eulerian and Lagrangian dynamics during the upwelling and relaxation wind conditions that characterize this event. The numerical model provides a framework in which to interpret and synthesize the WEST measurements. Previous modeling applications in this region have investigated the shelf circulation during upwelling and relaxation conditions (Gan and Allen, 2002a,b). However, emphasis was placed on the region immediately north of Bodega Bay for comparison with CODE measurements, and all of the analyses were of Eulerian fields. This study presents new Lagrangian analyses in the WEST region that are in support of WEST goals to determine the impact of the shelf circulation on water parcel displacements.

2. Model formulation

The regional ocean modeling system (ROMS) is a hydrostatic primitive equation model with terrain-following vertical coordinates. ROMS is based on the S-Coordinate Rutgers University Model (SCRUM) described by Song and Haidvogel (1994), but has been rewritten to include, for example, high-order advection schemes, accurate pressure-gradient algorithms, and several subgrid-scale parameterizations (Shchepetkin and McWilliams, 2005). We use the embedded Mellor-Yamada 2.5 level turbulence closure scheme (Mellor and Yamada, 1982) as modified in Galperin et al. (1988) and the third-order upstream bias advection scheme for momentum and tracers (Shchepetkin and McWilliams, 1998). The pressure-gradient scheme is a splines density Jacobian, one of three methods developed by Shchepetkin and McWilliams (2003) to minimize the errors associated with computing horizontal pressure gradients with terrain-following coordinates.

The model domain and bathymetry is shown in Fig. 2. The model bathymetry is a spatially-filtered version of that obtained from the National Geophysical Data Center GEODAS dataset with 3-s resolution. It is smoothed such that the maximum local slope factor, defined as $|\partial h/\partial z|$ (Mellor et al., 1994), is no greater than 0.2. This value has been shown to limit the errors associated with the pressure gradient calculation (Barnier et al., 1998; Beckmann and Haidvogel, 1993).

Periodic boundary conditions are used on the northern and southern boundaries, and the model bathymetry is adjusted to be equal at these boundaries. The alongshelf length of the domain is 400 km and the across-shelf length is approximately 135 km. The Cartesian $(x,y)$ grid has a resolution of 1.7 km in the alongshelf $y$ direction and between 0.8 and 1.8 km in the across-shelf $x$ direction, with highest resolution near the coast. In
the vertical, we use 40 $s$-levels with spacing such that there is higher resolution in the surface and bottom layers.

At the coastal boundary, conditions of zero normal velocity and of free slip for the tangential velocity component are applied. At the offshore boundary, we use a condition of zero normal depth-integrated velocity ($U = 0$) and free slip for the tangential depth-integrated velocity ($V_x = 0$). The free surface satisfies an implicit gravity wave radiation condition (GWI as described in Chapman, 1985), the depth-dependent velocities satisfy a radiation condition, and the tracers and turbulence quantities satisfy a no gradient condition ($T_x = S_x = q^2_x = 0$). The initial temperature and salinity profiles in Fig. 3 are horizontally uniform and are taken from the spatial average of the summer 2001 mean central D line CTD survey data. The temperature profile is very similar to that calculated as the mean of April 1981–1982 values from the CODE experiment central line. However, the salinity profile from the WEST measurements is less stratified than that from the CODE measurements, which has a value of approximately 33.3 psu near the surface and increases to 33.9 psu at 140 m depth (Huyer, 1984).

The model simulation occurs over 57 days during April 25–June 20, 2001. We regard the first 10 days (April 25–May 4), during which the WEST measurements are unavailable for model-data comparisons, as a spinup period. Therefore, we calculate time mean fields presented in Section 4 over 47 days from May 5 to June 20, 2001. The wind-stress and heat flux time series used to force the model (Fig. 4) are calculated from measurements at the central D090 mooring, then low-pass filtered and applied uniformly in space. The assumption of no spatial variability in the wind field is not accurate, as has been previously shown by wind measurements in the CODE and WEST regions (Beardsley et al., 1997; Dorman and Winant, 1995). However, generally favorable model-data comparisons in Section 3 show that forcing with spatially-uniform winds results in a successful representation of a substantial component of the observed shelf flow that is strongly influenced by interactions with topography. Future modeling studies utilizing wind-stress fields obtained from a high-resolution regional mesoscale atmospheric model, after these become available following analysis and model-data comparisons (D. Koracin and C. Dorman, personal communication), would be able to address
specifically the effects of spatially-variable wind stress in this region and should be informative.

3. Model-data comparisons

3.1. Moored velocity and temperature observations

In order to verify the model’s ability to represent the wind-driven shelf flow in this region, we conduct model-data comparisons with the available WEST observations. The first such comparison is with the moored velocity and temperature time series. Fig. 5 shows depth-averaged velocities from the observations and the model at the D Line moorings. The depth-averaged velocities are computed over the depths of the mooring instruments. The wind-stress vectors used to force the model are also plotted at the top. The time period shown is the entire duration of the model simulation (April 25–June 20, 2001), including the spinup. However, because the moored time series are not available until May 5, correlations between the model and mooring observations are computed over May 5–June 20. The complex correlation is highest at the inshore mooring (0.81); however, the phase angle shows a fairly significant counterclockwise rotation of the modeled currents as compared to the observed D040 currents (23.44°). The correlation is lowest at the central D090 mooring (0.68), but the phase angle is very small (1.07°), indicating that the model does get the direction of the velocity fluctuations correct at this location. Correlations at the other 90-m moorings, C090 and E090 (not shown) are 0.64 with 7.73° phase angle and 0.57 with 6° phase angle, respectively. The correlation and phase angle at D130 (0.74 and 12.34°) lie between those at D040 and D090, and the root mean square error (rmse) at this location is significantly smaller in the alongshelf direction than at the other two moorings.

The discrepancies between the model and observed time series are evidently due to the larger model velocity magnitudes during upwelling forcing (especially at D040 and D090) and the lack of a significant reversal in the modeled velocities when
upwelling winds relax (especially at D090 and D130). The model does exhibit a strong relaxation response that is highly dependent on spatial location as shown in Section 5.1, but along the D line, the model fails to predict the complete reversal of currents during wind relaxations. The model performs well at depth at the two offshore moorings due to decreased model velocities that match the...
observations more closely than at the surface (not shown). The difference in major axis orientations of model-data currents may be due to the use of spatially uniform winds. It will be interesting to see if future modeling studies with spatially variable winds can provide velocity fields in closer quantitative agreement with observed values in this region.

We also compare the near-surface temperature time series at the D Line moorings for the same time period (Fig. 6). The modeled temperature at the D040 and D090 moorings is 1°–4°C cooler than the observations, with this difference a maximum during wind relaxation events when the observed temperature increases by 1°–2°C. Thus, the model temperature has a weaker relaxation response than the observed temperature as expected from the weaker velocity relaxation response.

Time mean depth-averaged velocity vectors from the observations and the model at the WEST mooring locations show the mean modeled velocities are larger than the mean observed velocities at all mooring locations except D130 (Fig. 7). The first mode EOF, which explains 86% of the variance in both the model and observations, shows the model does better at matching the magnitude of the variability in the observations. The time series of the EOFs have a correlation of 0.88. The observed amplitude is consistently higher than the model, with the biggest difference when the sign is positive. This corresponds to the observed reversal of velocities when upwelling winds relax, a response that the model does not represent completely as discussed earlier.

Further comparison between the moored velocity observations and the model is made by computing space-lagged correlation coefficients between model velocity components and between observed velocity components at the mooring locations (Tables 1 and 2). The time period over which the correlations are computed is the same as for the model-data comparisons presented in Figs. 5–7. The observed and modeled depth-averaged velocities (averaged over the measurement depths) are rotated to their respective principal axes calculated from the depth-averaged currents to objectively determine local values of the alongshelf and across-shelf velocity.

![Figure 6](image-url)
components. The observed correlations are shown in bold. For both the model and the observations, across-shelf correlations are computed for D040, D090, and D130 (Table 1) and alongshelf correlations for C090, D090, and E090 (Table 2). The modeled and observed velocities have generally higher correlations alongshelf between the 90-m moorings than across-shelf on the D line. Alongshelf velocity correlations for \( v \) are high for both the model and observations at the 90-m moorings. The modeled \( u \) at E090 has low correlation with \( u \) at the southern moorings (C090 and D090), which have a high correlation with each other (0.44 for D-E, 0.08 for C-E, and 0.73 for C-D). In contrast, the observed \( u \) correlations are relatively uniform between all these pairs (0.51–0.61). Thus, the observed across-shelf flow shows a similar response

Table 1
Summary of space-lagged correlation coefficients for observed (bold text) and modeled depth-averaged alongshelf velocity \((v)\) and across-shelf velocity \((u)\) at the D line mooring locations

<table>
<thead>
<tr>
<th></th>
<th>D090</th>
<th>D130</th>
</tr>
</thead>
<tbody>
<tr>
<td>( v )</td>
<td>0.48</td>
<td>0.38</td>
</tr>
<tr>
<td>( u )</td>
<td>0.70</td>
<td>0.57</td>
</tr>
</tbody>
</table>

Table 2
Summary of space-lagged correlation coefficients for observed (bold text) and modeled depth-averaged alongshelf velocity \((v)\) and across-shelf velocity \((u)\) at the 90-m mooring locations

<table>
<thead>
<tr>
<th></th>
<th>D090</th>
<th>E090</th>
</tr>
</thead>
<tbody>
<tr>
<td>( v )</td>
<td>0.95</td>
<td>0.94</td>
</tr>
<tr>
<td>( u )</td>
<td>0.95</td>
<td>0.93</td>
</tr>
</tbody>
</table>

\( u \) and \( v \) are determined by rotating to the principal axes of the depth-averaged currents. The highest 95% significance level for the correlations shown in Tables 1 and 2 is 0.51.

Fig. 7. Mean (top) and mode 1 EOF (middle) of depth-averaged velocities from the mooring observations (left) and the model (right). The amplitude time series of the observed (solid line) and modeled (dashed line) mode 1 EOF is also shown at the bottom with the correlation coefficient. The 100-m and 200-m isobaths are provided.
between the C, D, and E mooring locations, but the model displays different $u$ behavior at the northern site, E090. Observed $u$ correlations between D040 and the offshore D moorings are small and negative; however, the model correlations are higher for the three D mooring pairs (0.54–0.66). This result illustrates a complexity in the observed flow at the inshore mooring location that is not captured by the model.

3.2. CODAR surface velocity observations

Daily averaged surface velocities in the CODAR region (see Fig. 1) are available from May 4 to June 20, 2001. We compare the modeled surface velocities over the same region in Fig. 8. Shown are the complex correlation and phase, the time mean surface velocities, and the mode 1 EOFs from the CODAR observations and the model. Note the 100-m and 200-m isobaths are plotted for reference. The correlations are between 0.6 and 0.8 over much of the coverage area, with the exception of a few patchy areas at the north and south and near Cordell Bank, the topographic feature offshore of Pt. Reyes where the 100-m contour is closed. Phase angles are small in the middle of the CODAR region, with higher values around the edges. The mean velocities are very similar offshore of

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**Fig. 8.** Magnitudes of complex correlation coefficients and phase angles between modeled and observed surface currents (left), time mean surface velocity vectors for CODAR and model (middle), and dominant mode 1 EOF of surface currents from CODAR and model (right). The percentage of total variance explained by the mode 1 EOF is displayed in the lower left corner of each panel. The amplitude time series of the observed (red) and modeled (blue) EOFs are also shown at the bottom with the correlation coefficient. The 100- and 200-m isobaths are shown in black. CODAR surface velocities were provided by D. Kaplan.
the 100-m isobath, but near the coast, the modeled velocities are stronger. This pattern is consistent with the discussion of Figs. 5 and 7 in which the depth-averaged velocity magnitudes at D130 are comparable, but the modeled velocities at the inshore moorings are stronger than the observed velocities. The root mean square errors between the observed and modeled mean surface velocities are 0.06 m s\(^{-1}\) and 0.11 m s\(^{-1}\) for the east/west and north/south components, respectively. The structure of the observed and modeled mode 1 EOFs is similar over the entire CODAR region.

The modal amplitudes (bottom) are highly correlated, with a correlation of 0.79. To attempt to determine if the mode 1 EOF represents the wind-driven mode, we compute cross-correlations between the wind-stress components and the observed and modeled mode 1 time series. Significant agreement is found between \(\tau_y\) and the mode 1 amplitudes, with zero-lag correlations of 0.88 and 0.63 for the observed and modeled modes, respectively. With the mode 1 amplitude lagging \(\tau_y\) by one day, the model correlation increases to 0.66. Cross-correlations of \(\tau_x\) with the mode 1 amplitudes are

Fig. 9. Potential density sections from CTD observations (top) and model (middle) along the D Line on May 20 (left) and the F Line on May 21 (right). Note the difference in the horizontal scales. At the bottom are scatter plots of all potential density observations compiled from five surveys during May/June for the D Line and one survey for the F Line. Provided below the scatter plots are the slope of the regression line (b), correlation coefficient (r), root mean square error (rmse), and variance from the observations (\(\sigma_o\)) and the model (\(\sigma_m\)). Hydrographic data were provided by N. Garfield.
small at all time lags. We conclude that the mode 1 EOFs represent the dominant coherent wind-driven variability.

3.3. Shipboard hydrographic observations

Shipboard surveys following the large-scale and small-scale survey plans shown in Fig. 1 allow an opportunity for model-data comparisons. The large-scale survey encompasses WEST lines A, D, and F, while the small-scale survey covers lines B, C, D, and E. In Fig. 9, we compare two lines from a single CTD survey to the modeled potential density $\sigma_0$ during the event period we discuss in Section 5. The modeled and observed $\sigma_0$ fields show similar characteristics. At the D line, upwelled isopycnals are found near the surface from the coast to 30 km offshore in both the model and observations. The sections at the F line also compare well qualitatively, with the exception of an observed bulge of low-density water extending from the surface to almost 100 m depth between 15 and 30 km offshore.

A quantitative comparison (Fig. 9) is given by scatter plots of all the observations available for each of these survey lines (five surveys in May/June on the D line and one survey on the F line) versus corresponding model values. The correlation is very high on the F line (0.97) and the regression line has a slope near one (0.91). The correlation is lower on the D line (0.76) and more scatter is found; however, the rmse values are similar at the two sites. The results for comparisons along lines A, C, and E (not shown), computed for one survey on lines A and E and two surveys on line C, have a high correlation (0.9) on the A line, while the C and E lines are much lower (0.51 and 0.44, respectively). A similar comparison between the modeled potential density and that observed with the tow-yo instrument, which encompasses the large-scale survey lines A, D, and F, yields a high correlation on the F line (0.93) and lower correlations on the A and D lines (0.54 and 0.53, respectively). The model clearly captures the density structure well at the northern end of the study region where the flow is relatively simple compared to that just north of Pt. Reyes and Cordell Bank, where the three interior survey lines are located. Somewhat surprising, perhaps, is the high correlation on the A line, where the model flow is quite complicated as we will illustrate.

4. Characteristics of upwelling season response

We are interested in the mean wind-driven response of the shelf flow over the model simulation period. The winds are dominantly upwelling-favorable (Fig. 4); however, several relaxation events lasting 1–3 days occur during May and June 2001. The modeled shelf response at the surface can be seen in the region from Pt. Reyes to Pt. Arena in Fig. 10. Clearly, the fields demonstrate that the upwelling response dominates the time mean. The surface velocity vectors show the location of the coastal upwelling jet, which separates slightly from the coast off Pt. Arena and significantly off Pt. Reyes, approximately following the 100-m isobath in both cases. Separation of the coastal upwelling jet from Pt. Reyes is a feature that has been previously discussed in the numerical model results of Gan and Allen (2002a). The largest variability in the velocity amplitude is off the capes and along the 200-m isobath. The variability is high off Pt. Arena and Pt. Reyes because the flows are complicated due to interactions with the variable coastline and bathymetry. (Gan and Allen, 2002a,b) found that, due to these interactions, the coastal capes in this area play a dominant role in the alongshelf variability of upwelling.

Another region of significant southward velocities is also apparent south of Pt. Reyes, at the surface offshore of the 200-m isobath. This corresponds to a bifurcation in the separated coastal jet with an inshore portion of the jet following the 100-m isobath and another portion turning offshore, crossing isobaths, and returning southward about 60 km offshore. This double jet structure will be discussed in further detail later in this section.

The mean surface temperature field displays an upwelling response, with the coldest water near the coast and increasing temperature offshore. An area of relatively cold water is also found on the shelf south of Pt. Reyes. The standard deviation of temperature shows high variability offshore and south of Pt. Reyes due to the relatively large signal observed during upwelling relaxation in this region. The coastal jet location is more obvious in the mean depth-averaged velocity than at the surface because velocities at depth decrease significantly offshore of the 200-m isobath (Fig. 11). Next to the depth-averaged velocity vectors are transport streamfunction contours, illustrating how the flow follows the coastline north of Pt. Reyes, where it veers offshore following the 100-m isobath. The
transport is greatly decreased inshore of the 100-m isobath south of Pt. Reyes and a cyclonic recirculation area is found 50 km south of Cordell Bank between the regions of southward velocities associated with the bifurcated jet.

Bottom velocities have an onshore component under the coastal jet due to the bottom Ekman flow. This onshore bottom flow is qualitatively consistent with mean profiles of the observed across-shelf velocity, defined by the principal axes of the depth-averaged flow, at the D040 and D090 mooring locations (not shown). Although the measurements omit the deepest 20% of the water column, onshore velocities are observed below 15 m at D040 and below 40 m at D090. Therefore, the time mean summer upwelling circulation from both the model and the observations includes onshore-directed flow below the coastal jet. Purely onshore bottom flow is found just south of Pt. Reyes. Consistent with the distribution of large bottom velocity, the bottom turbulent kinetic energy values are highest off Pt. Reyes and Pt. Arena. The turbulent kinetic energy variable \( (tke) \) is computed using the Mellor–Yamada level–2.5 closure scheme \( (\frac{1}{2}(u^2 + v^2)) \). Nonlinearity in the surface flow is indicated off Pt. Reyes and Pt. Arena by the relatively large magnitudes of the surface relative vorticity divided by the Coriolis parameter \( f \), which gives a measure of the local Rossby number. These results all point to the high spatial variability and complexity associated with the coastal jet separation. Mean potential density \( \sigma_0 \) at the bottom shows the isopycnals are approximately aligned with bathymetry contours, with somewhat lower values inshore of the 100-m isobath south of Pt. Reyes.

Sections of time mean alongshelf velocity \( v \) and potential density \( \sigma_0 \) at the locations shown in Fig. 2 illustrate the significant variability in structure due to the flow response to the variable shelf geometry and coastline shape at different alongshelf locations (Fig. 12). North of the F line (line 166), a coastal
jet with a defined core extends about 200 m deep and 30 km offshore, with slightly northward velocities inshore. Along the F line (line 148), the jet widens and shallows to about 100 m depth. Continuing south to the E line (line 130), the existence of two separate jet structures begins to develop, with stronger velocities in the inshore jet. Evidence in Figs. 10 and 11 of weaker surface and depth-averaged velocities between the 100-m and 200-m isobaths at this location was discussed previously. The double jet feature, however, is better defined and more visible in the sections than in the horizontal velocity fields. The double jet can be seen in Fig. 12 at each subsequent section south of line 130, with southward velocity increasing in the offshore jet and becoming larger than that inshore at Pt. Reyes and along the A line (lines 112 and 104). The jets move offshore between lines 112 and 104 due to separation off Pt. Reyes as shown in Fig. 11. Areas with zero or slightly northward velocity are found along the continental slope between the two jets at all sites. A region of northward velocity from 5 to 20 km offshore at the A line reflects the relaxation response that is strongest south of Pt. Reyes. The relaxation contribution to the time mean also causes slightly northward velocities near the coast just south of Pt. Arena (line 166), where a relaxation response is present, but much weaker than that south of Pt. Reyes.

The E line (line 130) is located about 10 km north of Cordell Bank, so it is possible that upstream influence of the bank causes this double jet velocity structure, which strengthens to the south as the isobaths are modified significantly by the presence of the bank (see Fig. 2). Other indications of upstream influence of Cordell Bank will be discussed later in connection with Lagrangian calculations presented in Section 5.2. The identifica-

Fig. 11. Time mean depth-averaged velocity vectors with magnitudes in color (upper left), contours of the transport streamfunction (upper middle), bottom-velocity vectors (upper right), surface relative vorticity $\zeta$ divided by $f$ (lower left), bottom potential density $\sigma_\theta$ in kg m$^{-3}$ (lower middle), and bottom turbulent kinetic energy (tke) in $10^{-4}$ m$^2$ s$^{-2}$ (lower right).
tion in the model results of this double jet structure is new. Attempts to find evidence of a similar jet structure in the observations, both from WEST and from other northern California programs such as CODE, did not lead to any conclusive results. It is unclear, however, whether an appropriate comparison could be made from these observations due to their limited coverage areas. The southernmost CODE survey line is just north of the E line, and the WEST CODAR region does not extend offshore of the 200-m isobath around Cordell Bank and Pt. Reyes. WEST shipboard ADCP surveys during May 2001 (M. Roughan, personal communication) are available at the A, D, and F lines, but only for a few days. The ADCP measurements do not always extend far enough onshore or offshore to capture the necessary range for a comparison. However, one ADCP survey on May 26 that extends offshore along the D line (line 124) shows negative $v$ velocity on the shelf, a region of near zero $v$ just offshore at the shelf break (30 km from the coast), and surface-intensified negative $v$ from about

![Fig. 12. Sections of time mean alongshelf velocity $v$ in m s$^{-1}$ (left) and potential density $\sigma_\theta$ in kg m$^{-3}$ (right) plotted as a function of depth (m) and distance offshore (km). The zero $v$ contour is shown in black. The alongshelf grid numbers shown in Fig. 2 are in the lower right corner of each $v$ section and corresponding WEST survey lines are in the lower right corner of each $\sigma_\theta$ section.](image)
32 to 50 km from the coast. The model $v$, sampled in the same manner as the observations, has a similar across-shelf structure. This is clearly a qualitative comparison, but one that shows that relevant available observations are not inconsistent with the model results.

Time mean density fields do not exhibit as noticeable a transformation between alongshelf locations, although the region of upwelled isopycnals extends farther offshore as the location progresses south from line 166 to line 104. This difference between sections is largest at the A line (line 104), where the 26.3 kg m$^{-3}$ isopycnal surfaces approximately 40 km offshore, as compared to 15–25 km offshore at lines 166–112. The contrast in across-shelf density front location illustrates that more dense water is upwelled at the surface over a larger expanse of the shelf south of Pt. Reyes, where the alongshelf coastal jet is displaced offshore relative to its location north of Pt. Reyes.

5. Dynamical analysis of upwelling and relaxation wind event

5.1. Eulerian analysis

We now focus our discussion on a specific wind event from May 17 to 23, 2001 during which the wind-stress begins as upwelling-favorable and greater than 0.1 N m$^{-2}$ until May 19, at which time it decreases to near zero on May 20. Both components of the wind-stress remain near zero until May 23 (see Fig. 4). The modeled upwelling and relaxation responses during this event are representative of those throughout the summer simulation period. This analysis allows a contrast between upwelling and relaxation conditions over a typical event time scale. Shown in Fig. 13 are surface velocity vectors, surface temperature, and surface elevation $\eta$ fields on May 18, 20, and 22. The wind-stress during the event is provided at the top for reference. The alongshelf coastal jet is displaced offshore relative to its location north of Pt. Reyes. Surface temperatures are coldest and surface elevations are lowest near the coast due to upwelling, with the coldest water north of Bodega Bay and a large region of cold water with low $\eta$ found south of Pt. Reyes from the coast to the 100-m isobath.

On May 20, surface velocities decrease everywhere except in the coastal jet region. Surface temperatures decrease between Pt. Reyes and Bodega Bay due to the southward advection of cold water from north of Bodega Bay. Surface elevations decrease near the coast by 2–3 cm. May 22 fields show the flow response to wind relaxation, with a decrease in surface velocities everywhere and higher surface temperatures and elevations near the coast. South of Pt. Reyes, model surface currents reverse direction near the coast and relatively warm surface waters are found. This warm water is advected up towards Pt. Reyes by the nearshore northward currents, as described in the relaxation response during CODE (Send et al., 1987). However, it appears that, in the model, strong southward currents deflected offshore by Pt. Reyes cause the warm water to be trapped off Pt. Reyes rather than going completely around the cape as described in Send et al. (1987). A possible reason for this difference is that spatial variations in upwelling intensity, caused by spatial variations in the wind field, play a role in the observed relaxation response inshore just north of Pt. Reyes.

In order to examine the dynamics of this event in more detail, we study fields of the depth-averaged velocity and alongshelf momentum term balance in Figs. 14 and 15. We write the momentum equation for the depth-averaged alongshelf flow in the $y$ direction as

$$\frac{\partial V}{\partial t} + U \frac{\partial V}{\partial x} + V \frac{\partial V}{\partial y} - G^y - F^y + fU + \frac{1}{\rho_0} \frac{\partial p}{\partial y} - \frac{1}{\rho_0 D} \tau^y_s + \frac{1}{\rho_0 D} \tau^y_b = 0,$$

where $(U, V)$ are the depth-averaged currents in the $(x, y)$ directions, $\tau^y_s$ and $\tau^y_b$ are the $y$ components of the surface and bottom-stress, respectively, $G^y$ is the dispersion term, $F^y$ is the horizontal diffusion term, $f$ is the Coriolis parameter, and $D$ is the water depth. We refer to the terms $\partial V/\partial t$ as acceleration, $U \partial V/\partial x + V \partial V/\partial y - G^y - F^y$ as advection, $fU + (1/\rho_0) \partial p/\partial y$ as the ageostrophic pressure gradient, $(1/\rho_0 D) \tau^y_s$ as wind-stress, and $(1/\rho_0 D) \tau^y_b$ as bottom-stress.

The depth-averaged velocity is southward with magnitudes of 0.2–0.3 m s$^{-1}$ in the coastal jet, which is shown to separate off Pt. Arena and Pt. Reyes. Weaker velocities are found in the lee of the capes. The bifurcation of the jet, as apparently caused by
upstream influence of Cordell Bank, is also shown with weak velocities directly over and north of the bank. Velocities increase near the capes on May 20, then decrease everywhere on May 22 following two days of near zero winds. On May 18, wind-stress values are relatively large everywhere on the shelf.
due to the spatial uniformity of the wind forcing. Bottom-stress values are high near the coast due to forcing of appreciable currents in the shallower water and off Pt. Reyes and Pt. Arena because bottom velocities are relatively large near the capes. As the winds decrease on May 20,
bottom-stress values decrease near the coast and the wind-stress is near zero everywhere. The regions of positive bottom stress south and offshore of Pt. Reyes on May 20 and 22 result from northward currents associated with the relaxation response.
In Fig. 15, large negative acceleration on May 18 is due to the continued strengthening of the southward coastal jet caused by upwelling wind-stress forcing. In the advection term, we see a spatial pattern of negative advection values offshore and north of Pt. Arena and Pt. Reyes and positive advection values to the south. This corresponds to southward spatial acceleration of the southward coastal jet off the capes and southward spatial deceleration south of the capes. The advection terms are balanced primarily by the opposite-signed ageostrophic pressure gradient term. Note the characteristic spatial variation of the advection and ageostrophic pressure gradient terms around Pt. Reyes and Pt. Arena also occurs on a smaller scale off Bodega Head, the small cape at the northern end of Bodega Bay in the center of the WEST study area.

As the winds decrease, the acceleration term switches sign on May 20 over the shelf and the alongshelf velocity field decelerates. The regions of large advection and ageostrophic pressure gradient off the capes increase in area and magnitude on May 20. All three terms in Fig. 15 then weaken on May 22.

The characteristics of the alongshelf pressure gradient pattern are of interest because the alongshelf topographic and coastline variability associated with Pt. Reyes and Pt. Arena are important factors in the dynamics of the shelf flow response. Gan and Allen (2002a) found a similar alongshelf pressure gradient distribution during upwelling-favorable winds in numerical modeling studies of this region. They explain the pattern as a response to decreased surface elevation south of the capes, which is related to increased velocities around the capes. In addition to a geostrophic balance in the across-shelf direction, which leads to lower surface elevation in regions of increased alongshelf velocities, nonlinear advection contributes significantly near Pt. Reyes and Pt. Arena. This leads to a gradient wind-like balance with a nonlinear centrifugal force term \( V^2/R \) that is important due to increased \( V \) and decreased radius of curvature \( R \) around the capes. Thus, the surface elevation response is tied to the alongshelf velocity through the across-shelf momentum balance, and the modified surface elevation field serves to intensify the alongshelf pressure gradient. In this study, we find similar results in daily plots of surface elevation \( \eta \) (Fig. 13) and in the ageostrophic pressure gradient and nonlinear advection terms from the across-shelf momentum balance (not shown). The surface elevation is low near the coast and south of Pt. Reyes and an across-shelf ageostrophic pressure gradient balances a relatively large nonlinear advection term near Pt. Reyes, and also near Pt. Arena. Throughout the event, surface elevation is a minimum in the region south of Pt. Reyes on May 20, and correspondingly, depth-averaged velocity vectors are largest just off Pt. Reyes and Pt. Arena on May 20 (Fig. 14). The changes in \( \eta \) and the depth-averaged velocities from May 18 to 20 are consistent with the strengthening of the alongshelf nonlinear advection and ageostrophic pressure gradient terms.

Also of interest is the coastal jet behavior across sections at the A, D, and F lines (Fig. 16). The structure of the alongshelf velocity and density during the event is quite different at these locations, as expected from the time mean fields discussed in Fig. 12. At the F line (line 148), the coastal jet weakens from May 20 to 22 with near-zero velocities inshore on May 22. The upwelled isopycnals also retreat slightly toward the coast following the relaxation of upwelling winds. At the D line (line 124), the velocity structure is more complex due to the strong inshore jet and weaker, but deeper, offshore jet with northward velocities at depth in between. The density field does not show a significant difference from the F line. At the southernmost A line (line 104), the complexity increases as velocities in the offshore jet are as strong as those in the shallower inshore jet. Both jets are displaced offshore relative to their D line locations due to separation off Pt. Reyes, and a region of significant northward velocities is found inshore near the coast. The density front is 30 km offshore and relatively dense water covers the broad shallow shelf inshore, illustrating the effectiveness of the upwelling shelf circulation at the A line. These sections clearly illustrate the alongshelf variability in the model upwelling and relaxation response. The coastal jet structure becomes more complex and the relaxation response more pronounced progressing southward. South of Pt. Reyes, upwelling occurs over the full extent of the shelf and significant northward flow is found inshore of 20 km during relaxation.

5.2. Lagrangian analysis

Lagrangian fluid motion is calculated using two different techniques. The first approach involves
Fig. 16. Sections of alongshelf velocity $v$ (left) and potential density $\sigma_y$ (right) along the F (line 148—top), D (line 124—middle), and A (line 104—bottom) lines on May 18, 20, and 22. The zero $v$ contour is shown in black.
computing parcel trajectories by solving the differential equations
\[
\frac{dx}{dt} = u, \quad \frac{dy}{dt} = v. \tag{2}
\]
Parcels are indexed by their initial positions at every \((x, y)\) grid point. In the vertical, parcels are initialized at the center of the grid cell next to the surface and remain at this depth throughout the simulation. Numerical solutions to (2) are calculated utilizing a fourth-order Milne Predictor–Hamming Corrector scheme. The time step used for the Predictor–Corrector scheme is the model internal time step. The updated positions obtained from the solutions to (2) are recorded daily for each initial parcel position.

In addition to tracking individual water parcels, we utilize a second technique that provides Lagrangian trajectories for a continuous field of parcels. This approach involves the definition of three Lagrangian label fields that are advected by the model velocities (Kuebel Cervantes et al., 2003). The Lagrangian labels, \(X(x, y, z, t)\), \(Y(x, y, z, t)\), and \(Z(x, y, z, t)\), satisfy the following equations:
\[
\frac{DX}{Dt} = 0, \quad \frac{DY}{Dt} = 0, \quad \frac{DZ}{Dt} = 0, \tag{3}
\]
where \(D/Dt = \partial/\partial t + u\partial/\partial x + v\partial/\partial y + w\partial/\partial z\). The initial conditions are
\[
X(x, y, z, t = 0) = x, \quad Y(x, y, z, t = 0) = y, \quad Z(x, y, z, t = 0) = z. \tag{4}
\]
For calculations without errors, i.e., those performed in the limit of vanishing spatial grid sizes and time steps, the Lagrangian parcel trajectories obtained from both techniques would be equivalent.

In this application of the Lagrangian labels, the initial \(Y\) and \(Z\) labels are a function of their initial alongshelf and vertical location, respectively, as in (4). To aid in the physical interpretation, however, the \(X\) label is initialized as the water depth \(h\),
\[
X(x, y, z, t = 0) = h(x, y). \tag{5}
\]
This allows a ready assessment of across-isobath displacement of water parcels, which we refer to as onshore/offshore displacement in the following discussion. The labels \(X\), \(Y\), and \(Z\) are calculated as fields on the model grid with the third-order upstream bias advection scheme. Since the model also includes a parameterization of small-scale turbulence, these Lagrangian parcel trajectories comprise only the resolved part of the full fluid motion represented in the model.

Examination of the time evolution of the label fields helps to provide insight into the Lagrangian characteristics of the flow. Surface distributions of the \(Y\) label and of the Lagrangian parcels, both initialized on May 17, are shown in Fig. 17. The color contours correspond to the initial alongshelf location of the \(Y\) labels, measured as distance from the southern boundary of the model domain. The colorbar indicates values from 100 to 400 km. The region shown in Fig. 17 is a sub-domain of the model and begins 150 km north of the southern boundary. The alongshelf locations of the surface parcels (bottom) have been color-coded to allow a comparison with the evolution of the surface \(Y\) label (top); therefore, the colors assigned to the parcels have the same meaning as the \(Y\) label and the colorbar applies to both fields. A striking feature in Fig. 17 is the significant southward advection in the region of the coastal jet, with parcels being advected up to 100 km southward by May 22. The structure shows a well-defined split in the trajectories that occurs near Cordell Bank, with significant southward advection both offshore of the bank and inshore along the 100-m isobath. In regions inshore of the 100-m isobath, on Cordell Bank, and south of Pt. Reyes, the southward surface advection is decreased. Just off Pt. Reyes, a cyclonic recirculation region, also evident in Fig. 13, is due to the interaction of northward flow south of the cape and the southward coastal jet.

The evolution throughout the event is shown most clearly in the \(Y\) label fields, but the parcel tracking shows a very similar response. Spatial correlations between the \(Y\) label and the parcels computed at the parcel locations range from 0.99 on May 18 to 0.95 on May 22. The comparison in Fig. 17 is evidence that the two techniques perform consistently and provides a quantitative check on the ability of the Lagrangian methods.

Depth-averaged values of the \(\Delta X\) and \(\Delta Y\) and surface \(\Delta Z\) Lagrangian label displacements are shown in Fig. 18 during the event. The label displacements are calculated as \(\Delta X(x, y, z, t) = X(x, y, z, t = 0) - X(x, y, z, t)\), \(\Delta Y(x, y, z, t) = Y(x, y, z, t = 0) - Y(x, y, z, t)\), and \(\Delta Z(x, y, z, t) = Z(x, y, z, t - 0) - Z(x, y, z, t)\) where the time of initialization \(t = 0\) corresponds to May 17. Recall that the \(X\) Lagrangian label is initialized as water depth (5) so that positive values of \(\Delta X\), \(\Delta Y\), and \(\Delta Z\) denote onshore, southward, and upward displacements,
respectively. The pattern of significant southward
$\Delta Y$ displacement on May 20–22 corresponds to the
coastal jet location during upwelling wind forcing.
$\Delta Y$ values are small inshore of the jet near the coast
and south of Pt. Reyes due to the coastal jet
separation. A large region of onshore displacement
(positive $\Delta X$) just offshore of the 200-m isobath is
apparent near the F line on May 18, with offshore
displacement (negative $\Delta X$) to the north and south.
Velocity fields on May 16 and 17 (not shown)
indicate evidence for a cyclonic eddy-like circulation
south of Pt. Arena that weakens as the southward
alongshelf currents intensify on May 18. Velocities
along the F line are onshore at the southern edge
of this eddy, apparently leading to the onshore
displacement seen in $\Delta X$ at this location. The
negative regions just off the shelf break are
caused by offshore velocities due to the coastal
jet separation off Pt. Reyes and Pt. Arena. The
positive $\Delta X$ patch is advected southward and
distorted on May 20–22. Inshore of the 100-m
isobath between Pt. Reyes and Pt. Arena, the $\Delta X$
values are generally positive, reflecting net onshore
displacement.

Fig. 17. Surface Y Lagrangian label in km (top) and Lagrangian parcels (bottom) initialized on May 17 and plotted on May 18, 20, and
22. Spatial correlation coefficients between the labels and the parcels are given for each day at the bottom. The colorbar indicates values
from 100 to 400 km, measured as distance from the southern boundary of the model domain. The initial $Y$ values at the southern and
northern boundaries of the region plotted are 150 and 320 km, respectively. The region shown is a sub-domain of the model and begins
150 km north of the southern boundary. The locations of the A, D, and F lines are indicated in the right $Y$ panel.
Surface $\Delta Z$ shows upwelling of water from about 40 m depth in a few locations near the coast on May 18. Significantly more water from depth is found on May 20 and 22, with initial depths greater than 60 m from the coast to about 10 km offshore everywhere north of Pt. Reyes. The response south of Pt. Reyes...
shows that the spatial pattern of upwelling is complex and extends offshore in response to the separated coastal jet, rather than being contained in a cohesive region near the coast, and that water upwells to the surface from shallower depths than north of Pt. Reyes. Note the effects of the cyclonic recirculation south of Pt. Reyes discussed previously with reference to Fig. 17. This feature is associated with upwelling of water from the north and offshore transport of water parcels (positive $\Delta Y$ and $\Delta Z$ and negative $\Delta X$).

A Lagrangian investigation corresponding to the discussion of Fig. 16 in Section 5.1 is presented with sections of all three Lagrangian label displacement fields along the A, D, and F lines during the upwelling and relaxation event (Fig. 19). The $\Delta Y$ Lagrangian label displacement shows a structure very similar to the alongshelf velocity, with the largest southward displacement corresponding to the jet location at all three sections. Upwelling of parcels occurs across the shelf, with upward displacements of up to 60 m in $\Delta Z$. Offshore displacements indicated by negative $\Delta X$ occur in the region of the upwelling jet, with the largest values at the surface due to the surface Ekman flow, and decreasing values with depth. Large patches of onshore displacement are found on the shelf at the F line (148) on May 18 and at the D line (124) on May 22, corresponding to the feature discussed in Fig. 18. At the A line (104), the region of offshore displacements is a striking feature that is related to the alongshelf flow. The $\Delta X$ response at the A line is caused by the coastal jet separation off Pt. Reyes, so that water in the jet is advected southward and offshore. Because the depth rapidly increases offshore of 100 m at the A line, the $\Delta X$ values are large just off the shelf break, although the actual offshore advection distance is 20 km or less. On the shelf at the A line, values of the label displacements are smaller than those at the D and F lines.

A conclusion of the analysis of Fig. 19 is that the source of upwelled water near the coast at the D and F lines is farther north and deeper than that at the A line, given the larger values of $\Delta Y$ and $\Delta Z$. Hence, water that is upwelled on the shelf just south of Pt. Reyes is from a more local source because the presence of the cape significantly impacts the upwelling dynamics through jet separation. This result is consistent with the idea that the lee of a large headland such as Pt. Reyes is a region of longer residence times, and thus, higher retention of planktonic organisms (Wing et al., 1995). As shown in Fig. 16, the upwelling circulation that occurs at the A line is effective at bringing more dense water to the surface across the entire shelf.

In order to explore further the question of how the Lagrangian dynamics vary as a function of across-shelf distance and the differences in upwelling source waters at the mooring locations, we plot contours of all three Lagrangian label displacements, as well as $v$ and $\sigma_t$, as a function of depth and time during the event at the D line (Fig. 20). Significant upwelling of deeper and more dense water from about 40 km to the north occurs at D040. At D090, upwelling of water from about 80 km to the north occurs, but the deeper water does not surface as at D040. The appearance of water from offshore is seen in $\Delta X$ at 40 m depth on May 19. This patch of offshore water corresponds to a smaller positive $\Delta X$ region at the D line on May 20 in Fig. 19 and the onshore spreading of the large positive $\Delta X$ region in Fig. 18. The water appears to be upwelled, as shown by its location at the upper edge of the upwelled water in $\Delta Z$. With regard to Lagrangian displacements, the upwelling response at D090 is clearly more complex than at D040.

At D130, there are distinct contrasts with the inshore moorings. Noticeable upwelling is not evident in either $\Delta Z$ or $\sigma_t$, but further evidence of the onshore patch of water discussed previously is found. $\Delta X$ shows that water from offshore is advected into the site at 40 m depth on May 21 and displaced downward to 100 m depth by May 23. Because there is no corresponding signal in $\Delta Z$, this water is not upwelled from depth. Water advected over 150 km south is seen near the surface at D130 on May 22, suggesting that upwelled water is from farther north at D130 than at the inshore moorings. Mooring observations of temperature and salinity also support a more northern source of upwelled water at D130 (E. Dever, personal communication).

A final illustration of the alongshelf variability of the Lagrangian characteristics is provided by time series of $v$, $\sigma_t$, $\Delta X$, $\Delta Y$, and $\Delta Z$ at 90 m depth along the A and F lines (Fig. 21). The alongshelf velocity $v$ is weaker than at D090, especially at the F line, where $v$ is near zero below about 20 m over much of the event period. Deeper isopycnals reach the surface at 90 m depth at the F line than at the D line, but upwelling in the isopycnals at the A line is much less evident. The A090 location is on the edge of the density front, just inshore of the signature of Cordell Bank (see Fig. 16). The density front is
farther offshore at the A line than at the northern lines, as discussed previously. Thus, upwelling occurs over much of the shelf region at the A line (Figs. 18 and 19), but the evolution throughout the event is relatively weak at this 90-m location. The structure of the $\Delta Y$ time series is similar at all three locations, especially at the A and D lines. Note that the location of A090 is offshore of the region of locally upwelled water on the A line discussed previously and it shows a significant signal in $\Delta Y$.
While ΔY displacements are comparable between the A, D, and F lines during the event, the ΔX and ΔZ responses are quite different. ΔZ displacements are generally complex at all three locations. ΔZ shows that water at depth after May 19 is upwelled from shallower depths on the F line as compared to the A and D lines. As shown in Fig. 19, deep water is upwelled at F; however, it does not reach as far inshore. ΔX shows that significantly less across-isobath motion occurs at 90-m depth at the A and F lines than at the D line. The A and F lines do not show evidence at 90 m of the large patch of onshore-displaced water that is prevalent at the D line in Figs. 18–20.

The four main results from the three-dimensional Lagrangian analyses discussed in this section can be summarized as follows: first, the influence of the double jet structure on the southward displacement of parcels over the shelf is clearly shown to produce greater southward displacement along the 100-m isobath and offshore of Cordell Bank. Directly over the bank and near the coast, southward displacement is decreased. As discussed in Section 4, although the bifurcation of the upwelling jet north of Cordell Bank was not directly seen in the observations due to limited spatial coverage, we believe this feature, which has significant impact on water parcel displacements in the model, may be
real. Second, alongshelf variability in fluid parcel displacement and, more specifically, in upwelling source waters, is introduced by the presence of Pt. Arena and Pt. Reyes. Offshore of these capes, parcels are advected off the shelf due to the offshore-directed velocities associated with the separation of the coastal upwelling jet. The result that upwelling source waters differ north and south of Pt. Reyes illustrates topographically-induced alongshore variability in Lagrangian aspects of the upwelling circulation that are not clearly derived from examination of the Eulerian fields alone. Smaller values of water parcel displacements on the shelf just south of Pt. Reyes indicate that water is upwelled locally. Thus, residence times in this region may be relatively high compared to surrounding areas, which could contribute to the retention of planktonic organisms. Roughan et al. (2005) found evidence that the presence of Bodega Head, which is a much smaller headland than Pt. Reyes, increases larval retention in Bodega Bay. VanderWoude et al. (2006) identified areas of
retention north and south of Pt. Reyes with satellite observations of sea-surface temperature and chlorophyll a. Third, a cyclonic recirculation feature detected south of Pt. Arena prior to the event leads to a large patch of onshore-displaced parcels that first appears near the F line and is displaced southward and distorted in time. Eddy-like structures do appear in satellite images of sea surface temperature and in CODAR measurements of surface velocity in this coastal region (Kaplan and Largier, 2006). The resulting Lagrangian behavior shows directly that eddy-like structures that develop as a result of flow-topography interactions can have a significant impact on across-shelf transport. Finally, across-shelf variability in the response of water parcels along the D line is evident. This includes decreased ΔZ and increased ΔY in the offshore direction, indicating that upwelled water at D130 is from farther north (north of Pt. Arena) than water that is upwelled at the inshore moorings.

6. Summary

We have presented a modeling application using the regional ocean modeling system (ROMS) with the objective of studying and understanding the wind-driven flows off Northern California in the region of the wind events and shelf transport (WEST) project. In this modeling effort, we set out to answer questions about flow variability across and along the shelf, flow complexities introduced by Cordell Bank and Pt. Reyes, and the source of upwelled water in different areas of the region. To address these questions, we used both an Eulerian and a Lagrangian approach and focused analysis on an upwelling and relaxation wind event that is representative of those occurring throughout the summer in this region.

Numerous model-data comparisons suggest that the model performs reasonably well in this region, with high correlations between modeled and measured depth-averaged velocities at the mooring locations, surface velocities in the CODAR region, and potential density along survey lines. A weakness of the model is the inability to fully capture the reversal of southward velocities that is observed following relaxation of upwelling winds.

The mean model circulation over the period of May 5–June 20, 2001 is dominated by an upwelling response, including a southward coastal jet of up to 1 m s⁻¹ that separates from the coast off Pt. Arena and Pt. Reyes. The capes are regions of nonlinearity in the surface flow and significant turbulence near the bottom. Sections of alongshelf velocity at several locations reveal significant alongshelf variability and the existence of a distinct double jet structure that develops just north of Bodega Bay. Upstream influence of Cordell Bank is postulated to be the cause for this double jet.

With regard to the time variability of the flow, we have focused attention on a specific upwelling and subsequent relaxation wind event. Flows around Pt. Reyes show a different response north and south due to interactions of the shelf flow with the highly variable coastline shape and shelf geometry. South of Pt. Reyes, the upwelling circulation is effective in bringing relatively cold, dense water to the surface over the entire shelf. Relaxation of upwelling winds causes the development of northward velocities south of Pt. Reyes and near the coast between Pt. Reyes and Pt. Arena. Depth-averaged momentum balance terms illustrate the importance of the alongshelf topographic and coastline variability in causing a negative nonlinear advection offshore of Pt. Reyes and Pt. Arena, balanced by positive ageostrophic pressure gradient. This pattern is due to the spatial acceleration of the velocity around the capes.

The Lagrangian flow characteristics and their relationship to the Eulerian fields throughout the wind event are also discussed. The Lagrangian displacement fields are a useful tool in determining the response of water parcels to upwelling wind forcing, thereby helping to address WEST objectives regarding the source of upwelled water in the region and the retention of water parcels on the shelf. Alongshelf sections and area maps of ΔX, ΔY, and ΔZ show the source of upwelled water near the coast at the A line is local, as compared to that north of Pt. Reyes, which has a signature of deeper water from farther north. This result once again points to a fundamentally different upwelling response north and south of Pt. Reyes, and gives further evidence of the significance of the coastline shape in the behavior of the shelf flow and resulting water parcel displacement. Additional results from the Lagrangian analysis include significant southward displacement in the double jet region, significant onshore displacement offshore of Pt. Arena due to a cyclonic eddy-like recirculation feature, and indications that water upwelled at the offshore D mooring came from north of Pt. Arena while water upwelled at the inshore moorings came from a more local source.
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