Coastal Atmospheric Circulation around a Cape and its Response to Wind-Driven Upwelling Studied Using a Coupled Ocean-Atmosphere Model

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February 1, 2010

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ABSTRACT

The study describes and analyzes coupled ocean-atmosphere regional model simulations of the coastal circulation in regions of orographically-intensified flow, during developing coastal upwelling. The domain resembled an eastern ocean boundary with a single cape protruding into the ocean in center of the coastline. The model simulated the formation of an expansion fan downwind of an idealized cape, featuring stronger wind stresses with high diurnal variability. Coastal upwelling develops during the simulation extending about 50 km offshore upwind of the cape, but widens twice as much on the lee side of the cape. Coldest sea surface temperatures (SST) are found near the coast at the southern (lee) side of the cape.

Orographic and diurnal modulations of the near-surface atmospheric flow on the lee side of the cape strongly affect the dependence of wind stress on underlying SST conditions. Strong orographic forcing weakens an average wind stress—SST coupling computed for the 100-km coastal zone, and indicates regions of high negative correlations next to the cape. Diurnal modulations in the expansion fan region, however, are found to enhance air-sea coupling, resulting in high wind stress –SST correlations.
1. Introduction

Coastal wind regimes along the west coast of North America are strongly influenced by terrain and sea-surface temperatures (SST) during the summer season. Terrain features are responsible for flow intensification downwind of major capes along the Oregon-California coastline. Compounding these wind variations are mesoscale boundary layer effects generated by air-sea interaction with coastal upwelling. These two phenomena are examined in this study using a coupled mesoscale ocean and atmosphere model (Perlin et al., 2007, Warner et al., 2008).

Atmospheric flow intensification downwind from coastal capes and points have been thoroughly studied using a combination of satellite and aircraft observations as well as numerical weather modeling and idealized mesoscale simulations (Beardsley et al., 1987; Winant et al., 1988; Burk and Thompson, 1996; Burk et al., 1999; Dorman et al., 2000; Koračin and Dorman, 2001; Edwards et al., 2001, 2002; Perlin, 2004, 2007). Flow intensification is typically explained by applying hydraulic theory in the form of a controlled flow expansion wave or expansion fan, within the trans- or supercritical flow in the marine atmospheric boundary layer. Alternative compression bulges are often found on the windward side of the coastal promontories (Beardsley et al., 1987; Winant et al., 1988; Rogerson, 1999; Samelson, 1992). The intensity of these mesoscale phenomena, their spatial extent, and effects on the coastal ocean circulation, however, show some uncertainty and variability in the near shore (~100 km), where coastal topography, the diurnal cycle, and underlying sea surface temperature field complicate analysis (Enriquez and Frihe, 1995; Dorman et al., 2000; Bielli et al., 2002; Haack et al., 2001, 2008).
Summer coastal conditions are characterized by persistent northerly winds over the Oregon-California coast, which cause divergence of the upper ocean Ekman layer and nearshore upwelling. Upwelling is often limited to distances of 10-50 km from shore in regions with relatively simple topography, but can be notably altered in strength and location around major coastal promontories (Strub et al., 1991; Barth and Smith, 1998; Barth et al., 2000; Perlin et al., 2004; Huyer et al., 2005, Castelao and Barth, 2007). Colder upwelled water tends to stabilize the overlying marine boundary layer, which in turn acts to reduce downward momentum flux and wind forcing of the coastal ocean (Vickers et al., 2001; Samelson, 2002; Bane, 2005; Skyllingstad et al., 2005, 2007, Perlin et al., 2007). Coupling between the upper ocean and low-level winds has been observed and reported on greater scales, and some attempts to quantify the relationship have been recently made (Chelton et al., 2001; Chelton, 2005; Chelton et al., 2007; Haack, 2008; Song et al., 2009, Jin et al., 2009). However, the extent to which this coupling is significant in regions of strong wind-topography interaction in the nearshore region and its implication for upwelling development are yet to be determined.

The objectives the present study are the following. First, to analyze an idealized coastal circulation and wind stress regime in presence of a coastal promontory and developing wind-driven upwelling, as simulated by the two-way coupled ocean-atmosphere model. Our second goal is to explore and determine the importance of the effects of air-sea coupling in the area surrounding the cape based on the relationships between wind stress, sea surface temperature, and their derivatives.
2. Model setup

In our coupled code, the Naval Research Laboratory (NRL) COAMPS™ mesoscale model (Hodur, 1997) is used as the atmospheric model and the Regional Ocean Modeling System (ROMS, Shchepetkin and McWilliams, 2005) is used as an ocean component. These two major components of the coupled code communicate and exchange data via the Model Coupling Toolkit (MCT, Larson et al., 2005; http://www-unix.mcs.anl.gov/mct). A description of the model coupling approach and simple testing results are presented in Warner et al., 2008.

COAMPS™ is a three-dimensional atmospheric prediction system based on the fully compressible form of the non-hydrostatic equations, solved using a time-splitting technique (Klemp and Wilhelmson, 1978). Surface fluxes are computed using a modified Louis (1982) parameterization, corrected using TOGA COARE data for water grid points as described by Wang et al. (2002). In our simulations, cloud and rain liquid water ratios are computed with explicit moist physics applied for clouds and precipitation. Sub-grid scale vertical mixing is treated following the Mellor and Yamada 2.5 level scheme (1974). The atmospheric model is run with a cloud microphysics package that parameterizes clouds and rain.

The ROMS ocean component is a free-surface, terrain-following hydrostatic model, based on primitive equations solved using a split-explicit time-stepping scheme, separating barotropic (fast) and baroclinic (slow) modes (Shchepetkin and McWilliams, 2005). For our study, vertical mixing is parameterized using Mellor-Yamada 2.5-level scheme. Horizontal harmonic mixing is chosen to be computed along vertical levels for momentum and two tracers (temperature and salinity). Both COAMPS and ROMS utilize an Arakawa C grid, where spatial horizontal momentum points \((u, v)\) are staggered relatively to mass points \(\rho\),
and all of the above vertically staggered relatively to vertical momentum component \((w)\). This facilitates the numerical algorithms for momentum and mass/tracer equations.

Horizontal model domain sizes of the control experiment are 310 x 410 points in latitudinal and longitudinal directions, respectively, and both ocean and atmosphere models use 3-km grid boxes (Fig. 1). The domain represents an Eastern ocean boundary with a straight coastline having a single cape. The cape is represented in the center of the domain by two successive linear coastal bends, angling about 26.6°, similar to the value used in Burk et al. (1999). The cape extends seaward for 90 km, and has an alongshore extent of about 350 km. Bathymetry for the simulations were derived from measured data by averaging values between 43°-46°N (region of predominately north-south oriented Oregon coastline with relatively simple bathymetry) and applying several linear fits for the coastal area as a function of offshore distance. The ocean depth in the model remains constant upon reaching 2835m. Along the NW Pacific coast, land topography varies from about 350 m (on average) along the central Oregon coast, to approximately 2000m for the coastal mountains in the northern California. In our model setup, land topography increases eastward with the rate of 25 m elevation per 1 km horizontal distance, and remains flat above 750 m.

Both the atmospheric and ocean models adopt terrain-following sigma coordinates. The atmospheric model has 47 vertical sigma-levels stretched from the ground up to 9300 m, with 15 levels below 200 m, in order to resolve the air-sea coupling effects. The ocean model has 40 vertical sigma-levels with vertical grid spacing concentrated near the surface and bottom. Horizontally homogeneous model initialization is applied based on vertical profiles of potential temperature and water vapor mixing ratio for the atmosphere, and temperature and salinity for the ocean (Fig. 2). The atmospheric marine boundary layer is not prescribed in the initial conditions, but develops in response to the diurnal forcing. Ocean stratification, however, is imposed at the beginning, because it is largely determined
by seasonal variability. Simulations start with the ocean at rest, whereas the wind is initially set to 15 m/s geostrophic northerly flow in the lowest 1.5 km, decreasing to 5 m/s above between 1.5 km and 2 km. A geostrophic pressure gradient is determined from this imposed flow and is used as a constant term in the momentum balance equations. To account for a typical summertime subsidence pattern in the Eastern Pacific region, incremental heating of the atmospheric profile is added at a rate of 1°C/day.

Choice of boundary conditions was given special attention due to the 14-day duration of simulations, which is an unusually long period for a regional atmospheric model to run in an idealized mode. Periodic north-south boundary conditions are employed in both the ocean and atmospheric models. Lateral boundary conditions for the ocean model consist of a coastal wall on the east, with radiation for momentum and tracers and gradient conditions for the free surface along the western boundary. In the atmospheric model, eastern and western lateral boundaries use radiation conditions that distinguish inflow and outflow points. At inflow points, all boundary variables are set to their initial values. At outflow points, normal velocity out of the domain, \( v_n \), is computed using the upstream differencing method by Miller and Thorpe (1981). The general approach of the radiation condition is to ensure that \( \frac{\partial v_n}{\partial t} + \hat{c} \cdot \left( \frac{\partial v_n}{\partial n} \right) = 0 \). Here, \( \hat{c} \) is a velocity that includes wave propagation and advection, \( \hat{c} = -(v_n + c*) \), where \( c* \) is the phase speed for fastest-moving gravity waves directed out of the domain. In the model, boundary values of \( v_n \) at the next time step are estimated from boundary values at time step \( t \), \( v_n^t \), and the value one grid point inside of the boundary at the same time step, \( v_{b-1}^t \), following:

\[
v_{b+1}^t = v_n^t - (\hat{c} \cdot \Delta t / \Delta x)(v_b^n - v_{b-1}^n),
\]

where \( \Delta t \) and \( \Delta x \) are model time step and grid spacing in the direction normal to the boundary, respectively. The modified gravity wave speed \( \hat{c} \) is either set to constant, or is
estimated at time step $t$ by solving Eq.(1) with respect to $\hat{e}$ and applying velocities at one grid point and two grid points from the boundary at the preceding time step, $v_{h-1}^{t-1}$ and $v_{h-2}^{t-1}$, respectively. All other variables, except velocity normal to the boundary, are linearly extrapolated at outflow points.

Time steps in the two models are 10 s in the atmospheric model and 300 s in the ocean model with data exchange occurring on the ocean model time step. In the data exchange, the ocean receives momentum flux, net heat flux, and solar radiation from the atmospheric model averaged over the data exchange time period. In return, the atmospheric model receives SST computed by the ocean model at each time step. Simulations are conducted for 336 hours (14 days of forecast). The model setup described above is referred to as the “control case” throughout the text.
3. Coastal ocean wind stresses and lower atmospheric regime

a. Temporal mean quantities

Modeled 10-m wind speed and wind stress values averaged over days 7-14 of the control simulation are shown in Figure 3 (a-b). Upwelling is well established at this time in response to the previous 7 days of wind forcing. Mean wind stress vectors are oriented south-southeastward, stemming from the northerly geostrophic flow rotated within the atmospheric Ekman layer. Over most of the domain in the offshore region, 50-100 km away from the western boundary and the coastline, 10-m winds are stronger than 12 m/s, and wind stresses are above 0.15 N m$^{-2}$. A broad wind stress maximum is produced west-southwestward from the point of the cape, extending 300-400 km offshore, as traced by the 0.2 N m$^{-2}$ and 13 m/s contours. Within this region, the strongest mean winds and wind stresses are found about 100 km off the coast on the southern side of the cape. Upwind and downwind of the cape, wind stresses and winds immediately adjacent to the coast are notably lower, up to 3-4 times, than their offshore values. As discussed below, the reduced winds in these areas are directly tied to ocean upwelling that develops during the simulation. Additional wind deceleration gradually occurs north of the cape, starting 200-300 km upwind of the cape and extending 150-200 km offshore; this deceleration is likely caused by topographic obstruction of the flow by the cape. The broad area of stronger wind stresses on the lee side of the cape roughly outlines an expansion fan region, and the deceleration on the windward side of the cape could be associated with a compression bulge.

Persistent north-northwesterly winds over the domain force coastal upwelling that transports colder water to the ocean surface in the nearshore region (Fig. 4a). Initial SSTs are 14$^\circ$C and increase about 1-2$^\circ$C because of solar radiative heating, whereas nearshore upwelling decreases SSTs by as much as 5$^\circ$C from their initial value. Persistent wind forcing
produces a surface ocean coastal jet directed southward alongshore, with highest currents indicated over the mid-shelf (between 90 and 400m depth). A separate manuscript focusing on an analysis and discussion of the coastal ocean circulation from this case study is currently in preparation.

Some aspects of the spatial patterns of mean sensible plus latent surface heat flux (positive upward), averaged over days 7-14 (Fig. 4b) are related to the upwelling-modified SST for the same period. The offshore extent of the negative heat flux approximately corresponds to the region with mean SSTs below 15°C, and this area is broader on the lee side of the cape. Total heat flux is below -60 W m⁻² inshore and next to the tip of the cape. Over most of the offshore ocean, latent heat flux is slightly negative and sensible heat flux is positive about 20-30 W m⁻² (not shown), whereas both fluxes become negative (into the ocean) in the upwelling region. Negative surface heat fluxes result above the cold upwelling waters, and act to stabilize the atmosphere in the near shore. Both average sensible and latent heat fluxes are positive and much stronger over the land.

Atmospheric boundary layer (ABL) heights for the same period vary notably across the domain, especially downwind of the cape (Fig. 4c). ABL height is a model-computed quantity; its computation is based on the flux Richardson number, the ratio of buoyant production of turbulent kinetic energy (TKE) to shear production of TKE. The boundary layer depth is determined as the lowest level where this ratio exceeds a critical value of 0.5. ABL heights over the ocean in our simulation could be referred to as the marine atmospheric boundary layer (MABL). Average MABL heights for the period are about 500-550 m over the open ocean and away from the western boundary. Inflow boundary conditions of the atmospheric model that maintain the initial atmospheric conditions are responsible for the lower MABL heights at the western edge of the domain, while their reduction near the coastline results from dynamic adjustment.
Generally, a gradual reduction of the MABL height to below 300 m occurs over the ocean in a strip adjacent to the coastline. Around the tip and on the lee side of the cape, the MABL decreases rapidly onshore, dropping below 100 m near the coastline, with an almost two-fold decrease in boundary layer heights occurring over short alongshore or cross-shore distances. This rapid thinning is associated with a hydraulic supercritical expansion fan response to the coastal bend, and is accompanied by flow acceleration as shown in Fig. 3. Low boundary layer heights are also found along the straight portion of the coastline, approximately corresponding to the offshore extent of the upwelling zone (compare to Fig. 4a), and where the wind stresses falls below 0.1 N m$^{-2}$. Spatial correlation of the mean SST and modeled PBL heights (Fig. 4a,c) within 300 km off the coast is 0.89. Some of this correlation arises because the strongest winds, and thus the strongest local upwelling response, are also found in the expansion fan where the PBL is shallow. Air -sea boundary layer coupling, as discussed below, also accounts for a portion of the high correlation. Correlation between the temporal time series of 24-h running average of SST and modeled PBL height is 0.83, for points within 300 km of the coast over the same time period, and for a similar combination of effects.

Flow acceleration around the cape caused by topography is illustrated in cross-sections of the corresponding $v$- and $u$-wind components (Fig. 5a-b, cross-section locations marked in Fig. 4d), complemented by the atmospheric potential temperature, turbulent kinetic energy (TKE) and cloud water mixing ratio (Fig. 5c-e). An elevated northerly jet is located on the lee side of the cape atop of the MABL. The alongshore wind structure shows a correspondence between the jet and the depth of the MABL: the jet strengthens as the boundary layer decreases (Fig. 5a). For example, in alongshore section A-B, the maximum $v$ component velocity is found near the low point in the MABL around $y=450$ km. Similar characteristics are seen in the offshore cross-section plot, except the nearshore 20 km, where
rapid shoreward decrease in MABL heights is associated with stabilizing effects of the upwelled water (see also the potential temperature plot). The strongest $u$-wind component values (Fig 5b) are found near the surface within the MABL, with a secondary increase at higher elevation approximately between 1000-2000 m, except in the vicinity of the cape. The origin of the elevated values above MABL becomes clearer after analyzing the thermal structure (Fig. 5c).

The potential temperature cross-section indicates a relatively deep signature that the coastal cape imposes on the coastal atmosphere: a doming of isentropes north of the cape with depressed isentropes in the expansion fan region. It also reveals the subsidence inversion starting just below 2000 m, which outlines the height of the planetary boundary layer (PBL) above the MABL. Westerly MABL flow at all levels weakens at the windward side of the cape ($y=610$ km in the alongshore plot, and $x=0$ in the cross-shore plot), and becomes easterly flow aloft near the tip of the cape. A minima in the potential temperature is evident in the offshore cross-section, with temperatures below 286 K over the cool upwelled water (Fig. 5d, right panel) suggesting the development of an internal boundary layer.

Highest parameterized turbulence kinetic energy values are found near the surface, and it appears that the TKE isoline of 0.1 m$^2$ s$^{-2}$ outlines reasonably well the MABL height over the ocean part of the domain. TKE values near the surface are higher on the lee side of the cape, despite lower MABL heights and lower temperatures, and are likely caused by strong vertical wind shear in that region (compare left panels of Fig.5a,b,d). Turbulence increase also occurs below the subsidence inversion, approximately in a layer with elevated $u$-winds. The cross-shore TKE plot indicates a drastic decrease in TKE within 20 km of the coast in the lower layers, which is associated with the stable MABL. TKE increases rapidly as atmospheric flow travels inland.
Elevated values of the westerly wind component over the ocean may originate from both wind veering of the imposed northerly geostrophic flow in the Ekman layer from local pressure gradients, and from the momentum flux divergence. Further analysis of the turbulence and momentum budget will be reported in a separate publication focusing on the atmospheric circulation.

Note the agreement of TKE aloft with cloud formation (Fig. 5e) below the inversion. Higher cloud water mixing ratios result on the windward side of the cape, with nearly cloud-free areas on the lee side. In the cross-shore direction, the average cloud layer is weaker on the lee side of the cape.

The atmospheric model uses a cloud microphysics package that parameterizes clouds and rain allowing for qualitative comparison with satellite observations. Cloud albedo, $A$, can be derived from the model using an optical path estimation method following Savic-Jovcic and Stevens (2008) (used in Skyllingstad and Edson, 2009),

$$A = \frac{\tau}{6.8 + \tau},$$

where

$$\tau = 0.19L^{5/6}N_c^{1/3},$$

is the optical depth, $N_c = 1 \cdot 10^8 m^{-3}$ is the cloud-droplet number concentration, $L$ is the liquid water path vertically integrated as

$$L = \int \rho q_c dz,$$

$q_c$ is the cloud water mixing ratio, and $\rho$ is the air density. The cloud albedo for the final simulation time and its 7-day average are shown in Fig. 6a,b. Cloud formation is highly dynamic and transient, varying in both space (Fig. 6a) and time, with significant diurnal variations. Nevertheless, the time-average albedo (Fig. 6c) shows that the cape circulation develops a cloud-free region south of the cape. Cloud formation is stronger on the windward side of the cape and in general on the windward slope of the coastal topography. This behavior is consistent with satellite observed cloud features typically seen along the northern California
coast, namely clearing on the lee side of capes, and cloud build-up on their windward side over the ocean and the coast (Fig. 6c).

b. **Diurnal variations**

Diurnal heating and cooling of the land causes a strong daily cycle of the surface heat fluxes and the PBL heights over land, in comparison with the notably smaller variation over the coastal ocean. The resulting contrast drives a significant diurnal cycle in wind and wind stress over the ocean south of the cape (Fig. 7). In the cross-shore direction, the strongest diurnal variation of the wind stress magnitude occurs beyond 40 km offshore (Fig. 7 top left), showing notably less inshore variation. SST is initially warm near the coast because of solar heating, with cooling from upwelling visible after about 3 days, as SST near the coast decreases. Temperatures in the upwelling zone decrease to about 9°C and the upwelled water expands offshore about 100 km by the end of the simulation. Note that the wind stress magnitude and diurnal variation decreases within 50 km of the coast during the last 7 days (second half) of the simulation in response to the cooler SSTs. Diurnal variations of SSTs reach about 0.5°C.

Diurnal variations in wind stress magnitudes along the E-F transect are found primarily along the southern part of the cape, where maximum values are 2.5-4 times greater than background stress values, and exceed 0.4 N m⁻². Diurnal periodicity can be found offshore and upstream of the cape as well, but the amplitudes of wind stress variations are notably smaller, with values of about 0.20-0.25 N m⁻². Daily propagation of the wind stress maximum along the coastline is evident between y = 500 km and y = 370 km; the maximum is strongest is the middle of this interval.
The spatial extent of the diurnal cycle can be illustrated by hourly average wind stress magnitudes at four times during the day at 6 h intervals (Fig. 8), which will be further referred to as “night” (0200 LST), “morning” (0800 LST), “daytime” (1400 LST), and “evening” (2000 LST). In the morning, wind stresses are low along the coastline, and stronger 50-200 km offshore. Wind stress reduction within ~200 km off the west boundary occurs due to the boundary conditions. During daytime, nearshore wind stress is reduced upwind of the cape, whereas wind stress increases on the lee side of the cape in the expansion fan region. By evening, the expansion fan maximum strengthens and propagates downstream along the coastline generating a peak in wind stress in the nearshore 50-100 km at the southern edge of the cape. This maximum weakens during nighttime, while in the offshore 200-400 km, wind stresses remain relatively high during the nighttime. Peak wind stresses in the “evening time” are over four times their “morning” values.

4. Wind-stress and SST relationships and statistics

a. Control case analysis

Analyses of SST - stress coupling such as that of Chelton (2001) have focused on three primary quantities, the correlation of wind stress and SST, of wind stress curl and cross-wind SST gradient, and of wind-stress divergence and downwind SST gradient. The overlay of time-average fields of wind stress and SST (Fig. 9a) provides an initial visual estimation of the correspondence between these fields. The spatial patterns of SST and wind stress are similar in the areas north of tip of the cape, and also within about 150 km off the southern domain boundary. On the lee side of the cape and about 300 km further downstream along the straight coastline, SSTs are still primarily aligned with the coastline, with some colder temperatures in the nearshore, and the entire upwelling area is wider than on the
upwind side. Average wind stress, as mentioned earlier, features a broad region associated with the expansion fan in the lee of the cape, and rapid weakening downwind and onshore from the region of stress maxima. Spatial correlation of these time-averaged SST and wind stress fields is 0.36 for the entire domain.

Based on the major spatial features of the SST and wind stress fields described above, we sub-divide the domain into the following two regions. The region further referred to as the “expansion fan” region stretches from the cape tip to about 300 km downwind of the cape. The other region extends from the northern boundary to the tip of the cape, and also includes about 150 km off the domain southern boundary. Due to the N-S periodicity of the domain, this region will be further referred to as the “upwind region”.

Time averages of wind stress curl \( \mathbf{k} \cdot (\mathbf{V} \times \mathbf{\tau}) = \frac{\partial \tau}{\partial x} - \frac{\partial \tau}{\partial y} \) and wind stress divergence \( \nabla \cdot \mathbf{\tau} = \frac{\partial \tau}{\partial x} + \frac{\partial \tau}{\partial y} \) are shown in Figs. 9(b-c), overlaid by the crosswind SST gradient (CWSST) and downwind SST gradient (DWSST), correspondingly. The CWSST is defined as the quantity \( |\nabla \text{SST}| \cdot \sin \Delta \theta \), and DWSST as \( |\nabla \text{SST}| \cdot \cos \Delta \theta \), where \( \Delta \theta \) is angle between the wind stress vector and the SST gradient vector. The values of CWSST and DWSST were computed from hourly SST output and hourly-averaged wind stresses, and then averaged over the same time period as wind stress derivatives.

Average wind stress curl and divergence are close to zero far offshore, beyond ~150 km off the coast. Similarly, CWSST and DWSST are negligible in the offshore, due to the lack of spatial SST structure. The wind stress curl field shows a well-defined coastal band of elevated values in the upwind region, which results from shoreward wind stress decrease over the upwelled colder water. In this region, the CWSST contours are aligned in
accordance with the curl field. A separate offshore feature is found 100-150 km from the windward side of the cape, aligned parallel to the coast; it is apparent in both wind stress curl and CWSST fields, but not entirely co-located. Further downstream, a narrow band of higher coastal curl pattern rapidly changes past the tip of the cape in the expansion fan region. Several finer structures of tens of kilometers in size are located within 50-100 km of the shoreline on the lee side of the cape, with some of them consisting of positive and negative wind stress curl couplets. The coastal band of high curl south of the cape and 300 km downstream widens, is oriented at an angle to the coastline with its maximum displaced as far as 50 km or more offshore (y=200-350 km). It is worth noting the lack of a strong wind stress curl signal in the expansion fan region beyond ~100 km offshore, in contrast to the mean wind stress. Note that in the CWSST in this region neither outlines the curl field nor shows consistent increase in values over or around the higher wind stress curl areas.

Wind stress divergence for the same period yields zero or slightly negative values (i.e., convergence) within the 50-km band offshore in the upwind region, and increased convergence in the coastal zone on the windward side of the cape. Two major spatial features appear in the expansion fan region: an area of strong divergence (positive) extending southward-southwestward from the tip of the cape, and area of negative divergence further downstream on the lee side, aligned at an angle to the coastline, similar to a feature seen in the curl field. Contours of negative DWSST mostly outline the divergence field in the upwind region. In the expansion fan region, there is a small enclosure of high positive DWSST, and other positive and negative DWSST areas further downstream in the nearshore 50 km, although these are not well correlated with the wind stress divergence field.

Overall, similarly to the results with wind stress and SST, both wind stress curl - CWSST and wind stress divergence - DWSST fields indicate strong interconnections
between the pairs of variables in the nearshore 50-100 km of the upwind region. In the expansion fan region, however, wind stress and its derivative fields are not tightly connected to the shoreline, exhibit smaller-scale structures, or aligned at an angle to the shoreline. CWSST and DWSST are no longer shoreline-oriented, or follow their corresponding WS derivative field.

Scatterplots and linear fit between the average wind stress magnitude and SST, WS curl and CWSST, WS divergence and DWSST are presented in Figure 10. Separate calculations and plots were made for the upwind and expansion fan regions. Data used in calculations were limited to grid points 100 km off the coastline to focus on the upwelling zone; two nearshore ocean grid points were excluded to avoid effects of numerical diffusion. Wind stress – SST relationships (Fig. 10a) show a slope of 0.023°C/(N m⁻²) for the upwind region, with relatively moderate scatter about the linear fit line. Note that mean values were retained during calculations of the linear fit, in order to demonstrate the range of SST and wind stress. The expansion fan region yields a slope half as large, 0.011°C/(N m⁻²), and the large scatter of the data suggests that this coefficient is not distinguishable from zero. Thus, the orographic and diurnal effects disrupt the SST-stress correlation in the coastal zone of the expansion fan region.

In linear fits of wind stress curl vs. CWSST and wind stress divergence vs. DWSST (Fig. 10b,c), the regression line slope is often referred to as the “coupling coefficient” The upwind region yields slopes of 1.75 and 1.06, for wind stress curl-CWSST and wind stress divergence – DWSST, respectively. The slopes of the corresponding wind stress derivatives to SST derivatives in the expansion fan region, are notably lower (0.27 and 0.16), with substantial scatter. This is in agreement with the previous finding, that the upwind region exhibits much stronger and more consistent SST-wind stress relationships than the expansion
fan region. Note the small range of DWSST values in the expansion fan area, which results from limited variability of the upwelling-associated SST gradient in the alongshore direction (lack of spatial structure). Thus, coupling coefficients for both wind stress curl-CWSST and wind stress divergence-DWSST are smaller in the expansion fan area, because of modulation of wind stress in this area by orographic and diurnal effects.

Temporal correlations between corresponding hourly time series of modeled wind stress, SSTs, and their derivatives were computed for 13 days (25-336 h) to include the first days of upwelling propagation in the nearshore, when the upwelling front develops and moves offshore and temporal changes of the SSTs are the greatest. Low-pass filtering using a 24-h running average (Fig. 11, upper panels) produced high correlations (mostly over 0.8) in the upwind region within 50 km of the shoreline for all pairs of variables. These high correlations reflect longer-term effects of upwelling on the atmospheric wind stress and its derivatives, and demonstrate air-sea coupling in the nearshore of the upwind region. Most of the correlations rapidly decrease beyond the 50-km zone, and become negative and significant for SST—wind stress pairs. Note that there are small offshore variations of the variables in the upwind region. Another spatial feature is apparent about 150-200 km off the coast parallel to the windward side, resulting in high positive correlations for SST—wind stress and wind stress divergence – DWSST, high negative correlations for wind stress curl – CWSST. This spatial feature is also evident in wind stress derivative fields (Fig.9 b-c). It is aligned with the shelf edge of the northern side of the cape (compare to Fig. 1), but extends southwestward beyond the bathymetric bend, suggesting an ocean circulation response to the bottom bathymetry changes outlining the coastal cape. Note that the SST gradients are minor in this area, and the correlations may have arisen from negligibly small variations.
The expansion fan region is characterized by notably less consistency in correlations of the low-pass filtered time series, especially for wind stress curl-CWSST and wind stress divergence-DWSST pairs, which show several intermittent areas of significant increased, reduced, or negative correlations (Fig. 11b-c, upper panels). Wind stress-SST correlations are high and positive within the nearshore 50 km, except in the southern end of the cape; a high correlation zone extends further offshore south of the cape, in agreement with widening of the upwelling zone. Nevertheless, there are substantial areas of reduced or negative correlations within the 100-km nearshore zone of the expansion fan.

Temporal correlations between the pairs of low-pass filtered time series of SST and wind stress fields and their derivatives illuminate the effects of the diurnal cycle on the air-sea coupling (Fig. 11, lower panels). The most prominent feature in these plots is the large zone of high positive correlations between the SST and wind stress in the expansion fan region stretching several hundreds of kilometers off the coast, corresponding to the diurnal variation of wind stress field (compare to Fig. 8). This is in contrast to the upwind region that shows weaker correlations between SST and wind stress within the 100-km nearshore zone, and only minor positive correlations further offshore. Wind stress curl-CWSST and wind stress divergence-DWSST correlation are mostly small or insignificant over most of the domain, except for higher correlations (~0.4) of wind stress curl-CWSST 50 km off the coast in the upwind region, and scattered areas of negative but significant correlations of wind stress divergence-DWSST within 50-80 km of the coast.

The above analysis of temporal correlations allows separating multi-day coupling effects from those on the diurnal scale, and illuminates the regions where these effects are stronger or weaker. For example, the decreased coupling coefficient estimated for the expansion fan area from temporal mean variables (Fig. 10, lower panels) can be explained by
decreased or negative correlations between low-passed time series. On the other hand, very strong coupling (significant high temporal correlations between SST and wind stress) on the diurnal scale results for the extended region on the lee side of the cape, but which is not evident for other pairs of time series.

\[ b. \textit{Simulation with fixed SST at the atmospheric lower boundary}\]

In the control case, upwelling generates reduced wind stress over the colder water adjacent to the coast. However, the winds are also strongly affected by the coastal terrain and diurnal forcing, which cause substantial wind stress gradients. To separate these effects, we conducted a second experiment by simulating a case with SST held invariant in time and space and at a fixed initial value (14°C), thus eliminating upwelling feedback to the atmosphere. Since the differences in wind stress fields in these two cases can result only from the evolution of the SST in the coupled case, the differences in results between the control case and this “fixed-SST” case provide a convenient framework for analyzing the wind stress—SST interaction.

Average wind stress magnitudes in the fixed-SST case are shown in Fig. 12(a), and average differences of hourly wind stresses and lower atmospheric boundary conditions (i.e., SSTs) between the control case and partially-coupled are shown in Fig. 12(b). The wind stress field qualitatively resembles that from the control case, with an expansion fan in the lee of the cape, but with stronger wind stresses. Mean wind stress magnitude differences between the two cases correspond well with the differences in SST; negative differences are commonly found in both variables inshore from the upwelling front (within ~50-80 km of the coast) and positive differences are evident further offshore. The exception to this general
picture occurs on the downwind side of the cape, where the expansion fan dominates the flow. Greater SST differences result on the lee side next to the coast, and the region of negative differences extends up to 100 km offshore. On the windward side of the cape, a band of weaker wind stresses is indicated about 150 km offshore, yielding positive differences. Spatial correlation of the temporal average of the differences in SST and wind stress fields is 0.84 for the entire domain.

Temporal average wind stress curl and wind stress divergence fields for the fixed-SST case are shown in Fig. 13(a,c). The differences in wind stress curl and wind stress divergence between the coupled and partially coupled cases are shown in Fig. 13(b,d); the overlaid contours are the “differences” in CWSST and DWSST between the same cases. Note that due to the fixed-SST case being spatially uniform, both CWSST and DWSST are zero. Thus, the differences in CWSST and DWSST in Fig. 13(b,d) are determined by their corresponding values from the coupled case only. Wind stress curl and divergence fields differ from those in the control case in that greater curl and stronger divergence/convergence results in the expansion fan area, and also in the lack of nearshore variability in the upwind zone. Increased correspondence is visible between the differences in wind stress curl – CWSST and wind stress divergence – DWSST, than between their nominal values.

Scatterplots and linear fits for separate regions (inshore 100 km) for the three pairs of differences, namely, wind stress – SST, wind stress curl – CWSST, and wind stress divergence – DWSST, are shown in Figure 14. The scatterplot for the upwind region indicates a consistent response of wind stress to SST, which is an increase of about 0.13 Nm-2 in wind stress per 10oC of temperature change, while the expansion fan region yields only 0.07 N m-2 per 10oC; the total region results in 0.14 N m-2 per 10oC. The wind stress curl – CWSST difference plot results in linear slopes of 1.41 and 0.41 for the upwind and
expansion fan region, respectively, and 1.25 for the entire region. Note that the coefficients are smaller, and the scatter of values in the expansion fan region are noticeably greater for wind stress—SST and wind stress curl–CWSST pairs. The wind stress divergence–DWSST differences produced linear slopes of 1.21 and 1.94 for the corresponding regions. Despite higher coefficients for the expansion fan region, greater scatter of the values and their concentration around the center of the plot do not support a concept of increased correlation. Due to the lack of SST gradients offshore, all the scatterplots are limited to within 100 km of the coastline, focusing on the upwelling zone. The area of high temporal correlations on a diurnal scale in the expansion fan region further offshore (Fig. 11a, bottom) is therefore not considered in the above calculations. Nevertheless, the high correlation zone has interesting implications for the ocean circulations, which will be discussed in the subsequent manuscript.

5. Summary and Discussion

This study describes and analyzes coupled ocean-atmosphere regional model simulations of the coastal circulation in regions of orographically-intensified flow during the development of coastal upwelling. The model domain simulates an eastern ocean boundary with a single cape protruding into the ocean in center of the coastline. This setting resembles the U.S. West Coast, where northerly or northwesterly atmospheric flow often generates a sequence of expansion fans and compression bulges along major capes and points. When such a flow persists over a longer period, as during summertime conditions, it also causes nearshore Ekman divergence resulting in coastal ocean upwelling. Our coupled simulation lasted for 14 days, and included both diurnal effects and multi-day upwelling development. The model simulated the formation of an expansion fan downwind of an idealized cape,
featuring higher wind stresses with significant diurnal variability. The atmospheric flow reflected the formation of an elevated southward jet on the lee side of the cape, in which the strongest winds were found near the top of the MABL. This jet weakened southward with the distance from the tip of the cape. The inshore region of the expansion fan was also characterized by rapid shoreward drop in the MABL heights.

The general realism of these features is supported by observations and simulations from earlier studies (Haack, et al., 2001, Kindle et al., 2002). The presence of strong diurnal modulation in the vertical structure of the atmospheric boundary layer wind in the coastal zone was also reported by Bielli et al., 2002. In particular, Haack et al. (2001) investigated real-data forecasts of the summertime marine layer flow between Cape Blanco and Cape Mendocino, and found that supercritical flow features and their degree of interaction vary diurnally. These diurnal oscillations, driven by sea-land-breeze circulation may enhance or diminish the expansion fan in the lee of Cape Blanco. Our study demonstrates a strong diurnal cycle in the lee side of an idealized cape; wind stresses are strongest and show greatest spatial gradients in the local evening, while this area of wind intensification almost disappears in the local morning.

Simulated coastal upwelling produced a drop in SST of over 4°C SST drop near the coast, whereas offshore temperatures increased by over 2°C due to solar heating. The upwelling front was found about 50 km off the coast upwind of the cape or away from the expansion fan region. The upwelling front doubled in width on the lee side of the cape with the lowest SSTs found near the coast at the southern (lee) side of the cape. Negative surface fluxes produced over the coldest SSTs contributed to increased static stability of the lower tropospheric flow, which was marked by a rapid drop in the atmospheric TKE. Modeled
cloud regimes compared well with typical conditions, reflecting formation of a cloud-free region behind the cape and cloud build-up on the upwind side of the cape.

Studying the wind stress – SST coupling effects included several techniques. First, wind stresses and their derivatives from the control coupled case were analyzed and compared with the corresponding SSTs and its derivatives from the same simulation. An additional simulation with fixed SST, eliminating an ocean feedback to the atmosphere, was conducted in order to isolate the effects of ocean coupling.

Temporal means of modeled quantities from the control simulations were analyzed as follows: wind stress was mapped vs. average SST for the 100-km coastal area, and similarly for the average wind stress curl vs. average cross-wind SST gradient (CWSST), and for wind stress divergence vs. downwind SST gradient (DWSST). Linear regression slopes, or coupling coefficients, were calculated separately for the expansion fan area and for the upwind region and showed a significant decrease in SST-wind stress coupling within the expansion fan, where the coefficients were two or more times smaller than in the outer region. Higher higher scatter (and as a result, higher standard deviations) of the values in the expansion fan area supported the assumption of reduced correlations in that region.

Temporal correlations were also computed for similar pairs of modeled time series, wind stress—SST, wind stress curl—CWSST , and wind stress divergence—DWSST. The time series were split into “low-pass” and “high-pass” signals by computing 24-h running averages and departures from these moving averages, respectively. Temporal correlations revealed spatial extent and temporal scale of the coupling, which generally agreed and complemented the earlier assumptions about the coupling in upwind vs. expansion fan regions. However, one of the major findings was the high correlation between wind stress
and SST on a diurnal scale in the expansion fan region extending 300-400 km offshore. Reduced correlations of the modeled SST-wind stress suggested by lower coupling coefficients within the nearshore 100 km therefore occur solely due to orographic control of the wind stress, and not the diurnal modulations.

Comparison of our temporal correlation results to real data simulations conducted by Haack et al. (2008) suggests good correspondence for coupling coefficients. In their study, temporal correlations were computed between overlapping 29-day averages of wind stress derivatives and SST derivatives from real-data COAMPS model predictions, and they found negative correlations downwind of all major capes and headlines along southern Oregon – California coast. Along the entire coastline in our simulation, the wind stress curl—CWSST coupling coefficient (1.25) was within 10% of the real-data COAMPS forecasts for a 100-km coastal swath zone (1.39); our wind stress divergence—DWSST coefficient (0.84) was almost 40% lower than in the same study (1.37). Lower wind stress divergence—DWSST coupling ratios in our study likely resulted from the lack of spatial SST structure in the downwind direction.

The correlation between the time series of differences in SST and differences in wind stress, taken between the control and “fixed-SST” cases confirmed previous findings on their strong relationship within the upwelling zone in the upwind region, and lower correlations in the lee of the cape. Our estimations of modeled wind stress change relative to SST change indicated an average reduction 0.14 Nm-2 in wind stress per 10oC of SST decrease.

A major result of this study is that orographic and diurnal modulations of the near-surface atmospheric flow on the lee side of the cape strongly affect the wind stress dependence on underlying SST conditions. Strong orographic forcing is found to weaken the
average wind stress—SST coupling computed for the 100-km coastal zone, and indicates high negative correlations next to the tip of the cape. On the other hand, diurnal modulations of the wind stress in the expansion fan region are found to correlate well with diurnal SST changes, and to enhance the air-sea coupling. The study suggests that use of a fully coupled model is necessary for accurate prediction in the nearshore zone, where the coupling effects appeared to be the strongest and highly dependent on both orographic and diurnal forcing.

Subsequent publications focused on these coupled results are expected to detail the MABL momentum and stability budgets and address the role of the local pressure gradient, including its spatial and temporal variability, on the MABL wind regime in the vicinity of coastal capes. A separate publication is also planned for the ocean model analysis, the role of diurnal modulations, and ocean feedback to the atmosphere on resulting coastal circulation about the irregular coastline.
Acknowledgements

This research has been supported by the Office of Naval Research Grant N00014-08-1-0933. This work has been also supported in part by a grant of computer time from the DoD High Performance Computing Modernization Program at Maui High Performance Computing Center.
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