Sensitivity of Simulated Coastal Upwelling to Wind Resolution

Changming Dong*, Alex Hall, Mimi Hughes, and James McWilliams

Department of Atmospheric and Oceanic Sciences, University of California, Los Angeles, CA 90095

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*Corresponding author address: Dr. Changming Dong, IGPP, University of California, Los Angeles, 405 Hilgard Avenue, Los Angeles, CA, 90095-1567, email: cdong@atmos.ucla.edu
In March 2002 an upwelling event was observed in the Southern California Bight, marked around the middle of the month by precipitous cooling of at least 4°C within 10 20-km of the coast, and slow recovery to pre-event temperatures by month’s end. Imposing winds of a 54-km resolution regional atmospheric model constrained by larger-scale observed conditions on a 1-km-resolution regional oceanic model, we simulate this event qualitatively; yet the SST anomalies associated with it are very weak—only about a third of the observed. When atmospheric model resolution is increased to 18-km and the oceanic simulation is repeated, the oceanic realism improves, and cooling increases to roughly half the observed. The cooling increases further to about 2/3 of the observed with an increase of atmospheric resolution to 6-km. A final resolution increase to 2-km gives nearly the same result, due to convergence in the simulated wind fields. The cooling increases so much as atmospheric resolution increases to 6-km because the upwelling-favorable flow—clearly visible even in the 54-km atmospheric solution—is increasingly displaced toward the coastal ocean by the newly resolved coastal topography, intensifying wind stress in the upwelling zone. Moreover, the displacement of the flow has the effect of channeling the large-scale flow so that it follows the meanders of the coastline more precisely, increasing the favorability of the winds for upwelling all along the coast. This process does not occur with the final increase in resolution to 2-km because the 6-km simulation resolves nearly all the region’s topographic variance, as revealed by an estimate of the characteristic length scales of the region’s nearshore topography. These length scales likely determine the atmospheric resolution necessary for realistic simulation of coastal upwelling, not only in Southern California, but in any mountainous coastal region.
1 The Spatial Scales of Coastal Upwelling

Coastal upwelling is one of the principal means by which nutrients critical for marine organisms make their way to surface ocean, and it also dramatically affects the temporal and spatial distribution of sea surface temperatures (SSTs) in coastal zones worldwide. Atmospheric subtropical highs generate upwelling-favorable winds along the eastern boundaries of the subtropical oceans, leading to semi-permanent upwelling zones there. Upwelling is also common along the rest of the world’s coasts, and is typically associated with episodic alignments of winds with the coast on the left (right) in the northern (southern) hemisphere.

In spite of its global significance for oceanic physics and biology, coastal upwelling is an inherently local process, and as a result is not well-simulated in current global oceanic models. Some obvious reasons for this can be traced directly to the coarse resolution of these models, typically on the order of 100-km. Coastal upwelling is usually confined to a narrow coastal strip approximately one oceanic baroclinic deformation radius wide (about 20 km-50 km). Meanwhile, coastlines are often complex and meandering, so that the direction of upwelling-favorable winds can be a highly-sensitive function of coastal position. More subtle is the possibility that the atmospheric processes affecting upwelling may also exhibit small-scale variability, introducing a resolution requirement in atmospheric models as well. The topography of coastal land areas is often complex, and the flow of large-scale winds over this topography introduces fine-scale structures into coastal winds fields. The implications for upwelling are difficult to predict \textit{a priori}. However, they are certainly characterized by anomalies in wind stress and wind stress curl, both of which affect upwelling distributions in the coastal ocean (Capet et al., 2004). To examine the affect of these anomalies, we simulate atmospheric winds at four different resolutions by downscaling an atmospheric reanalysis product to a regional atmospheric model. We focus on a period when upwelling occurred in the region, as registered by buoy measurements of SST. We then impose these winds one at a time on a regional oceanic model with 1-km horizontal grid resolution, giving a sensitivity of simulated
upwelling to the spatial scales resolved in the wind products.

Our test bed for this study is the Southern California Bight (SCB), Fig. 1. Upwelling events are less common here than elsewhere along the U.S. west coast. The reason is that the area is often sheltered from the upwelling-favorable northerly winds often found further north (Caldeira and Marchesiello 2002), and usually exhibits a significant shoreward decrease in wind intensity. However, during spring, this gradient weakens episodically (Winant and Dorman, 1997), a state often associated with upwelling in the SCB. Evidence of such upwelling events is visible in the time series of daily-mean SST anomalies at National Data Buoy Center (NDBC) buoy 46025 (Fig. 2), located about 50-km offshore from the coast. This is one of the longest available time series of SST in the SCB’s coastal waters. Once or twice nearly every spring, SST dips by 2-4°C for a week or two at a time. Based on this time series, upwelling in the SCB appears to be highly episodic. This makes the region well-suited to the study of the role of fine-scale atmospheric structures in generating upwelling, because events can be unequivocally attributed to specific wind forcing, that can then be analyzed.

A typical upwelling episode in the SCB occurred in the middle of March 2002, with a roughly 3°C decrease in SST at NDBD buoy 46025. An observed SST time series from a mooring at the outer edge of the Santa Monica Bay (SMB mooring) also registered a sharp drop in SST beginning about March 13, 2002 (Fig. 5, dashed blue line), the same date as the onset of the March 2002 event recorded in buoy 46025. From then until March 18, SST decreased at the SMB mooring by about 4°C, after which SSTs slowly recovered. The event was also recorded by NDBC buoy 46053, which showed a drop of about 3°C from March 13 to 18 (Fig. 6). The fact that this upwelling event was recorded by three different buoys suggests it was a significant regional-scale upwelling event. This is confirmed by SST data from the satellite Pathfinder. Comparing the SST images on March 13 (upper panel in Fig. 3) with that on March 18 (lower panel in Fig. 3), one can see the sharp drop in SST from Point Conception all the way to San Diego.
The sensitivity of oceanic responses to wind resolution has also been noticed previously, for example, Pullen et al. (2003) and Gan and Allen (2005). Previous studies in the SCB also hint at the importance of wind variations with small spatial scales in forcing the SCB circulation. Analyzing observational data, Harms and Winant (1998) found a correlation between the local wind and circulation in Santa Barbara Channel (SBC), the northern part of SCB. Using three wind products with different resolutions to simulate the seasonal cycle of the SCB circulation, Di Lorenzo (2003) showed that the model solutions vary significantly depending on which wind product is imposed. Capet et al. (2004) pointed out the upwelling strength along the US west coast is sensitive to the cross-shore gradient of the wind stress, which is the major part of the wind curl. And finally applying three types of winds, including two different model-generated winds and one observation-merged wind to simulate the surface circulation in SBC, Dong and Oey (2005) suggested that if the fine structure can be resolved in the wind stress field, such as wind curl around headlands, the oceanic response is much closer to observation.

In section 2 we describe in detail the atmospheric and oceanic models and the experimental design employed in this case study. Then in section 3, we present the simulated oceanic response to the four atmospheric simulations, showing that upwelling is highly sensitive to the resolution of the imposed winds, but that the oceanic solution converges as the highest resolution is approached. Then, in section 4, we relate this convergence to the characteristic spatial scales of the coastal topography, and in so doing provide a physical basis for determining the atmospheric model resolution necessary to simulate coastal upwelling in the region. In section 5 we discuss our results and their implications for realistic simulation of coastal upwelling worldwide.

2 Experimental Design

The Regional Oceanic Modeling System (ROMS) was used to simulate the oceanic response to atmospheric forcing in the SCB during March 2002, the month when the large upwelling event
was recorded. ROMS solves the rotating primitive equations (Shchepetkin and McWilliams, 2005), and uses a generalized sigma-coordinate system in the vertical direction and curvilinear grid in the horizontal plane. It is a split-explicit, free-surface oceanic model, where short time steps are used to advance the surface elevation and barotropic momentum equations, with a larger time step used for temperature, salinity, and baroclinic momentum. A third-order, upstream-biased advection operator allows the generation of steep gradients in the solution, enhancing the effective resolution of the solution for a given grid size when the explicit viscosity is small. The numerical diffusion implicit in the third-order upstream-biased operator allows the explicit horizontal viscosity to be set to zero without excessive computational noise or instability. The vertical viscosity is parameterized using a K-profile parameterization (KPP) scheme (Large et al., 1994). The no-slip lateral boundary condition is also imposed through the momentum advection operator and yields an implicit lateral stress (see Dong et al., 2006, for a full discussion).

The ROMS model domain used in this study is plotted in Fig. 1. It has a grid spacing of 1-km horizontally and 40 levels vertically. The grid resolves all eight islands in SCB. Mixed boundary conditions are used along the open boundaries, i.e., the Orlanski radiation condition in the tangential direction and the Flather condition with adaptive restoration of material properties to imposed data under inflow conditions (Marchesiello et al., 2001). The restoring data for the lateral open-boundary conditions and the initial conditions are extracted from a 1996-2003 ROMS solution in a larger U.S. West Coast domain (Dong et al., 2006), using procedures similar to those in Marchesiello et al. (2003). The solid boundary around the islands and land has zero-normal and no-slip flow implemented through a standard land-mask algorithm. The same model configuration has been applied to an island wake study in the SCB (Dong and McWilliams, 2006).

Four wind products of varying resolution were imposed on this ROMS configuration, creating four distinct oceanic simulations whose only difference is the resolution of the wind forcing. The Climatological Oceanic and Atmospheric Data Set (COADS) heat flux and freshwater flux are
applied (DaSilva et al., 1994). The wind products were generated with the regional atmospheric model MM5 (5th generation Pennsylvania State University-National Center for Atmospheric Research Mesoscale Model, Grell et al., 1994). Four nested grids of resolution 54, 18, 6, and 2-km were implemented with MM5. Two-way communication takes place with the parent nest at the lateral boundaries of the three innermost domains. The coarsest 54-km resolution domain covers the Western U. S. and an equivalent-sized portion of the Pacific Ocean, while the highest 2-km resolution domain covers the SCB. Each nest therefore includes the SCB and the ROMS domain shown in Fig. 1, and records its own simulation of the region. The outermost 54-km domain was forced at its lateral and surface boundaries with data from the ETA model reanalysis for the entire month of March 2002. The lateral boundary conditions are available every 3 hours from this archive, and we interpolated them in time prior to imposing them. SSTs were updated every 3 days. This simulation can be thought of as a reconstruction of regional atmospheric conditions over March 2002, consistent with our best estimate of the large-scale conditions, the resolved topography, and MM5 model physics. Conil and Hall (2006) analyzed an atmospheric simulation identical to the present one except that it did not include a 2-km domain and covers a much longer time period (1996-2003). They provide further details about model parameterizations.

Conil and Hall (2006) also verified MM5 winds against observations for the entire time period of their simulation. Comparing the 6-km simulated daily mean wind anomalies with the daily mean wind anomalies observed at 16 stations over land and two buoys over the ocean (NDBC 46025 and 46053), they found simulated and observed winds are highly coherent throughout the 6-km domain. For example, correlations between observed anomalies in wind direction and those simulated at the nearest model grid points are greater than 0.5 and are generally around 0.7. For wind speed, correlations are above 0.4 at all 18 stations. At 10 locations they are above 0.6, with the highest correlation reaching almost 0.8. For the two ocean buoys, the direction correlations are about 0.7 and the speed correlations are about 0.7 and 0.5 for 46025 and 46053, respectively.
We performed further verification of MM5 winds over the ocean during the March 2002 period by projecting the winds at the buoys onto their principal axis of variability and computing correlations between simulated and observed winds for all four resolutions (Table 1). Though this approach is different from that of Conil and Hall (2006), the results are qualitatively similar. While the coarse-resolution winds are not so realistic, correlations reach 0.5-0.7 at 46025 and 46053 for the 6-km and 2-km winds. (However the correlation at the SMB buoy is generally lower even at the highest resolution.) Finally, we also compared a snapshot of 25-km-resolution QuickSCAT winds with the 2-km resolution MM5 winds at a moment of intense wind forcing during March 2002 (Fig. 4). The magnitudes of the winds agree nearly perfectly, though the model winds tend to be rotated a few degrees clockwise of QuickSCAT. In summary, MM5 does a reasonable job capturing the magnitude, direction, and variability of the winds, particularly at 2-km and 6-km resolution.

3 Simulation of the March 2002 event

All four simulated wind fields indicate unusual meteorological conditions developing in the SCB just prior to the March 2002 upwelling event. This is evident in the strong peaks in wind stress and wind stress curl averaged over the SCB beginning around March 12 (Figs. 7a and c). (To convert MM5 winds to wind stress, the formula of Large and Pond (1981) was used.) These mark the occurrence of unusually strong northwesterly surface winds throughout the SCB in all four wind products. On March 13 for example, simulated surface winds from the northwest reached 16 m/s to the south and west of the Channel Islands in the 2-km resolution product. This strong alongshore flow would force coastal upwelling of cold water from depth.

This is precisely what occurred in the ROMS simulations. The four SST time series at the ROMS grid point nearest the SMB mooring (Fig. 5) show a reduction in SST beginning on March 13, as in the observations. The relaxation time scales to pre-event conditions seen in all four simulations are also similar to that of the observed time series. However, the upwelling event’s
magnitude becomes increasingly realistic as resolution in wind forcing increases. The maximum
cooling relative to pre-event conditions more than doubles from about 1.0°C with 54-km forcing to
2.4°C with 2-km forcing, and with intermediate maximum cooling values for intermediate forcing
resolutions. Though the cooling with 2-km forcing is slightly larger than that associated with 6-km
forcing, 2-km and 6-km forcing give very similar SST evolutions, suggesting convergence as 2-km
resolution is approached. The four simulated SST time series at the grid point nearest NDBC Buoy
46053 (Fig. 6) show qualitatively similar results. At the SMB mooring, the observed magnitude
of the cooling is larger than in all four ROMS simulations, both because the ROMS SSTs are too
cool prior to the upwelling event, a difference also visible in the NDBC buoy 46053 time series,
and because they do not drop enough during it.

To shed light on the upwelling event’s geographical extent and put simulated and observed SST
time series in context, Fig. 8 shows a snapshot of the spatial distribution of simulated SST when
2-km wind forcing is applied. The date of the snapshot is March 18, roughly when the simu-
lated and observed SMB mooring SSTs reach their minima. The simulated upwelling is narrowly
confined to the coast, with abnormally cold waters (light blue shading) within about 20-km of the
coast stretching from Point Conception to San Diego, and even more extreme conditions (dark blue
shading) within just a few km of the coast at most locations. The SMB mooring is clearly located
within the zone of intense simulated upwelling, though SSTs even closer to the coast are approx-
imately 0.5°C colder. So the simulated SST times series at the SMB mooring location shown in
Fig. 5 probably represent close to the maximum simulated SST anomalies associated with this
event. Protruding filaments in the front separating the upwelling waters from the waters of the rest
of the SCB are also seen in this submesoscale-resolving ocean model, including at the locations
of the other two moorings. The location of mooring 46053 is particularly impacted by a plume
emanating from the large pool of cold, upwelled water just to the southeast of Point Conception.

Fig. 8 also demonstrates that the simulated upwelling event is regional in scale. In fact, the same
convergence in the simulated upwelling as 2-km-resolution wind forcing is approached not only
the SST time series at the grid points nearest the moorings, but all along the coast, as reflected in
the time series of SST averaged over the coastal waters of the SCB (Fig. 9a), and this same quantity
averaged over the course of the upwelling event (Fig. 9b).

The event’s simulated oceanic signature can also be seen in surface kinetic energy (KE) and
enstrophy (EN) fields. The surface-integrated KE and EN are defined as follows:

\[
KE = \int \int \frac{(u^2 + v^2)}{2} dA
\]

\[
EN = \int \int \left( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right)^2 dA
\]

where \(u\) and \(v\) is the eastern and northern velocities at the sea surface, respectively, and \(A\) the area.
The KE reflects the momentum of surface waters, and the enstrophy represents small-scale oceanic
surface variability. Time series of these two quantities integrated over the SCB ROMS domain are
shown in Figs. 9c and e. For all wind resolutions, significant increases in both enstrophy and KE
take place beginning around March 13. The KE in all four simulations reaches its maximum at
about the same time as the wind stress and wind stress curl reach their peak (Figs. 7a and c), and
then slowly decreases, consistent with the ocean receiving an energy pulse from the atmosphere
and then slowly dissipating it. The enstrophy maximum, on the other hand, is delayed by about 6
days relative to the wind forcing in all four simulations. This delay occurs because the time scale
of energy transfer to the small spatial scales represented by the enstrophy is about the same as the
advection time scale. Given the SCB’s 100-km length scale and a mean current speed of 0.2 m/s,
the advection time scale in this case is about 6 days, consistent with the delay time. However it
is possible that the time scale is an instability development time as a particular form of energy
transfers to small scales.

Prior to the large increases in KE and enstrophy around March 13, there is virtually no depen-
dence of these quantities on wind resolution. In addition, the wind stress and wind stress curl
integrated over the SCB (Figs. 7a and c) are nearly zero during this period. These facts suggest that prior to event onset, both the ocean’s surface momentum and small-scale variability are driven to great extent by features of the wind field much larger in scale than the SCB, and hence easily resolved at 54-km. To the extent that there are smaller-scale features in the ocean, they result from the ocean’s internal dynamics. However, both the KE and enstrophy increase noticeably with wind resolution once the event begins, reflected also in the time-averaged values of KE and enstrophy (Figs. 9d and f). This gives further evidence that wind features much smaller in scale than 54-km are largely responsible for the upwelling event.

4 Convergence with Resolution

Not surprisingly, the convergence in oceanic fields as 2-km wind forcing is approached is the result of convergence in the wind forcing fields themselves. Going from 54-km to 6-km resolution, both wind stress and wind stress curl averaged over the period of greatest wind forcing nearly double (Figs. 7b and d). Only slight increases in wind stress and wind stress curl are then seen going from 6-km to 2-km resolution, indicating little additional spatial variability in wind fields would result if resolution were increased beyond 2-km. This is consistent with the general convergence in the correlations between the MM5 wind and buoys as 2-km resolution is approached (Table 1). Here we examine the spatial distributions of wind in the four atmospheric simulations and address why convergence in the atmospheric and oceanic simulations occurs around 2-km wind resolution.

We begin by examining the patterns during the period of greatest wind forcing (Fig. 10). In all four simulations wind stress decreases shoreward, and a maximum is seen to the south and west of the Channel Islands. With increasing resolution wind stress increases and its maximum moves towards the coast. Fine structures appear in the 6-km and 2-km solutions. These structures are qualitatively similar, though they exhibit sharper gradients in the 2-km simulation. The corresponding wind stress curl patterns are shown in Fig. 11. With the coarsest 54-km grid resolution,
the only feature is a weak onshore gradient. When resolution increases to 18-km, the gradient becomes stronger, particularly southeast of Point Conception in the Santa Barbara Channel. (This is similar to the SCB wind stress curl pattern of Koracin et al. (2004), who used a 9-km version of MM5 for another time period.) With 6-km resolution wind wakes around islands are exposed. Finally, when resolution reaches 2-km, these wakes increase in intensity and remarkably narrow streaks are seen in the Santa Barbara Channel and in the lee of islands. Like their wind-stress counterparts, the 6-km and 2-km resolution curl patterns agree qualitatively, though the fine-scale structures are clearly better resolved and generally have larger magnitudes in the 2-km product.

To assess why the oceanic response converges as wind resolution approaches 2-km, it is helpful to consider whether the upwelling is caused more by anomalies in wind stress or in wind stress curl. Both alongshore wind with land on the left or positive wind curl can induce coastal upwelling. Figs. 10 and 11 show a general increase with resolution in both upwelling-favorable alongshore wind intensity and wind stress curl. This makes it challenging to separate these two mechanisms (Enriquez and Frihe, 1995). However some progress can be made by examining the incremental changes in wind stress and wind stress curl as resolution increases (Fig. 12), focusing first on the case where wind stress increases without an accompanying increase in wind stress curl: When resolution increases from 18-km to 6-km, there is a consistent increase in alongshore wind stress all along the coast from Point Conception to San Diego (Fig. 12c); however there is no systematic increase in wind stress curl (Fig. 12d). This suggests that most of the increase in simulated upwelling when wind resolution increases from 18-km to 6-km is due to a rise in alongshore wind stress.

It is not so straightforward to extend this conclusion to the case where wind resolution increases from 54-km to 18-km, since here both wind stress and curl increase systematically (Figs. 12a and b). However, the increases in wind stress going from 54-km to 18-km are at least as large as those going from 18-km to 6-km in areas critical for upwelling, and the corresponding de-
creases in simulated SST along the coast are also comparable (Fig. 9a). So it is plausible that the increase in upwelling going from 54-km to 18-km resolution winds is also due primarily to an increase in alongshore wind stress, assuming the upwelling-induced SST anomaly scales with alongshore wind stress. The changes in both wind stress and curl are small going from 6-km to 2-km resolution, consistent with the small change in oceanic response. Though it is not possible to disentangle the effects of winds stress and wind stress curl completely, on balance it seems likely that the general increase in upwelling with increasing wind resolution is mostly due to an increase in alongshore wind stress.

Given its likely importance in generating coastal upwelling, what then is the origin of the increase in alongshore wind stress with resolution? We shed light on this question by examining the variance of the observed nearshore terrain in Southern California when four spatial smoothers are applied to it whose length scales correspond to the topographic length scales resolved by the four MM5 simulations (Fig. 13). (Note that the effective cutoff spatial scale of the topography in MM5 is roughly twice the model resolution, so that the topography in the 54-km-resolution simulation does not include any terrain with length scales less than 108-km, the 18-km simulation does not include any terrain with length scales less than 36-km, etc.) When terrain is smoothed to correspond to the 18-km-resolution MM5 simulation, the variance is nearly 15% larger than when it is smoothed to correspond to the 54-km simulation. Including the topography of the 6-km simulation gives a further increase in terrain variance of a few percent. Including the length scales of the 2-km simulation yields very little increase in terrain variance, indicating most of the region’s terrain is resolved at 6-km.

The signatures of the northwesterly wind anomaly that initiated the event are striking even at 54-km resolution, indicating that this wind event is regional in scale, if not larger. However, the topography within just a few tens of km of the coast significantly distorts this flow, and greatly amplifies its power to generate coastal upwelling, both because it intensifies the flow in the up-
welling zone and changes its direction to be more favorable to upwelling. Introducing additional coastal topography in the 18-km simulation displaces the land component of this flow out over the ocean. The distribution of the incremental increase in wind stress when resolution is increased to 18-km (Fig. 12a) extends about 100-km from the coast, which corresponds quite intuitively with the length scales of the added coastal topography (between 36-km and 108-km). Likewise, the distribution of the wind stress increase when resolution is further increased to 6-km (Fig. 12c) extends roughly 30-km from the coast, and corresponds with the length scales of the added coastal topography with this increase in resolution (between 12-km and 36-km). Perhaps more importantly, the change in wind stress going from 18-km to 6-km resolution enhances the alignment of the wind stress with the complex coastline, transforming a large-scale northwesterly wind anomaly roughly parallel to the coast into a flow that follows the meanderings of the coastline more precisely. Very little incremental change in wind stress is seen going from 6-km to 2-km resolution (Fig. 12e), consistent with the fact that most of the region’s terrain is resolved in the 6-km MM5 simulation.

5 Summary and Discussion

In this study, we examine the possibility that atmospheric processes affecting coastal upwelling exhibit small-scale variability due to the complex topography of coastal land areas, introducing a resolution requirement in atmospheric models for the realistic simulation of coastal upwelling. Our approach is to simulate winds at four different resolutions in a coastal region by downscaling an atmospheric reanalysis product to a regional atmospheric model, focusing on a period when upwelling occurred in the region. We then impose these winds one at a time on a 1-km regional oceanic model, giving a sensitivity of simulated upwelling to the spatial scales resolved in the wind products.

Our test bed for this problem is the SCB, a region of highly episodic coastal upwelling. An upwelling episode typical of the region occurred in the middle of March 2002, marked by decreases
in SST of roughly 3-4°C, as recorded at three buoys along the Southern California coast. Imposing winds of a 54-km resolution regional atmospheric model constrained by larger-scale observed conditions on a 1-km-resolution regional oceanic model, we simulate this event qualitatively; yet the SST anomalies associated with it are very weak—only about a third of the observed. When atmospheric model resolution is increased to 18-km and the oceanic simulation is repeated, the oceanic realism improves, and cooling increases to roughly half the observed. The cooling increases further to about 2/3 of the observed with an increase of atmospheric resolution to 6-km. A final resolution increase to 2-km gives nearly the same result, due to convergence in the simulated wind fields. The cooling probably increases so much as atmospheric resolution increases to 6-km because the upwelling-favorable flow—clearly apparent even in the 54-km atmospheric solution—is increasingly displaced toward the ocean by the newly-resolved coastal topography, intensifying upwelling-favorable wind stress in the upwelling zone. This process does not occur with the final increase in resolution to 2-km because the 6-km simulation resolves nearly all the coastal region’s topographic variance, as revealed by an estimate of the characteristic length scales of the region’s nearshore topography.

These length scales likely determine the atmospheric resolution necessary for realistic simulation of coastal upwelling, not only in Southern California, but in any mountainous coastal area. This generalization applies to much of the world’s coastline, including the entire west coast of the Americas, the east coasts of Canada and central America, and eastern Brazil, the Scandanavian and Mediterranean coasts, much of the west coast of Asia, the Indonesian coast, the east and west coasts of Australia, southern African coast, and the coasts of Antarctica and Greenland. When large-scale atmospheric flows are roughly aligned with the coastline in an upwelling-favorable direction, topography adjacent to the coast will channel and displace the flow over the ocean, increasing its intensity in the upwelling zone. Perhaps more importantly, it will realign the flow to follow the meanderings of the coastline more precisely, and enhance its favorability for upwelling at all
locations along the coast. This link between the coastal topography and upwelling processes is physically intuitive: The characteristic length scales of the meanderings of the coastline and those of coastal mountain complexes ought to be roughly comparable, as the coastline can be thought of as an elevation isoline of the solid earth’s surface. Our results indicate that if these length scales are not resolved in both atmospheric and oceanic solutions, upwelling will be seriously undersimulated.

The ROMS model forced by 2-km-resolution MM5 winds simulates the timing and magnitude of this upwelling event with remarkable fidelity. However, differences between this simulation and available observations remain. The simulated SST anomaly at the SMB mooring is weaker than observed both because simulated SSTs are about 0.75°C too cold prior to event onset, and because SSTs during the event are about 0.75°C too warm (Fig. 5). These biases may be caused by errors in solar radiation fields imposed on ROMS. These fields are very difficult to simulate because they are strongly modulated by the presence of stratocumulus cloud. Current atmospheric models misrepresent boundary layer processes generating stratocumulus at the regional scale, including high resolution atmospheric models such as the one used here (Bretherton et al. 2004; McCaa et al. 2004). For this reason we imposed an observed 1° by 1° climatological solar radiation field (DaSilva et al., 1994) on the ocean model. The values in this data set may differ from those actually occurring in the SCB during March 2002 both because of its coarse resolution and because it is a climatology. The convergence of the ROMS solution when wind forcing resolution reaches 2-km eliminates the possibility that lack of resolution in wind forcing is responsible for these differences. They could arise for many complicated reasons, such as inaccuracy in initial condition, open lateral boundary flux, and surface fluxes, some of which can not be eliminated without data assimilation.

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References


Table 1 Correlation between MM5 and Buoy Winds of March 2002 along the principle axis.
FIGURE CAPTIONS

Fig. 1 Bathymetry of the Southern California Bight. The variable in color is depth (m). The red solid circles indicate the locations of three buoys (SMB-SCOOS, NDBC-46025 and NDBC-46053) used in this study. This region also corresponds to the ROMS domain discussed in section 2.

Fig. 2 Time series of the SST anomaly relative to the annual-mean at NDBC buoy 46025 for 10 years (1995-2004).

Fig. 3 SST distribution on March 13 (upper) 18 (lower) in 2002 in SCB, acquired by AVHRR/Pathfinder.

Fig. 4 Comparison between QuickSCAT wind and MM5 wind (2km) at 10:00 AM (GMT), March 16: MM5(blue) and QuickSCAT(red).

Fig. 5 SST time series during March 2002 at the SMB mooring (33.9°N, 118.70°W) and nearest ocean model grid point when the ocean model is forced by the four wind fields of varying resolution.

Fig. 6 SST time series during March 2002 at the NDBC Buoy 46053 (34.24°N, 119.85°W) and nearest ocean model grid point when the ocean model is forced by the four wind fields of varying resolution.

Fig. 7 March 2002 wind stress and wind stress curl magnitudes integrated over the ROMS SCB model domain: (a) wind stress time series at four atmospheric model resolutions, (b) wind stress averaged over the whole month as a function of atmospheric model grid spacing, (c) and (d) are as in (a) and (b), except for wind stress curl.

Fig. 8 A snapshot of the simulated sea surface temperature (°C) distribution on March 18, 2002 when 2-km resolution winds are imposed on ROMS.

Fig. 9 Oceanic response convergence with the increase in wind resolution. (a) SST time series integrated over the area within 50-km of the coast when ROMS is forced by winds of four atmospheric model resolutions, (b) SST in (a) averaged over the whole month as a function of
atmospheric model grid spacing, (c) time series of domain-integrated oceanic kinetic energy when ROMS is forced by winds of four atmospheric model resolutions, (d) domain-integrated oceanic kinetic energy averaged over the whole month as a function of atmospheric model grid spacing, (e) and (f) as in (c) and (d), except for sea surface enstrophy.

Fig. 10 The time-averaged surface wind stress magnitude (Pa) in the SCB in March 2002 for the four atmospheric grids: (a) 54 km, (b) 18 km, (c) 6 km, (d) 2 km.

Fig. 11 The time-averaged surface wind stress curl (Pa/100-km) in the SCB averaged over March 2002 for the four atmospheric grids: (a) 54 km, (b) 18 km, (c) 6 km, (d) 2km.

Fig. 12 Spatial distribution of wind stress and curl changes as atmospheric model resolution is increased, with the increments in resolution increase noted as text inside each panel. On the left are wind stress, and on the right are wind stress curl, averaged over the upwelling period.

Fig. 13 The observed spatial variance of the terrain in coastal Southern California (i.e. the land areas of Fig. 1) when the terrain is smoothed with filters corresponding to the resolved topographic length scales in the four MM5 simulations. The grid spacing of the atmospheric simulations is the abscissa.
Table 1: Correlation between MM5 and Buoy Wind along the principle Axis

<table>
<thead>
<tr>
<th>Buoys</th>
<th>54km</th>
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<th>6km</th>
<th>2km</th>
</tr>
</thead>
<tbody>
<tr>
<td>46025</td>
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<td>0.61</td>
<td>0.50</td>
<td>0.65</td>
</tr>
<tr>
<td>46053</td>
<td>-0.01</td>
<td>0.46</td>
<td>0.68</td>
<td>0.68</td>
</tr>
<tr>
<td>SMB</td>
<td>-0.18</td>
<td>0.24</td>
<td>0.26</td>
<td>0.26</td>
</tr>
</tbody>
</table>
Figure 1: Bathymetry of the Southern California Bight. The variable in color is depth (m). The red solid circles indicate the locations of three buoys (SMB-SCOOOS, NDBC-46025 and NDBC-46053) used in this study. This region also corresponds to the ROMS domain discussed in section 2.
Figure 2: Time series of the SST anomaly relative to the annual-mean at NDBC buoy 46025 for 10 years (1995-2004).
Figure 3: SST distribution on March 13 (upper) 18 (lower) in 2002 in SCB, acquired by AVHRR/Pathfinder.
Figure 4: Comparison between QuickSCAT wind and MM5 wind (2km) at 10:00 AM, March 16 (GMT): MM5(blue) and QuickSCAT(red).
Figure 5: SST time series during March 2002 at the SMB mooring (33.94°N, 118.70°W) and nearest ocean model grid point when the ocean model is forced by the four wind fields of varying resolution.
Figure 6: SST time series during March 2002 at the NDBC Buoy 46053 (34.24°N, 119.85°W) and nearest ocean model grid point when the ocean model is forced by the four wind fields of varying resolution.
Figure 7: March 2002 wind stress and wind stress curl magnitudes integrated over the ROMS SCB model domain: (a) wind stress time series at four atmospheric model resolutions, (b) wind stress averaged over the whole month as a function of atmospheric model grid spacing, (c) and (d) are as in (a) and (b), except for wind stress curl.
Figure 8: A snapshot of the simulated sea surface temperature (°C) distribution on March 18, 2002 when 2-km resolution winds are imposed on ROMS.
Figure 9: Oceanic response convergence with the increase in wind resolution. (a) SST time series integrated over the area within 50km of the coast when ROMS is forced by winds of four atmospheric model resolutions, (b) SST in (a) averaged over the whole month as a function of atmospheric model grid spacing, (c) time series of domain-integrated oceanic kinetic energy when ROMS is forced by winds of four atmospheric model resolutions, (d) domain-integrated oceanic kinetic energy averaged over the whole month as a function of atmospheric model grid spacing, (e) and (f) as in (c) and (d), except for sea surface enstrophy.
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