Idealized Simulations of Atmospheric Coastal Flow along the Central Coast of California

ZHIGIANG CUI,† MICHAEL TJERNSTRÖM, AND BRANKO GRISOGONO

Department of Meteorology, Uppsala University, Uppsala, Sweden

(Manuscript received 10 February 1997, in final form 15 December 1997)

ABSTRACT

A fully nonlinear, primitive equation hydrostatic numerical model is utilized to study coastal flow along central California, combining a realistic atmospheric model, with a higher-order turbulence closure, with highly simplified background flow. Local terrain and surface forcing of the model are treated realistically, while the synoptic-scale forcing is constant in time and space. Several different simulations with different background wind directions were performed. The motivation is to isolate the main properties of the local flow dependent on the coastal mesoscale influence only and to facilitate a study of the structure of the coastal atmospheric boundary layer, the mean momentum budget, and the atmospheric forcing on the coastal ocean for simplified quasi-stationary but still typical conditions. The model results feature the expected summertime flow phenomena, even with this simplified forcing. A coastal jet occurs in all simulations, and its diurnal variability is realistically simulated. The coastal topography serves as a barrier, and the low-level coastal flow is essentially coast parallel.

Among the conclusions are the following. (i) The boundary layer for a northerly jet is more shallow and more variable than that for a southerly jet. One reason is an interaction between waves generated by the coastal mountains and the boundary layer. A realistic inclusion of the Sierra Nevada is important, even for the near-surface coastal atmosphere. (ii) The transition from southerly to northerly flow, when changing the background flow direction, is abrupt for a change in the latter from west to northwest and more gradual for a change east to south. (iii) The low-level flow is in general semigeostrophic. The across-coast momentum balance is geostrophic, while the along-coast momentum balance is dominated by the pressure gradient. Local acceleration and spatial variability close to the coast arise as a consequence of the balance among the remaining terms. For southeasterly background flow, the across-coast momentum balance is dominated by the background synoptic-scale and the mesoscale pressure gradients, sometimes canceling the forcing, thus making this case transitional. (iv) Smaller-scale flow transitions arise for some background flow directions, including an early morning jet reversal north of Monterey, California, and a morning-to-noon low-level eddy formation in the Southern Californian Bight. (v) The model turbulence parameterization provides realistic patterns of the atmospheric forcing on the coastal ocean. (vi) Characteristic signals measured in propagating wind reversals related to boundary layer depth and inversion structure here are seen to correspond to different quasi-stationary conditions.

1. Introduction

A coastline represents a step change in surface characteristics, that is, roughness, temperature, moisture, and orography and, as a consequence, a variety of mesoscale phenomena occur, for example, land–sea-breeze systems and coastal jets. A special class of coastal flows appears when the coastal terrain acts as a barrier. This happens on any coast provided that the coastal marine atmosphere (MABL) is sufficiently stably stratified, shallow, and/or that the coastal terrain is high enough. With sufficient stability even moderate terrain may become hydrodynamically steep (Tjernström and Grisogono 1996; Grisogono and Tjernström 1996) and can block the flow. Areas with high coastal mountains are subject to such blocking more frequently; this is the case for much of the west coast of North America. Although terrain heights vary significantly along this coast, the depth of the MABL is often less than the height of the terrain, in particular during summer (cf., e.g., Dorman and Winant 1995). Flow phenomena that develop as a consequence of this blocking have a significant impact on local weather.

Overland (1984) scales the momentum equations under such conditions. He introduces four nondimensional numbers in addition to the Rossby radius of deformation, \( l_R = NDf^{-1} \) (\( N \) is the Brunt–Väisälä frequency, \( D \) is the MABL depth, and \( f \) is the Coriolis parameter).
and the Froude number, \( Fr = V[DN]^{-1} \). There are separate along- and across-coast Rossby numbers, \( R_l = V[fl]^{-1} \) and \( R_2 = U[fl]^{-1} \) (where \( l \) and \( L \) are typical across-coast and along-coast length scales, while \( U \) and \( V \) are typical velocity scales, respectively); a drag parameter, \( C_D = C_D LD^{-1} \) (where \( C_D \) is a drag coefficient); and a stratification parameter, \( S = [ND]^2[fl]^{-2} \). With typical values for the central coast of California, \( R_l \gg R_2 \), the flow is often semigeostrophic; geostrophic in the across-coast and ageostrophic in the along-coast balance (Overland and Bond 1993, 1995) and the flow is quasi-parallel to the coast (Beardsley et al. 1987; Dorman and Winant 1995). Stability and turbulent stress are often but not always factors since \( S \) is about 0.1–10 and \( C_D \) is about 0.1–1. Based on the typical range of these numbers, the coastal terrain almost always influences the coastal MABL, and perturbation effects on this state can be significant for distances within 10–100 km of the coast.

The sea breeze is probably the most studied mesoscale circulation. However, in the presence of vertical stability and blocking by coastal orography, it becomes complex and the pressure gradient associated with the horizontal temperature gradient may feed directly into the coastal-parallel wind component. Thus, even in cases when variations in the across-coast wind is small, there is often a significant diurnal cycle in the along-coast wind speed close to the coastline (Beardsley et al. 1987; Winant et al. 1988). In areas where the coastal terrain is lower or where there are gaps in the barrier, the sea breeze may penetrate. Due to the highly three-dimensional flow that arises, such circulations often deviate significantly from the idealized theory. Height-to-length scale ratios are different, multilayer structures occur, and the expected return flow aloft is often not observed (Banta et al. 1993; Banta 1995). Other features include short-term temporal and small-scale spatial variability, with horizontal shear and convergence zones (Tjemström and Grisogono 1996; Grisogono and Tjemström 1996; Svensson and Klemm 1996). The structure of the seaward component of such flows is largely undocumented, even for ideal conditions.

During summer, winds along the U.S. West Coast are normally quasi-steady from the northwest due to the North Pacific subtropical high and a thermal low present over the southwestern United States. Close to the coast this flow is enhanced by coastal baroclinicity and often takes the form of a jet at the top of the MABL, which is capped by a strong inversion (Zamba and Frieh 1997; Beardsley et al. 1987). Due to the diurnal variability of the coastal baroclinicity, the coastal jet has a diurnal cycle with the most pronounced and strongest jet close to the coast during mid- to late afternoon. This flow may reverse locally during the night (C.A. Frieh 1996, personal communication). Winant et al. (1988) showed that when the MABL depth is below the coastal terrain height, the flow can be described as a single-layer reduced-gravity flow bounded by a sidewall. If the flow speed is sufficiently high and/or the MABL is sufficiently shallow, the value of the Froude number exceeds unity, and for such supercritical flow, the shape of the coastline has a profound influence on the flow. When the coastline turns away from the flow, an expansion fan forms, the MABL depth decreases, and the wind speed increases. As the coastline turns into the flow again, the MABL depth grows and the flow speed is reduced; if it becomes subcritical, a hydraulic jump appears with a local increase in turbulence. Winant et al. (1988) report in detail on one case and find a good correspondence to a single-layer hydraulic model. Conditions favorable for such phenomena occur frequently during summer along much of the U.S. West Coast (Dorman and Winant 1995).

Sometimes, however, this pattern of consistent northwesterly flow reverses abruptly to a southerly flow. In cases with a strong transition, the low-level flow may change from northerlies at 1–10 m s\(^{-1}\) to southerlies at about 10 m s\(^{-1}\) in a matter of hours or less (Mass and Albright 1987; Bond et al. 1996). Bond et al. (1996) assembled a composite of the phenomena, showing that the along-coast wind reversal is preceded by a slight drop in surface pressure and followed by increasing pressure, and sometimes the events are accompanied by a drop in temperature. These events are often accompanied by a tongue of stratus clouds or fog, propagating north along the coast and are often seen in satellite imagery. This phenomena, sometimes called a coastally trapped disturbance (CTD), has been studied intensive- ly, but origin, flow structure, and dissipation are not very well understood. They have been described as coastal Kelvin waves (Dorman 1985, 1988), as topographically trapped density currents (Dorman 1987; Mass and Albright 1987), and as a mesoscale response to along-coast pressure gradients forced by the synoptic-scale flow (Mass et al. 1986; Mass and Albright 1987). The most intense debate has been on its relationship to developing synoptic conditions. Mass and Bond (1996) show convincingly that there is a significant anomaly in the synoptic-scale flow preceding and accompanying these events, but their analysis fails to indicate a triggering mechanism. Recent model results by Skamarock et al. (1998) indicate that temporal offshore advection of warm continental air is important. It suppresses the MABL locally and causes an adjustment of the coastal atmosphere that is impeded by the barrier, thus forming a disturbance that propagates north.

Theoretical studies addressing these issues (Reason and Steyn 1992; Klemp et al. 1994; Rogerson and Samelson 1995, 1996) are restricted by lack of observations on an adequate scale, while intensive field experiments are made difficult by the relatively infrequent occurrence of the events—one or two events per month during the summer (Bond et al. 1996). Detailed in situ observations exist only from very few cases. Airborne observations (Thompson and Bane 1998) do not seem to favor any of the suggested theories. In their limited data (four cases), the tongue of the stratus clouds follows...
several hours after the wind reversal. The reversal is strongly heterogeneous in the vertical and occurs aloft several hours earlier than at the surface. The thermal structure of the MABL resembles a multilayer fluid; most of the perturbation remains within the inversion, leaving the structure of the MABL proper relatively unperturbed.

Most theoretical work of the flow along the U.S. West Coast have been performed with simplified models, for example, shallow water-type models, while systematic simulations with more realistic atmospheric models (cf., e.g., Burk and Thompson 1996) are few. Furthermore, such work is by and large based on specific events rather than attempting a more general interpretation of the impact of the coastal terrain on the MABL flow. High-resolution atmospheric models are limited by assumptions made in the formulation of the model. Nevertheless, a realistic three-dimensional (3D), time-dependent, nonlinear primitive equation model, as applied in the present study, may address some of the deficiencies in simpler models and can provide time–history information with high spatiotemporal resolution. Such a model can provide full 3D coverage of all terms in the mean budgets and most of the terms in the turbulence budgets with sufficient accuracy and realism. Model budgets can then be compared to their counterparts derived, for example, from buoy observations during CODE (Samelson and Lentz 1994).

Coastal effects on land-falling synoptic-scale weather systems are not considered in this paper. The focus is on the structure and diurnal variability of along-coast coastal flow during the warm season. Sea breezes will sometimes be considered, as they are intimately linked to the diurnal cycle of the along-coast flow. One difficulty while studying the phenomena summarized above is to determine interactions between different scales of motion and forcing. The aim here is to study these flows with the synoptic variability removed. Several simulations are thus performed for a generic summertime MABL with stationary and homogeneous forcing on the synoptic scale. Surface forcing is also deliberately kept highly simplified, while the terrain is included realistically. In the rest of this text such stationary background conditions are referred to as “stationary background flow” or “quasi-stationary flow.” The word “transitory” is then used meaning a flow that changes its main character, for example, from a northerly to a southerly flow. Note that quasi-stationary or nontransitory flows also vary in time (e.g., diurnally); however, their main property remains the same. We hope to isolate transitions in the flow structure that can be induced without changing the synoptic-scale flow and to see if structures or signals observed in transitory flows, and also in quasi-stationary flows, can be found.

2. The experiment
   a. The numerical model

   The model has previously been applied in a variety of applications including terrain-induced flow (Tjernström 1987a, 1988a, 1989; Enger and Tjernström 1991; Enger 1990a; Enger et al. 1993; Koracin and Enger 1994; Grisogono 1995), coastal flows (Tjernström and Grisogono 1996; Grisogono and Tjernström 1996), dispersion calculations (Enger 1983, 1986, 1990b), marine stratocumulus (Tjernström 1988b; Tjernström and Koracin 1995), and air chemistry (Svensson 1996a, b, 1998; Svensson and Klemm 1998). It has thus been thoroughly examined for a variety of flows and is well documented. Detailed descriptions are found in Tjernström (1987a,b). Shorter descriptions are found in Tjernström (1988a) and Enger (1990a). Also, see Andrén (1990) for a detailed discussion of the turbulence closure.

   The model is a 3D hydrostatic meso-γ-scale model with a consistent higher-order turbulence closure. The vertical coordinate is transformed into a terrain-influenced coordinate system (Pielke 1984). The grid expands toward the lateral boundaries and from the surface toward the top to achieve maximum resolution close to the surface and in the central parts of the domain, while moving lateral boundaries far from the area of interest. The turbulence closure is an improved, consistent version of the level 2.5 closure (Yamada and Mellor 1979) in the hierarchy of closures introduced by Mellor and Yamada (1974). It carries an improved description for the pressure redistribution terms (the “near-wall” correction) and an algorithm to keep second-order moments realizable (Andrén 1990). This closure is a viable compromise between numerical efficiency and realistic treatment of turbulence. The model also includes routines for subgrid-scale condensation and radiation, as well as surface energy balance; these routines are not used here for the sake of simplicity.

b. The numerical experiment

   The terrain was extracted from the approximately 500-m resolution terrain database available at the National Center for Atmospheric Research (NCAR)1 and was averaged to the model grid. Spectral analysis of the terrain data (Young and Pielke 1983) was used to optimize resolution, domain size, and the number of grid points. The model grid was rotated to align the north–south coordinate of the model with the coast. Thus, the zone of maximum resolution in the central domain will coincide with the coast. The domain size was chosen to 400 × 450 × 5 km3. In this domain, the terrain was resolved by 41 × 41 grid points with 30 points in the vertical. The maximum resolution is Δz = Δy = 4 km at the domain center (somewhat inland, south of Monterey Bay) and Δz = 4 m near the surface, expanding log-linearly toward the model top. With this horizontal resolution, the smallest resolvable horizontal scale is

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1 NCAR is supported by the National Science Foundation.
about 8 km; the hydrostatic approximation will thus be invalid only for circulations deeper than the model domain (cf., e.g., Pielke 1984, 33 and 87). An additional restriction, due to the terrain-influenced coordinate system, requires that the terrain slopes $<45^\circ$ (Pielke 1984, 118–121). While the real terrain certainly slopes more locally, the model-resolved terrain came close to this limit only at a few locations. The resulting terrain, as seen by the model, is shown in Fig. 1. The resolution is adequate along most of the central coast, deteriorating north of Monterey Bay and in the Southern California Bight. To the east, the slope of the Sierras was included up to the crest (see section 2c). To the west, the model extends some 150 km west of the coast.

The sea surface temperature was set constant at $15^\circ C$. The temperature of the soil surface was prescribed to vary sinusoidally with a diurnal amplitude of $\pm 10^\circ C$ and an average value of $20^\circ C$ at sea level, decreasing with terrain height by $6 \times 10^{-3} ^\circ C\text{m}^{-1}$. These conditions were transferred onto the lower boundary of the model, matching surface layer and roughness sublayer similarity theory. Similarly, specific humidity was specified at the surface using potential evaporation over the sea and a fraction (15%) of this value over land. Together, this allows coastal land surfaces to be influenced by local advection of moist and cold air from the sea, while allowing for a strong control over the surface forcing.

The pressure gradient force in this type of model is decomposed into a resolved mesoscale part and a part representing the background (synoptic scale) forcing (see Pielke 1984, 124). The model thus requires an assigned background pressure field, as it would in the absence of the resolved terrain. To simplify further, the synoptic-scale flow, specified as a background geostrophic wind, was kept constant in time and space with a magnitude of $7 \text{m s}^{-1}$ (Nelson and Husby 1983) for all simulations, while its direction was varied for eight different simulations. These are referred to as the N, NE, E, SE, S, SW, W, and NW cases, referring to the direction of the background synoptic-scale flow, always in relation to true north.

The initial temperature and humidity profiles were chosen to be similar to a typical summertime MABL profile for this region. In the lowest 500 m, potential temperature and specific humidity were set constant at $\Theta = 10^\circ C$ and $q = 5 \text{g kg}^{-1}$, respectively. This layer was capped by a $15^\circ C$ inversion and aloft $\Theta$ increases.
by about $3 \times 10^{-3}$ K m$^{-1}$, while $q = 1$ g kg$^{-1}$. This profile was used initially in all the simulations, assuming that the inland PBL will develop a realistic structure by itself, given the proper boundary conditions, while initiation of the subsidence inversion capping the MABL takes place over a significantly longer timescale and must be established initially.

These conditions are a simplification of the actual weather along the central coast of California. In particular, the assumption of stationary and homogeneous background flow and the prescribed surface temperature and humidity are crude. The presence of marine stratocumulus is in reality also often a factor but is neglected here. However, the purpose of this study is not to simulate any particular event. The simplifications are introduced to limit the number of degrees of freedom of the model atmosphere. This allows us to study the impact of a real coastline on a realistic coastal MABL for different background flow directions, while including a reasonably realistic thermal contrast to the land, including the diurnal variability. To include all other possible processes would result in an intrinsically large number of simulations, obscuring the motive of this study. The model is thus used more as a “numerical laboratory” than as an atmospheric forecasting tool. One advantage of using a numerical model for this type of study is actually the possibility of having some control over the forcing. The results may thus not coincide with any particular case, but conclusions based on these simplified simulations may nevertheless contribute to an understanding of observed phenomena.

Two technical issues remain to be solved: the minimum size of the model domain to realistically include the most relevant forcing and the initialization procedure. The Sierra Nevada are expected to generate wave motions when the background flow has a component across the mountains. Rogers et al. (1995a,b) show that gravity waves, induced by inland mountains, can modulate the MABL depth and thus cause changes to the wind speed in the MABL. Waves generated by different sources, here by the coastal mountains and by the Sierras, may interfere. Within linear theory, Grisogono et al. (1993) show that the horizontally integrated mountain wave drag is nonlinearly dependent on the ridge-to-ridge separation, besides the dependence on background wind speed, stability, and maximum terrain elevation squared (e.g., Smith 1979). The present study also allows for time-varying lower boundary conditions in a fully nonlinear environment. These will alter the stability, creating a seemingly transient behavior as waves with varying wavelengths are generated, trapped, or released. The slope of the Sierras will also change the stability at the eastern boundary of the domain and induce additional baroclinicity.

The impact of the of the Sierra Nevada on the coastal MABL was investigated with 2D versions of the model for a cross section through Monterey Bay. The background flow was from the northeast, across the Sierras, using the same conditions as in the full 3D simulations. The results are illustrated in Figs. 2 and 3. First, two simulations were carried out: one without the Sierras and one including only the western slope of the Sierras. In the second experiment, the area east of the Sierras was included, and the shape of the mountain crest was varied. To make this possible, the model domain was extended, while maintaining the resolution in the MABL. The model top was lifted to 11 km and the number of grid points was increased, so that the resolution above the MABL was increased by a factor of 2. A sponge layer was also included above 8 km (Grisogono 1995).

The results from the first experiment show differences both in temporal development (not shown) and in spatial structure (Fig. 2). Significant differences in the inland PBL structure were expected. These are seen in the inland wind field, where the homogeneous flow from northeast ($\sim 5$ m s$^{-1}$) is turned to the northwest over the slope. Over the coastal mountains, the effect is smaller. The depth of the inland PBL is reduced in the lee of the Sierras (Burk and Thompson 1996). The nearshore MABL across-shore wind speed is similar with or without the Sierras. Further out, however, the low-level offshore flow is weakened by the presence of the Sierras. The differences are most pronounced in the along-coast MABL wind. Without the Sierras, a local low-level jet, with wind speeds reaching about 10 m s$^{-1}$, forms at the top of the MABL, where the inversion slopes down toward the coast. Including the Sierras increases the depth of the far offshore MABL from about 100 to about 150 m. This generates an additional inversion slope over a larger distance, leading to a broadening of the jet. The wind speed now reaches a maximum of about 14 m s$^{-1}$ in a wide band.

In the second experiment, the result differed insignificantly from the previous results when the Sierras were extended as a plateau (not shown). A mountain-peak-like shape of the Sierras was then introduced and Fig. 3 shows the wind fields at midnight. The flow with a more realistic representation of the Sierras clearly exhibits a different structure, although the differences are less pronounced in the coastal MABL. The most significant difference centers on the lee slope of the Sierras, where a hydraulic jump develops when the crests are included. Even though this is out of the coastal region, the gravity–inertia waves have a large enough amplitude not only to drastically alter the downslope structure at the Sierras but also to gradually modify the coastal and offshore flow. Significant effects on the MABL were observed; the strength of the jet decreases with 30%–50%, the depth of the MABL decreases from 200 to 100 m, the depth of the sea breeze during the day is about 100 m lower than in the control run (not shown), and the inland penetration is different.

It is far from straightforward to initialize a model on the scales considered here, in a region as complex as the California coast. Even with a theory for filtering
gravity wave noise in this multiscale environment, the resolution and variability of the terrain and the resolution of the available observations may prevent success of any such scheme in obtaining a proper balance between the mass and the motion fields. It is uncertain if quasigeostrophic initialization approaches are at all valid on this scale. The initialization procedure utilized is dynamic initialization, a de facto standard for simulations on this scale. The model is given homogeneous balanced initial mass and wind fields, along with assumptions for the PBL. It is then run for some time, while all variables adjust to a quasi balance. No data from the simulation are extracted until this initialization is over. This method is difficult to apply when simulating particular events since much of the initial information pertinent to a particular event may be destroyed in the process. For generic simulations that is not a problem. However, there remains the problem of judging the proper length of the preintegration period. This relates to the time for gravity wave noise, due to initially unbalanced fields, to be advected out of the domain and also to the possible presence of more long-lasting gravity-inertia oscillations. To address these problems, two fully 3D simulations for NW background flow were started at 1800 and 2400 LST. Also, following the analysis by Thompson et al. (1976), the two momentum equations were decomposed into an inertial and a non-inertial part. Thus, $F = (\partial U/\partial t) - fV$ and $G = (\partial V/\partial t)$.
$fU$ were calculated, where $F$ and $G$ may be interpreted as forcing functions including all realistic forcing and acceleration, except the Coriolis acceleration. In this way, the inertial forcing can be separated out as a function of time directly from the model results. Although minor details vary between the two simulations, it is obvious that the MABL coastal jet varies diurnally as a function of the local time regardless of initialization time (not shown). Figure 4 shows the noninertial forcing. In the across-coast momentum, inertial forcing dominates up until about 6–7 h after initialization. Later, the noninertial forcing becomes quasi-steady. The noninertial forcing in the along-coast momentum is smaller as it also becomes quasi-steady but not until after about 10–12 h into the simulation.

In summary, simulations of the near-coast MABL, even close to the ocean surface, are significantly influenced by the Sierra Nevada. The results presented here (Figs. 2 and 3) are worst-case scenarios in as far as the background wind was chosen to come across the mountains and that the times are those with largest differences. Other wind directions and times show smaller effects and, furthermore, wave effects in 3D flow are weaker, by a factor of about 2 (Miranda and James 1992; Nappo and Chimonas 1992). In the interest of reducing the calculations and maintaining the horizontal resolution near the coast we choose to include the Sierras but only as a sloping wall. Also, the results indicate that all the gravity wave oscillations triggered by the initialization have propagated out of the model domain after a dynamic initialization of 12 h.
3. Results

a. Mean conditions

Horizontal low-level (25 m) wind fields are illustrated in Figs. 5–8. The time is 1600 LST (22 h into the simulation), about the time when the observed along-coast northwesterly low-level flow is at its maximum (Beardsley et al. 1987). Note again that the model grid is rotated about 45°, and that when the background wind direction is discussed, it always refers to true north. The NW and SE cases are those where the background wind is aligned with the coast in the plots.

The coastal flow is predominantly along the coast, either from northwest or from southeast. Only the case with background flow from SE is transitory in the whole domain. The results for the most common summertime background flow directions, northwest and north (Beardsley et al. 1987; Mass and Bond 1996), are shown in Fig. 5. In the NW case (Fig. 5a, background flow aligned with this coast), the flow is predominantly along coast from the northwest, but the Ekman turning of the MABL winds induces an onshore component, which is locally strengthened by a sea breeze. The along-coast wind speed has a pronounced maximum close to the coast south of Monterey, southwest of the main Santa Lucia range, where the slope of the coastal terrain is especially prominent. Farther south, toward Point Conception and into the Southern California Bight, the flow speed is again decreasing. The sea breeze has two maxima where the coastal mountains are lower: one in Monterey Bay and one in the Morro Bay area (y = 120–200 km). From Monterey Bay the sea breeze is channeled into the Salinas River Valley, extending south and converging with the other sea breeze. Some of it then penetrates into the San Joaquin Valley through gaps in the El Diablo range. Figure 6 shows the across-coast wind speed component with the maxima at the coast and subsequent maxima farther inland as the sea breeze is channeled through gaps in the coastal mountains. Note also the minimum of onshore wind speed at the coast along the Santa Lucia range (y = 220–290 km). This is where the along-coast flow is instead accelerated southward. The N case (Fig. 5b) reveals a surprisingly different structure compared to the NW case. The flow is more aligned with the coast and the strongest flow is pushed offshore. This case has the strongest northwesterly flow with maximum wind speeds some 2.5 times stronger than the assigned background wind. There is relatively little crossing of the shoreline and no proper sea breeze is present. On closer inspection, the wind turns onshore at the very coastline; however, inland penetration is less than 10 km, except for in the Monterey Bay area. The flow offshore has less structure, and the maxima in the wind speed extends almost through the entire domain, from north down to Point Conception, while slowing down into the Southern California Bight. Right at the coast, there is a local wind speed minimum at Point Sur. This minimum occurs in several simulations and is also indicated in measurements (C. E. Dorman 1997, personal communication).

The remaining two cases with consistent northwesterly along-coast flow, the NE and E cases, are shown in Figs. 7a,b. Here the along-coast variability is significantly larger and the winds are weaker. In both cases there are significant maxima along the coast between the northern Santa Lucia range and Point Conception, although at slightly different position. In the NE case (background flow perpendicular to the coast), the inland winds are very weak from east, but the coastal wind speed maximum at 24 m is greater than 10 m s⁻¹ from the northwest, larger than the assigned background wind speed. For the E case, the inland winds are stronger, but the coastal wind maximum is weaker, about 5 m s⁻¹, and located farther north. In this case, there is also a more pronounced sea breeze, again predominantly into Monterey Bay and the Morro Bay area. In both cases, there is again a minimum in the wind speed along the Point Sur area.

Fig. 4. Contour plots of the noninertial forcing, G and F (m s⁻²), in the coastal zone for (a) the across-coast wind component (×10⁻¹) and for (b) the along-coast wind component (×10⁻¹). Data were extracted from a fully 3D simulation.
The remaining cases with a consistent along-coast flow are the W, SW, and S cases, all with a strong southeasterly coastal flow. They are all similar, much more similar than the different cases with consistent northwesterly flow; only the W case is shown (Fig. 7c). They all have a minimum (7–11 m s\(^{-1}\)) in wind speed around and south of Point Sur and a pronounced maximum into Monterey Bay. A similar but less pronounced maximum is located north (downstream) of the highest terrain at Point Conception. Both maxima are downslope flow phenomena associated with height changes in the terrain along the flow due to the fact that the MABL flow is here confined to a layer of similar depth as the terrain height. These cases also show significant across-coastline flow at the same location as previously described—into Monterey Bay and the Morro Bay area. In the W case, the southeasterly flow reverses into Monterey Bay. The sea breeze turns the low-level flow about 120° from southeasterly to westerly, around the northern Santa Lucia range into the bay and farther into the Salinas River Valley. The strongest winds occur in the SW case, up to 17 m s\(^{-1}\), again about 2.5 times the assigned background wind.

The four cases with consistent northwesterly flow are distinctly different from each other, presumably due to differences in the interaction between the MABL and waves generated by the upstream terrain, as the assigned background offshore flow across the Sierras and coastal mountains is varied. Similar effects do not exist for the different cases with consistent southeasterly flow; while varying the onshore background flow, they are quite similar in structure. The transition between consistent flow up and down the coast as the assigned background flow is changed from west to northwest is abrupt and requires only a 45° change in background flow direction. The corresponding change, while changing the background wind from east to south is much less abrupt and the SE case is transitory. The 25-m flow field for the SE case is shown in Fig. 8 for 0900 and 1600 LST. During the night, morning, and early afternoon, MABL winds are quite weak from the east or the southeast. As time progresses, the southward acceleration associated to the coastal baroclinicity takes over and the flow becomes westerly or northwesterly. This case is much more transient than the other, which can be explained by an inspection of the momentum budget (see section 3c).

The reason for the different flow structures is revealed by the vertical structure of the MABL, illustrated in Figs. 9–11. Figure 9 shows the thermal structure for a
cross-coast cross section from southwest to northeast at 1600 LST, taken south of where the Santa Lucia range is the steepest, for the NW, N, SE, and S cases, respectively. The MABL becomes more shallow than the offshore component in the background flow increases. In the lowest atmosphere, within the MABL and the inversion, the isotherms slope down toward the coast. This generates the baroclinicity (thermal wind) that drives the northwesterly jet. This slope varies with time and is strongest some hours after the inland heating has its maximum due to the cross-coast flow induced by the thermal contrast (Beardsley et al. 1987; Burk and Thompson 1996). It is present regardless if a sea breeze is formed or if the thermal perturbation is confined only to a narrow coastal zone. In all these cases, except for the NW case, the slope of the isotherms down toward the coast extends above the MABL. In contrast, the MABL depth for the W, SW, and S cases is the same, essentially at the initial depth. Here, the slope of the isotherms reverses in the inversion such that above about 600 m, the isotherms slope upward approaching the coast. This causes the thermal wind to reverse, thus the southeasterly jet. This is also true for the NW case in a shallow layer above the inversion (see section 3b). Such “fanning” of the isotherms was also observed in CTDs (Thompson and Bane 1998). The present results show that this structure is consistent also with a quasi-stationary southerly along-coast flow. It suggests that the fanning in a CTD might be a consequence of the flow in the CTD itself and not a signature from the triggering mechanism.

The along-coast winds corresponding to Fig. 9 are shown in Fig. 10. In the three first cases there is an MABL jet from the northwest, strongest for northerly background flow at about 18 m s$^{-1}$. The wind maxima are located in a horizontal band within or at the base of the MABL inversions. The absolute maxima forms downstream of the steepest coastal terrain (Figs. 10a,b), and the jets are always at a lower height than the topography. In contrast, the maximum southeasterly wind for the S case (Fig. 10d) is located above the MABL at approximately crest height over the terrain slope rather than offshore. In this case also, the along-coast wind speed is about 18 m s$^{-1}$. This difference in position is significant and related to the thermal structure in Fig. 9. In all the cases with a down-coast flow, the jet resides in the offshore MABL below the terrain, while in all cases with up-coast flow, the wind speed maximum is located above the MABL, over the slope of the coastal terrain.

Figure 11 shows the along-coast structure of the coastal jet, in a south–north cross section just off the coast, for the NW, NE, SE, and SW cases, again at 1600 LST. For the NW and SW cases, the wind is consistently from northwest and southeast, respectively, throughout the lower atmosphere, while for the two other cases there is a significant vertical wind direction shear. There are several maxima and minima, as the flow responds to differences in the coastline shape and/or topography. For the northwesterly jet, there are two maxima: one along the sloping terrain north of Monterey Bay and a stronger one at the southern Santa Lucia range. The depressed coastal MABL depth forces the flow around the terrain; it is then accelerated where the coastal terrain is steep. The maxima tend to move south when the MABL shrinks, as the offshore component of the background wind increases. For the southeasterly jet (the SW case), wind speed maxima are connected to the downstream slopes of the Sierra Madre (north of Point Conception) and the Santa Lucia range, south of Monterey. and are
Fig. 7. Same as Fig. 5 but for background flow from (a) NE, (b) E, and (c) W.
even stronger somewhat inland. This jet has a wind speed minimum at the location where the northwesterly jet has a maximum. It is higher, at the height the terrain, and a notable downslope acceleration occurs north of the northern Santa Lucia range. The wind speed minima at Point Sur is again present for both flow directions.

The temporal development of the jets in Fig. 11 is illustrated in Fig. 12. The first two cases (Figs. 12a,b) show a rather typical diurnal behavior of the northwesterly flow with a late afternoon maxima (1500–1900 LST) and a morning minima (0600–0900 LST). This corresponds well with observations (Beardsley et al. 1987). The maximum solar heating occurs around midday and the highest inland temperatures a few hours later. As long as the heating continues, the across-coast flow will be accelerated. The maximum in the across-coast flow will thus lag the heating; at the shoreline this occurs between 1300 and 1600 LST. The along-coast wind is always stronger in the coastal zone than maxima as late as 2000 LST occur locally, more frequently as the background flow is turned northeast to southeast. A more typical time for the maximum of the northwest jet is 1600–1800 LST. Even locally the maximum does not occur after 2000 LST. The southeasterly jet shows an opposite diurnal structure (Fig. 12d): a minimum in the late afternoon and a maximum during late morning, as the coastal baroclinicity now opposes the jet. The remaining transient SE case (Fig. 12c) is different. The timing of the wind maximum is variable and not cyclic, occurring as early as 1600 LST until late at night at different locations. At the location for Fig. 12, the flow is weak during the morning but accelerates from northwest from noon into the night; the maximum wind speed occurs as late as early the following morning.

Figure 13 shows the along- and across-coast wind components at about 25 m for four different offshore distances from the coast at 1600 LST. Noteworthy in the northwesterly flow cases is the acceleration of the along-coast wind near the coast south of Monterey and the retardation into the Californian Bight. As the background flow turns from north to southeast, the zone of coastal influence increases, and in the NE, E, and SE cases, the entire model domain is affected. The along-coast wind is always stronger in the coastal zone than

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**Figure 8.** Same as Fig. 5 but for background flow from SE, the transitional case at (a) 0900 and (b) 1600 LST.
Fig. 9. West–east vertical cross section across the coast of the potential temperature (°C) for four different cases at 1600 LST: (a) NW, (b) N, (c) SE, and (d) S. The location of the cross section is south of the highest peak in the Santa Lucia range, south of Monterey, at y ~ 250 km (see Fig. 1 for the exact location).

farther offshore, somewhere along the coast. For the three cases with southeasterly flow, there are pronounced structures in the along-coast wind over the coast associated with the highest terrain, at Point Conception and the northern Santa Lucia range, with significant minima downslope from these points. This is particularly marked for the W case with a slight reversal into Monterey Bay. Also clear is the highly selective sea breeze (Fig. 13b) for the northwesterly jet cases, in particular for the NW and SE cases, where it only penetrates where the terrain is lower. It is less selective for the southeasterly jet cases since the MABL is deeper. The blocking at the highest terrain south of Monterey is an exception. The far offshore flow for the SE case is very weak and the coastal flow is totally dominated by the coastal baroclinicity. The wind speed minimum at or close to the coast in the Point Sur area (south of Monterey, y ~ 250 km), which has also been observed (C. E. Dorman 1996, personal communication), is seen in both components. It is clear that summarizing coastal wind conditions from measurements that are taken at different distances from the coast (slightly inland, at the coast or from buoys) may become misleading.

Figure 14 summarizes information from all eight runs. The data is averaged from three southwest–northeast cross sections (see Fig. 1). Figure 14a shows the maximum strength of the jet inshore and offshore. As the background flow turns from northwest to southeast, the
strength of the northwesterly flow gradually decreases. The strongest jets, at about 2.5 times the background wind speed, occur for background flow from the north and southwest, respectively. Figure 14b compares the height of the near-coast wind speed maximum to the MABL depth (the base of the inversion). First, it is clear that the MABL depth is decreased significantly for background flow with an offshore component (north, northeast, and east), while the MABL depth for cases with up-coast flow (south, southwest, and west) remains relatively constant. The boundary layer is thus in general significantly more shallow for the northwesterly jet cases, gradually more so for the N, NE, and E case than for the NW case. This enhances the likelihood for supercritical flow. In fact, while the NW and E cases are nearly supercritical ($Fr \sim 1$), the remaining cases are clearly supercritical ($Fr \sim 2$) (Fig. 14c). Second, for northwesterly flow the jet is located at the MABL top (associated with the inversion slope), while the southeasterly jet is located well above the boundary layer (Fig. 14b, also see Fig. 10). The northwest jet thus has its maximum offshore and the southeast jet has its maximum over the coastal terrain slope. Figure 14d summarize some sea-breeze characteristics at these three cross sections directly inland from the coast zone. A sea breeze is here defined, somewhat arbitrarily, as a diurnal crossing of the coast by the wind, regardless of inland penetration of the front. Note that the values are averages for three locations and a nonzero value only requires a sea breeze (with this definition) at one of the

Fig. 10. Same as Fig. 9 but for the along-coast wind speed component, $V$ (m s$^{-1}$).
locations. The systems are deeper and stronger for the cases with southeasterly flow. The flow crosses the coast in all simulations, but for the N and NE cases essentially it crosses only at a few selected places with quite weak flow. For example, south of Monterey, where the terrain is high, there is no crossing at all for these cases.

In summary, a change in the background flow by only 45° from the northwest to west causes a 180° shift in the winds, from northwest at about 10 m s\(^{-1}\) to southeast at about 13 m s\(^{-1}\), and changes the typical boundary layer depths from about 50–200 to about 400–500 m. The magnitude of these changes are similar to those observed in connection with coastal CTDs (Bond et al. 1996; Dorman 1987, 1988). Again, these simulations indicate that such changes are consistent also with the
differences between northwesterly and southeasterly flow and also for stationary background flow. They may thus be a consequence of the changing flow direction in the CTD rather than the signature of the causal mechanisms.

b. Local transitions

In addition to the transitory behavior of the SE case, local transitions appear in two more cases, as illustrated in Figs. 15 and 16. It was noted in Fig. 10 that the southeasterly flow is associated with a reversal of the coastal baroclinicity with height. In the NW case (Fig. 10a) there is a slight reversal in a layer from the MABL top up to about 1200 m, but it is not strong enough to
reverse the northwesterly flow. However, for a period during the morning, when the low-level coastal baroclinicity is not sufficiently strong, there is a secondary reversing jet north of Monterey Bay (Fig. 15). It is similar in position and height to the regular southeasterly jets, but weaker. Such jet reversals have been observed, for example, during the Shelf Mixed Layer Experiment (C. A. Friehe 1996, personal communication). This generates a double structure with a down-coast flow in the MABL at about 8 m s^{-1}, temporarily weakened to about 5 m s^{-1}, and an up-coast flow over the coastal terrain slope at 2–4 m s^{-1}. It is clear that this requires a delicate balance between the strength and direction of the background flow and the baroclinicity of the coastal zone, controlled here by a prescribed surface heating. Nevertheless, it occurred here without manipulation, which seems to indicate that it could be a common phenomenon.

The second reversal, occurring for the E case, is shown in Fig. 16. At 0400 LST the flow is predominantly from the northwest, when the local flow in the Californian Bight is starting to reverse. At 0800 LST, an eddy with a width of about 100 km has formed in the bight region, extending almost to Point Conception where it seems to form a wind front, pushing the northerly flow offshore. At noon the reversal halted at Point Conception, and the eddy moved westward and now extends to the western edge of the model domain, while still about 100 km in size. Also noteworthy, is the very strong horizontal wind shear along the coast north of

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**Fig. 12.** Time–height cross sections of the along-coast wind component, $V$ (m s$^{-1}$), for the same cases as in Fig. 11 at a location just outside the central coast. (See Fig. 1 for the exact location.)
Fig. 13. (a) Along-coast and (b) across-coast wind speed components at different distances from the local coastline as a function of south–north distance for all the simulated cases. The model data here is extracted at roughly equal distances from the local coast, rather than along constant y lines in the model. The dotted vertical lines indicate the location of Monterey Bay and Point Conception. The various lines show winds at different distances from the coastline: (solid) an average over ~0–10 km inland, (dashed–dotted) ~20 km out from the coast, (dotted) ~80 km out from the coast, and (dashed) at the western model boundary, 130–170 km out from the coast.
Figure 14. Statistics generated from all the cases, averaging three locations: Monterey Bay, along the southern Santa Lucia range, and at the central coast (see Fig. 1 for the exact locations). The plots show (a) the magnitude maximum of the along-coast wind speed component in the jet close to the coast (solid) and offshore (dashed); (b) the boundary layer depth (dashed) and the height to the jet core (solid) close to the coast; (c) the Froude number close to the coast (solid), offshore (dashed), and for the prescribed background condition (dotted); and (d) the maximum sea-breeze intensity (solid) and depth (dashed) inside the coastline.

Point Conception. Inland in the convective PBL, the background flow is mixed down to the surface, while the offshore MABL is controlled by local forcing. In the afternoon, at 1600 LST, the eddy dissipates and the stronger flow to the north enters the bight. Eddies often form in this area, and the stronger ones are often referred to as Catalina eddies; these occur more seldom and require specific synoptic conditions (Mass and Albright 1989). It is also well known that coastal wind reversals have a tendency to stop or to be delayed at points or capes along the coast, as it does here. This case also is presumably a consequence of a delicate balance. Again the fact that it occurs in a generic simulation, intended to resemble typical conditions, indicates the potential for such reversals/eddies to occur.

c. Average momentum budgets

One way to investigate the physics of the flows available in a model study is a direct examination of relevant budgets. All the budget terms are available within the accuracy of the assumptions made in the model design. The momentum budget equations for the across-coast and along-coast directions can be written as

\[ U_x + UU_x + VU_y + WU_z + P_x + C_U + \Pi_x + D_U = 0 \]

and

\[ V_x + UV_x + VV_y + WV_z + P_y + C_V + \Pi_y + D_V = 0, \]

where \( U, V, \) and \( W \) are the mean wind components; \( P \) represents the large-scale pressure (responsible for the background flow and prescribed here); \( \Pi \) is the mesoscale pressure perturbation; \( D \) is the diffusion of momentum through turbulent friction; and \( C \) is the Coriolis acceleration (note that \( C_U \) is a function of \( V \) and vice versa). A lowercase subscript represents the derivative with respect to that variable. Model results were extracted every hour of the simulation, and budget terms were calculated and then averaged over six time periods, according to the different periods of diurnal maxima and minima in the wind speed. The cases with a persistent northerly and southerly jet, respectively, were
then also averaged together. The transitory SE case was treated separately. Only ocean grid points were used.

Figures 17 and 18 summarize the across- and along-coast momentum budgets for all ocean grid points at $z \approx 25$ m, averaged from 1400 to 1700 LST, while Fig. 19 shows the across-coast momentum budget for the transient SE case. In Fig. 17a, the budget is dominated by $C_U$ and $(P_x + \Pi_x)$ to a first-order approximation. However, the pressure forcing varies substantially more in space than the Coriolis forcing. While most of the points vary within $\pm 20\%$, a few are scattered more ($\pm 50\%-100\%$), more often for the southerly flow than for the northerly flow. Since the background flow is always stationary and homogeneous, this must be due to the mesoscale coastal baroclinicity and terrain variations. There is also a good correlation between $(P_x + \Pi_x)$ and $D_U$, although the magnitude of the latter is a factor of about 4 smaller. The first response to the mesoscale pressure forcing is an increasing across-coast flow, which leads to increasing turbulent drag. Adding $D_U$ and $C_U$ reduces the scatter significantly (Fig. 17c). Ageostrophic advection of across-coast momentum is smaller still, typically a factor of about 2–4 smaller than $D_U$, but it is significant locally along the coast; thus the scatter is reduced significantly in Fig. 17d. The remaining small imbalance is the time tendency. The across-coast flow is thus a quasigeostrophic boundary layer flow; the dominating balance is between $C_U$ and $P_x + \Pi_x$, with a significant modification by turbulent friction ($D_U$). In contrast, the along-coast momentum is
FIG. 16. An example of a weak eddy wind reversal in the Californian Bight at (a) 0400, (b) 0800, (c) 1200, and (d) 1600 LST. The plots show horizontal wind fields, as in Fig. 5, with wind vectors and scalar wind speed (in m s$^{-1}$).

not at all in geostrophic balance; there is no correlation between $(P_x + \Pi_x)$ and $C_v$ (Fig. 18a). The dominating balance is instead between $(P_x + \Pi_x)$ and $D_v$ (Fig. 18b). The inclusion of the Coriolis term contributes to decrease the balance for the southerly jet cases (Fig. 18c) but does not improve the scatter, while the ageostrophic advection reduces the scatter significantly but leaves some imbalance, in particular for the southerly jet. There remains a negative time tendency for both flows, and the northerly and southerly jets are accelerated and decelerated, respectively, during this time period.

The SE case (Fig. 19) is different as there is no correlation at all between $(P_x + \Pi_x)$ and $C_v$ (Fig. 19a). Instead the mesoscale and background pressure gradients, $\Pi_x$ and $P_x$, balance each other to a first-order approximation, except where the local pressure forcing is larger (Fig. 19b). Inclusion of friction, $D_v$, improves the balance marginally (Fig. 19c), while inclusion of ageostrophic advection (Fig. 19d) decreases the imbalance without reducing the scatter significantly. This balance is the reason for the transient behavior in the along-coast wind component for this case, as discussed earlier. The flow results from a balance between the Coriolis force and the small difference between two opposing across-coast pressure gradients. As the mesoscale across-coast pressure gradient varies through the day and is sometimes larger and sometimes smaller than the imposed large-scale gradient, the along-coast geostrophic wind will fluctuate.

Figure 20 shows the across-coast variability in the momentum budgets on an offshore line west of the Santa Lucia range. Budget terms are evaluated at $z \sim 25$ m and averaged between 1400 and 1700 LST. Offshore, the flow behaves as the ageostrophic downgradient theory suggests (Overland 1984; Mass and Albright 1987; Overland and Bond 1993). The winds are primarily along the coast and the flow is semigeostrophic with a balance between $(P_x + \Pi_x)$ and $(C_v + D_v)$ in the across-coast momentum equation. In the along-coast equation the balance is dominated by the pressure gradients and the decelerating effect of the turbulent diffusion, $D_v$. For northerly flow (Fig. 20a), the magnitude of $(P_x + \Pi_x)$ increases within about 20–40 km from the coast. This increase is mainly balanced by increased friction, $D_v$, and ageostrophic acceleration. In the along-coast component, turbulent friction varies less, while the ageostrophic acceleration varies more. This is in contrast to the southerly flow (Fig. 20b), where $C_v$ is bal-
balanced mainly by $D_v$ and (less) by ageostrophic acceleration. The systematic across-coast variability is generally smaller than for the northerly flow. In summary, the presence of the mountain barrier has two effects. On the larger scale, the flow is generally semigeostrophic, modified by surface friction, offshore. Due to local non-stationarity and the nonhomogeneous coastline, ageostrophic perturbations occur within a zone about 20–50 km from the local coast, similar for both northwesterly and southeasterly flow. Here, the local increase in the across-coast pressure gradient ($P_x + \Pi_x$) causes the wind to increase in the coastal zone. The local increase in the along-coast pressure gradient is balanced by advection of momentum, which is manifested as local maxima and minima in the flow along the coast, and by turbulent friction, as a response also to varying wind speed.

The along-coast variability is illustrated in Fig. 21. The budget terms here were averaged over a band about 30–40 km wide (four grid points), immediately off the coast at $z \sim 25$ m, and the result is plotted versus north–south distance. The plots show northerly (Figs. 21a,c) and southerly (Figs. 21b,d) flow for afternoon (Figs. 21a,b) and nighttime (Figs. 21c,d) flow. It is obvious that the along-coast variability is significantly larger than the across-coast variability. Also, it appears that although the variations in pressure forcing along the coast are large, the wind speed (manifested by the Coriolis terms) varies much less. Both the ageostrophic acceleration and turbulent friction are important for the balance. For northerly daytime flow, the across-coast momentum balance is still mostly quasigeostrophic, modified by turbulence. Along the northern Santa Lucia range and northward, in particular in Monterey Bay, the ageostrophic acceleration is 30%–60% of the pressure gradient term. In the along-coast momentum the balance between the pressure gradient and the turbulent friction actually only holds south of Monterey Bay (the location for the cross-coast plots in Fig. 20), where the pressure forcing is large. Elsewhere, ageostrophic acceleration is required to balance the budget. The main maximum in the pressure forcing in the two budgets are roughly collocated. In contrast, for the southerly flow, these maxima do not occur at the same locations along the coast. The across-coast momentum is less in balance, in particular from the Santa Lucia range and north. The acceleration
term is large only locally, while turbulent friction is large from Monterey Bay and north. For much of the central coast south of Monterey, the balance in the along-coast momentum is between ageostrophic acceleration and turbulent friction. In general, the picture is quite complicated. The nighttime results are even more complicated and, in particular, the northerly flow is not in geostrophic balance.

The vertical and temporal variability of the jets are illustrated in the vertical profiles of along-coast momentum budgets outside of the immediate coastal zone but within a Rossby radius of deformation, as shown in Fig. 22. For the northerly flow (Fig. 22a) the on-coast flow, indicated by $C_v$, is increasing during the morning and into the afternoon and decelerating during evening and night. This is intimately coupled to the temporal variability of the jet. The time tendency of the along-coast wind component in the morning, when the flow is offshore, is positive below about 1000 m; the northerly flow decreases. The dominating terms near the surface are $(P_y + P_C)\, D_v$ and $V_v$. As the importance of the turbulence, $D_v$, becomes negligible with height, the Coriolis term becomes more important and the flow is geostrophic. This is consistent with the argument by Zemba and Friehe (1987) that the jet is driven by a thermal wind, increasing with decreasing altitude, which becomes balanced by turbulence in the MABL. This implies that the wind speed must have a local maximum in the vertical. As the sea breeze develops, the across-coast wind component becomes positive or small below about 400 m. The Coriolis acceleration increases, and the MABL time tendency becomes negative in the early afternoon; the southward jet accelerates. In the evening, the across-coast flow again turns offshore, and the northerly jet must weaken in the lowest MABL. During the night, the along-coast momentum budget again becomes similar to that in the morning and the jet continues to decelerate. The along-coast momentum budgets for the southerly jets cases are shown in Fig. 22b. The dominating balance is similar to that for the northerly jets cases. The turbulent layer is thicker, up to about 500 m in the afternoon, since the MABL is deeper, while the flow aloft is essentially geostrophic. In the morning, the across-coast component is offshore below 300 m, while the time tendency is mostly positive below 800 m; thus the jet accelerates. As the sea breeze turns onshore, the time tendency must become negative. In the evening, the decrease of the along-coast component is confined below 100 m—deceleration is mainly through turbulent diffusion—while the jet is essentially stationary. Into the night, ageostrophic acceleration and $C_v$ decrease.
Fig. 19. Same as Fig. 17, but showing the balance between different terms ($\times 10^{-2}$ m s$^{-2}$) in the across-coast momentum budget for the transient SE case only: (a) $[\Pi_i + P_x] \times [-C_U]$, (b) $[\Pi_i] \times [-P_x + C_U]$, (c) $[\Pi_i] \times -[P_x + C_U + D_U]$, and (d) $[\Pi_i] \times -[P_x + C_U + D_U + \text{SUM}_{\text{advection}}]$. 

Again, while the tendency term becomes positive again between about 500 and 1000 m, that is, the jet increases.

To summarize, the across-coast momentum equation is essentially quasigeostrophically modified by MABL turbulence with a balance among the pressure gradients (large scale and/or mesoscale), the turbulent friction, and the Coriolis term, while the along-coast momentum equation is not. The dominating balance in the latter is between the pressure gradient and the turbulent diffusion of momentum or ageostrophic acceleration; this forms a special class of semigeostrophic flow that governs the flow to large distances offshore. This is essentially in agreement with Samelson and Lentz (1994). While analyzing CODE measurements around Point Arena, they also found that this balance dominates. During periods of strong southward flow, they also found a balance between ageostrophic acceleration, associated with the curvature around Point Arena, and the across-coast pressure gradient. Also here, along-coast variations in the pressure gradients are most often balanced by ageostrophic advection rather than by variations in the flow velocity (the Coriolis term). For locations with strong coastline curvature or steep terrain, ageostrophic advection and time tendency become important within about 50 km of the coast. The near-coast variability in the budgets along the coast is in general quite large, and the flow becomes less geostrophic at night, particularly for the northwesterly flow. The presence of the coastline affects the mesoscale pressure gradient through heating of the land and the slope of the terrain; both feed primarily into the across-coast momentum equation. Obviously, the across-coast and along-coast momentum equations are coupled by the Coriolis terms, and the cross-coast variability in the mesoscale pressure gradient, $\Pi_{cc}$, is responsible for the increase in the along-coast wind through geostrophic balance. The jet-shaped flow is formed by a balance between increasing thermal wind and gradually increasing turbulence deceleration while approaching the surface (Zemba and Friehe 1987). As the across-coast mesoscale pressure gradient increases, the geostrophic balance is perturbed and an onshore flow forms. This perturbs the along-coast momentum equation via the Coriolis term, accelerating northerly and decelerating southerly jets, slightly out of phase with the sea-breeze cycle. At the same time, the effect of the flow perturbations also generates along-coast perturbations in the potential temperature field, for example, through along-coast variations in height or slope of the coastal terrain. The resulting imbalance in the along-coast momentum equation seems to be balanced by the
Fig. 20. A plot of the terms in the momentum budgets (m s\(^{-2}\)) at z = 25 m, as a function of offshore distance (the coast is at x = 0) for (a) northerly flow and (b) southerly flow. The location of the cross section is south of the highest peak in the Santa Lucia range, south of Monterey, at y = 250 km (see Fig. 1 for the exact location). The plots show along- and across-coast momentum balance (see the labels in the upper left corner in each subplot). The terms are (1) the sum of the pressure gradients \([P_x + P_y]\), or \([P_y + P_y]\), (2) the Coriolis term \([C_U \text{ or } C_V]\), (3) turbulent friction \([D_U \text{ or } D_V]\), and (x) ageostrophic acceleration.

d. Turbulence structure and forcing on the ocean surface

Although many field experiments include turbulent measurements, these are either performed continuously at only a few sites or by aircraft that cover larger areas but only during a limited time. Provided that the turbulence closure is reliable, temporal and spatial coverage can be obtained from a model. With the closure used here (Andrén 1990), explicit calculation of second-order moments is in principle possible. Figure 23a shows the turbulent kinetic energy (TKE, here a prognostic variable) close to the sea surface for four different simulations (NW, N, NE, and W). Note that the gray scale had to be adjusted, to be able to show these in one plot. The maximum varies from 0.2 m\(^2\) s\(^{-2}\) (NE), through 0.4 m\(^2\) s\(^{-2}\) (NW and W), to 0.8 m\(^2\) s\(^{-2}\) (N). The variability between the cases with a northerly flow is significant. In contrast, all cases with a southerly flow, represented by the W case, are similar. In general, there is some correlation between wind speed and TKE, but the maximum turbulence does not always coincide with the maximum wind speed and the TKE fields are more complex. The variability is determined both by the wind field and by the vertical stability, and wind shear close to the surface becomes more important as the jet is lowered. TKE maxima are found at points and capes along the coast, but larger offshore maxima and minima have other origins. For the NW case, the weak TKE maximum off the central coast is collocated with a wind speed maximum (Fig. 5a) but is less pronounced. In contrast, the TKE minimum along the coast to the north coincides with higher wind speeds. Into Monterey Bay, both wind speed and TKE are large. The maximum along the southwest of the Santa Lucia range in the N case has no correspondence with an extreme in the wind field, while the pronounced maximum northwest of Point Conception in the NE case is collocated with a wind speed maximum (Fig. 7a). For the southerly flow, the main maximum is connected to the downslope flow into Monterey Bay.

For background flow from the eastern sector perturbations in \(\Pi_x\), are displaced seaward by offshore advection of temperature and formation of gravity waves, causing the larger sensitivity in the MABL structure to these flow directions. Dynamically, the northerly and the southerly flows are nevertheless similar in the sense that the same semigeostrophic balance rules, although the causes are different, as discussed in section 3b. While the along-coast wind component changes sign, the diurnal variability is opposite for the northerly and southerly flow, respectively. The transitory flow in the SE case is caused by the balance between the mesoscale and the large-scale pressure gradient. Here, these came to have the same order of magnitude for much of the day, canceling the dynamic forcing. As inland heating increases, the mesoscale pressure forcing dominates and the northerly flow prevails.

along-coast advection terms, causing along-coast variability. For background flow from the eastern sector perturbations in \(\Pi_x\), are displaced seaward by offshore advection of temperature and formation of gravity waves, causing the larger sensitivity in the MABL structure to these flow directions. Dynamically, the northerly and the southerly flows are nevertheless similar in the sense that the same semigeostrophic balance rules, although the causes are different, as discussed in section 3b. While the along-coast wind component changes sign, the diurnal variability is opposite for the northerly and southerly flow, respectively. The transitory flow in the SE case is caused by the balance between the mesoscale and the large-scale pressure gradient. Here, these came to have the same order of magnitude for much of the day, canceling the dynamic forcing. As inland heat-
the stress on the coastal ocean using wind speed measurements, as when using bulk formulations.

One significant effect of spatial variability in the wind stress in the coastal region is its impact on coastal upwelling. Upwelling occurs when the wind stress forcing pushes surface water seaward and warm surface water is replaced by cool water from below. From an atmospheric point of view, this changes the surface (boundary) conditions, affecting both static stability and coastal baroclinicity. One question is therefore if there is a significant feedback on the wind field from the upwelling (Burk and Thomson 1996; Enriques and Friehe 1997). The wind field drives the upwelling, cooling the surface at the coast. This changes the stability and the mesoscale pressure gradient, which then affects the local wind field, etc. Kelly (1985) analyzed measurements of wind, wind stress, and SST along the California coast and conclude that strong upwelling generally require northerly flow and that patterns of sea surface temperature (SST) anomalies are connected to the coastline geometry, while the magnitude of these anomalies depend on wind speed.

A useful indicator for upwelling-favorable conditions is the curl of the wind stress vector, \( \xi = \mathbf{k} \cdot \nabla \tau \) (cf., e.g., Gill 1982), where \( \tau \) is the wind stress vector. Note that \( \xi \) is only a “source function” for upwelling; moreover, it is only truly valid for conditions when the ocean depth is sufficient for a developed Ekman layer. The cold surface water is subsequently transported by ocean circulation processes; \( \xi \) fields can therefore not be compared directly to SST fields. Figure 24 shows \( \xi \) calculated from model data for the cases in Fig. 23. It appears that \( \xi \) is even more sensitive to the background flow field than either wind speed, TKE, or \( u^* \) since it is related to the gradient of the stress (loosely connected to the gradient of the wind). Larger areas of significant upwelling conditions are only indicated for the N and NE cases. However, the NW case shows small areas with weak downwelling close to the coast, although all three cases have northerly flow. The most pronounced upwelling is indicated for the N case, which is about three times stronger than for the NE case, developing a “dipole” pattern. Small local areas with upwelling are also indicated for the W case, in spite of the southerly

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**Fig. 21.** Same as Fig. 20 but showing all the terms in the momentum balance equations \( \times 10^{-3} \text{ m s}^{-2} \) as a function of along-coast distance from south to north for (a), (c) northerly flow, and (b), (d) southerly flow, averaged for (a), (b) 1400–1700 LST, and (c), (d) 2300–0200 LST. The data are averaged over a \(-30\)-km zone from the coast at \(-25\) m.
On the other hand, for the NW case the gradients are very localized at the coast, and the downwelling is associated with the gradient seaward of the wind speed maxima and with the sea breeze in Monterey Bay. At points and capes along the coast, the jet may become locally detached from the coast, thus downstream of such features upwelling-favorable conditions also appear between the jet and the coast. Preliminary sensitivity studies with the model, however, indicate that there is no strong feedback, lowering the SST at locations with indicated upwelling had only marginal effects on the flow (not shown).

4. Discussion and conclusions

A fully nonlinear, primitive equation hydrostatic numerical model is utilized to study coastal flow along the central parts of California. This combines a realistic atmospheric model, including a higher-order turbulence closure, with highly simplified background flow resembling local climatology; local terrain and surface forcing of the model were treated realistically, while the synoptic-scale forcing was constant in time and space. Several simulations with different background wind direction were performed. The motivation is to study the main properties of the local flow during semistationary conditions in order to isolate MABL structures related to the coastal forcing from those that depend on the temporal variations in the large-scale background flow. For these simplified but typical conditions we study the structure of the coastal atmospheric boundary layer, the mean momentum budgets, and the dynamic atmospheric forcing on the coastal ocean. Care was taken in setting up the numerical experiment to minimize the impact from boundary and initial conditions. The general results agree well with experimental studies in this area. All cases feature a jetlike flow, quasi-parallel to the coast, while the coast acts as a partial barrier to the flow. Significant sea breezes occur for some background flow directions, mainly in gaps in the terrain, most predominant in the Monterey Bay area. The timing of the diurnal variations of the jet and the sea breezes are realistic. The analysis of the momentum budgets reveals a semigeostrophic flow with a geostrophic across-coast momentum balance, modified by turbulent friction, and a balance between the pressure gradient and the turbulent diffusion in the along-coast momentum, which is in agreement with earlier studies. Some specific conclusions include the following.

1) Simulations of the local coastal MABL off the California coast are sensitive to the representation of the Sierra Nevada, some 200 km to the east. This is due both to the interaction between the MABL and waves induced by the coastal topography and by the Sierras, for offshore across-mountain flow, and to the advection of warm inland air out over the MABL. This perturbs the structure of the MABL, depressing it so
Fig. 23. Gray-scale plots of (a) TKE (m$^2$ s$^{-2}$) and (b) friction velocity (m s$^{-1}$) for four different cases showing the offshore spatial structure. Gray scaling is different in the different subplots; white is maximum for each case, while black is zero. For TKE, the maxima are NW, 0.4; N, 0.8; NE, 0.2; and W, 0.4. For $u_*$, the maxima are NW, 0.1; N, 0.2; NE, 0.05; and W, 0.2. Inland values are not shown, while the thick solid white line is the model coastline and the dashed white line indicate the terrain.
Fig. 23. (Continued)
FIG. 24. Grayscale plots of the curl of the stress vector for the cases in Fig. 23, indicating the offshore spatial structure of up- and downwelling favorable conditions. Gray scaling is again different. White indicates maximum upwelling favorable conditions, while black indicates maximum downwelling-favorable conditions for each case. Gray is centered at zero. The range in values for the N case is $\pm 3.0 \times 10^{-3}$ m s$^{-2}$, while for the other cases it is $\pm 1.0 \times 10^{-3}$ m s$^{-2}$. 
that it becomes more shallow as the background flow turns from northwest to east over north.

2) The coastal flow is consistently from northwest for background winds from the northwest, north, northeast, and east, and from southeast for background winds from the south, southwest, and west. The switch between these two flows is abrupt as the background wind is turned from northwest to west, while the corresponding transition for a turning from south to southeast to east is less well defined and the SE case is transitory.

3) Both the northwesterly and the southeasterly jets are thermal winds but have distinctly different structures. The northwesterly jet is located at the MABL top and is generated by the slope of the inversion, which is a consequence of the local across-coast flow. The southeasterly jet, in contrast, is located above the slope of the coastal terrain and above the MABL. It is also driven by coastal baroclinicity as the isotherms in and above the MABL inversion slope upward as a result of the general lifting of the air for background flow with an onshore component.

4) The momentum budgets for both the along- and across-coast components are highly variable along the coast. Within a Rossby radius of deformation the flow is in general semigeostrophic, while ageostrophic acceleration is significant locally within about 20–50 km. The across-coast variation in the across-coast mesoscale pressure gradient balanced by the Coriolis force causes the jet shape. Temporary imbalance in this balance drives the across-coast flow. In the along-coast momentum balance, the Coriolis effect from this across-coast flow is balanced by the time tendency, thus the diurnal cycle of the jet. Along-coast variability in the along-coast mesoscale pressure gradient is balanced by advection of momentum; thus, local maxima and minima in the jet strength appear along the coast. Local across-coast advection terms in addition respond to the curvature of the coastline. The conditions hold similarly for both up- and down-coast flow. Although the mechanisms forming the jets are different, they have dynamic similarities.

5) The dominating balance in the across-coast momentum equation for the SE case is between the mesoscale and background pressure gradients. The mesoscale pressure gradient is modulated by the diurnal heating of the land and the net pressure gradient can be of either sense. This makes the flow transient for the southeast background flow. Other local transitions—reversing northwest to southeast wind direction—occur in two cases, even with this stationary background flow. These include an early morning jet reversal along the northern coast for the NW case and a morning-to-noon low-level eddy formation/wind reversal in the Southern California Bight region for the SE case. Observations support the existence of such features.

6) Upwelling-favorable conditions were examined. For all the up-coast flow cases there is generally a weak offshore pattern of downwelling-favorable conditions. For the down-coast flow cases, there are large differences. Only the N and NE cases have significant patterns of upwelling-favorable conditions close to the coast. Of the remaining northerly cases with northerly flow, the NW case has weak downwelling close to the coast, while the E and SE cases show weak upwelling conditions along the central coast only. The pattern of upwelling-favorable conditions is very sensitive to coastline geometry, via the effect of such variability on the wind structure, while the magnitude is sensitive to wind speed.

7) Among the signatures observed in CTDs (when the northwesterly flow is abruptly reversed to southeast and the reversal propagates up the coast) are an increase in the MABL depth within the perturbation, compared to the MABL in front of it, and a “fanning” of the isotherms in the MABL inversion toward the coast, also within the CTD. Both these features are here shown to be inherent signatures of the different flow directions also during semistationary conditions. It is thus possible that these signatures, when observed in CTDs, are the result of the flow reversal rather than being related to its cause.

Acknowledgments. This study was sponsored by the Office of Naval Research, Grants N00014-95-1-0827 and N00014-96-1-0002. Author ZC was also sponsored by the State Education Commission of the People’s Republic of China. Thanks to Dr. David Rogers and Dr. Gunilla Svensson for help and suggestions. Detailed comments by Dr. Clive Dorman are greatly appreciated as are suggestions from anonymous reviewers. This paper was written while MT was visiting the California Institute of Technology, Pasadena, California, on sabbatical and MT is grateful to Professor John Seinfeld for providing the necessary facilities.

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