Adjustment of the Marine Atmospheric Boundary Layer to a Coastal Cape

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ABSTRACT

During summer, significant changes in marine atmospheric boundary layer (MABL) speed and depth occur over small spatial scales (<100 km) downstream from topographic features along the California coast. In June and July 1996, the Coastal Waves 96 project collected observations of such changes at capes with an instrumented aircraft. This paper presents observations from the 7 June flight, when the layer-averaged speed increased 9 m s⁻¹ and depth decreased by 500 m over a 75-km downdraft from Cape Mendocino, accompanied by enhanced surface fluxes and local cloud clearing. The acceleration and thinning are reproduced when the flow is modeled as a shallow transcritical layer of fluid impinging the bends of a coastal wall, leading to the interpretation that they are produced by an expansion fan. Model runs were produced with different coastlines and imposed pressure gradients, with the best match provided by a coastline in which the cape protruded into the flow and forced a response in the subcritical region upstream of the cape. A hydraulic jump was produced at a second bend, near where the aircraft’s lidar observed the MABL height to increase. Light variable winds observed within Shelter Cove were replicated in model flows in which the flow separated from the coastline. Though highly idealized, the shallow-water model provided a satisfactory representation of the main features of the observed flow.

1. Introduction

The West Coast marine atmospheric boundary layer (MABL) has several characteristics that explain its strong response to coastal topography in summer. Along most of the California coast, an inversion layer persists to cap the MABL in summer, representing a 10²–20°C increase in temperature (Neiburger et al. 1961). From 2000 m in depth at Hawaii, the MABL thins to several hundred meters near the coast, where it is shallower than the typical 1000-m height of the coastal mountains. The slope results both from subsidence associated with the Northern Pacific high and a geostrophic response to equatorward winds. Adjacent to the coast, winds are typically strong. Based on 114 yr of ship observations, Nelson (1977) produced a wind stress climatology that exceeded 0.1 Pa along much of California in June. Aircraft studies, for example, Brost et al. (1982) and Lester (1985), have revealed variations on smaller spatial scales. Based on a 14 July 1982 Coastal Ocean Dynamics Experiment (CODE) aircraft survey, Dorman (1985) and Winant et al. (1988) showed layer-averaged wind speeds that increased 7 m s⁻¹ over 20 km downdraft from Pt. Arena, accompanied by a 300-m decrease in MABL depth.

Due to the strength of the inversion and the blocking of flow in the lower troposphere, the MABL is effectively channeled by a sidewall whose bends are coastal headlands. When the flow moves faster than the gravity wave speed, it is said to be supercritical [classical references are Shapiro (1953) or Henderson (1966)]. If the wall opens away from the oncoming supercritical flow, an expansion fan develops to smoothly turn the flow around the bend, accompanied by acceleration and thinning. If the wall bends inward upon the supercritical flow, the flow abruptly thickens and decelerates in a hydraulic jump. These concepts were first applied to the marine atmosphere by Freeman (1950, 1951), who showed that the acceleration and thinning of the MABL as it rounded Peru’s coastline was due to a supercritical
expansion fan. Dorman (1985) and Winant et al. (1988) attributed acceleration and thinning downwind of Pt. Arena to an expansion fan, while a jump was observed at Stewart’s Point. For the same flight, Enriquez and Frieh (1997) showed that a maximum in wind stress (over 6 Pa) and wind stress curl [4 Pa (100 km)] coincided with the jump location, far exceeding the local stress and Grisogono (2000). Before reaching the cape, flight showed strong modification of the MABL at the California in June (Casey and Cornillon 1999). The 7 June flow showed strong modification of the MABL at the cape. Thirty-meter winds accelerated and veered around the cape (Ström et al. 1997), as they did at 100 m (Tjernström and Grisogono 2000). Before reaching the cape, the MABL was fairly deep (about 600 m) and had a strong cross-shore slope accompanied by a narrow wind jet exceeding 25 m s$^{-1}$ (Tjernström and Grisogono 2000). Lidar based estimates of the MABL height (Ström et al. 2001, hereafter STR) showed marked thinning immediately downwind of the cape with a milder slope elsewhere.

The response of the MABL to the cape and other coastline features has been analyzed numerically. Whether the cape was flattened, upstream winds were reduced, or uniform versus spatially varying SST fields were used, an expansion fan at the cape persists in the hydrostatic MIUU model (Tjernström 1999). Across a range of background conditions, an expansion fan and jump were produced by the nonhydrostatic, compressible Coastal Ocean–Atmosphere Mesoscale Prediction System (COAMPS) with a smooth representation of the cape (Burk et al. 1999). Above the inversion, isentropes dapped (rose) over the MABL fan (jump). The flow departed most from upstream conditions when subcritical flow became supercritical at the cape, as was found in Tjernström (1999). Tjernström and Grisogono (2000) reproduced acceleration and thinning at the cape on 7 June and at Pt. Sur for another flight day, using the MIUU model with realistic topography. Fan and jump features were observed at Pt. Sur in aircraft and COAMPS results (Dorman et al. 1999). On the larger scale, fast shallow flow persists from the cape to Pt. Conception, based on a June–July 1996 COAMPS average (Dorman et al. 2000). Local divergence at the cape and other West Coast headlands was produced by COAMPS as initialized with observations in Burk and Thompson (1996). Flow is blocked by the coast to about 100 km off of California (D. Koracin 2000, personal communication) and supercritical to over 400 km from shore in a June 1996 average of the nonhydrostatic Mesoscale Model 5 (MM5) (Koracin and Dorman 2001). These simulations represented topography, stratification, or turbulence in more detail than we will employ here.

In contrast to these studies, we test whether a minimal, shallow-water model captures the July observed features. Despite simplifying thermodynamics, forcing, topography, and background conditions, Tjernström (1999) and Burk et al. (1999) modeled flows that resembled Cape Mendocino observations. Considering the MABL as a shallow channel flow captured Pt. Arena observations in Winant et al. (1988), though agreement improved with the inclusion of friction, rotation, and synoptic-scale forcing (Samelson 1992; Samelson and Lentz 1994). We will also seek a minimal-physics description using a model suited to the observed transcritical conditions. However, it will be found that neglecting a seemingly minor coastal feature, namely the offshore protrusion of Cape Mendocino, markedly changes the flow response and requires altered initial conditions. In the next section, 7 June observations are presented in layer-averaged form and simulated numerically in section 3. Section 4 compares the model results to the observations. The study’s findings are evaluated in section 5. A neutral drag coefficient is calculated from observations in the appendix.

2. Observations from Cape Mendocino on 7 June 1996

a. Data analysis

On 7 June, the C130 aircraft conducted a survey of Cape Mendocino that included of 35 aircraft profiles, 10 dropsonde profiles (6 with reliable winds), and several level-altitude runs including 7 at 30 m (Fig. 1). Data were recorded at 25 Hz so that turbulence analysis could be performed and at 1 Hz for vertical profiling of meteorological variables (Ruth 1998). From these profiles, layer averages of quantities such as speed ($U$) and virtual potential temperature ($\theta_v$) were calculated using the height of the inversion base, $H$, for the depth of the MABL. The inversion base and top were visually selected from each temperature profile. The Froude number for the layer was calculated as,

$$Fr = \frac{U}{c} = \frac{U}{(g'\theta')^{1/2}},$$

(1)

where $c$ is the long wave speed, $g' = g(\Delta\theta/\theta_0)$ is the reduced gravity, and $\Delta\theta$ is the virtual potential temperature difference between the inversion base and top.

A Scanning Aerosol Backscatter Lidar (Morley et al. 1996) allowed the MABL height to be identified as a sharp gradient in backscatter, typically located at the inversion base based on profile/lidar comparisons. Outside of the profile locations, legs were flown above the...
Laboratory radars profiled virtual temperature and atmospheric flow fields away from profile sites. To distinguish profile to which the maps of profile data would agree at the potential variation. The goal was to obtain a background field used for the downward-looking legs, neglecting its spatial variation. The most notable feature in the observed MABL depth field (Fig. 5a) was a drastic thinning downwind of Cape Mendocino. North of the cape, the MABL was 600 m deep. By 75 km downstream, it had dropped to less than 100 m. Farther south, the MABL deepened again to 300 m but did not return to its upstream height. There was a more gradual cross-shore slope of the MABL north of the cape, from 600 m nearshore to over 800 m offshore. The thinning of the MABL is colocated with the location of cloud clearing. In Fig. 5a, an optically thick stratus cloud (lightest shade) cleared (dark) south of the cape where the MABL was thin, attributed to thinning beneath the lifting condensation level. As well, clearing is found downwind of Pt. Arena (38.96°N, -123.74°W), site of the CODE
expansion fan, where the profiler recorded low boundary layer heights (Fig. 3). The cloud image is from the morning; all cloud cleared by the early afternoon when the aircraft surveyed the region. Clearing in the lee of major West Coast capes is found in a June–July 1996 average (Dorman et al. 2000) and is related to MABL divergence in MM5 model results (Koracin and Dorman 2001). Divergence in the lee of the capes was earlier modeled by Burk and Thompson (1996).

2) Wind Speed and Direction

South of the cape, layer-averaged winds accelerate and turn (Fig. 5b). Moderately fast winds (14 m s\(^{-1}\)) north of the cape reached a peak of 23 m s\(^{-1}\) downwind of the cape with slower winds offshore. The gradient in speed between this maximum and light nearshore winds within Shelter Cove was 15 m s\(^{-1}\) over 40 km (Fig. 6a). The model results presented in the next section will show that the nearshore light winds could be due to flow separation. One would then expect Shelter Cove to be filled with stagnant marine-origin air. The 7 June Shelter Cove station data is not available, but over mid-June to mid-October 1996, offshore flow was associated with warm, fairly dry air (Fig. 6b). The temperatures in Shelter Cove on 7 June are lower than expected for offshore flow, but mixing ratios less than 7 g kg\(^{-1}\) were observed. Tjernström and Grisogono (2000), Tjernström (1999), and Ström et al. (2001) have shown that a lee wave over the cape, seen as warm, low-momentum air
3) Froude number

The layer-averaged Froude number (Fig. 5c) was close to critical upstream, increasing to supercritical approaching the cape, and high (Fr = 4) within the expansion fan. Downstream, the Froude number decreased while remaining higher than upstream values. Offshore where the flow was slower and the MABL thicker, its Froude number was subcritical. The lowest Froude numbers were found in Shelter Cove, where the flow becomes subcritical nearshore. The gradient between the Froude number maximum in the fan and the minimum in Shelter Cove was significant. Other studies (Tjernström and Grisogono 2000; Tjernström 1999; STR) have assessed that the observations showed slightly supercritical, rather than slightly subcritical, flow upstream. The many approximations in estimating a layer Froude number for the MABL account for this difference.

3. Modeling study for 7 June MABL flow

a. Model description

To aid in the interpretation of the observations, the flow was modeled numerically as a shallow layer of fluid topped by an inactive deep layer. It is forced by synoptic-scale pressure gradients and slowed by bottom stress at the sea surface, as described in the layer-averaged momentum and continuity equations,

\[ \mathbf{u}_t + \mathbf{u} \cdot \nabla \mathbf{u} + f(k \times \mathbf{u}) = -g' \nabla h - \frac{1}{\rho} \nabla p - \frac{1}{\rho h} \tau_b \]  

\[ h_t + \nabla \cdot (\mathbf{uh}) = 0, \]  

where \( \mathbf{u} = (u, v) \) is the horizontal velocity vector, \( h \) is the depth of the layer, \( \rho \) is the density of the layer, \( g' = g \Delta \rho / \rho \) is the reduced gravity, \( f \) is the Coriolis parameter, \( \tau_b \) is the stress at the bottom of the layer, and \( \nabla p \) is the imposed pressure gradient forcing. The latter is separate from the pressure gradient forcing due to the slope in the layer depth. The density of the MABL was set to \( \rho = 1.25 \text{ kg m}^{-3} \), and the Coriolis parameter to \( f = 10^{-4} \text{ s}^{-1} \). Although the reduced gravity varies spatially in the observed flow (see Fig. 5d), and inversion strength has been shown to vary downwind of coastal caps (Burk et al. 1999), we fix \( g' \) to a median value of 0.3 m s\(^{-2}\) for simplicity but emphasize that this is a strong simplification. The last term in Eq. (2) is the vertical average of the stress divergence, assuming free slip at the layer top.

The bottom stress is parameterized as

\[ \tau_b = \rho C_D |\mathbf{u}| \mathbf{u}, \]

where \( C_D \) is a neutral drag coefficient computed from all low-level flights during Coastal Waves 96 (appendix). The estimated 10-m drag coefficient is given by

\[ 10^3 C_{D_{10}} = \begin{cases} -0.26U_{10} + 2.4, & U_{10} < 6 \text{ m s}^{-1} \\ 0.065U_{10} + 0.44, & U_{10} \geq 6 \text{ m s}^{-1} \end{cases} \]  

where \( U_{10} \) denotes the 10-m wind speed. Since layer-averaged flow fields are used in the model equations, (4) is reformulated in terms of the layer-averaged velocity. During Coastal Waves 96, the observed 10-m wind speed was found to be approximately 75% of the vertically averaged boundary layer wind speed (Edwards and Winant 1999). Therefore we relate \( U_{10} \) to the model wind speed as

\[ U_{10} = 0.75 |\mathbf{u}| \]

giving

\[ 10^3 C_D = (0.75)^2 10^3 C_{D_{10}} \]

\[ = \begin{cases} -0.1097|\mathbf{u}| + 1.350, & |\mathbf{u}| < 6 \text{ m s}^{-1} \\ 0.0274|\mathbf{u}| + 0.247, & |\mathbf{u}| \geq 6 \text{ m s}^{-1} \end{cases} \]
as the drag coefficient expressed in terms of layer-averaged velocity.

The initial height and velocity field are selected from a match to upstream observations and are assumed to be steady. The imposed pressure gradients are found as follows. At initial time (subscript 0) north of the bend (superscript N), there is no cross-shore flow. Cross-shore, the sum of the imposed pressure gradient and that due to the slope of the layer geostrophically balance the alongshore flow. Alongshore, the imposed pressure gradient balances friction:

\[ u_0^N = 0 \]  
\[ g' \frac{\partial h_0^N}{\partial x} = f_0 u_0^N = -\frac{1}{\rho} \frac{\partial p}{\partial x} \]  
\[ 0 = -\frac{1}{\rho} \frac{\partial p}{\partial y} = \frac{C_D}{h_0^N} |u_0^N| v_0^N. \]  (8)

The imposed pressure gradients from this balance differ for each set of initial flow fields. South of the first bend, the initial height and depth fields evolve in response to the pressure gradients and topography.

Initial conditions, and thus imposed pressure gradient, differed between numerical experiments. In the first experiment (run A1), the initial flow north of the bend is uniform (Fig. 7) with

\[ u_0^N = -14 \text{ m s}^{-1} \]  
\[ h_0^N = 700 \text{ m}. \]  (9)  (10)

According to Eqs. (6)–(8), this requires the constant pressure gradients

\[ \frac{\partial p}{\partial x} = -0.175 \times 10^{-2} \text{ kPa km}^{-1} \]  (11)  
\[ \frac{\partial p}{\partial y} = 0.022071 \times 10^{-2} \text{ kPa km}^{-1} \]  (12)

to balance the upstream flow (Fig. 7). The initial upstream flow is close to critical (Fr = 0.97).

In the second set (runs A2 and B2), the initial upstream layer thickness slopes downward toward the coast (Fig. 7), which is more representative of the observations. The layer thickness increases from \[ h_0^N = 600 \text{ m} \] at the coast to \[ h_0^N = 800 \text{ m} \] at an offshore distance of 250 km. We specify a constant alongshore pressure gradient and calculate an alongshore velocity that satisfies Eq. (8). In this case, \[ \frac{\partial p}{\partial y} = 0.021875 \times 10^{-2} \text{ kPa km}^{-1} \] results in faster \( u_0^N \) offshore (14.5 m s\(^{-1}\)) than at the coast (13 m s\(^{-1}\)), in fair agreement with the observations. Again, the flow is initially subcritical upstream. The cross-shore pressure gradient required to balance the initial upstream flow is computed from Eq.
\(\frac{\partial p}{\partial x}\) varies in the cross-shore direction and is the same order of magnitude as that derived from the height of 850-mb surface produced by the 8 June 0000 UTC Eta model analysis (not shown).

Two coastline geometries are used, one in which the cape is a simple bend and one in which it protrudes into the flow (Fig. 8). The observed nearshore layer thickness upstream of the cape is on the order of 600 m. Coastline A approximates the topography at this elevation as a simple bend. In contrast, the cape in coastline B, which roughly represents 500-m topography, protrudes to the west. A conformal mapping program discretizes the domain onto an orthogonal curvilinear grid (Fig. 8) using 100 × 160 grid points (Rogerson 1999b). The minimum grid spacing is located near the bends in the coastline, and is on the order of \((\Delta x, \Delta y) = (0.75, 1.5)\) km in the cross-shore and alongshore directions, respectively. Far offshore the grid spacing is on the order of \((\Delta x, \Delta y) = (6, 3)\) km. The boundary conditions on the domain are no normal flow along the free-slip coastal wall and open radiation conditions at its the north and south edges. A free-slip wall at the western edge of the computational domain is sufficiently far offshore that it has minimal effect.

We anticipate that the model flows will be transcritical, permitting hydraulic jumps, which destabilize a traditional centered difference scheme that does not include strong numerical and/or physical dissipation. Al-

![Fig. 5. (a) MABL depth from profiles, sondes, lidar. Higher cloud albedo values are lighter in GOES image from four hours before survey. (b) Layer-averaged speed (solid lines near profiles/sondes, fading to SSM/I estimate). (c) Froude number from profiles, fading to estimate based on lidar and SSM/I. (d) Reduced gravity.](http://journals.ametsoc.org/doi/pdf/10.1175/1520-0469(2001)058<1511:AOTMAB>2.0.CO;2)
Fig. 6. (a) Shelter Cove wind vectors recorded at 25–50 m aircraft elevation. Temperature (°C, upper number) and mixing ratio (g kg\(^{-1}\), lower number) are labeled. (b) Mixing ratio (solid) and air temperature (dash) vs wind direction (mathematical convention) from Shelter Cove station from 8 Jun to 14 Oct 1996, the longest available time series, averaged within 5-degree bins.

Fig. 7. (a) Initial MABL height along 41.2 N for cross-shore-uniform (solid) and cross-shore-varying upstream conditions (dash), and after spinup of run B2 (dot). The Rossby radius (Ro) based on run B2 offshore height is labeled. (b) Speed. The cross-shore (c) and alongshore (d) components of the pressure gradients which drove the model. The two lines in (d) are on top of each other.
ternatively, the finite-difference shock-capturing Essentially Non-Oscillatory (ENO) scheme [Shu and Osher (1988) and Shu and Osher (1989)] produces highly accurate, numerically stable solutions containing jumps through use of adaptive stencils. In calculating spatial gradients, the stencils are shifted away from regions where discontinuities are present. Developed for gas dynamics problems, the ENO scheme was employed in the shallow-water model of Rogerson (1999a) for a recent study of the MABL. The same model is used in the current study. A complete description of its formulation and implementation can be found in Rogerson (1999a) and Rogerson (1999b).

### b. Model results

With the numerical model, we can test the effect of varying background conditions and coastline geometry. Both a straight and a protruding representation of Cape Mendocino are tested (Fig. 8). As well, uniform and cross-shore-varying initial conditions are tested (Fig. 7). Table 1 summarizes the model runs.

We first consider the flow past the nonprotruding cape under uniform upstream initial conditions, run A1 (Fig. 9). Immediately downstream from the cape (the northernmost bend in the coastline), the supercritical expansion fan is seen as a region in which the layer thickness decreases from its upstream value of $h = 700$ m to less than $200$ m while the winds increase from $v = 14$ m s$^{-1}$ to more than $21$ m s$^{-1}$. Emanating from the second, concave bend in the coastline is a jump in which the flow rapidly decelerates while the layer thickness increases. South of the jump near the coast, there is a small region of the flow that is subcritical in which the winds diminish to less than $13$ m s$^{-1}$. During the nonlinear adjustment of the model flow from the initial conditions to steady state (not shown), this nearshore region of subcriticality is somewhat larger and connected to the second and third bends. The jump is classified as strong since the flow becomes subcritical downwind of it. It is attached to the wall at a larger oblique angle than if the jump were weak and the flow remained supercritical downwind [e.g., Ippen 1951]. As the flow adjusts, the jump detaches from the vertex of the bend and moves slightly upstream, first noted in transonic flows in compressible gasdynamics [e.g., Liepmann and Roshko (1957)] and in prior transcritical simulations of the MABL (Burk et al. 1999). As the flow adjusts to the steady state shown in Fig. 9, the northern transition to supercritical flow remains slightly upstream of the first bend while the subcritical region south of the jump shrinks. At the third (southernmost) bend, a second weaker expansion fan is present.

For the same coastline but with initial conditions that vary more realistically cross-shore (run A2, Fig. 10), the region upstream of the cape resembles the initial flow fields (dashed lines in Fig. 7). The layer thickness increases offshore from $h = 600$ m to $h = 800$ m and the magnitude of the alongshore velocity increases from...
\( v \approx 13 \text{ m s}^{-1} \) to \( v \approx 14.5 \text{ m s}^{-1} \). Near the cape, the flow becomes supercritical. Expansion fans at the first and third (convex) bends and a jump at the second (concave) bend are again evident within a supercritical region, which extends to the western and southern edges of the domain. Near the coast, the cross-shore pressure gradient forcing is weaker in run A2 than A1, resulting in a weaker response, which is most noticeable in the wind speed. Because the nonlinear interaction with the topography dominates the model response, however, the differences between A2 and A1 downstream of the cape are not dramatic.

Previous work indicates how varying the friction, rotation, imposed pressure gradient, and inversion strength would alter these solutions. In modeling the MABL, Samelson (1992) found that including surface friction produced a “bull’s eye” shaped expansion fan at a coastline bend, limiting the downstream extent of the accelerated winds. Rotation weakly intensified the expansion fan by forcing alongshore flow offshore. The angle between a jump at a second, concave bend and the downstream coastline decreased with larger upstream Froude number or smaller coastline bend angle, as expected from classic gas dynamics. A weaker expansion fan resulted if the imposed pressure gradient was more aligned with the coast downstream of the bend, pushing the flow around the bend. Rogerson (1999a) varied the upstream Froude number, or equivalently imposed pressure gradient, in a model of transcritical MABL flow. Unless the initial flow was quite subcritical, locally supercritical regions formed at coastline bends that produced expansion fans and jumps. As the upstream Froude number increased, the extent of the supercritical regions grew. Based on the study, for nearly supercritical flows and moderately large coastline angles, we expect supercritical expansion fans. The effect of changing inversion strength would be more complicated. If MABL depth and speed are unchanged, weakening the inversion would increase the Froude number by lowering the denominator of Eq. (1), as was done in Burk et al. (1999). However, Tjernström (1999) found that weakening the inversion actually lowered the Froude number slightly.
because vertical mixing and thus MABL depth increased.

We now consider the effect of altering the shape of the cape so that it protrudes into the flow, rather than a simple bend (Fig. 8). Run B2 (Fig. 11) has the same cross-shore-varying initial conditions as A2 (dashed lines in Fig. 7). However, the protruding cape changes the solution significantly, especially upstream. The protruding cape produces a similar effect to that of a bump on the floor of a channel (Baines 1995), which partially blocks the flow so that it deepens upstream. As the model flow spins up, a jump forms north of the cape, which partially blocks the flow. The initial and steady-state depth and speed at 41.2°N for run B2 can be compared in Figs. 7a and 7b. Even a barely protruding representation of the cape (not shown) showed these effects and differed strongly from the straight case.

The supercritical flow south of the cape in run B2 supports an expansion fan at the first and third (convex) bends and a jump at the second (concave) bend, as before. However, the larger effective bend angle of the cape produces a new feature. Within the second, concave bend corresponding to Shelter Cove (Fig. 12), the MABL is very thin (about 50 m) but so slow (<5 m s⁻¹) that the flow becomes subcritical (Fig. 12c). Within this subcritical region, wind direction is light and variable, suggesting it is a region of flow separation. As it turns the cape via the expansion fan, the flow becomes increasingly fast and thin. The bend angle is so large that MABL thins to nearly zero depth; the flow can turn no more, and it separates from the coast, as described in Ippen (1951) for channel flow. Inshore of the line of separation, a stagnant region is left within the model representation of Shelter Cove. The MABL
depth never reaches zero because surface friction becomes very large at the thinnest, fastest edge of the expansion fan.

4. Model–data comparison

The previous parameter study demonstrated that a protruding cape alters conditions upstream, requiring cross-shore-varying initial conditions to compensate. We now assess how closely the model captured the observed flow fields, using the most realistic run, B2. In the point-by-point comparison of model to observed MABL depth (Fig. 13a) they are within about 100–200 m of each other. Overall, the model was too deep, especially within Shelter Cove and the deepened flow upstream of the cape. While no aircraft observations exist within the smaller expansion fan modeled at the third, convex bend, a stratus cloud cleared locally downwind of Pt. Arena (about 39°N, −124°W in Fig. 5a), suggesting that the MABL there was shallower than the lifting condensation level.

There is limited observational support for the model jump emanating from the second, concave bend. At its sharpest point, 14 km from the model coastline, the model jump represented a height increase of 126 m and a speed decrease of 3 m s⁻¹ over 9 km. The aircraft survey did not cross the model jump location. However, 20 km south of the model jump, the MABL increased 90 m over 6 km as seen in the lidar data (Fig. 14a). Where the lidar track would have cut across the model jump, the model height increased 59 m and speed decreased 2 m s⁻¹ over 14 km (Fig. 14c). Collocated velocity observations were not available along the track, but SSM/I speeds (Fig. 14b) were available 40 km away from the lidar track. The layer speed estimate based on the SSM/I data decreased about 1 m s⁻¹ over a horizontal distance similar to the model jump. The location of these features can be compared in Fig. 14d. As in the observations, slower model winds upstream accelerated strongly downwind of Cape Mendocino, then slowed equatorward. Outside of the expansion fan,
the model wind magnitudes were typically within 2 m s\(^{-1}\) of the observed speeds (Fig. 13b) and slightly too slow, most markedly at the center of the expansion fan. In Shelter Cove, where both observed and model winds were light, the model speed exceeded the observations, perhaps due to the fact that the collocated observations were taken closer to the real shore than were the model points, since the model wall aligns with midlevel topography rather than the shoreline. Model winds were similar in orientation to the observed winds; both were dominated by turning through the expansion fan.

Both observations and model show nearly critical flow becoming supercritical at Cape Mendocino and remaining so to the edge of the domain. A comparison of the observations with the model Froude number at gridpoints within 5 km of the observation locations (Fig. 13c) shows that the model Froude number was typically within 0.5 of the observed Froude number and slightly too low, except in the expansion fan where the model Froude number more significantly underestimated the observations. As well, in Shelter Cove the model Froude numbers are notably higher than the observed Froude number at collocated grid points.

5. Discussion

Aircraft observations on 7 June 1996 showed strong modification of the MABL on the scale of local coastline features. Layer-averaged speed increased 10 m s\(^{-1}\) while depth decreased 500 m downwind of Cape Mendocino, where a marine stratus cloud had previously undergone local clearing. The acceleration and thinning were consistent with a supercritical expansion fan that developed to turn the flow around the cape but could not negotiate.
the entire bend, leaving light, variable winds within Shelter Cove. The temperature in Shelter Cove indicated a marine origin for the air, but this was not confirmed by its mixing ratios. The gradient between weak Shelter Cove winds and strong winds within the adjacent expansion fan was pronounced (15 m s\(^{-1}\) over 40 km). Though these changes take place over short distances, they are important for air–sea exchanges. As found for Pt. Arena in CODE (Enriquez and Friehe 1995), strong winds in the expansion fan resulted in high values of stress at the surface (Fig. 15a). In summer, the coldest SSTs along the West Coast are found at Cape Mendocino due to strong wind-driven upwelling (Casey and Cornillon 1999). For the 7 June case, Tjernström and Grisoni (2000) found that model wind stress curl corresponded well with SST. Other fluxes respond to the cold SST and strong winds at the cape. The spatial distribution of surface buoyancy flux on 7 June is similar to that of SST (Fig. 15b). Heat loss and moisture fluxes were also enhanced within the expansion fan (Figs. 15c,d). Finally, we note that the transcritical flow observed on 7 June may not be atypical of flow near West Coast capes in summer. In Dorman et al. (2000), the authors show that on 12 and 29 June and 1 July the MABL was transcritical, as it was in a June–July 1996 Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS) model average. However, the 7 June flow is nearly critical upstream. Based on the results of Burk et al. (1999), we would expect a stronger response if lower-Froude number flow had become supercritical at the cape.

The 7 June observations were simulated with a transcritical shallow-water model with upstream conditions based on observations. Many cases were tested before the three shown in this study were chosen; all with nearly critical Froude numbers upstream produced a supercritical region marked by a major expansion fan at the cape and a jump. Similarly, Tjernström (1999) and Burk et al. (1999) found that these features were persistent across a range of forcing for the Cape Mendocino coastline. The coastline and initial conditions, or equiv-
alently pressure forcing, were varied in a parameter study. When the modeled cape was made to protrude into the oncoming flow, MABL depth upstream thickened from its initial state, similar to transcritical flow over a bump. In this way, the cape exerts upstream influence on the flow. Some of the model’s offset from the observations is attributed to the resulting difficulty in selecting initial conditions based on observations along an upstream line. The best match with the observations (run B2) required cross-shore-varying initial conditions to compensate for the effect of the protruding cape. The model captured the major speed and depth features of the observed flow: a fast, shallow region downstream of the cape corresponding to an expansion fan; and stagnant flow in Shelter Cove, due in the model to flow separation. A modeled hydraulic jump was supported by nearby, though not collocated, data. The observed increase in depth was offset from the model jump location by 20 km, perhaps due to differences in the real and model coastline.

Different cross-shore length scales appeared to be at work in the sub- and supercritical regions. The strongest upstream departure from the initial depth (Fig. 7a) and speed (Fig. 7b) was found within the Rossby radius based on offshore depth, around 170 km, while the cross-shore width of the supercritical region exceeded the 250-km width of the domain. For subcritical flow around a bend, Burk et al. (1999) found that the cross-shore balance remained nearly geostrophic, suggesting the Rossby radius would apply, while within the supercritical expansion fan, it does not (Burk et al. 1999; Tjernström and Grisogono 2000; Samelson and Lentz 1994). On a longer timescale, D. Koracin (2000, personal communication) noted that the supercritical region

Fig. 14. (a) Lidar backscatter, MABL height. (b) Speed estimate from SSM/I. (c) Height (solid), speed (dash) from model run B2 along section approximating lidar track. (d) Cape Mendocino, model coastline B. Locations where height increase exceeds 5 m per km are stippled to indicate jump location. Maximum MABL height from (a) = solid circle; lidar track = dash; SSM/I observations = open circles.
in a June 1996 MM5 average exceeded the local Rossby radius, also true for a June–July 1996 COAMPS average (Dorman et al. 2000).

For insight into how our results might change with continuous stratification and vertical mixing, we consult previous studies. Flow above the MABL responded similarly to the underlying expansion fan and jump in the continuously stratified simulations of Tjernström and Grisogono (2000) and Burk et al. (1999). A lee wave excited by the terrain of the cape, simulated by Tjernström and Grisogono (2000), would enhance slow, shallow flow within Shelter Cove. The turbulent response of the MABL to changes in surface forcing may be small relative to the topographic response. In a model that permitted vertical mixing, Tjernström and Grisogono (2000) found only marginal differences between runs with uniform and spatially varying SST. In conjunction with such studies, the shallow-layer model has given a first-order explanation of observed features of the wind and depth fields on the 7 June flight.
Fig. A1. Neutral drag coefficient \( C_{dn} \) vs 10-m wind speed \( U_{10} \) based on all Coastal Waves 96 low-altitude, high-rate data. Line represents fit to 1 m s\(^{-1} \) averages (circles), 1 standard deviation shown.

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APPENDIX

The Neutral Drag Coefficient

Drag over the coastal ocean might be expected to exceed that over the open ocean due to shorter fetch or differences in wave state (Smith et al. 1992; Davidson et al. 1992), while long fetches experienced by coast-channeled winds might produce a drag coefficient resembling open-ocean conditions (Enriquez and Friehe 1997). To determine which was the case, the neutral drag coefficient was calculated following Enriquez and Friehe (1997).

Fluxes of momentum and buoyancy \( \left( \langle u'w' \rangle, \langle v'w' \rangle, \theta'w' \right) \) were directly calculated using the 25-Hz data recorded during all 30-m runs of Coastal Waves 96. A 2-min averaging interval was selected based on ogive curves (cumulative sum of contribution to fluxes by frequency). The Monin-Obukhov surface layer scaling scheme (Monin and Obukhov 1957) scales velocity as

\[
\begin{align*}
\frac{\langle u'w' \rangle}{\langle v'w' \rangle} = \frac{1}{2} \frac{u}{k},
\end{align*}
\]

where the fluctuating horizontal velocity components are \( u' \) and \( v' \), and \( w' \) is the fluctuating vertical velocity. Length is scaled as

\[
L = \left( -\frac{u}{k} \right) (kgw' \theta'),
\]

where \( \theta' \) is the mean virtual potential temperature, \( g \) is gravity, and \( k \) is the von Kármán constant, here set to be 0.4. These scales were used to adjust the measured winds \( (u_{obs}) \) at the aircraft altitude \( (z_{obs}) \) to 10 m:

\[
U_{10} = u_{obs} + \frac{u_{obs}}{k} \log \left( \frac{10}{z_{obs}} \right) - \Psi_m \left( \frac{10}{L} \right) + \Psi_n \left( \frac{z_{obs}}{L} \right),
\]

(A1)

For stable conditions \( (L > 0) \), the stability function \( \Psi_m \) is

\[
\Psi_m = -8.1 \frac{z}{L}.
\]

(A2)

For unstable conditions \( (L < 0) \), \( \Psi_m \) is

\[
\Psi_m = 2 \log \left( \frac{1 + \Phi^{-1}}{2} \right) + 2 \log \left( \frac{1 + \Phi^{-2}}{2} \right) - 2 \arctan \Phi^{-1} + \frac{\pi}{2},
\]

(A3)

where

\[
\Phi = \left( \frac{1 - 15z}{L} \right)^{-1/4}.
\]

(A4)

The assumption of constant fluxes between 30 and 10 m is a strong one since Zemba and Friehe (1987) showed large vertical gradients for data taken above 33 m, in the same geographic region. The winds at 10 m were used to compute the drag coefficient

\[
C_{dn10} = \left( \frac{u_{obs}}{U_{10}} \right)^2.
\]

(A5)

The drag coefficient was adjusted to neutral by iterating

\[
C_{dn10} = C_{dn10}^{-1/2} + 1 \frac{1}{k} \Psi \left( \frac{10}{L} \right) - \frac{1}{k} \log \left( \frac{C_{dn}}{C_{dn10}} \right)^2.
\]

(A6)

Averaging within 1 m s\(^{-1} \) bins, a line fit above and below the minimum at low speeds (Fig. A1), attributed to a change in sea state (Yelland and Taylor 1996) or a change from smooth to rough flow (Liu et al. 1979), gives

\[
C_{dn10} \approx \begin{cases} -0.26U_{10} + 2.4, & U_{10} < 6 \text{ m s}^{-1} \\ 0.065U_{10} + 0.44, & U_{10} \geq 6 \text{ m s}^{-1} \end{cases}
\]

(A7)

which is similar to that found by Enriquez and Friehe (1997) for data taken at Pt. Arena, except for the increase at low speeds, which has been attributed to a change in sea state (Yelland and Taylor 1996) or a change from smooth to rough flow (Liu et al. 1979).
The significant scatter may be due to dependence on such parameters as wave state. It is concluded that a drag coefficient that resembles those from open ocean conditions (Large and Pond 1981; Smith 1980) is appropriate within this coastal region.

REFERENCES


