Wind relaxation and a coastal buoyant plume north of Pt. Conception, CA: observations, simulations, and scalings

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Abstract. In upwelling regions, wind relaxations lead to poleward propagating warm water plumes that are important to coastal ecosystems. The coastal ocean response to wind relaxation around Pt. Conception, CA is simulated with a Regional Ocean Model (ROMS) forced by realistic surface and lateral boundary conditions including tidal processes. The model reproduces well the statistics of observed subtidal water column temperature and velocity at both outer- and inner-shelf mooring locations throughout the study. A poleward-propagating plume of Southern California Bight water that increases shelf water temperatures by \( \approx 5^\circ C \) is also reproduced. Modeled plume propagation speed, spatial scales, and flow structure are consistent with a theoretical scaling for coastal buoyant plumes with both surface-trapped and slope-controlled dynamics. Plume momentum balances are distinct between the offshore (> 30-m depth) region where the plume is surface-trapped, and onshore of the 30-m isobath (within 5 km from shore) where the plume water mass extends to the bottom and is slope-controlled. In the onshore region, bottom stress is important in the alongshore momentum equation and generates vertical vorticity that is an order of magnitude larger than the vorticity in the plume core. Numerical experiments neglecting tidal forcing show that modeled surface temperatures are biased 0.5\(^\circ\)C high, potentially affecting plume propagation distance and persistence.
1. Introduction

Point Conception, California marks an abrupt change in coastline orientation separating the Southern California Bight (SCB) and the Santa Maria Basin (SMB) (Figure 1). Many observational studies have captured the seasonal and interannual regional-scale ocean dynamics due to this topographic variability and wind-driven processes in the area [e.g., Brink and Muench, 1986; Harms and Winant, 1998; Auad et al., 1999; Winant et al., 2003; Hickey et al., 2003; Dever, 2004; Cudaback et al., 2005; Fewings et al., 2015]. Between June – August, the mean wind direction is equatorward (upwelling-favorable), with maximum wind speed at the western edge of Pt. Conception rapidly decreasing in strength towards the center of the Santa Barbara Channel [Dorman and Winant, 2000]. During this period, the SMB continental shelf typically has recently upwelled cold surface water, adjacent to warmer waters within the SCB (Figure 1). These two disparate water masses mix in the Santa Barbara Channel (SBC) [Hendershott and Winant, 1996].

Event-scale (3 – 5 day) relaxations of the mean wind frequently occur in summer [Dorman and Winant, 2000]. The ocean response to these events is characterized by the poleward propagation of SBC water over the SMB shelf [e.g., Melton et al., 2009; Washburn et al., 2011]. Transport mechanisms are important as connectivity pathways across Pt. Conception for marine species such as giant kelp [Johansson et al., 2015]. Because relaxation events are episodic, observational studies have relied on moorings or snapshots of surface scales from satellite sea surface temperature and HF radar currents when available [Cudaback et al., 2005; Melton et al., 2009; Washburn et al., 2011]. Although differences in mean circulation have been identified between the outer (100 m water depth) and inner...
shelf (5 – 30 m water depth) [e.g., Winant et al., 2003; Cudaback et al., 2005], differences between the outer and inner shelf during event-scale wind relaxation are not well understood. Cross-shore-vertical ($x, z$) cross-sections and dynamical analysis during events are also lacking and difficult to observe without a permanent high spatial resolution in situ presence. Here, we analyze the response to wind relaxation in a realistic coastal ocean numerical model, expanding prior observational results on thermally-buoyant pumes in a geophysical setting.

Numerical models have been used to understand the regional-scale $O(100 \text{ km})$ processes. For example, the large-scale seasonal spatial gradients in sea surface temperature (SST) [e.g., Veneziani et al., 2009], wind stress, and sea level pressure [e.g., Wang, 1997; Oey et al., 2004; Dong and Oey, 2005; Hsu et al., 2007] have been successfully reproduced and their dynamics elucidated by numerical models. In these studies, the horizontal grid resolution $\Delta x = 5 \text{ km}$ is sufficient for regional scales but too coarse to explore shelf processes because the shelf is only $\approx 15 \text{ km}$ wide. Additional modeling effort has focused on processes within the SCB [e.g., Dong et al., 2009], and recently include higher-frequency processes such as tidal-band processes through model nesting with horizontal resolutions of $\Delta x = 250 \text{ m}$ [e.g., Buijsman et al., 2011; Romero et al., 2013] and up to $\Delta x = 15 \text{ m}$ to resolve surfzone processes [Kumar et al., 2015].

Aspects of the coastal ocean response to wind relaxation, including the role of local topographic and bathymetric variability in creating along-shore pressure gradients have been described with numerical models [Gan and Allen, 2002a, b]. However, due to periodic boundary conditions, these studies did not include the effect of water masses originating beyond the simulated grid nor large-scale pressure gradient forcing [e.g., Oke et al.,
2002; Oey et al., 2004; Pringle and Dever, 2009]. These considerations are particularly important around Pt. Conception, thus a multi-nested modeling approach will be used here.

Previously compiled observations of many regional wind relaxation events \( (n = 40) \) showed that the poleward-propagating ocean response on the SMB shelf is consistent with buoyant gravity current theory [Washburn et al., 2011]. Much of the theory for buoyant coastal current geometry and dynamics has emerged from scaling idealized laboratory and numerical experiments [\textit{e.g.}, Yankovsky and Chapman, 1997; Lentz and Helfrich, 2002; Pimenta et al., 2010]. These scalings provide estimates of propagation speed, geometry, and flow field characteristics as functions of coastal current transport, density contrasts \( (\Delta \rho) \), latitude, and bottom slope [Lentz and Helfrich, 2002]. Observations and models have focused on salinity-driven coastal currents with relatively large \( \Delta \rho \) between the current and ambient ocean waters. For example, \( \Delta \rho \) from the Chesapeake River current is 2–3 kg m\(^{-3}\) [Lentz et al., 2003]. Thermally-buoyant coastal currents have a much smaller \( \Delta \rho \) (\textit{e.g.}, \( \Delta \rho = 0.1–0.9 \text{ kg m}^{-3} \)), lack modeling attention on geophysical scales, and are less well-understood [Woodson et al., 2009; Washburn et al., 2011].

In this paper, a multi-nested non-data-assimilative model is configured for the SBC and SMB shelf and validated with available observations. The observations and numerical setup are described in Section 2. Section 3 presents a general model-data time series and statistical comparison. Section 4 examines the observed and modeled kinematic response to a wind reversal event. In Section 5, theoretical scalings used to interpret buoyant coastal currents are presented [Lentz and Helfrich, 2002], followed by methods used to characterize the modeled current, and a comparison between the model, observations,
and theory. The modeled coastal current is then used to provide a first approximation of the three-dimensional, time-varying structure of the SBC coastal current to expand on observational results (Section 6). These include the evolving spatial scales, a contrast between offshore and onshore dynamics near the nose of the coastal current, and the effects of tidal mixing on its structure. Results are summarized in Section 7.

2. Methods

2.1. Observations

Moored observations and model results are reported in a local coordinate system where $x$ is the cross-shore, onshore positive, with the coast located at $x = 0$. The $y$ coordinate is in the along-shelf direction, positive poleward. The mean sea surface is located at the vertical coordinate $z = 0$, positive upward with the ocean bottom at $z = -h$, where $h$ is the local water depth. Modeled and observed currents are rotated into their principal axes as defined by the depth-averaged subtidal flow at each location. Once rotated, currents along the major axis are considered to be in the along-shelf ($v$) direction, while currents along the minor axes are oriented in the cross-shore ($u$) direction.

A range of in situ temperature and velocity observations are used for model-data comparison. The focus is on subtidal-frequencies, thus observations and modeled temperature and velocity are low-pass filtered (retaining periods $> 33$ hours). Over the outer shelf (100 – 200 m water depth), four moorings deployed from the Santa Barbara Channel–Santa Maria Basin Coastal Circulation Study measured temperature ($z = -1, -5, -25, -45, -65$ and $-100$ m) and currents ($z = -5$ m) [Harms and Winant, 1998; Winant et al., 2003] (Figure 2). In this work, the northernmost outer-shelf mooring is referred to as the Santa Maria Basin (SMB) mooring. The naming convention for loca-
tions at the western (SMIN, SMOF) and eastern (ANMI) entrances of the Santa Barbara Channel follows previous studies.

Inner-shelf (15 m water depth) temperature and velocity observations are from the Partnership for Interdisciplinary Studies of Coastal Oceans (PISCO) moorings [e.g., Cudaback et al., 2005; Melton et al., 2009; Fewings et al., 2015; Aristizbal et al., 2016]. Temperature observations were collected at three depths ($z = -4, -10$ and $-15$ m) from ARG, PUR and SAL locations (Figure 2). Velocity measurements at PUR were available from an upward-looking RDI 600 kHz Acoustic Doppler Current Profiler (ADCP) with good data from $z = -3$ m to $z = -14$ m.

Supplementary observations include winds from NDBC buoy 46011 and sea level from NOAA Port San Luis and Santa Monica tide gauges (Figure 2). Additionally, satellite observations of sea surface temperature (SST) from the Advanced Very High Resolution Radiometer (AVHRR) satellite are used to provide the regional oceanographic setting and qualitative comparison to modeled results (Section 4.1).

### 2.2. Numerical Model

The model used is the Rutgers Regional Ocean Modeling System (ROMS), a three-dimensional, terrain-following, open source numerical model that solves the Reynolds-averaged Navier-Stokes equations with hydrostatic and Boussinesq approximations [Shchepetkin and McWilliams, 2005; Haidvogel et al., 2008; Shchepetkin and McWilliams, 2009; Warner et al., 2010]. The modeling system is configured to run a series of nested offline simulations whereby the largest relevant forcing scales are simulated on a coarse grid and variability is transmitted to smaller domains of higher horizontal resolution through open boundary conditions [e.g., Wilkin, 2006; Penven et al., 2006; Springer et al., 2009;...
Dong et al., 2009; Ganju et al., 2011; Romero et al., 2013; Kumar et al., 2015, 2016]. At subsequent levels of nesting, additional smaller-scale forcing mechanisms (i.e., tides) are then added.

Model bathymetry is from the NOAA NGDC database (1 arc second resolution - https://www.ngdc.noaa.gov/). Multiple nested grids are configured to simulate the SMB continental shelf (Figure 2). Both bathymetric and coastline resolution are increased at each level of nesting. The lowest level, the outermost grid (L0, Δx = 3 km, 556 × 541 grid cells), extends from the Baja peninsula to Washington State spanning 18° of latitude and longitude encompassing the eastern Pacific basin. Subsequent higher level child grids have increased horizontal resolution (L1, Δx = 1 km, 770 × 392 grid cells and L2, Δx = 600 m, 546 × 386 grid cells) until a final grid which resolves inner-shelf processes (L3, Δx = 200 m, 194 × 362 grid cells). Standard nesting techniques are used to ensure volume conservation and bathymetry at child grid boundaries match the parent grid [e.g., Mason et al., 2010].

The outermost simulation is run on L0, forced by daily-averaged, realistic atmospheric fields (see 2.2.2 below) and climatological lateral boundary conditions from the World Ocean Atlas. The model is similar to that described and evaluated in Veneziani et al. [2009], although with three times higher horizontal resolution in both zonal and meridional directions. The simulation has 42 vertical levels, and applies the Generic Length Scale (GLS) vertical mixing scheme with $k−\omega$ parameters to solve for vertical eddy viscosity and diffusivity [Warner et al., 2005]. Radiation boundary conditions are applied to barotropic fields to allow outgoing energy [Chapman, 1985; Flather, 1976; Mason et al., 2010], and both radiation and nudging are used for baroclinic boundary conditions [Marchesiello
Model boundary values are nudged to incoming information with a 1-day time scale ($\Delta \tau = 1$ day) and are weakly nudged to outgoing information ($\Delta \tau = 365$ days). A roughly 130 km wide sponge layer with horizontal eddy viscosity, that increases from the interior value of 1 m$^2$s$^{-1}$ to a boundary value of 200 m$^2$s$^{-1}$, is placed along the open boundaries. The simulation produces a California Current System with statistically reasonable mean and mesoscale variability [Veneziani et al., 2009]. Simulations used below follow 6 years of model spinup from climatological boundary and surface forcing conditions and a 1-year-long integration with realistic surface forcing.

To downscale this L0 simulation, sea level, tracer, and momentum fields are interpolated at each time step to the L1 grid boundaries (boundary conditions) and are interpolated to L1 interior grid points for the first time step of the nested simulation (initial conditions). The L1 simulation starts on May 1, 2000, with baroclinic velocity and tracer values clamped at the boundaries and no sponge layer. Subsequently, simulations on the L2 grid use results from L1 and simulations on L3 use results from L2 as boundary conditions.

The L2 and L3 simulations are integrated for a 60 day period covering June–July, 2000. All nested grids share the same 42 vertical levels and sigma-coordinate parameter settings as the parent simulation. To accommodate the addition of tides (see 2.2.1), in L2 and L3 tracer and incoming baroclinic velocities are strongly nudged ($\Delta \tau = 6$ hours on L2 and $\Delta \tau = 1$ hours on L3) and outgoing velocities are weakly nudged ($\Delta \tau = 365$ days on all grids). Similar to Kumar et al. [2015] in L2 and L3 a horizontal eddy viscosity of 0.1 m$^2$s$^{-1}$ is applied. Additional simulations on higher nests with grids capable of resolving surface gravity wave forcing were also conducted and will be described in future manuscripts.
2.2.1. Tides

Tidal forcing is not included in the outermost L0 or first nested L1 simulation. Harmonic sea level and barotropic velocities from eight astronomical tidal constituents (K\textsubscript{2}, S\textsubscript{2}, M\textsubscript{2}, N\textsubscript{2}, K\textsubscript{1}, P\textsubscript{1}, O\textsubscript{1}, Q\textsubscript{1}) and two overtides (M\textsubscript{6}, M\textsubscript{4}) are applied as boundary forcing on the L2 simulations from the ADCIRC tidal model [Mark et al., 2004]. Within the L2 simulation, the interaction of tidal forcing, variable bottom topography, and stratification can produce internal waves of tidal periodicity (internal tides) which propagate throughout the numerical domain and are transmitted to the higher-resolution domain L3 via the lateral boundary conditions as described above.

Barotropic tides are validated by comparison to observations of sea level from the Port San Luis tide gauge and depth-averaged velocity from PUR (Table 1). Harmonic analysis conducted with the T.TIDE package [Pawlowicz et al., 2002] shows that the four largest-amplitude tidal constituents of sea level and major-axis velocity compare well (Section 3 and Table 1). The phases of the diurnal velocity constituents and the amplitude of O\textsubscript{1} velocity are poorly modeled, potentially due to diurnal sea breeze effects not included in the model.

2.2.2. Model surface forcing

The Coupled Ocean-Atmosphere Mesoscale Prediction System (COAMPS) model was run in a quadruply-nested configuration over the northeast Pacific and western North America domain [Hodur et al., 2002; Doyle et al., 2009]. Output from the four grids are daily-averaged and combined to produce a ROMS forcing file on L0. The resolution of the atmospheric forcing in the SMB–Pt. Conception region is 9 km, which is interpolated to higher horizontal resolution for all ROMS child grid simulations. Although previous
regional studies suggest the importance of wind forcing with resolution higher than 9 km [e.g., Oey et al., 2004; Dong and Oey, 2005], our results show very good model-data agreement over the SMB shelf regions during the period of study (Section 3).

The direction, amplitude, and temporal variability of the COAMPS wind compares well ($r^2 > 0.87$) to the observed wind at NDBC buoy 46011 during the simulation period (Fig. 3a). Observed (modeled) principal axis winds are $-54^\circ (-52^\circ)$ clockwise from north and are used in Section 3 for correlation to observed and modeled ocean fields. Consistent with typical summer conditions in the area [e.g., Dorman and Winant, 2000; Winant et al., 2003; Melton et al., 2009], winds were predominantly upwelling-favorable with an extended wind reversal between 06/13 - 06/20 and a wind relaxation around 07/13.

### 2.2.3. Large-scale sea level variations

In both models and observations, regional sea level variations induce along-shelf pressure gradient forces that control poleward flows over the continental shelf from seasonal [e.g., Harms and Winant, 1998; Fewings et al., 2015; Connolly et al., 2014] to subtidal [e.g., Brink and Muench, 1986] time-scales. Along-shelf sea level variations are generated over scales ranging from the entire (1000 km) North American west coast [e.g., McCreary, 1981] to regional (10–100 km) coastline variability [e.g., Gan and Allen, 2002a; Oey et al., 2004]. For example, over 2000 km of the North American west coast, seasonal sea level differences $\Delta \eta \approx 0.10$ m were well reproduced in a year-long 9-km resolution model nested within a global model [Connolly et al., 2014]. On smaller spatial and time-scales, a nested simulation of the northern California shelf produced sea level differences $\Delta \eta \approx 0.03$ m over 150 km at 20 day time-scale [Pringle and Dever, 2009], although not compared to
observations. Accurate model simulation of sea-level differences at subtidal time-scales and a range of spatial-scales requires both large-scale and local dynamics to be resolved.

Here, the ability of the multi-nested shelf model to reproduce observed larger scale sea level differences $\Delta \eta$ is tested from the Santa Monica and Port San Luis tide gauges, separated by an along-shelf distance of $\approx$ 300 km (white triangles in Figure 2). Model sea level differences $\Delta \eta$ are estimated at modeled grid points corresponding to tide gauge locations (Figure 3b) where Santa Monica is within $L_2$ and Port San Luis is within $L_3$. Observed and modeled $\Delta \eta$ are generated by removing the time-mean (over the 60 day simulation) sea-level at each gauge, de-tiding with harmonic analysis, low-pass filtering at 33 h, and then differencing Port San Luis from Santa Monica. Observed $\Delta \eta$ includes local atmospheric pressure differences which are not included in the model. The observed and modeled $\Delta \eta$ compare well (Figure 3b) and are well-correlated ($r^2 = 0.62$) with similar variability ranging from 0.05 m to $-0.03$ m. This demonstrates that the multi-nested approach accurately generates sea-level differences over the larger SCB scale of 300 km. However, these $\Delta \eta$ should not specifically be interpreted as local pressure gradients surrounding individual headlands such as Pt. Conception (10 km).

3. Model-data comparisons

For model-data comparison, modeled $L_3$ temperature and velocity are extracted from the location of the outer- (SMB) and inner-shelf (ARG, PUR, SAL) moorings. Modeled $L_2$ results are extracted at SMIN, SMOF, and ANMI moorings located outside the $L_3$ domain. Unless otherwise stated, all correlations reported are significant at 95% confidence.
Observed and modeled temperature time series are compared and discussed in relation to NDBC 46011 daily-averaged principal axis wind forcing, oriented to the south-east. At the outer-shelf SMB mooring, modeled and observed near surface temperature time series are in good agreement and are correlated \( r^2 > 0.3 \) to the principal axis wind (Figure 4a). The most pronounced warming signal in both model and observations is a near-surface \( \approx 5^\circ \text{C} \) temperature increase during a reversal of the predominantly northwesterly wind (06/12 - 06/19). Modeled temperature is biased high \( \approx 2^\circ \text{C} \) due to a coastal summer-time bias in the COAMPS heat flux applied to all model nests [Veneziani et al., 2009]. At the mid-water column of the outer-shelf moorings \((z = -65 \text{ m})\), neither observed nor modeled temperature is strongly correlated to the principal axis wind, indicative of other forcing mechanisms of temperature at depth at these locations (Figure 4a, c). Although observed surface temperature is not as strongly correlated to the principal axis wind at the southern SMIN mooring, the pronounced warming occurs 2–3 days earlier as evidenced by the smaller lag to the wind (Figure 4c). Inner-shelf mooring time series also show good agreement and a similar sense of the southern mooring time series ARG (Figure 4d) leading the northern mooring SAL (Figure 4b) [Melton et al., 2009; Washburn et al., 2011]. In contrast to the outer-shelf moorings, water temperatures at all depths at all 3 inner-shelf moorings are correlated to the wind \( r^2 > 0.5 \) and experience the marked warming during wind reversal.

Observed major-axis (along-shelf) currents are well-reproduced by the model at both the outer-shelf SMB and inner-shelf PUR moorings (Figure 5a). Modeled flow is stronger than observations and more strongly correlated to the principal axis wind \( r^2 > 0.6 \) in model, \( r^2 > 0.3 \) in observations). The largest poleward flow signals \((> 0.2 \text{ m s}^{-1})\) are seen
around 06/14 in both observations and the modeled flow (Figure 5a). Although near-bottom along-shelf flows are mostly weak (< 0.005 m s\(^{-1}\)), during poleward flow events inner-shelf flows can be directed poleward throughout the water column (Figure 5b and Washburn et al. [2011]).

In the cross-shore direction, SMB currents are varied and only weakly correlated to the principal axis winds. The near-surface flow on the inner shelf is directed offshore over most of the study period and is opposite at the near-bottom, consistent with the wind forcing (surface wind correlations \(r^2 > 0.4\) in both model and observations). Near-surface cross-shore flows are onshore in the surface during the relaxation period. Observed correlations between the flow in this location and the wind are consistent with previously reported values [Fewings et al., 2015].

**Vertical structure of mean and subtidal standard deviations**

At both outer- and inner-shelf locations, the modeled and observed time-mean temperatures are similar except that model temperatures are biased \(\approx 1.5^\circ C\) high (Figure 6a, c). Both the observed and modeled mean vertical temperature gradient is about 0.06\(^\circ C\) m\(^{-1}\).

At SMB, low-frequency standard deviations are surface intensified, although modeled values are 0.5\(^\circ C\) smaller (Figure 6b). Subtidal temperature standard deviations compare well at PUR (Figure 6d) with a magnitude of 1.5\(^\circ C\). Comparisons at other mooring locations yield similar results (not shown), a summary of model-data temperature error statistics are given in Appendix A.

Modeled and observed velocity statistics at PUR compare favorably in structure but with differences in magnitude (Figure 6; e, f). In the direction of the major axis \((v)\), both observed and modeled mean flows are surface intensified and equatorward (< 0) in
the near-surface, although the modeled mean is \( \approx 0.05 \, \text{m} \, \text{s}^{-1} \) stronger. In observations, subtidal current standard deviations are larger than the mean [Fewings et al., 2015]. Modeled velocities similarly follow this trend (Figure 6f). Both modeled and observed major-axis variability is much larger than minor axis, though modeled variability is \( \approx 0.03 \, \text{m} \, \text{s}^{-1} \) larger than observed throughout the water column. Observed and modeled differences in major-axis (along-shore oriented), low-frequency dynamics are possibly due to unresolved processes in the lee of PUR. Specifically, the 200 m horizontal resolution bathymetry could be too smooth. Additionally, the COAMPS wind field under-resolves spatial variability of the wind in the lee of coastal headlands [Dong and Oey, 2005; Hsu et al., 2007].

4. Evolving response to wind relaxation

4.1. Regional scale: satellite-model comparison

After an increase in the strength of upwelling-favorable winds between 9 June – 11 June, winds begin to relax and completely reverse direction between the 14 and 18 June (Figure 3a). On regional scales, the modeled response to changing wind patterns is qualitatively similar to available satellite SST observations (Figure 7). Consistent with the in situ model bias (Section 3), 1.5 °C is subtracted from all modeled SST panels to facilitate the comparison (Figure 7b, d, f). Prior to the wind relaxation, warm water (> 14 °C) is found to the east of Pt. Conception with a well-developed cyclonic eddy in the center of the Santa Barbara Channel in both observations and the model (Figure 7a, b). As winds briefly increase (11 June), warm SBC water advances towards Pt. Conception and the coastal upwelling signature remains over the SMB shelf (Figure 7c, d). Entering the reversal period (14 June), SBC water rounds Pt. Conception and appears to split into two
branches: a portion that flows poleward onto the SMB shelf (discussed below) and one that spreads southward to eventually recirculate back into SBC. Further satellite imagery of the event was not available due to cloud contamination.

4.2. SMB shelf scale

During wind reversal, the observed [Washburn et al., 2011] and modeled coastal ocean response is the poleward-propagating flow of warm water as seen in modeled sea surface temperature (SST) and cross-shore temperature transect snapshots (Figure 8). Preceeding the reversal (13 June), modeled SST patterns are typical of upwelling conditions on the SMB shelf including colder water in the lee of headlands, and smaller-scale filaments stretching seaward from the coastline (Figure 8a, d). One day later (14 June), warmer water ($\approx 14^\circ C$) has begun to fill the near-surface of the SMB shelf potentially due to a combination of increased surface heating and the return of offshore waters in response to relaxing winds (Figure 8b). An even warmer water mass ($> 16^\circ C$) can be seen to the south of ARG, not yet reaching the PUR $x-z$ transect (Figure 8e). Two days later, the front has passed PUR, $17^\circ C$ water extends across the entire shelf and warm water also fills the upper 20-m of the water column (Figure 8c, f).

5. A coastally-trapped buoyant plume

5.1. Theory

An idealized buoyant coastal current attached to a sloping shelf in a fixed reference frame [Lentz and Helfrich, 2002; Washburn et al., 2011] is illustrated in Figure 9. Away from the source and bounded by a coast, a buoyant plume of fluid with density $\rho$ discharging into a rotating fluid of higher density $\rho + \Delta \rho$ takes the limiting form of either a surface-
trapped or a bottom-slope controlled coastal current [Yankovsky and Chapman, 1997]. In
the vertical wall limit (located at \(x = 0\)), the front (nose) of a purely surface-trapped
plume propagates in the direction of a Kelvin wave \((y > 0)\) at the internal wave speed,
\[
c_w = (g' h_p)^{1/2}. \tag{1}
\]
Here, \(g'\) is the plume reduced gravity \(g \Delta \rho / \rho\) and \(h_p\) is the depth where the plume intersects
the sloping shelf (Figure 9). This occurs at an offshore location \(x = x_p\), termed the 'foot'
of the plume. Combining volume conservation with the assumption that the equilibrium
along-shelf frontal velocity is geostrophically balanced, Yankovsky and Chapman [1997]
derive an expression for \(h_p\),
\[
h_p = \left( \frac{2 Q f}{g'} \right)^{1/2}. \tag{2}
\]
The additional parameters are the Coriolis frequency \(f\) and the plume volume transport
\(Q\) (m\(^3\) s\(^{-1}\)).

A bottom-slope controlled plume propagates at a speed
\[
c_\alpha = \theta g' / f, \tag{3}
\]
similar to a topographic wave where \(\theta\) is the bottom slope [Lentz and Helfrich, 2002].
These results are generalized to give the propagation speed of intermediate plumes,
\[
c_p \propto \frac{c_w}{1 + c_w / c_\alpha}. \tag{4}
\]
The ratio \(c_w / c_\alpha\) determines whether plumes are governed by primarily surface-trapped
\((c_w / c_\alpha \ll 1)\) or slope-controlled \((c_w / c_\alpha \gg 1)\) dynamics [Lentz and Helfrich, 2002]. Ther-
ma!ly buoyant plumes such as the relaxation flows at Pt. Conception are in an interme-
diate parameter regime \((c_w / c_\alpha \approx 1)\) [Washburn et al., 2011] because density differences
with the ambient fluid are much smaller than in river plumes. The generalized interme-
mediate plume has an offshore region that is surface-trapped, and an onshore region that is slope-controlled. Scalings for the width of these plume regions, the surface-trapped portion \( W_w \propto c_w/f \), and slope-attached portion \( W_\alpha \propto h_p/\theta \), are combined to provide a scale relation for the width of the entire plume \( W_p = W_w + W_\alpha \) [Lentz and Helfrich, 2002],

\[
W_p \propto \frac{c_w}{f} (1 + c_w/c_\alpha) .
\] (5)

### 5.2. Observed and modeled plume characteristics

#### 5.2.1. Plume arrival and characteristics

Buoyant plume arrival at alongcoast locations on the SMB shelf is identified by the rapid increase in depth-averaged temperature \( \bar{T} \) (Figure 10a, b). From observations, plume arrival at each mooring \( t_a(y) \) is determined from local maxima in the first time derivative of \( \bar{T} \) (see Washburn et al. [2011] for details). The \( \bar{T} \) increase propagates northward as it is first observed at ARG, followed by PUR and finally SAL (Figure 10a). The modeled front similarly propagates northward, however is not as sharp as observed (Figure 10b).

Seen from a time-latitude plot of \( \bar{T}^{(m)} \), \( t_a(y)^{(m)} \) from these modeled moorings roughly coincide with the arrival of the 14.5\(^\circ\)C isotherm. The arrival of this isotherm herein determines \( t_a(y)^{(m)} \) (Figure 10c). In addition to the poleward-propagating plume, \( \bar{T}^{(m)} \) is highly variable including cold water pockets that are advected past the virtual moorings on tidal timescales. The linear regression slope between latitude and time yields an average propagation speed \( c_p^{(m)} = 0.20 \pm 0.03 \text{ m s}^{-1} \) of the modeled plume (Figure 10c). This value is well within the range of propagation speeds noted in Washburn et al. [2011]. The confidence interval in modeled speed is due to differences in isotherm choice (13 - 16\(^\circ\)C).

Confidence intervals of the regression slope are smaller.
The change in water column temperature and stratification associated with the arrival of the plume are determined by

\[ \Delta X = \langle X \rangle_+ - \langle X \rangle_- . \]  

Here, \( X \) is either temperature \((T)\) or vertical thermal stratification \((\partial T/\partial z)\) and \( \langle . \rangle \) represents a temporal average applied over the 24 hours prior to (−) and after (+) plume arrival. For example, an average of the three inner-shelf moorings shows that the plume depth-averaged temperature increased \( \Delta \bar{T}^{(m)} = 2.4^\circ C \). However, the model average over ARG, PUR and SAL is smaller \( \Delta \bar{T}^{(m)} = 1.3^\circ C \), consistent with the less sharp modeled plume.

At the shallow inner-shelf mooring locations, plume arrival increased water temperature throughout the water column. This is not the case in deeper water as evidenced by an across shelf transect of temperature increase \( \Delta T^{(m)} \) (Figure 11a). In deeper water, the region of large temperature increase \( (\Delta T > 1.3^\circ C) \) is confined to the near surface \((<25\) m depth). At the outer-shelf SMB mooring (Latitude of 34.80°N), similar temperature increases associated with the plume are present in the observations and model where at depth temperature loggers \((z \leq -45\) m) do not register the increase.

In addition to temperature increases at \( z > -25\) m, the plume also alters the thermal stratification across the shelf (Figure 11b). A similar modeled and observed structure to the stratification variations can be seen. In the near-surface, the water column becomes less stratified with plume arrival, indicating a thermal plume that is well-mixed in this region. Below the well-mixed layer, the plume increases stratification across the shelf with a maximum increase in stratification at about 10 m water depth (Figure 11b).
\[ \Delta T = 1.3^\circ C \] contour encompasses the majority of the stratification increase and will subsequently be considered the bounding region of the plume (Figure 11b).

Modeled plume dimensions \( W_w, W_\alpha \) and \( h_p \) are extracted from transects of \( \Delta T \) at each latitude (e.g., Figure 11). At the surface, the offshore distance where plume temperature increase \( \Delta T = 1.3^\circ C \) denotes the total width of the plume \( W_p \). Similarly, the offshore distance where near-bottom \( \Delta T = 1.3^\circ C \) denotes the slope-controlled width of the plume \( W_\alpha \). The water depth at this location is the plume depth \( h_p \). The width of the surface-trapped portion of the plume is determined from (5), \( W_w = W_p - W_\alpha \).

### 5.2.2. Plume-following reference frame

Plume dimensions and characteristics are extracted in a poleward-propagating reference frame moving with the plume (Figure 12). To remove tidal fluctuations, plume characteristics are determined from a 24-hour average of temperature and surface poleward velocity after arrival \( \langle T_p \rangle = \langle T \rangle + \) and \( \langle v_{sp} \rangle = \langle v_s \rangle + \). At each latitude, the largest surface plume temperature \( T_{sp} \) is found offshore of the 50-m isobath (Figure 12a). As the plume propagates poleward, the width of the high \( T_{sp} \) region decreases and the \( T_{sp} \) values also decrease as evidenced by the termination of the 17\(^\circ\)C isotherm around SAL. Plume width \( W_p \) (marked by magenta dots in Figure 12a) also decreases northward. The identified offshore extent of the plume initially follows the 16\(^\circ\)C isotherm (until \( \approx 34.9^\circ \)N), where the plume region departs from the 16\(^\circ\)C isotherm farther north indicating the importance of other sources of thermal variability at these times relative to the plume (Figure 12a).

In contrast, bottom temperature \( f (\approx 14^\circ \)C) and cross-shore location (consistently found in water depths \( 20 \text{ m} \leq h_p \leq 30 \text{ m} \)) remain fairly constant following the plume (not shown). Cross-shore transects of plume temperature \( T_p \) show a decrease in the near-surface...
temperature as the plume propagates northward (Figure 12c, e). Near the surface, at the offshore edge of the modeled plume, the cross-shore momentum balance is approximately geostrophic as the maxima in northward velocity coincides with the strongest cross-shore surface temperature gradient (Figure 12b - e). A plume momentum balance is given in Section 6.2.1. Unlike idealized models and laboratory situations which have fairly simple velocity profile decay as the coast is approached [Lentz and Helfrich, 2002], over the realistic and complex coastline topography, the surface velocity profiles have a complicated cross-shore decay.

Observed plume properties of Washburn et al. [2011] are compared to modeled plume properties (Table 2). Observed (o) values reported for the change in depth-averaged temperature $\Delta T$, and the thermally induced reduced gravity $g' = \alpha g \Delta T$, where $\alpha$ is the coefficient of thermal expansion of seawater, are estimated for this specific event. Observed $c_o$ is calculated using (3) with a bottom slope $\theta = 7.2 \times 10^{-3}$, representative of the SMB shelf. The range of observed phase speed $c_p$ is taken from the composite of all reported events and observed plume width is taken from a composite of 5 HF radar observed events [Washburn et al., 2011]. Volume transport $Q$ and plume depth $h_p$ are not directly observed and instead are estimated from the scaling equations (1 - 4). Methods to estimate modeled plume properties (m), except for $Q$ are described in Section 5.2. Modeled $Q$ is given by the integrated northward flow of plume water,

$$Q(t_a)^{\text{(m)}} = \int_{-L}^{0} \int_{-h}^{0} v_p dz dx; v_p = \begin{cases} v(x, z, t_a), & \Delta T \geq 1.3^\circ C \\ 0, & \Delta T < 1.3^\circ C \end{cases}$$

where $x = -L$ is the offshore boundary of the L3 domain (Figure 2). Modeled plume transport is temporally averaged over the event duration (12 – 18 June).
The propagation speed for the modeled front is similar to observed and to the theoretical phase speed of a buoyant coastal plume with both surface-trapped and slope-controlled dynamics (Table 2). As noted earlier, modeled temperature increases $\Delta T^{(m)}$ and $g'^{(m)}$ are less than the observed values, which are also reflected in the reduced phase speed estimates of the surface-trapped $c_w$ and slope-controlled $c_\alpha$ plumes relative to those observed. Modeled plume depth $h_p^{(m)}$ is also shallower than inferred $h_p^{(o)}$. Differences between modeled and inferred plume transports and depth indicate that direct observations are needed. Although individual scaled propagation speed values differ between the observations and the model, the relevant parameter governing plume dynamics, the $c_w/c_\alpha$ ratio is slightly above 1 (Table 2). This is consistent with a buoyant plume that has both surface-trapped and slope-controlled characteristics and suggests that the degree to which buoyant plume dynamics are controlled by one or the other is similar in both the observations and the model.

6. Discussion
6.1. Plume narrowing

As the modeled plume propagates poleward over the SMB shelf, it undergoes substantial evolution including a decrease in plume reduced gravity $g'$ (Figure 13a) and a decrease in plume width (“narrowing”). These behaviors have also been noted in a many–event composite of plume observations [Melton et al., 2009; Washburn et al., 2011]. Two possible mechanisms for plume narrowing are explored for the single modeled event; a predicted response due to evolving plume reduced gravity $g'$, and the plume response to downwelling-favorable wind forcing [Lentz and Largier, 2006; Moffat and Lentz, 2012; Mazzini et al., 2014].
From estimates of $c_w^{(m)}$ and $c_{\alpha}^{(m)}$, the plume width decrease can be predicted by the scaling equation (5). The surface-trapped width $W_w \propto \sqrt{g' h_p / f}$ [Yankovsky and Chapman, 1997; Lentz and Helfrich, 2002] is a function of both $g'$ and $h_p$, thus a buoyant plume with slowly decreasing $g'$ is expected to narrow barring changes in $h_p$. The applicability of this scaling to the modeled plume is tested by finding a scaled plume width $W_{sc}$,

$$W_{sc} = M \left[ \frac{c_w^{(m)}}{f} \left( 1 + \frac{c_w^{(m)}}{c_{\alpha}^{(m)}} \right) \right].$$

(8)

where M is a constant solved for by linear regression. The regression between $W_p^{(n)}$ and $W_{sc}$ predicts 40% of the along-shelf $W_p^{(m)}$ variance, with a best-fit $M = 2.4$ (Figure 13b) consistent with the expected O(1) scaling [Lentz and Helfrich, 2002]. Good agreement with the scaling indicates that the quasi-steady assumptions of the theory are valid for modeled plume width evolution during its poleward propagation.

During the reversal event, downwelling-favorable wind is fairly weak ($\approx 2$ m s$^{-1}$) but persistent (Figure 3a). Whitney and Garvine [2005] developed a wind strength index to assess the relative importance of wind forcing to buoyancy, through a ratio of velocity scales,

$$S_{wg} = \frac{v_{wind}}{v_p}.$$  

(9)

Here, $v_{wind}$ is an estimate of the surface current driven by the principal axis wind speed $V$, approximated by equating the quadratic surface and bottom along-shelf stresses $v_{wind} = 2.65 \times 10^{-2} V$ [Whitney and Garvine, 2005]. The buoyancy velocity scale is estimated as the time-varying along-shelf velocity associated with the plume $v_p$. Throughout the period of poleward-flow $S_{wg} \approx 1$, suggesting that the weak downwelling-favorable winds cannot be ruled out as a driver of decreasing plume width (Figure 13c) [e.g., Lentz and Largier, 2006; Moffat and Lentz, 2012].
6.2. Surface-trapped and slope-controlled regions

Laboratory studies suggest the importance of bottom stress in the plume along-shelf momentum equation is confined to the onshore, slope-controlled region \( (h < h_p) \) [e.g., Lentz and Helfrich, 2002]. Direct observations of these dynamics in realistic environments are limited as most observations detail surface-trapped plume dynamics [e.g., Lentz et al., 2003]. Contrasting onshore and offshore dynamics are illuminated by the model results here.

6.2.1. Momentum balances

From observations, the near-nose region of the Chesapeake coastal plume had a geostrophic cross-shore momentum balance and included nonlinearity and pressure gradients in the alongshore balance with minimal contribution from bottom stress in 8 m water depth, resulting in a surface-trapped plume [Lentz et al., 2003]. The observed and modeled SBC plume has features consistent with both surface and bottom-slope controlled plumes and thus the momentum balance should be distinct in the surface-trapped \( (x < x_p) \) and bottom-slope controlled \( (x > x_p) \) portions of the plume. Terms in the depth-averaged momentum balance are rotated into the local along- and cross-shore directions as determined from the principal axes of the depth-averaged flow. The root-mean-square magnitudes of the different terms are calculated from the 24-hr period surrounding plume arrival \( (t_a) \).

Depth-averaged momentum balances are first examined in 50 m water depth, near the plume core, where isobaths are mostly oriented north-south with little curvature (Figure 14a, c). The dominant cross-shore momentum balance is geostrophic, with little contribution from the other terms in the momentum equation (Figure 14a). The alongshore balance has contributions from the nonlinear advective terms and local acceleration,
although these are both smaller than the pressure gradient and Coriolis terms, and all
terms are a factor of three smaller than those in the cross-shelf balance (Figure 14c).
These are consistent with the momentum balance expected for surface-trapped plumes
[Lentz et al., 2003; Lentz and Helfrich, 2002].

At 15 m water depth, where isobaths follow the curved coastline around the capes
(Figure 2), although the cross-shore momentum balance is still primarily geostrophic,
there are non-negligible contributions from the nonlinear and local acceleration terms
(Figure 14b). In the along-shelf momentum equation, the pressure gradient, nonlinear,
and advective acceleration terms are the largest contributors indicating the importance of
non-steady advection (Figure 14d). Root-mean-square of the bottom stress term is about
one-third the magnitude of pressure gradient, nonlinear, and advective acceleration, but
still important when compared to the other momentum equation terms and consistent
with a bottom-slope controlled plume [Lentz and Helfrich, 2002].

6.2.2. Shallow-water vorticity

A flow characteristic easily calculated from numerical model output, although difficult
to observe, is plume vorticity. The barotropic plume vertical vorticity $\zeta_p$,

$$\zeta_p = \partial_x \overline{v}_p - \partial_y \overline{u}_p$$

(10)
is calculated from the depth-averaged horizontal velocities associated with plume arrival
($\overline{u}_p, \overline{v}_p$). The barotropic plume vorticity has a strong signal near the shoreline which is
$\approx 5$ times stronger than the time-averaged vorticity (Figure 15). Following the plume,
the onshore vorticity signal is similar to that of a transient jet with anticyclonic ($\zeta_p < 0$)
vorticity on the shoreward side and cyclonic ($\zeta_p < 0$) vorticity seaward (Figure 15). Nor-
malized by the Coriolis frequency and interpreted as a Rossby number, the dimensionless
vorticity is $O(1)$, indicating a submesoscale flow feature \cite[e.g.,][]{Capet2008}. The high vorticity flow follows the coastline topography and extends $\approx 5$ km from the coast, on the order of the internal Rossby deformation radius ($\sqrt{g\bar{h}_p/f}$). Offshore, near the 50-m isobath, a much smaller $O(0.1)$ nondimensional plume vorticity is visible due to horizontal shear in the predominantly geostrophic plume core (Figure 15).

Although vorticity sources widely vary and can include the advection of vorticity and vortex stretching due to flow separation, a portion of this onshore vorticity is a consequence of a region with negligible bottom stress (the surface-trapped plume portion, $x < x_p$) adjacent to a region where bottom stress is important (the slope-controlled portion, $x > x_p$). Where coastal flows encounter spatially variable bottom stress, transient vertical vorticity can be generated by two additional mechanisms: slope torque, which generates vorticity due a greater depth-distributed bottom stress in shallower water, and speed torque due to the quadratic formulation of bottom stress, whereby stronger flows experience stronger bottom stress \cite[e.g.,][]{Signell1991}. Over the sloping SMB shelf, the root-mean-squared bottom stress associated with the buoyant plume has significant cross-shore structure ranging from a maximum at about $4 \times 10^{-6}$ ms$^{-2}$ in 8 m water depth decreasing rapidly to $< 0.5 \times 10^{-6}$ ms$^{-2}$ over a distance of approximately 6 km (not shown). Along with vorticity advection and vortex stretching, both bottom stress mechanisms appear to contribute to generating the onshore vorticity structure.

6.3. The effect of excluding tides

To isolate the effect of tides on the modeled plume, an additional simulation (denoted NT) is downscaled from L1 to L3 without tidal boundary forcing. Although the larger-scale temperature and flow features are qualitatively similar in with tide (WT) and with
out tide (NT) simulations, there are significant differences. The NT simulation is 0.5°C warmer near the surface due to decreased mixing. Model mixing is diagnosed by comparing the time-, depth-, and cross-shelf (water depth < 100 m), averaged vertical eddy diffusivity $\langle K_T \rangle_{t,z,x}$ between the WT and NT cases. The NT $\langle K_T \rangle_{t,z,x} = 10^{-3} \text{ m}^2 \text{s}^{-1}$, about 1.5 times smaller than WT. Idealized two-dimensional ($x,z$) models have shown that the interaction of tidal and subtidal wind-driven processes increases mixing coefficients in ROMS simulations due to both barotropic [Castelao et al., 2010] and baroclinic tides [Kurapov et al., 2010].

Neglecting tidal forcing also affects some characteristics of the poleward-propagating buoyant plume. NT plume surface temperature is warmer because of less vertical mixing within the L3 domain, as well as a less mixed water mass incoming from the southern boundary (mixing is also reduced within the L2 simulation). The persistence of the plume, as diagnosed by the rate of reduction in plume-induced depth-averaged temperature increase $\Delta T$ as it propagates northward, also differs between the two simulations. With tides, the rate of decrease in $\Delta T$ was about 0.5°C per day, almost twice the reduction in $\Delta T$ for the NT simulation. This indicates that without tides, a simulated thermal plume potentially propagates farther poleward along the coast and/or persists on the shelf for longer duration.

7. Summary

A nested realistic Regional Ocean Model (ROMS) is configured to simulate the coastal ocean response to wind relaxation around Pt. Conception, CA. The model reproduces reasonably well the statistics of observed in-situ subtidal water column temperature and velocity at both outer- and inner-shelf mooring locations. The model also reproduces
kinematic properties including the \( \approx 5^\circ \text{C} \) temperature change, stratification variation, and flow structures associated with a poleward-propagating buoyant warm plume originating from the Santa Barbara Channel. The modeled plume provides a first approximation of how difficult-to-observe features, such as cross-shore-depth sections \((x, z)\), momentum balances, and vorticity can be manifested.

The modeled plume agrees with a theoretical scaling for intermediate buoyant plumes [Lentz and Helfrich, 2002], with distinct onshore and offshore dynamics. In the offshore (> 30 m depth) region, where the plume velocities are strongest and controlled by surface-trapped dynamics, the depth-averaged cross-shore momentum balance is geostrophic and bottom stress is unimportant. Within 5 km from shore, onshore of the 30-m isobath, the plume water mass extends to the bottom. Here, the dynamics are consistent with slope-controlled plumes and bottom stress is important in the alongshore momentum equation contributing to the generation of vertical vorticity. This vorticity is an order of magnitude larger than the vorticity due to horizontal shear in the geostrophic plume core.

Additional simulations neglecting tidal forcing show that modeled surface temperatures are biased 0.5\(^\circ\text{C}\) higher when excluding tidal mixing processes. This has important implications for the along-shore propagation distance and persistence of the modeled plume on the SMB shelf.

Appendix A: Model-data differences

Model-data differences in subtidal temperature are quantified by root-mean-squared error, partitioned into separate components [e.g., Oke et al., 2002; Wilkin, 2006; Liu et al., 2009]. The root-mean-squared error is \( \text{RMSE} = \langle (m_i - o_i)^2 \rangle^{1/2} \) where \( m_i \) and \( o_i \) are the i-th modeled and observed subtidal temperature \( T(t) \) and angle brackets denote
a time-mean. RMSE is written as the sum of three terms such that:

\[
RMSE = \left[ \frac{(\langle m \rangle - \langle o \rangle)^2}{MBE^2} + \frac{(S_m - S_o)^2}{SDE^2} + \frac{2S_mS_o(1 - CC)}{CCE^2} \right]^{1/2}.
\]

Terms in (A1) are (1) a mean bias error, MBE, (2) standard deviation error, SDE, where \( S_m \) and \( S_o \) are the respective model and observed subtidal standard deviations, and (3) the cross-correlation error CCE, proportional to the correlation between observed and modeled temperature.

At all moorings except SMOF and the ANMI middle water column temperature loggers, MBE is the largest source of model-data misfit (Figure A1). Taking an average of all temperature errors, CCE is about 40% of MBE and SDE is the smallest term (20% of MBE). Small SDE indicate that subtidal temperature variability is well-reproduced by nested models which resolve processes across the outer and inner continental shelf [Kumar et al., 2015]. With improved surface flux and lateral boundary conditions (perhaps through data assimilation), MBE is potentially reduced.

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References


Figure 1. Sea surface temperature (SST) snapshot on 11 June 2000 observed by the Advanced Very High Resolution Radiometer (AVHRR). Regional locations are labeled; Santa Maria Basin (SMB), Southern California Bight (SCB), Santa Barbara Channel (SBC), and Pt. Conception. The dashed black–white contour is the 200 m isobath, approximately denoting the continental shelf.
Figure 2. Four simulation grids and observation locations: (a) Second level of nested grid (L2, see Section 2.2 for grid dimensions and resolution). Yellow circle marks NDBC buoy 46011. Lower-level grids are shown in inset (L0, and L1). Colorbar is the vertical coordinate $z$ in meters. (b) L3 grid. White triangles mark the Port San Luis (within L3) and Santa Monica tide gauges. In both panels, red circles are moored observation locations. Outer-shelf moorings are in 100-m depth (SMIN, SMOF, SMB), except for ANMI (200-m depth). From south to north, inner-shelf (15-m depth) moorings are Pt. Arguello (ARG), Pt. Purisma (PUR), and Pt. Sal (SAL).
Table 1. Observed and modeled tidal constituents. \(^a\)

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Sea level</th>
<th>Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>amp (m)</td>
<td>ph (°)</td>
</tr>
<tr>
<td>S2</td>
<td>0.10 (0.12)</td>
<td>168.3 (164.4)</td>
</tr>
<tr>
<td>M2</td>
<td>0.49 (0.47)</td>
<td>170.7 (167.6)</td>
</tr>
<tr>
<td>O1</td>
<td>0.21 (0.19)</td>
<td>186.2 (189.8)</td>
</tr>
<tr>
<td>K1</td>
<td>0.42 (0.43)</td>
<td>226.1 (221.5)</td>
</tr>
</tbody>
</table>

\(^a\) Comparison of observed to modeled (in parentheses) major tidal constituents for sea level and major-axis velocity at PUR (15-m water depth). Amplitude (amp) and phase (ph) are reported from harmonic analysis with ph = 0° marking the beginning of the time series.
Figure 3. Time-series of observed (black) and modeled (red) (a) daily-averaged vector winds from NDBC buoy 46011 and COAMPS modeled (offset by 0.5 day), and (b) Santa Monica to Port San Luis subtidal sea level differences $\Delta \eta$. NDBC buoy 46011 and tide gauge locations are noted with yellow circle and white triangles, respectively in Figure 2a. The period of propagating poleward flow analyzed in Section 4 is shaded.
Figure 4. Model-data comparison of low-pass filtered temperature at a) SMB, b) SAL, c) SMIN, and d) ARG moorings. Observations (black) and model (red) results are shown at the surface-most (thick lines, $z = -1$ m in panels a and c, $z = -4$ m in panel b and d), mid-water column (thin lines, $z = -65$ m in panels a and c) and near-bottom (thin lines, $z = -15$ m in panels b and d) measurement locations. Squared correlation coefficients ($r^2$) and the lag which maximizes the correlation between near surface temperature to principal axis winds are noted in each panel. The period of propagating poleward flow analyzed in Section 4 is shaded.
Figure 5. Model-data comparison of low-pass filtered velocity at SMB and PUR moorings. Velocities are oriented in the direction of their principal axes such that along-shelf poleward is $+v$ and cross-shelf shoreward is $+u$. (a) Near surface along-shelf velocity $v_s$. (b) Near bottom along-shelf velocity $v_b$. (c) Near surface cross-shelf velocity $u_s$. (d) Near bottom cross-shelf velocity $u_b$. Period of propagating poleward flow analyzed in Section 4 is shaded.
Figure 6. Observed (black) and modeled (red) profiles of subtidal temperature and velocity statistics. Time mean (left column) and standard deviation (right column) of temperature at (a, b) SMB and (c, d) PUR, and velocity at (e, f) PUR. Velocities are in the direction of the major principal axis, \( v \) (solid lines) and the minor axis, \( u \) (dashed).
Figure 7. Regional SST snapshots around wind reversal period. SST observed by AVHRR (a, c, e) and corresponding modeled SST (b, d, f) from nearest hour to observations. In the satellite imagery, there has been no attempt at either cloud contamination masking nor temperature calibration. Portions of panels c and e were shown in [Melton et al., 2009]. Mean bias (1.5 °C) has been removed from modeled results.
Figure 8. Three snapshots of modeled (a - c) SST and (d - f) cross-shore and vertical temperature transects at latitude of Pt. Purisma mooring (34.73°N, white dashed line in a - c). Bathymetry contours are in 50 m intervals, and the 15°C and 17°C isotherms are contoured black.
Figure 9. Schematic of idealized buoyant coastal current attached to a shelf with slope $\theta$ in a fixed reference frame [Lentz and Helfrich, 2002; Washburn et al., 2011]. The coastal current has density $\rho$, separated from ambient ocean water of density $\rho + \Delta \rho$. The portion of the current attached to the slope has width $W_\alpha$, at water depth $h_p$. The offshore (surface-trapped) portion of the current has width $W_w$, and the total plume width is $W_p = W_w + W_\alpha$. Currents are denoted by gray arrows, including a geostrophic along-shelf flow near the offshore edge, and an onshore flow at the leading edge of the coastal current. The leading edge propagates at speed $c_p$ (red arrow) oriented towards the $+y$ direction.
Figure 10.  (a) Observed $\bar{T}^{(o)}$ and (b) modeled $\bar{T}^{(m)}$ depth-averaged temperature from three inner-shelf mooring locations spanning $\approx 35$ km. Vertical dashed lines mark the warm front arrival as determined by Washburn et al. [2011] methods (see 5.2.1), applied separately to model and observations. (c) Modeled time-latitude contour plot of depth averaged temperature following the 15-m isobath. Locations of headland moorings are noted by horizontal dashed lines. The 14.5 degree isotherm is noted (black contour), with a linear regression between latitude and time (magenta dashed). Propagation speeds (white-black dashed) correspond to the range of Washburn et al. [2011] observations (0.04 - 0.46 m/s) and the approximate modeled plume propagation (0.20 m/s).
Figure 11. Cross shelf transect of (a) plume temperature increase ($\Delta T$) and (b) stratification increase ($\Delta dT/dz$) at $t_a$ for the latitude of the SMB mooring (34.80°N). In both panels, modeled results are contoured and observations from the SMB mooring are shown as filled circles at their deployed longitude and depth. The Black contour line denotes $\Delta T = 1.3^\circ\text{C}$, that is overlaid on both panels.
Figure 12. Plume characteristics in plume-following coordinates. (a) Plume surface temperature $T_{sp}$ versus latitude or time, and longitude with the 16 and 17 °C isotherms contoured black. Magenta dots delineate plume width as described in Section 5.2.2. Isobaths are contoured gray in 50-m increments. (b, d) Transects of plume surface northward velocity $v_{sp}$; (c, e) transects of plume temperature $T_p$ at the (b, c) northern and (d, e) southern transects denoted by black dashed lines in panel (a). In (b, d) yellow dot denotes maximum $v_{sp}$. Magenta dots denote the surface and bottom extent of the plume as described in Section 5.2.2. Color range used in (c, e) is the same as in (a). Vertical red dashed line denotes location of maximum surface cross-shore temperature gradient.
Table 2. Comparison of modeled (m) to observed (o) plume characteristics.  

<table>
<thead>
<tr>
<th>Parameter</th>
<th>(m s(^{-1}))</th>
<th>(°C)</th>
<th>(×10(^{-3}) m(^2) s(^{-1}))</th>
<th>(10(^5) m(^3) s(^{-1}))</th>
<th>(m s(^{-1}))</th>
<th>(m s(^{-1}))</th>
<th>(m s(^{-1}))</th>
<th>(km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(o)</td>
<td>0.04 - 0.46</td>
<td>2.4</td>
<td>4.8</td>
<td>0.4 - 2</td>
<td>0.41 ((h_p = 40) m)</td>
<td>0.36</td>
<td>1.14</td>
<td>8 - 16</td>
</tr>
<tr>
<td>(m)</td>
<td>0.17 - 0.23</td>
<td>1.3</td>
<td>2.6</td>
<td>0.2</td>
<td>0.23 ((h_p = 20) m)</td>
<td>0.20</td>
<td>1.08</td>
<td>10 - 23</td>
</tr>
</tbody>
</table>

\(\Delta T\) are described in Section 5.2. Equations to estimate \(c_w\) and \(c_\alpha\) are (1) and (3) respectively. \(W_p\) is plume width (5).
Figure 13. (a) Reduced gravity $g'$, (b) modeled plume width $W_p = W_w + W_\alpha$, (c) non-dimensional wind-strength parameter $S_{wg}$ (9) following Whitney and Garvine [2005], as functions of latitude or time. In (a), vertical dashed lines denote the latitudes of inner-shelf moorings ARG, PUR and SAL, and colored circles denote mooring-observed $g'$. In (b), $W_p$ (magenta dots) are extracted from modeled transects (Section 5.2.2), and compared to scaled width $W_{sc}$ (black x) from (5).
Figure 14. Root mean square (RMS) of depth-averaged (upper) cross-shore and (lower) along-shore momentum balance terms at the PUR latitude (34.73°N). In (a, c) outer-shelf (h = 50 m) terms in cyan. In (b, d) inner-shelf (h = 15 m) terms in red. The RMS momentum balance terms are calculated over a ±12 hour period surrounding plume arrival and include the local acceleration (ACC), pressure gradient (PRS), Coriolis acceleration (COR), nonlinear advective (NLN), bottom stress (BSR), and surface stress (SSR).
Figure 15. Depth-averaged plume vorticity $\zeta_p$ normalized by the Coriolis frequency $f$ as a function of latitude or time and longitude in plume-following coordinates as described in text and Figure 12. Gray shading is land.
Figure A1. Evaluation of subtidal-frequency model-data differences as in (Equation A1). Values are reported as the square of each term (units are in °C). From left to right, panels are (a) mean bias error (MBE), (b) standard deviation error (SDE), (c) model-data cross correlation error (CCE), and (d) root mean square error (RMSE). Triangles (circles) denote outer (inner) shelf mooring locations.