# A modeling study of shelf circulation off northern California in the region of the Coastal Ocean Dynamics Experiment 2. Simulations and comparisons with observations

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[1] This is the second part of a modeling study of wind-forced flow on the continental shelf off northern California in the region  $(37^{\circ}-40^{\circ}N)$  of the Coastal Ocean Dynamics Experiment (CODE). Gan and Allen [2002] analyzed the shelf flow response to idealized wind stress forcing in a process-oriented study. The study here applies forcing from observed winds and heat flux for April-May 1982 and compares the model results with moored current and temperature measurements. The Princeton Ocean Model (POM) is used in a three-dimensional limited area domain with a high-resolution curvilinear grid (approximately 1 km horizontal spacing, 60 vertical levels) and realistic coastline and bottom topography. The objectives of the study are to simulate the response of the shelf circulation field to time-varying observed wind stress and heat flux, to compare model results with oceanographic observations to establish confidence in the model, and to subsequently analyze the model fields and the model dynamical balances to help understand the behavior of the observed flow. The model variables show overall good agreement with corresponding observations. Similar to the conclusions by Gan and Allen [2002], it is found that the alongshore variability of upwelling is mainly controlled by the interaction of the wind-forced shelf flow with the coastline and bottom topography. Different dynamical regimes in the regions north and south of the coastal capes formed by Pt. Reves and Pt. Arena and in the more uniform region between these capes are identified and investigated. The results demonstrate that the coastal capes play a dominant role in causing alongshore variability of the upwelling flow, including the setup of an alongshore pressure gradient that forces northward currents during relaxation of southward upwelling favorable winds. An analysis of the balance of terms in the equation for potential temperature indicates that across-shore temperature advection is the major contributor to the cooling of coastal water during upwelling, with a larger magnitude to the south of the capes. To the north of the capes, however, alongshore temperature advection is the dominant contributor to the colder water near the coast. INDEX TERMS: 4255 Oceanography: General: Numerical modeling; 4219 Oceanography: General: Continental shelf processes; 4512 Oceanography: Physical: Currents; KEYWORDS: upwelling, relaxation, shelf circulation, numerical modeling

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# 1. Introduction

[2] The Coastal Oceanic Dynamics Experiment (CODE, 1981–1982) was designed to understand wind-driven dynamical processes in the coastal ocean using high-resolution observations [e.g., *Beardsley and Lentz*, 1987]. Although the shelf topography in the central CODE region is fairly uniform, strong coastline and bottom topography variations associated with Pt. Arena and Pt. Reyes are found to the north and south of the domain (Figure 1). During the

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spring and summer time period of the CODE field experiments, winds are predominantly southward and favorable for upwelling. CODE observations [*Winant et al.*, 1987; *Huyer and Kosro*, 1987; *Kosro*; 1987; *Kosro and Huyer*, 1986] show strong spatial and temporal variability in the upwelling circulation. In response to time-varying wind forcing and to spatially variable coastal topography, the upwelling response exhibits a complex time- and spacedependent three-dimensional circulation. South of Pt. Arena, the upwelling coastal jet tends to flow offshore, cross isobaths and to form a cyclonic eddy [*Kosro and Huyer*, 1986]. At the C line (Figure 1) where the coastline is relatively uniform, the response is closer to two-dimen-



Figure 1. Map of the CODE region showing the topography and the locations of the current meter moorings (solid dots) in 1982. Meteorological measurements were obtained from surface buoys at  $C_2$ ,  $C_3$ ,  $C_4$ ,  $C_5$ ,  $N_3$ ,  $R_3$ , and at some of the stations shown as open circles [adapted from *Limeburner*, 1985].

sional with weaker alongshore variations. The observations show that the flow field is markedly altered by the coastline variations near Pt. Arena. Although limited observations were made near Pt. Reyes during CODE, a similar response is expected to occur there and that will be examined in this study.

[3] During the weakening or reversal of southward winds, southward currents over the shelf weaken or reverse direc-

tion to flow northward near the coast. This phenomenon, termed upwelling relaxation, is clearly present in the observations [Send et al., 1987; Kosro, 1987]. The process study by Gan and Allen [2002] shows that the northward currents during wind relaxation are driven by a negative (northward) pressure gradient that is set up by the interaction between the time-varying wind-forced coastal currents and the coastal topography during upwelling. The pressure gradient variations near the coast are associated with the dynamical response of the flow to coastline and bottom topography. In particular, the acceleration of the geostrophically balanced southward alongshore current in the vicinity of coastal capes results in lower pressure at the coast. In addition, the enhancement of nonlinear advective effects further reduces the pressure through the gradient wind balance and this strengthens the negative pressure gradient south of the capes. This negative pressure gradient geostrophically balances onshore flow at depth, which intensifies local upwelling. As the winds relax and the onshore flow subsides, the negative pressure gradient drives northward currents.

[4] In this paper, the shelf flow response in the CODE region is studied further by using observed atmospheric data to force the model, by comparing the model fields with CODE observational data, and by analyzing dynamical balances. In addition, the time-varying and time-averaged thermal response is analyzed to better understand the relative contributions of alongshore and across-shore advection to upwelling and how these contributions vary along the coast. A period with a relatively complete set of measurements and atmospheric forcing fields in April and May 1982 is chosen for this modeling study (Figure 2). Winds are predominantly upwelling favorable during this time, with a short period of downwelling from April 10-15. Two typical upwelling relaxation events occur around April 19 and May 4. Strong southward winds are rapidly reduced to no winds or to weak northward winds on a timescale of 3-4 days during the relaxation time periods. Using idealized wind stress, the complex dynamics involved in the upwelling relaxation have been investigated in Part 1. The goal of this study is to utilize the circulation model with time-dependent atmospheric forcing, together with the CODE measurements, to further explore the mechanisms that govern the time and space variability of the observed upwelling and the relaxation processes. The outline of the paper is as follows. The ocean model and its implementation are introduced in section 2.



**Figure 2.** Time series of the wind stress components calculated from wind measurements at buoy C<sub>3</sub>. The heavy line is the alongshore component  $\tau^{\nu}$  (positive northward), and the light line is the across-shore component  $\tau^{x}$  (positive eastward). The alongshore direction is 317°T. The mean values and standard deviation of ( $\tau^{\nu}$ ,  $\tau^{x}$ ) are (-0.098, 0.022) Pa and (0.123, 0.014) Pa, respectively.



Figure 3. (left) Model curvilinear grid and (right) topography with the 100, 500, 1000, and 1500 m isobaths shown.

Comparisons of model results and observation are given in sections 3 and 4. Analyses of dynamical and thermal balances are discussed in section 5. A summary of the findings is presented in section 6.

# 2. Ocean Model

[5] The model used is the Princeton Ocean Model (POM) [Blumberg and Mellor, 1987] for three-dimensional, time-dependent, oceanographic flows governed by the hydrostatic primitive equations. The model incorporates the Mellor and Yamada [1982] level 2.5 turbulent closure scheme to parameterize vertical mixing. The model domain used here is 469 km in the alongshore direction and 155 km in the across-shore direction (Figure 3). The alongshore direction has been rotated counterclockwise from north by 24.5° so as to be better aligned with the coastline. In order to include only locally generated alongshore pressure gradients and to avoid the set up of any artificial alongshore pressure gradients, periodic boundary conditions are chosen at the north and south acrossshore boundaries so the domain is similar to a periodic channel. To construct this periodic domain, regions with straight coastline of alongshore extent 70 km in the north and 49 km in the south are utilized to gradually adjust the shelf and slope topography so that it agrees at the north and south boundaries. A high-resolution orthogonal curvilinear horizontal grid is used, which allows for a smooth and accurate representation of the variable shoreline. The grid sizes are approximately 1 and 1.3 km in the across-

shore and alongshore directions, respectively. The domain includes  $151 \times 359$  grid cells and ranges in latitude from  $36^{\circ}$  to  $40^{\circ}$ N and in longitude from  $122^{\circ}$  to  $126^{\circ}$ W. The vertical coordinate has  $60 \sigma$  levels which resolves both the surface and the bottom boundary layers over the continental shelf. The bottom topography (Figure 3) in the region of deep water is interpolated from ETOPO5 data (National Geodetic Center, Boulder, Colorado). Data from the USGS (CD ROM) and digitized data from a highresolution (1/240 degree) bathymetry map (NOAA, NOS 1307N-18B, 1974) are used to construct the water depths in the near coastal region. The bathymetry is slightly smoothed to reduce truncation errors. Since we are mainly focusing on the flow field over the continental shelf, the maximum water depth is chosen to be 2000 m. On the western open boundary, no normal flow and radiation conditions are applied to the external and internal normal velocity components, respectively.

[6] The model is initialized as by *Gan and Allen* [2002] with horizontally uniform climatological temperature and salinity profiles and with zero velocities. The wind stress is calculated from the wind measurement at buoy C<sub>3</sub> (Figure 1) following *Large and Pond* [1980] and is assumed to be spatially uniform. The wind stress time series is filtered with a 36 hour low-pass filter. Using observed meteorological variables, surface heat flux is obtained from bulk aerodynamic formulae, as described in Appendix A. The horizontal kinematic eddy viscosity and diffusivity coefficients are constant and chosen to be small 5 m<sup>2</sup> s<sup>-1</sup> so that the effects of horizontal diffusion are minimal.



**Figure 4.** Time series of observed (dotted lines) and modeled (solid lines) daily averaged alongshore velocities at moorings  $N_2$ ,  $C_2$ ,  $R_2$ , and  $C_3$ . Correlation coefficients between corresponding observed and modeled velocity components are listed on the right hand side.

[7] The wind field in the CODE region is known to have substantial spatial variability [e.g., *Winant et al.*, 1988]. Winds at buoy  $C_3$ , for example, are typically stronger than those south of Pt. Reyes. Consequently, the assumption here of spatially uniform wind stress is not strictly accurate. We make that assumption because spatially variable wind would complicate the interpretation of results from the periodic channel geometry. One advantage of the utilization of spatially uniform winds in these initial studies is that the mechanisms involved in the interaction of wind-forced shelf flow with alongshore variable shelf topography may be studied in isolation without additional influence from alongshore variability in the atmospheric forcing.

#### 3. Model-Data Comparisons

#### 3.1. Comparisons With Mooring Data

[8] Time series of model and observed alongshore velocity v from the listed measurement depths at several mooring locations are shown in Figure 4. Corresponding time series for the temperature T at C<sub>3</sub> are shown in Figure 5. The temperature variability and the model-data comparisons at the other moorings locations in Figure 4 are similar to that shown for C<sub>3</sub> in Figure 5. All time series and statistical functions for both model and observations are calculated from daily averaged variables. The currents at all depths vary in response to the wind stress (Figure 2). Colder water temperatures are found in the water column after strong upwelling on April 19. At the near coast locations, N<sub>2</sub>, C<sub>2</sub> and R<sub>2</sub>, model and observed alongshore currents are highly correlated, with larger correlation coefficients (CC) found for the upper part of the water

column. At the midshelf mooring  $C_3$ , the CC values are smaller for the upper part of the water column, but gradually increase at depth. The quantitative agreement of the model and observed temperatures and the high values (>0.9) of the CC indicate that the simulated thermal field is close to the conditions in the ocean. The magnitudes of the alongshore velocity from model outputs are reasonably close to those observed during upwelling and during relaxation. Both modeled and observed results indicate that the northward currents during relaxation start from the deeper water at all stations, with the first appearance at the C line. The model, however, has stronger southward alongshore velocities during upwelling and weaker northward velocities during relaxation. The large southward alongshore velocities are presumably because the model includes only the locally generated response from a spatially uniform wind field. The effects, for example, of weaker winds south of Pt. Reyes are not represented. In addition, a northward pressure gradient force formed as part of the interior ocean response over larger alongshore scales would not be included in the present study.

[9] The agreement between the model and observed results can also be seen from the time mean horizontal velocities at water depths of 10 m and 53 m (Figure 6). Both modeled and observed velocities show the existence of a southward coastal jet located at the midshelf. Weaker or northward mean currents at nearshore locations between Pt. Arena and Pt. Reyes reflect weaker southward flow during upwelling or northward currents during wind relaxation. North of Pt. Arena at station I, the mean coastal jet in both the model and observations is directed offshore



**Figure 5.** Time series of observed (dotted lines) and modeled (solid lines) daily averaged temperatures at mooring  $C_3$ . Correlation coefficients between corresponding observed and modeled temperatures are listed on the right hand side.

suggesting coastal jet separation at the cape. The model tends to generate a stronger southward flow compared to that observed, but overall appears to have qualitatively similar spatial variability. The systematically larger southward currents in the model lead to a small time mean southward flow near the coast, instead of to the mean northward currents shown in the observations. The values of the standard deviation (std) of the velocity components resolved in local principal axes (Table 1) from both observed and model fields indicate good agreement at near coast locations. The standard deviations of the modeled flow field at the offshore stations on the shelf and upper slope, however, are larger than observed. The reasons for this are unclear, but again are presumably partly related to the influence on the model response of the limited area alongshore-periodic coastal domain forced with spatially uniform winds.

[10] Figures 7 and 8 show the time mean modeled and observed alongshore velocity v and temperature T and the corresponding *std*, as a function of water depth at the mooring locations along the C line. Values from the model and from the observations are reasonably close. Better agreement is found for mean v at the offshore locations while better agreement is found for the *std* near the coast. Modeled mean temperatures in the whole water column across the C line are quantitatively similar to the observations (Figure 8). The *std* values of temperature from the model are generally larger than those from observations, with the largest difference of 1 C found at the nearshore

station  $C_2$ . Both simulated and observed *std* values, however, show the same tendency toward depth independence.

[11] A comparison of the spatial characteristics of the modeled and observed flow fields can be seen from the cross-correlation coefficients (CC) between the alongshore



**Figure 6.** Time mean modeled and observed velocity vectors (m  $s^{-1}$ ) at 10 m and 53 m depths from the CODE mooring locations.

|                      | v at     | 10 m    | <i>u</i> at | 10 m    | v at 5   | 53 m    | u at a   | 53 m    |
|----------------------|----------|---------|-------------|---------|----------|---------|----------|---------|
| Station, m           | Observed | Modeled | Observed    | Modeled | Observed | Modeled | Observed | Modeled |
| I <sub>3</sub> (90)  | 0.28,    | 0.45    | 0.05,       | 0.05    | 0.13,    | 0.21    | 0.03,    | 0.02    |
| $I_4(130)$           | 0.21,    | 0.43    | 0.08,       | 0.05    | 0.10,    | 0.33    | 0.05,    | 0.02    |
| $N_2$ (60)           | 0.29,    | 0.22    | 0.03,       | 0.02    | 0.15,    | 0.08    | 0.02,    | 0.01    |
| $N_3(90)$            | 0.35,    | 0.27    | 0.05,       | 0.04    | 0.17,    | 0.07    | 0.03,    | 0.02    |
| N <sub>4</sub> (130) | 0.23,    | 0.15    | 0.08,       | 0.01    | 0.15,    | 0.23    | 0.04,    | 0.04    |
| $C_2$ (60)           | 0.19,    | 0.15    | 0.02,       | 0.01    | 0.11,    | 0.09    | 0.01,    | 0.02    |
| $C_{3}(90)$          | 0.22,    | 0.19    | 0.05,       | 0.03    | 0.14,    | 0.14    | 0.03,    | 0.02    |
| C <sub>4</sub> (130) | 0.22,    | 0.36    | 0.09,       | 0.08    | 0.13,    | 0.24    | 0.04,    | 0.04    |
| $C_5$ (400)          | 0.16,    | 0.34    | 0.09,       | 0.05    | 0.11,    | 0.29    | 0.06,    | 0.03    |
| R <sub>2</sub> (60)  | 0.16,    | 0.17    | 0.03,       | 0.03    | 0.08,    | 0.08    | 0.02,    | 0.02    |
| R <sub>3</sub> (90)  | 0.14,    | 0.24    | 0.04,       | 0.04    | 0.11,    | 0.13    | 0.02,    | 0.03    |
| R <sub>4</sub> (130) | 0.20,    | 0.35    | 0.06,       | 0.05    | 0.10,    | 0.22    | 0.04,    | 0.03    |

**Table 1.** Standard Deviation (m s<sup>-1</sup>) of Observed and Model Velocity Components (*v*, *u*) Resolved in Local (Major, Minor) Principal Axes at Approximately 10 and 53 m Depths From the Indicated Mooring Locations<sup>a</sup>

<sup>a</sup> The water depth (m) at the mooring locations is given in parentheses.

 $(v_p, \text{ major principal axis})$  and between the across-shore  $(u_p, \text{minor principal axis})$  velocity components from different alongshore locations. Figure 9 shows the CC between  $v_p$  and between  $u_p$  from 35 m water depth at C<sub>2</sub> with the corresponding variables from R<sub>2</sub> and N<sub>2</sub> and similarly for C<sub>3</sub>, R<sub>3</sub>, and N<sub>3</sub> at 35, 53 and 70 m water depths. In general, at all these stations and depths the correlations of  $v_p$  are relatively high while those of  $u_p$  are relatively low for both modeled and observed flow fields, implying much shorter alongshore correlation scales for the across-shelf velocity  $u_p$ 

than for the alongshelf velocity  $v_p$ . Dever [1997] found similar results from the CODE currents measurements. A further analysis of the dynamical balances related to the different alongshore spatial correlation scales for  $u_p$  and  $v_p$ will be presented in section 5.1. Figure 9 also shows that the CC values of  $v_p$  between R and C from modeled and observed are very close. The CC of  $v_p$  between C and N from the model, however, have larger values than those from the observations. It is reasonable to believe that the CC between C and N may be smaller than the CC between C and R due to stronger eddy





**Figure 7.** (left) Time mean values and (right) standard deviations of the observed (green) and modeled (blue) alongshore velocities v (m s<sup>-1</sup>) as a function of depth at stations C<sub>2</sub>, C<sub>3</sub>, C<sub>4</sub>, and C<sub>5</sub>.

**Figure 8.** (left) Time mean values and (right) standard deviations of the observed (green) and modeled (blue) temperatures  $T(^{\circ}C)$  as a function of depth at stations  $C_2$ ,  $C_3$ ,  $C_4$ , and  $C_5$ .



**Figure 9.** Cross-correlation coefficients between alongshore velocity components (solid) and between across-shore velocity components (dashed) from  $R_2$ ,  $C_2$ , and  $N_2$  at water depth 35 m and from  $R_3$ ,  $C_3$ , and  $N_3$  at water depths 35, 53, and 70 m. Model-model correlation coefficients are blue, and observed-observed are green. For these calculations, local alongshore and across-shore directions are found from the major and minor principal axes, respectively, for each individual observed or modeled current location.

activity in the region near the N line south of Pt. Arena. Therefore, large CC between C and N from the model implies that model may not resolve the eddy activity there well.

# **3.2.** Comparison With Observed Upwelling Relaxation Events

[12] The generation of northward currents near the coast as southward upwelling favorable winds abate or cease is one of the major dynamical phenomena observed in the CODE region during the upwelling season. During CODE-2 1982, there were a total of five relaxation events [Send et al., 1987]. Two of them occurred in our study period in April and May. Figure 10 shows the observed and modeled current vectors at 10 m depth from the mooring locations, as well as the corresponding modeled surface temperature field for the days before and during these two upwelling relaxation events. During upwelling (April 19 and May 3), the southward coastal jet is interrupted by the coastline topography and is directed offshore at Pt. Arena. Observations show that the largest southward velocities at the mooring locations between Pt. Arena and Pt. Reyes are found on the 130 m isobath. A similar qualitative feature is found in the model results except on April 19 at the R line, where relatively large southward velocities in the jet core exist at 60 m water depth. Stronger upwelling is found south of the capes as indicated by colder water at the surface in those locations. During the abatement of southward winds (April 22-24 and May 5-7), the currents near the coast become northward, particularly at the stations south of Pt. Arena at

 $N_2$ ,  $N_3$  and  $C_2$ ,  $C_3$ . Combined with a strong southward jet offshore, a cyclonic circulation is formed near the coast in that region. At the R line, northward currents are found mainly at  $R_2$  and shoreward of  $R_2$ . The northward advection of warmer water from south of Pt. Reyes by northward currents during the second event and the general increase in surface temperature during wind relaxation (e.g., from May 5 to May 7) are clearly shown. More details concerning the thermal response will be presented in section 5.2.

[13] A circulation pattern similar to that found in the model and in the observations around Pt. Arena may be expected to occur at Pt. Reyes. Figure 11 shows the response of the depth-averaged velocity and the surface elevation around Pt. Reyes during the upwelling relaxation events. Similar to the flow at Pt. Arena, the coastal jet is directed offshore as it approaches Pt. Reyes. The cyclonic circulation formed south of the cape is a result of the jet separation. Lower pressure, as indicated in the surface elevation field, is seen near the coast south of the cape. As pointed out by Gan and Allen [2002], lower pressure at the capes results from the geostrophic balance of the alongshore velocity v and the increase in magnitude of vas the southward coastal jet approaches the cape, which reduces the pressure near the coast. In addition, the increase in magnitude of v at the cape is sufficiently large to change the local balance in the across-shelf momentum equation from geostrophic to gradient wind and this leads to a further reduction in pressure. With higher pressure farther south and north of the cape, this process forms negative/positive



**Figure 10.** (left) Modeled and (right) observed near surface (10 m depth) velocity vectors (m s<sup>-1</sup>) from the CODE mooring locations during upwelling (April 19, May 3) and during relaxation (April 22 and 24, May 5 and 7). The modeled surface temperature (°C, color contours) and the wind stress vectors  $\tau$  for each day are also shown.

pressure gradients south/north of Pt. Reyes, respectively. During relaxation, northward currents develop and the cyclonic eddy is strengthened south of the cape. Some of the currents near the coast are able to flow northward past Pt. Reyes. The northward currents are driven by a northward ageostrophic pressure gradient force south of the cape. Details of the time evolution of the dynamical balances during upwelling relaxation will be given in the next section. As shown in Figure 11, a stronger response to southward wind relaxation is found for the upwelling relaxation event in May, with larger northward currents south of the cape. This is probably associated with the stronger deceleration rate of southward winds during this event and hence with a larger northward ageostrophic pressure gradient force during this time.

[14] Dramatically different responses in the velocities and density fields occur to the north and to the south of Pt. Reyes during the upwelling relaxation events. We select two across-shore sections, one north (line 143) and one south (line 113) of Pt. Reyes (Figure 3) to illustrate the difference. Figure 12 shows daily averaged potential density  $\sigma_{\theta}$  and alongshore velocity v at these two sections during upwelling (April 19) and during relaxation (April 22). On April 19, stronger upwelling is found at section 113, south of Pt. Reyes. Surface values of  $\sigma_{\theta} = 25 \text{ kg m}^{-3}$  are located 30 km offshore, 15 km farther offshore than at 143. At the same time, the alongshore coastal jet core is also located farther offshore at 113 due to separation. The alongshore currents are stronger at 143 during upwelling. During the southward wind relaxation on April 22, the jet separation at 113 is intensified, the jet core drifts farther offshore, and northward currents are found over the inner shelf. At 143, northward currents are found only within a few kilometers of the coast. The potential density at these two sections reacts consistently during the relaxation. A stronger downwelling-like response in density is found at 113. These results further demonstrate the important effects of coastline topography on the variability of the upwelling and the upwelling relaxation flow fields.

#### 4. Mean Circulation and Temperature Fields

[15] Time mean surface velocity and surface temperature fields during the study period contain the averaged responses of upwelling, downwelling, and relaxation. The mean field, however, is dominated by upwelling features due to the mean upwelling favorable wind stress (Figure 2). Figure 13 shows the time mean surface velocity vectors together with the standard deviations of the vector amplitudes and also the time mean surface temperature field and corresponding standard deviations. These fields characterize the response and variability of the coastal ocean off northern



**Figure 11.** Depth-averaged velocity vectors (m s<sup>-1</sup>) and surface elevation (cm, color contours) in the vicinity of Pt. Reyes during upwelling (April 19, May 3) and during relaxation (April 22 and 24, May 5 and 7).

California in the CODE region during the upwelling season. The core of the mean coastal jet and the largest magnitude of the velocity *std* are at the same location, near the 200 m isobath, indicating strong variability in the jet. The offshore position of the jet core varies along the coast in response to the coastline and to the wind stress variations. During the relaxation of southward winds, northward currents develop near the coast, the southward jet moves offshore and separation near the capes intensifies (Figure 12). On the other hand, the jet tends to follow the coastline during weak upwelling. The variations of the jet core position and of the local magnitude in v, therefore, are determined by both the temporal variability in the wind and the dynamic response of the jet to the coastline and the shelf topography.

[16] Similar to the upwelling relaxation events presented in the previous section, the mean surface temperature field also shows generally colder water south of coastal capes compared to that directly off the capes. Higher surface temperature variability is found in the region between Pt. Arena and south of Pt. Reyes, reflecting more effective upwelling and upwelling relaxation due to the existence of coastal capes in that region [*Gan and Allen*, 2002].

[17] To help obtain a three-dimensional picture of the mean fields and their variability, we plot in Figure 14 the time mean alongshore velocity and density at sections 113, 143 and C (Figure 3) where different dynamical and thermal responses are found. In the v field, the mean jet core

gradually shifts offshore southward from section C to section 113. The relatively steep shelf at C allows the jet to penetrate into deeper water. The velocity of the southward jet is weaker, of greater horizontal extent, and farther offshore over the wider shelf topography at 113 and 143. More effective upwelling can be seen in the density section at 113 as indicated by the larger values of  $\sigma_{\theta}$  at the surface. Similar to the results shown in Figure 13, higher std values of v and density are found in the core of the jet and near the coast, respectively. The sections of  $q^2$ , twice the turbulent kinetic energy, also display distinct features. At 113, the turbulence near the bottom is stronger around the shelf break. Weaker v over the inner shelf reduces the strength of the turbulence in the bottom boundary layers there. Stronger v near the bottom at 143 leads to a larger bottom  $q^2$  values compared to C. The steeper shelf and smaller v at the bottom at C generate a weaker bottom boundary layer. The surface mean turbulent layer is about 30 m deep at 143 and C. Over the shelf at 113, it is shallower. Evidently, stronger variability in the density due to upwelling over the shelf at 113 reduces the turbulence in the surface layer.

[18] The field of time mean  $q^2$  at the bottom and the associated time mean bottom velocity vectors and standard deviations of the vector amplitudes are shown in Figure 15. The values of mean  $q^2$  at the bottom are related to the bottom stress which is determined by the bottom velocity [*Blumberg and Mellor*, 1987]. The fields in Figure 15 reflect this relationship and show the regions of the shelf where the bottom frictional processes are largest. The relatively large bottom velocities found in the vicinity of Pt. Arena and Pt. Reyes produce stronger turbulence there. The larger bottom velocities near these capes is consistent with the behavior of the shelf currents discussed previously.

#### 5. Momentum and Thermal Balances

[19] To help identify the dynamical processes that determine the shelf flow during upwelling and upwelling relaxation, we examine term balances in the momentum equations and in the potential temperature equation.

#### 5.1. Momentum Balances

[20] The alongshore depth-integrated and the alongshore depth-dependent momentum equations from the model are written below as equations (1) and (2), respectively. The across-shore depth-dependent momentum balance is recorded below in equation (3).

$$\frac{\partial \overline{V_b D}}{\partial t}^{1} + \frac{\partial \overline{U_b V_b D}}{\partial x} + \frac{\partial \overline{V_b^2 D}}{\partial y} - F_{by} - \overline{G_y}^{2} + \overline{fU_b D}^{3} - \overline{\tau_{ys}}^{4} + \overline{\tau_{yb}}^{5} + \overline{gD} \frac{\partial \overline{\eta}}{\partial y} + \frac{gD}{\rho_0} \int_{-1}^{0} \int_{\sigma}^{0} \left( D \frac{\partial \rho'}{\partial y} - \frac{\partial D}{\partial y} \sigma' \frac{\partial \rho'}{\partial \sigma} \right) d\sigma' d\sigma^{6} = 0, \quad (1)$$

$$\frac{\partial vD}{\partial t}^{1} + \frac{\partial uvD}{\partial x} + \frac{\partial v^{2}D}{\partial y} + \frac{\partial v\omega}{\partial \sigma} - F_{y}^{2} + fuD^{3} - \frac{\partial}{\partial \sigma} \left(\frac{K_{M}}{D} \frac{\partial v}{\partial \sigma}\right)^{4} + gD\frac{\partial \eta}{\partial y} + \frac{gD^{2}}{\rho_{0}} \int_{\sigma}^{0} \left(\frac{\partial \rho'}{\partial y} - \frac{\sigma'}{D} \frac{\partial D}{\partial y} \frac{\partial \rho'}{\partial \sigma'}\right) d\sigma'^{5} = 0, \quad (2)$$

**5** - 10



**Figure 12.** Across-shore sections of daily averaged alongshore velocity  $v \text{ (m s}^{-1})$  and potential density  $\sigma_{\theta}$  (kg m<sup>-3</sup>) at lines 113 and 143 during upwelling on April 19 and during relaxation on April 22. The contour intervals are 0.1 m s<sup>-1</sup> for v, with the 0.5 m s<sup>-1</sup> and 0 contours bold, and 0.2 kg m<sup>-3</sup> for  $\sigma_{\theta}$ , with the 25 and 26 kg m<sup>-3</sup> contours bold.

$$\frac{\partial uD}{\partial t}^{1} + \frac{\partial u^{2}D}{\partial x} + \frac{\partial uvD}{\partial y} + \frac{\partial u\omega}{\partial \sigma} - F_{x}^{2} - fvD^{3} - \frac{\partial}{\partial \sigma} \left(\frac{K_{M}}{D} \frac{\partial u}{\partial \sigma}\right)^{4} + gD\frac{\partial \eta}{\partial x} + \frac{gD^{2}}{\rho_{0}} \int_{\sigma}^{0} \left(\frac{\partial \rho'}{\partial x} - \frac{\sigma'}{D} \frac{\partial D}{\partial x} \frac{\partial \rho'}{\partial \sigma'}\right) d\sigma'^{5} = 0, \quad (3)$$

where  $(V_b, v)$  and  $(U_b, u)$  are the alongshore and acrossshore (depth-averaged, depth-dependent) velocity components, respectively,  $D = H + \eta$  is the water depth, H is the undisturbed water depth,  $\eta$  is the surface elevation,  $F_{by}$ ,  $F_y$ and  $F_x$  are the corresponding horizontal viscosity terms,  $G_y$ is the dispersion term [*Blumberg and Mellor*, 1987],  $\tau_{ys}$ and  $\tau_{yb}$  are alongshore components of the surface and bottom stress divided by  $\rho_0$ , respectively,  $\omega$  is a velocity normal to  $\sigma$  surfaces, and  $K_M$  is the vertical viscosity coefficient.

[21] The terms in equation (1) are referred to as 1 acceleration, 2 nonlinear advection, 3 Coriolis force, 4 wind stress, 5 bottom stress, and 6 pressure gradient  $P_y/\rho_0$ . In equations (2) and (3), terms 1–3 have the same designation while term 4 is vertical diffusion and term 5 is the pressure gradient ( $p_y$ ,  $p_x$ )/ $\rho_0$ . The numerical model equations are written in horizontal curvilinear coordinates. For simplicity in notation, we write the equations here in locally Cartesian form, but the variables are evaluated with respect to the curvilinear coordinates. In addition, before evaluation of terms in equations (1), (2), and (3), we divide by *H* so that, assuming *D* is approximately equal to *H*, equation (1) corresponds to the depth-averaged momentum equation and the terms in equations (2) and (3) likewise have units m s<sup>-2</sup>.



**Figure 13.** (a) Time mean surface velocity vectors (m s<sup>-1</sup>) and standard deviation of the vector amplitudes (m s<sup>-1</sup>, color contours), (b) time mean surface temperature (°C), and (c) standard deviations of the surface temperature (°C).

[22] Time series of daily averaged terms in the depthaveraged alongshore momentum equation (equation (1) divided by H) and the corresponding velocity vectors from locations about 5.5 km offshore near C2 and near R2 are shown in Figure 16. During both upwelling relaxation events on April 19-24 and May 3-7, the magnitude of the southward currents initially increase in response to upwelling favorable winds, followed by a decrease in magnitude and eventual reversal of current direction to northward as southward winds relax. At the C line during these events, the positive southward wind stress is initially balanced mainly by negative southward acceleration. A negative pressure gradient develops along with an increase in the negative nonlinear advection term. After the wind stress decreases, the negative pressure gradient is essentially the only forcing mechanism that is able to accelerate northward currents. At the R line, the response is more complex, but the behavior of the acceleration and pressure gradient terms is similar to that found at the C line. Although at both locations the term balances exhibit considerable time variability with all of the terms important at different times, the presence of a negative pressure gradient appears to be a consistent factor in the transition between upwelling and relaxation.

[23] The temporal and alongshore spatial variability of terms in the depth-averaged alongshore momentum equation (equation (1) divided by H) and the corresponding depth-averaged velocity and ageostrophic pressure gradient (term 3 + term 6) at locations approximately 2.5 km from the coast are shown in Figure 17. During the period of downwelling winds around April 10, the alongshore variability of the terms in equation (1) is relatively weak. The dynamical balances are mainly between wind stress and the

bottom stress. In contrast, significant alongshore variability is found in all of the terms in equation (1) during upwelling. Enhanced time and spatial variability of the dynamical balances occur around the capes. Strong southward v is found north of capes, where positive pressure gradients are primarily balanced by the large negative nonlinear advection and bottom stress terms. South of capes, the pressure gradient is negative and is balanced by a positive Coriolis force, reflecting a net onshore flow in the water column. During the entire period, the ageostrophic pressure gradient is negative almost everywhere except north of capes. In particular, the presence of a negative ageostrophic pressure gradient during the relaxation supports the conclusion that the northward currents are driven by a northward pressure gradient force.

[24] The set up of the pressure gradient is a result of the interaction between the wind-forced flow and the coastal topography [Gan and Allen, 2002]. Figure 18 shows the time mean fields of terms in the depth-averaged alongshore momentum equation including the nonlinear advection (term 2), the sum of surface wind stress and bottom stress (term 4 + term 5) and the ageostrophic pressure gradient (term 3 + term 6). Dominant features are the following. Near the coast over most of region, except directly north of the capes, the sum of the wind stress and bottom stress is positive, i.e., the southward wind stress is dominant and is not balanced by the bottom stress. That positive net forcing term is balanced by both a negative, i.e., northward, ageostrophic pressure gradient and negative nonlinear advection. The latter evidently reflects the mean offshore transport near the surface of a southward alongshore momentum deficit on the shoreward side of the coastal jet. The balances are different just north of Pt. Reyes and Pt.



**Figure 14.** Across-shore sections of time mean values (black contour lines) of the alongshore velocity (m s<sup>-1</sup>), potential density  $\sigma_{\theta}$  (kg m<sup>-3</sup>) and twice the turbulent kinetic energy  $q^2$  (m<sup>2</sup> s<sup>-2</sup> 10<sup>-4</sup>) at lines 113, 143, and C. The corresponding standard deviations are shown by the color contours. The contour intervals are 0.05 m s<sup>-1</sup> for v, 0.2 kg m<sup>-3</sup> for  $\sigma_{\theta}$ , and 1 m<sup>2</sup> s<sup>-2</sup> 10<sup>-4</sup> for  $q^2$ .

Arena. In these regions, a positive ageostrophic pressure gradient balances large negative nonlinear advection, which presumably results from the spatial acceleration of the southward mean flow as it approaches the capes from the north, and a negative contribution from the relatively large bottom stress in this region (as indicated in Figure 15). The main point here is that, in the presence of coastal topography, both the ageostrophic pressure gradient and nonlinear advection terms play an important role in balancing the mean wind stress forcing.

[25] As we have seen, coastline topography is a major factor in inducing alongshore variability in the upwelling dynamics. The alongshore variability in the response of currents and density to upwelling winds can be examined further by comparing the time mean z-dependent dynamical term balance in the alongshore momentum equation (equation (2) divided by H) on across-shore sections at lines 113, 143 and C (Figure 19). At 113, a strong negative pressure gradient and corresponding geostrophically balanced onshore currents under the surface Ekman layer essentially hold over the entire shelf. In contrast, a positive pressure gradient and offshore currents spread over the shelf and slope at 143. A negative nonlinear advection term is relatively large at the surface near the coast at 143. It contributes to the balance of the local positive pressure gradient. At 113 and C, negative values of the nonlinear advection term exist near the surface next to the coast while positive values are found offshore. At 113, the offshore positive advection effects are large on the inshore side of the mean coastal jet (Figure 14) while at C they are large offshore of the jet core.

[26] The alongshore variability of the z-dependent time mean dynamical balance can be seen from Figure 20. The potential density and the terms in the z-dependent alongshore momentum equation are plotted as a function of depth z and alongshore coordinate y at a distance from coast of approximately 2.5 km. Consistent with the results in Figures 17-19, positive pressure gradients are found to the north of the coastal capes and negative pressure gradient are found to the south. These pressure gradients are mainly balanced to the north by a negative nonlinear advection term, and to the south by vertical diffusion in the surface layer and by the Coriolis force below the surface layer. Relatively large values of potential density are found below the surface layer in the regions south of capes and between Pt. Arena and Pt. Reyes, generally coincident with a positive Coriolis



**Figure 15.** (a) Time mean values of twice the turbulent kinetic energy  $q^2$  (m<sup>2</sup> s<sup>-2</sup> 10<sup>-4</sup>) at the bottom and (b) time mean bottom velocity vectors (m s<sup>-1</sup>) and standard deviation of the vector amplitudes (m s<sup>-1</sup>, color contours).

force corresponding to onshore velocities. The pressure gradient is basically depth-independent, which explains why the northward currents appear first near the bottom during relaxation (Figure 4). During upwelling, the alongshore velocity is stronger southward near the surface and is weaker or even northward near the bottom (Figure 12). Since the response to the relaxation is forced by a nearly depth-independent pressure gradient, northward currents will develop first close to the bottom near the coast where the southward currents are smaller.

[27] The correlation coefficients discussed previously and shown in Figure 9 between variables at different alongshore locations indicate, for both the observed and model variables, relatively short alongshore correlation scales for the across-shelf velocity component u and correspondingly larger scales for the alongshore velocity v. This characteristic of west coast shelf flow fields was first pointed out by Kundu and Allen [1976] from analysis of current measurement off Oregon and was found from the CODE observations by Dever [1997], but to our knowledge the associated dynamics have not been previously examined with shelf circulation models. In order to help understand the reasons behind these differing correlation scales, we examine similar alongshore spatially lagged correlations of respective terms in the alongshore and across-shore momentum balances. To make the calculation more meaningful, the local alongshore direction is determined by the major principal axes of the velocity components in the Coriolis force terms

in equations (2) and (3). Time series of daily averaged terms in the local across-shore (x) and alongshore (y) momentum equation from 35 m depth at C<sub>3</sub> are shown in Figure 21. As expected, in the x equation a geostrophic balance of the alongshore velocity and the pressure gradient is dominant. In the alongshore y equation, the Coriolis force, pressure gradient, nonlinear advection and acceleration terms are all appreciable, but a tendency for geostrophic balance of the across-shelf velocity at this middepth location is evident. It should be noticed, in addition, from the relatively small magnitude of the vertical diffusion term over most of the time period except around May 9, that this 35 m depth is generally below the model surface boundary layer.

[28] Correlation coefficients between terms in the x and y equation at C<sub>3</sub> with the corresponding terms at R<sub>3</sub> and N<sub>3</sub> are also shown in Figure 21. In the x equation, the relatively high correlation coefficients of the Coriolis force (-fv) and pressure gradient terms are consistent with the results in Figure 9. In the y equation, the correlation coefficients for the Coriolis force, pressure gradient, and nonlinear advection terms are low, while those for the acceleration term  $(v_t)$  are notably higher. Again, the low correlations of the Coriolis force eterm (fu) are consistent with the results in Figure 9.

[29] In connection with this analysis, it is useful to compare the correlation coefficients of the model velocity components (v, u), resolved in local principal axes at these locations, and the alongshore component of the wind stress  $\tau^{v}$  with the corresponding correlation coefficients calculated



**Figure 16.** Time series of terms in the depth-averaged alongshore momentum equation (m s<sup>-2</sup>, averaged over 24 hours, multiplied by 10<sup>6</sup>) from locations about 5.5 km offshore at the C (water depth 58 m) and the R (water depth 55 m) lines. Time series of corresponding daily averaged depth-averaged velocity vectors (m s<sup>-1</sup>) from the same locations are also shown. The velocity vectors are plotted such that the vertical axis in the figure is aligned with the local *y* coordinate direction.

for the observed velocity components (Table 2). In general, the model and observed  $(\tau^{v}, v)$  maximum magnitude correlations have similar values with small difference in lag. The  $(\tau^{v}, u)$  correlation are consistently smaller than those for  $(\tau^{v}, v)$  for both model and observed velocities, except for the observed  $(\tau^{v}, u)$  correlation at N<sub>3</sub>. The counterintuitive positive model  $(\tau^{v}, u)$  correlation at C<sub>3</sub> evidently reflects the tendency for positive onshore flow there during wind relaxation (Figure 21). Overall, the  $(\tau^{v}, v)$  correlations with model and observed velocities compare favorably. The  $(\tau^{v}, u)$ correlations show more disagreement, but these are generally smaller in magnitude and presumably affected by the short alongshore correlation scales for u.

[30] As a result of these calculations, we conclude that, because of the tendency for geostrophic balance of fu

(Figure 21), the low correlations of u are directly related to low correlations for the pressure gradient term  $p_y$ . We have seen many clear indications of alongshore pressure gradient  $p_y$  set up by the interaction of the wind-forced shelf currents with coastal topography on somewhat larger scales. Although the results of this analysis are not definitive, they suggest that small alongshore scale perturbations in the alongshore pressure gradient induced by the time-dependent interaction of the shelf flow with alongshore topographic variations are the cause of the short alongshore correlation scales for u.

#### 5.2. Thermal Balances

[31] The alongshore variability of the temperature field in response to upwelling and relaxation is investigated in this



**Figure 17.** Depth-averaged alongshore velocity v (m s<sup>-1</sup>), and terms in the depth-averaged alongshore momentum equation. *Ace* acceleration, *Adv* nonlinear advection, *Cor* Coriolis force, *Pre* pressure gradient, *Wind* surface stress, and *Bot* bottom stress (m s<sup>-2</sup>, averaged over 24 hours, multiplied by 10<sup>6</sup>) as a function of time and distance along the coast at locations approximately 2.5 km offshore. The corresponding ageostrophic pressure gradient (*Age* = *Pre* + *Cor*) is also shown.

section. Figure 22 shows the time- and depth-averaged terms in the potential temperature equation (Appendix B). In order to identify the relative contributions to the time rate of change of temperature from alongshore and from across-shore advection, the advection terms, which are written in conservation form in POM, are rewritten to remove the contribution of the continuity equation. To help in the interpretation, the net advective effects only are plotted as explained in Appendix B.

[32] Consistent with the mean temperature field (Figure 13), large negative time-integrated temperature changes  $\langle dT/dt \rangle$ , reflecting cooling, are found near the coast between Pt. Arena and south of Pt. Reyes. The major contribution to the cooling is from across-shelf advection  $\langle ADVX \rangle$ , particularly south of the capes. This agrees with the dynamical results discussed in the previous sections. On the other hand, appreciable contributions to cooling from alongshore advection  $\langle ADVX \rangle$  occur on the northern side of the capes

5 - 16



Figure 18. Time mean fields of terms from the depthaveraged alongshore momentum equation. Adv nonlinear advection, *Wind* + *Bot* sum of surface stress and bottom stress, and *Age* ageostrophic pressure gradient. Solid contours are positive values, and dashed contours are negative.

and in the region farther south of Pt. Reyes. The contribution from atmospheric heat flux  $\langle Q \rangle$  is strongest near the coast where colder upwelled water leads to a larger air-sea temperature difference and thus to larger surface heat flux.

[33] Time series of daily averaged terms in the depthaveraged temperature equation (Appendix B) at locations 4.5 km offshore on the 113 and 143 lines (Figure 23) show the contributions of each individual term to the change of temperature in response to the wind stress (also shown) during downwelling, upwelling, and relaxation events. During the downwelling event on April 10-12, dT/dt is positive, indicating a net increase of temperature in the water column at both lines 113 and 143. Larger negative values of dT/dt, corresponding to cooling, occur at both stations during upwelling. The terms that balance dT/dt, however, are different at the two lines. The warming due to downwelling is mainly from advection in the across-shore direction ADVX at 113 while both advection terms are important at 143. At 113 during upwelling, cooling is generally the result of across-shore advection ADVX plus a much smaller contribution from alongshore advection ADVY. At 143, however, advection from upstream ADVY, is the primary source of local cooling during upwelling. As the winds relax, both advection terms decrease and become negative at the two stations which leads to a positive dT/dt. Clearly, the processes controlling the temperature changes



**Figure 19.** Across-shore sections of time mean values of terms in the alongshore momentum equation (2) divided by the water depth at lines 113, 143, and C (in  $ms^{-2}$ , multiplied by 10<sup>6</sup>). *Pre* pressure gradient; *DIFF* vertical diffusion; *COR* Coriolis force; and *NL* nonlinear advection.



**Figure 20.** Time mean potential density (kg m<sup>-3</sup>) and time mean values of terms in the alongshore momentum equation (2) divided by water depth: pressure gradient, vertical diffusion, Coriolis force, and nonlinear advection (m s<sup>-2</sup>, multiplied by 10<sup>6</sup>) as a function of depth and distance along the coast at locations approximately 2.5 km offshore.

differ substantially in the different dynamical regimes north and south of Pt. Reyes.

### 6. Summary

[34] A high-resolution three-dimensional coastal ocean model has been used to successfully simulate the upwelling and upwelling relaxation events observed off northern California during the CODE experiment. Forced with observed wind stress and heat flux, model fields are found to compare favorably with the observations. The model appears to capture the overall characteristics of the observed upwelling relaxation events. Reasonably good quantitative agreement is found between the modeled and observed alongshore currents. The results of statistical model-data comparisons indicate that mean and standard deviations of modeled alongshore currents agree better with those from the measurements on the inner shelf at 50 m water depth than farther



**Figure 21.** (top) Time series of terms (m s<sup>-2</sup>, averaged over 24 hours, multiplied by 10<sup>6</sup>) in the acrossshore (*x*) and alongshore (*y*) momentum equations at 35 m depth at mooring C<sub>3</sub>. (bottom) Crosscorrelation coefficients between corresponding terms in the across-shore (*x*) and alongshore (*y*) momentum equations at 35 m depth at moorings R<sub>3</sub>, C<sub>3</sub>, and N<sub>3</sub>. For these calculations, the local alongshore direction is defined by the major principal axes of the velocity component in the Coriolis force terms at each mooring.

offshore. Inclusion of more accurate spatially variable atmospheric forcing in future studies may quantitatively improve model-data comparisons.

[35] The modeling analysis is directed at examining the spatial and temporal variability of the upwelling and upwelling relaxation circulation and the corresponding dynamical mechanisms governing the circulation processes. Weakening of upwelling favorable southward winds not only reduces the upwelling strength, but also leads to the generation of northward currents near the coast. The northward currents that develop during relaxation are the consequence of the pressure gradients set up by interaction of wind-induced coastal currents with alongshore variations in coastal topography. Coastal capes, in particular, play a significant role in inducing alongshore variability in local dynamical balances as demonstrated through an analysis of terms in both the depth-integrated and depth-dependent alongshore momentum equations. The stronger upwelling and the northward currents during relaxation south of the capes are associated with local negative pressure gradients that result from the interaction of the wind-forced flow and the topography of coastal capes. An examination of the balance of terms in the equation for potential temperature shows that the alongshore variations of surface temperature and of upwelling are primarily generated by the alongshore variations of across-shore temperature advection. The alongshore advection of temperature plays an important role north of capes. The effectiveness of the across-shore temperature advection appears to be directly related to the setup of the

**Table 2.** Maximum Magnitude of the Correlation Coefficients Between the Alongshore Component of the Wind Stress  $\tau^{\nu}$  (Figure 2) and the Observed and Model Velocity (v, u) Resolved in Local (Major, Minor) Principal Axes at 35 m Depth From the Indicated Locations<sup>a</sup>

|                | v at 3   | 35 m    | <i>u</i> at 35 m |          |  |
|----------------|----------|---------|------------------|----------|--|
| Location       | Observed | Modeled | Observed         | Modeled  |  |
| N <sub>3</sub> | 0.58(0)  | 0.64(2) | -0.70(0)         | -0.37(0) |  |
| $C_3$          | 0.59(0)  | 0.68(1) | -0.15(1)         | 0.25(1)  |  |
| R <sub>3</sub> | 0.71(1)  | 0.71(1) | -0.40(2)         | -0.35(1) |  |

 $^{a}$  The lag in days for the maximum magnitude correlation coefficient is given in parentheses. Positive lags imply that the wind stress leads. The water depth (m) at the mooring locations is 90 m.



**Figure 22.** Fields of time mean values of terms (C s<sup>-1</sup>, multiplied by  $0.289 \times 10^{-6}$ ) in the depth-averaged temperature equation (Appendix B).  $\langle ADVY \rangle$  alongshore advection (equation (B16)),  $\langle ADVX \rangle$  across-shore advection (equation (B17)),  $\langle dT/dt \rangle$  (equation (B18)), and  $\langle Q \rangle$  surface heat flux (equation (B19)).

negative alongshore pressure gradients that allow geostrophic balance of the onshore flow below the surface layer.

# **Appendix A: Heat Flux Formulation**

[36] Using observed solar radiation and air temperature, the surface heat flux is obtained by bulk aerodynamic formulae:

$$Q_1 = Q_o(1 - \alpha) - L_a - H_a - LE_a, \tag{A1}$$

where  $Q_1$  is surface heat flux,  $Q_o$  is the measured incoming shortwave radiation at the surface,  $L_a$  is long wave radiation and  $\alpha$  (=0.1) is the albedo of the sea surface. L is the latent heat of vaporization (2.5 × 10<sup>6</sup> J kg<sup>-1</sup>).  $H_a$  and  $LE_a$  are the sensible and latent heat, respectively, and are calculated from

$$H_a = \rho_a C_{p,air} C_a |W| (T_1 - T_a), \tag{A2}$$

$$E_a = \rho_a C_L |W| (e_{sat}(T_1) - re_{sat}(T_a)) \left(\frac{0.622}{p_a}\right), \qquad (A3)$$

where  $\rho_a$  is air density (1.3 kg m<sup>-3</sup>).  $C_{p,air} = 1004 \text{ Jkg}^{-1}$  K<sup>-1</sup> is the specific heat of air,  $C_a$  (=10<sup>-3</sup>) and  $C_L$  (=10<sup>-3</sup>) are the turbulent exchange coefficients,  $p_a$  is the sea level pressure, |W| is the wind magnitude,  $T_a(C)$  is air temperature,  $T_1$  is sea surface temperature from model and  $e_{sat}$  is the saturation vapor pressure which is computed by the empirical formula given by *Bolton* [1980]:

$$e_{sat}(T) = 6.112 exp[17.67T/(T+243.5)],$$
 (A4)

The values of  $Q_o$  and  $T_a$  are interpolated from measurements [*Beardsley et al.*, 1998]. Based on shipboard measurements, we assume that the relative humidity requals a constant value of 0.85 [*Rosenfeld*, 1988].  $L_a$  is long wave radiation, which is calculated following the Berliand formula [*Budyko*, 1974],

$$L_a = \epsilon \sigma T_1^4 \left( 0.39 - 0.05 e_a^{1/2} \left( 1 - 0.71 C^2 \right) \right) + 4 \epsilon \sigma T_a^3 (T_1 - T_a),$$
(A5)

where  $\varepsilon$  is the emissivity of the ocean (0.97),  $\sigma$  is the Stefan–Boltzmann constant. *C* is the cloud fraction approximated as 0.5.  $e_a$  is the atmospheric vapor pressure, which can be defined as  $e_a = re_{sat}(T)$ .

#### Appendix B: Potential Temperature Equation

[37] The depth-integrated equation for potential temperature is

$$\int_{-1}^{0} \frac{\partial \Theta D}{\partial t} d\sigma + \int_{-1}^{0} \frac{\partial \Theta u D}{\partial x} d\sigma + \int_{-1}^{0} \frac{\partial \Theta v D}{\partial y} d\sigma + \int_{-1}^{0} \frac{\partial \Theta \omega}{\partial \sigma} d\sigma$$
$$- \int_{-1}^{0} \frac{\partial}{\partial \sigma} \left( \frac{K_{H}}{D} \frac{\partial \Theta}{\partial \sigma} \right) d\sigma - \int_{-1}^{0} F_{\Theta} d\sigma - \int_{-1}^{0} \frac{\partial R}{\partial \sigma} d\sigma = 0, \quad (B1)$$

where  $\Theta$  is potential temperature,  $D = H + \eta$  is the water depth, H is the undisturbed water depth,  $\eta$  is the surface elevation,  $K_H$  is the vertical diffusivity coefficient,  $F_{\Theta}$ represents horizontal diffusion, R is short wave radiation flux, and the velocity components are defined in section 5.1. The nonlinear advection terms are written in conservation (or divergence) form,

$$DADV = \int_{-1}^{0} \frac{\partial \Theta uD}{\partial x} d\sigma + \int_{-1}^{0} \frac{\partial \Theta vD}{\partial y} d\sigma + \int_{-1}^{0} \frac{\partial \Theta \omega}{\partial \sigma} d\sigma.$$
(B2)

[38] To evaluate the relative contribution of alongshore and across-shore temperature advection, it is necessary to remove terms in the continuity equation from (B2). Since, for daily time averages, the contribution of  $\eta_t$  in the



*Q*: Surface heat flux

**Figure 23.** Time series of terms (C s<sup>-1</sup>, multiplied by  $10^{-7}$ ) in the depth-averaged temperature equation (Appendix B) at about 4.5 km offshore at lines 113 and 143. *ADVY* alongshore advection (equation (B13)), *ADVX* across-shore advection (equation (B14)), dT/dt (equation (B10)), and Q surface heat flux (equation (B11)). Time series of the wind stress components as in Figure 2.

continuity equation is relatively small, that balance is well approximated by

$$\frac{\partial uD}{\partial x} + \frac{\partial vD}{\partial y} + \frac{\partial \omega}{\partial \sigma} = 0.$$
 (B3)

The depth-averaged form of equation (B3) is

$$\frac{\partial U_b D}{\partial x} + \frac{\partial V_b D}{\partial y} = 0. \tag{B4}$$

[39] We rewrite the individual advection terms in (B2) as

$$DADVY = \int_{-1}^{0} \left( \frac{\partial \Theta v D}{\partial y} - \Theta \frac{\partial V_b D}{\partial y} \right) d\sigma, \tag{B5}$$

$$DADVX = \int_{-1}^{0} \left( \frac{\partial \Theta uD}{\partial x} - \Theta \frac{\partial U_b D}{\partial x} \right) d\sigma.$$
 (B6)

As a result of the boundary condition  $\omega = 0$  at  $\sigma = 0, -1$ , the last term on the right-hand side of equation (B2) is zero so that with (B4),

$$DADV = DADVY + DADVX.$$
 (B7)

[40] In Figure 23 we plot time series of the terms in equation (B1) (converted to depth averages by dividing by H) at two locations. We start by defining

$$ADVY_I = H^{-1}DADVY, (B8)$$

$$ADVX_I = H^{-1}DADVX, (B9)$$

$$\frac{dT}{dt} = H^{-1} \int_{-1}^{0} \frac{\partial \Theta D}{\partial t} d\sigma, \qquad (B10)$$

$$Q = -H^{-1} \left( \int_{-1}^{0} \frac{\partial}{\partial \sigma} \left( \frac{K_{H}}{D} \frac{\partial \Theta}{\partial \sigma} \right) d\sigma + \int_{-1}^{0} \frac{\partial R}{\partial \sigma} d\sigma \right), \quad (B11)$$

where *DADVY* and *DADVX* are given by equations (B5) and (B6), respectively, and where we assume *H* approximately equals *D*. To help identify the net contributions of  $ADVY_I$  and  $ADVX_I$  to dT/dt, we further remove any common part of opposite sign that cancels in their sum,

$$ADV = ADVY_I + ADVX_I. \tag{B12}$$

For example, for ADV > 0, we define

$$ADVX = \frac{1}{2}(ADVX_{I} + |ADVX_{I}|) + \frac{1}{2}(ADVY_{I} - |ADVY_{I}|),$$
(B13)

$$ADVY = \frac{1}{2}(ADVY_I + |ADVY_I|) + \frac{1}{2}(ADVX_I - |ADVX_I|).$$
(B14)

As a result, the sum is preserved, i.e.,

$$ADV = ADVY + ADVX, \tag{B15}$$

and the common part of opposite sign that would cancel in the sum is removed. For ADV < 0, equations (B13) and (B14) are altered by reversing the signs of the terms with absolute values.

[41] In Figure 22 we plot the spatial distribution of the time-averaged values of the same terms, designated by an angle bracket,

$$\langle ADVY \rangle = t_f^{-1} \int_0^{t_f} ADVY \, dt,$$
 (B16)

$$\langle ADVX \rangle = t_f^{-1} \int_0^{t_f} ADVX \, dt,$$
 (B17)

$$\left\langle \frac{dT}{dt} \right\rangle = t_f^{-1} \int_0^{t_f} \frac{dT}{dt} dt,$$
 (B18)

$$\langle Q \rangle = t_f^{-1} \int_0^{t_f} Q dt, \qquad (B19)$$

where  $t_f$  is equal to the 40 day time period of the simulation.

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