Effects of Coastal Orography on Landfalling Cold Fronts.  
Part II: Effects of Surface Friction

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ABSTRACT

The role of surface friction in modifying cold fronts as they make landfall in regions of steep coastal orography is examined by means of idealized simulations. Both the effects of the surface-roughness change at the coast in the absence of orography and the effects of orography are considered. Flow over a large and abrupt change in surface roughness generates an inertia–gravity wave above the boundary layer with characteristics similar to that associated with flow over a plateau. Deceleration of the cross-coast flow occurs over land, as well as for a short distance upstream, and causes retardation of frontal motion. A prescribed northerly postfrontal jet weakens rapidly after landfall. Maximum vertical motions are several centimeters per second; however, only small rainfall enhancement is expected since the updraft is very narrow and produces only small vertical displacements.

Friction modifies the flow over the orography by increasing the upstream flow deceleration and reducing the magnitude of the barrier jet. The reduction of the barrier-jet strength (when compared to inviscid simulations) by surface friction becomes more pronounced as the mountain forcing of the jet increases. With surface friction, frontal-motion retardation by the orography is strong and upstream frontogenesis is enhanced. The frontal updraft is strongest at the coast and remains strong for a short distance inland along the lower portion of the windward slope. The coastal enhancement of the frontal updraft results from the combined effects of the orography and the surface-roughness change, but large parcel displacements are due mainly to the orographic forcing. Along-coast winds in the coastal zone during frontal passage are approximately determined by a superposition of the southerly barrier jet and the frontal jets.

1. Introduction

Landfalling fronts contribute to most of the severe weather (e.g., high winds, floods) along the western coast of North America. Observational and numerical modeling case studies of landfalling fronts have demonstrated important effects of orography on coastal weather, including enhanced precipitation and winds (Parish 1982; Marwitz 1987; Overland and Bond 1993 1995; Doyle 1997) and the retardation of frontal motion (Braun et al. 1997; Doyle 1997). In nature, the evolution of fronts as they move over coastal orography is controlled by the complex interaction of many physical processes. Such complexity can impede understanding of the basic governing dynamics. Idealized models can provide the basic building blocks to a more complete understanding of the phenomenon. For the idealized model presented herein, the important physical processes of interest are those associated with the orography and with the change in surface roughness at the coast.

Studies by Echols and Wagner (1972), Rao et al. (1974), and Schwiesow and Lawrence (1982) have examined the growth of an internal boundary layer from a discontinuity in surface roughness, but they focused on small-scale adjustments of the wind field and neglected Coriolis effects. Taylor (1969) considered Coriolis effects and examined horizontal scales ranging up to larger (~1000 km) scales. His results indicated that the adjustment of the wind speed occurred much sooner than that of the wind direction and that damped oscillations in the wind speed occurred far downstream of the surface-roughness discontinuity as a result of the Coriolis force. Roeloffzen et al. (1986) determined secondary circulations associated with neutrally stratified flow over a coast (surface-roughness discontinuity). Considerable variability of the secondary circulation with direction of the ambient flow was found, with southwesterly to south-southwesterly flow generally resulting in the strongest confluence and vertical motions near the coast.

Previous studies have addressed boundary layer ef-
fects and mountain effects on fronts, but not simultaneously. Eliassen (1959) demonstrated analytically that surface friction can lead to frontogenesis and ascending motion near the region of a surface front. The ascending motion is produced by the convergence of the surface stress induced by the cyclonic shear in the frontal zone. However, Eliassen (1966) pointed out that the upper part of the circulation associated with Ekman pumping is frontolytic so that the net frontogenetic effect of friction would be much weaker than he previously proposed. The results of Roeloffzen et al. (1986) and Økland (1990) suggest that a surface-roughness discontinuity at a coastline can contribute to frontogenesis, particularly when the flow is directed nearly parallel to the coast and the progression of the front across the coast is slow. The numerical models of Keyser and Anthes (1982) and Dunst and Rhodin (1990) showed that friction can produce frontogenesis as frictional convergence drives the temperature gradient toward smaller scales, followed by a period of steady or weakening temperature gradient.

Blumen and Gross (1987) and Zehnder and Bannon (1988) examined mountain effects by considering generally broad, symmetrically shaped mountains with relatively gentle slopes. These studies, which assumed seomigeostrophic balance, demonstrated that the ageostrophic cross-mountain circulation tends to be characterized by weak upstream deceleration near the base of the windward slope, strong acceleration (deceleration) along the upper portion of the windward (lee) slopes, and weak acceleration near the base of the lee slope. The implied divergence pattern contributes to weak frontogenesis upstream near the base of the windward slope, stronger frontolysis (frontogenesis) over the windward (lee) slopes, and weak frontolysis near the base of the lee slope. Tilting of the isentropes over the mountain also contributes to frontolysis (frontogenesis) over the windward (lee) slopes. The presence of mountain waves over the terrain produces only minor changes to this general distribution of frontogenesis (Keuler et al. 1992; Williams et al. 1992). These previous studies were generally restricted to conditions in which the retardation of fronts along the windward slopes tended to be weak. Braun et al. (1999, hereafter BRK) showed that the steep slopes of the coastal terrain along the U.S. west coast (hereafter referred to as the West Coast) produce stronger retardation of frontal motion and upstream frontogenesis than documented in previous modeling studies. The purpose of the present study is to examine the simultaneous effects of surface friction and steep orography on landfalling frontal systems.

Frontogenesis produced by a surface roughness change or topography can result in a secondary circulation about the front. However, Zehnder and Bannon’s (1988) results suggest that the mountain circulation tends to dominate the secondary circulation associated with the front in the region of orography. Furthermore, BRK suggested that for steeper terrain, the mountain circulation dominates that of the front and that changes in the frontal circulation associated with mountain-induced frontogenesis are much smaller than the changes in the mountain circulation caused by the stability perturbations associated with the front. Therefore, we will not attempt to diagnose the secondary circulation associated with the front, but instead focus on other characteristics of the mountain’s effects on fronts.

BRK investigated the effects of steep orography on landfalling fronts via idealized numerical experiments. They considered inviscid flow over very broad, plateau-shaped terrain with a steep windward slope, similar to orography in the western United States and Canada, and described the dynamics of the flow response. They found that the small horizontal scale of the windward slope contributed to strong deceleration of fronts, weak upstream frontogenesis, and strong windward-slope frontolysis. The broad horizontal scale of the terrain contributed to strong along-mountain winds, or a barrier jet, over the windward slope. Also, during frontal passage, the along-mountain winds in the coastal zone were approximately a superposition of the winds associated with the front (i.e., prefrontal and postfrontal jets) and the barrier jet.

In BRK, the term “barrier jet” was used to identify a maximum in the along-mountain winds near the surface above the windward slope that resulted strictly from the inviscid mountain-wave dynamics. Another mechanism that contributes to jets along barriers is cold-air damming, such as that seen along the eastern slopes of the Appalachian (Bell and Bosart 1988) and Rocky (Colle and Mass 1995) Mountains. A characteristic of the environment of eastern-slope barrier jets is a low-level along-mountain temperature gradient with cold air to the north. The cold air is advected southward along the slope and is trapped against the topography by the combined effects of friction, the Coriolis force, and the high static stability at the top of the cold dome. However, this type of cold-air damming mechanism will not generally be responsible for jets along the West Coast since it requires an along-mountain temperature gradient with the cooler air to the south, which is the opposite of typical cool-season conditions. The generation of along-mountain winds by the inviscid mountain–wave dynamics is one mechanism for the development of barrier jets along the West Coast. This process is likely modified substantially by the effects of surface friction and boundary layer processes, moisture and microphysics, three-dimensionality, and vertical wind shear, particularly when they result in the complete blocking of the low-level air. In this study, we extend the analysis of BRK by considering the impact of surface friction on the development and structure of the barrier jet and on the evolution of fronts crossing coastal topography.

This study is organized as follows. Section 2 provides a description of the incorporation of surface friction effects into the model of BRK and summarizes the various simulations. Section 3 describes the characteristics of flow across the surface-roughness change at the coast.
(in the absence of topography) and the subsequent modification of landfalling fronts. Section 4 describes the modification of the mountain flow by surface friction and the characteristics of frontal evolution as fronts cross the topography. Conclusions are provided in section 5.

2. Model description

a. Governing equations

We use a two-dimensional, nonhydrostatic, Boussinesq terrain-following coordinate numerical model. BRK provide a complete description of the inviscid model. This section describes the incorporation of surface friction effects into the model. The governing equations for momentum, pressure, and potential temperature are given by

\[
\frac{du}{dt} = -c_p \theta_o \frac{\partial u}{\partial x} + f v + F_u, \\
\frac{dv}{dt} = -f u + F_v, \\
\frac{dw}{dt} = -c_p \theta_o \frac{\partial v}{\partial z} + \frac{(\theta - \theta_o)}{\theta_o} + F_w, \\
\frac{\partial \theta}{\partial t} = \frac{c_p}{\rho_o} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial z} \right), \\
\frac{d\theta}{dt} = F_\theta,
\]

where \(d/dt = \delta \delta t + (u + U) \delta \delta x + w \delta \delta z, \rho = \rho_o c_p, \theta_o(\Pi + \pi'), p = \rho_o c_p, \theta_o, \Pi, \pi = -fU.\) The coordinate system is defined such that \(z\) is height, \(x\) is the zonal (west to east) direction, and \(y\) is the meridional (south to north) direction. In the equations above, \(u\) is the perturbation zonal velocity; \(U\) the background zonal flow; \(v\) the meridional flow; \(w\) the vertical velocity; \(\theta\) the potential temperature; \(\rho\) the pressure; \(c_p\) and \(c_v\) the specific heats at constant pressure and volume, respectively; \(R\) the gas constant for dry air; \(g\) gravity; \(f\) the Coriolis parameter; and \(\rho_o, \rho_v, \theta_o\) reference values of pressure, density, and potential temperature equal to 1000 mb, 1 g kg\(^{-1}\), and 300 K, respectively. The pressure is expressed in terms of basic-state and perturbation Exner functions, \(\Pi\) and \(\pi',\) respectively.

The terms \(F_z, F_x, F_y, and \theta\) are the boundary layer forcing terms:

\[
F_x = \frac{\partial}{\partial x} K_m \frac{\partial X}{\partial z},
\]

where \(K_m\) is the eddy exchange coefficient. The eddy exchange coefficients for heat and momentum are assumed to be identical. Following Troen and Mahrt (1986), the exchange coefficient for neutral conditions is specified in the terrain-following coordinate system as

\[
K_m = u^*k(z - z^* \left(1 - \frac{z - z^*}{h}\right),
\]

In (7), \(k\) is the von Kármán constant \((k = 0.4), u^*\) is the friction velocity, \(z_i\) is the height of the terrain, and \(h\) is the depth of the boundary layer. The boundary layer depth is determined through an iterative calculation of the height below which the boundary layer Richardson number is less than 0.5. This Richardson number is defined as in Vogeleyzan and Holtsl (1996):

\[
\text{Ri} = \frac{(g/\theta_o)(\theta - \theta_i)(h - z_i)}{(u_h - u_i)^2 + (v_h - v_i)^2 + bu_i^2},
\]

where \(\theta_h, u_h,\) and \(v_h\) are the potential temperature and velocity components at the top of the boundary layer; \(z_i, \theta_i, u_i,\) and \(v_i\) are the height, potential temperature, and velocity components, respectively, at the top of the surface layer; and \(b = 100.\) The top of the surface layer is taken as the lowest grid level (typically 40–50 m). The friction velocity is given by \(u^* = u_{\theta}/\ln(z_i/z_o),\) where \(V_{\theta}^2 = (U + u_1)^2 + v_1^2\) and \(z_o\) is the surface roughness parameter.

b. Simulation parameters

As in BRK, the conceptual framework for the simulations consists of a north–south-oriented cold front approaching a similarly oriented coastline. The background flow above the boundary layer consists of a uniform zonal current \(U\) balanced by a large-scale meridional pressure gradient. The coast is assumed to be associated with a rapid and large increase in the surface roughness, with \(z_o = 10^{-3} m (C_D \sim 0.0014)\) corresponding to the ocean surface and \(z_0 = 0.5 m (C_D \sim 0.008)\) corresponding to a forested land surface. Since the focus is on surface friction effects, no surface heat fluxes are considered, which causes the boundary layer parameterization to produce a near-neutral boundary layer.

The terrain shape is specified by an arctangent profile of variable height \(H\) and windward-slope half-width \(L_i:\)

\[
z_s(x) = \max \left\{ \frac{6H}{5} \frac{1}{\pi \tan^{-1}} \left( \frac{2(x - x)}{L_i} \right) + \frac{1}{3}, 0 \right\},
\]

The parameter \(x\) specifies the location where the windward slope is most steep. The model domain is 3000 km wide and varies in depth from 12 to 16 km, depending on the magnitude of \(U.\) For the results presented in the figures, only a portion of the domain is shown. The horizontal grid spacing is fixed at 6 km. A stretched vertical grid of 61 levels is used with grid spacings varying from 80 to 100 m at low levels and 400 to 530 m at upper levels. Experiments with double the vertical resolution showed no differences from the coarser simulations. Unless otherwise specified, \(H = 1\) km, \(L_i = 78\) km, and \(x_c = 2000\) km, which implies a total plateau.
Table 1. Summary of the idealized simulations. Acronyms are composed of OC, ocean; z, change; P, plateau; and F, front.

<table>
<thead>
<tr>
<th>Name</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>OC110</td>
<td>Flow over ocean surface</td>
</tr>
<tr>
<td>OCF110</td>
<td>Frontal simulation over ocean, initialized from (10)</td>
</tr>
<tr>
<td>OCF210</td>
<td>Frontal simulation over ocean, initialized from Fig. 2</td>
</tr>
<tr>
<td>CST110</td>
<td>Flow over coast (z, change)</td>
</tr>
<tr>
<td>CSTF110</td>
<td>Frontal simulation over coast, initialized from Fig. 2</td>
</tr>
<tr>
<td>P110</td>
<td>Flow over plateau, no boundary layer</td>
</tr>
<tr>
<td>P25</td>
<td>Flow over plateau, no boundary layer</td>
</tr>
<tr>
<td>P320</td>
<td>Flow over plateau, no boundary layer</td>
</tr>
<tr>
<td>PFS10</td>
<td>Frontal simulation over plateau, no boundary layer</td>
</tr>
<tr>
<td>OCP110</td>
<td>Flow over plateau, uniform z, change at coast</td>
</tr>
<tr>
<td>OCP25</td>
<td>Flow over plateau, uniform z, change at coast</td>
</tr>
<tr>
<td>OCP320</td>
<td>Flow over plateau, uniform z, change at coast</td>
</tr>
<tr>
<td>CSTP110</td>
<td>Flow over plateau, with z, change at coast</td>
</tr>
<tr>
<td>CSTP25</td>
<td>Flow over plateau, with z, change at coast</td>
</tr>
<tr>
<td>CSTP320</td>
<td>Flow over plateau, with z, change at coast</td>
</tr>
<tr>
<td>CSTPP10</td>
<td>Frontal simulation over coastal plateau, initialized from Fig. 2</td>
</tr>
</tbody>
</table>

width within the domain of 1000 km. As discussed in BRK, because the plateau intersects the downstream lateral boundary, the model behaves as if the effective width of the terrain is roughly twice that included within the domain. For the cases in this study, then, the effective width of the plateau, designated Lx, is approximately 2000 km.

Leapfrog time differencing is used along with an Asselin time filter. The coefficient of the time filter is 0.1. The large and small time steps in the split explicit time stepping are 30 and 6 s, respectively, for cases with U = 5 and 10 m s⁻¹, and 20 and 4 s for cases with U = 20 m s⁻¹. Horizontal and vertical derivatives are approximated by second-order-centered differences. Fourth-order numerical damping is applied every time step in the horizontal, while second-order damping is used in the vertical. Smoothing coefficients are small and are set to 10⁻² and 10⁻³ s⁻¹ in the horizontal and vertical directions, respectively. Radiative boundary conditions are applied at the lateral (Klemp and Wilhelmson 1978) and upper (Klemp and Durran 1983) boundaries.

Table 1 lists the simulations that were conducted. The simulations consist of five basic types, both with and without fronts: 1) ocean (OC), with z, = 10⁻³ m; 2) coast (CST), with a z, jump from 10⁻³ to 0.5 m, but no mountain; 3) coastal plateau (CSTP), with a z, change at the coast; 4) uniform plateau (OCP), with z, fixed at the ocean value; and 5) plateau (P), but no boundary layer. For some simulation types, U was varied from 5 to 20 m s⁻¹. Frontal simulations were conducted primarily with U = 10 m s⁻¹ and are indicated in Table 1 by "F."

For cases without fronts, the background flow was increased linearly in time from zero to U over the first 10 h of simulation to allow for a gradual development of the boundary layer. The initial potential temperature distribution was specified as \( \theta = \theta_0(1 + N^2 z/g) \). The value of \( N \) and the Coriolis parameter \( f \) were set to 10⁻² and 10⁻⁴ s⁻¹, respectively. Simulations with fronts were initialized by superimposing frontal potential temperature and velocity perturbations on the steady (t = 60 h) solutions of the no-front simulations. The derivation of the initial frontal perturbations is described next.

c. Initialization of the frontal simulations

To aid the discussion of the initialization of the frontal simulations, we first describe some of the basic characteristics of boundary layer effects on the cold fronts in the model. For a more complete discussion of the dynamics of this problem, see the studies by Keyser and Anthes (1982) and Dunst and Rhodin (1990). An illustrative example is obtained from a simulation (case OCF1) in which the following frontal temperature and velocity perturbations were superimposed upon the steady flow over an ocean surface (case OC1). The initial potential temperature perturbation (\( \theta_f \)) associated with the front was prescribed following BRK as

\[
\theta_f(x, z) = \frac{2\Delta\theta}{\pi} \tan^{-1}\lambda \cos\left(\frac{\pi z}{2H_f}\right) \\
\times \exp[-\max(0, \lambda^{6.5})],
\]

where \( \Delta\theta \) is the cross-frontal temperature gradient, \( \lambda = \delta \zeta_1(x-x_f) + \delta \zeta_2 \), \( \delta_1 \) and \( \delta_2 \) together specify the slope and width of the front, \( x_f \) is the initial horizontal position of the surface front, and \( H_f \) is the height of the front. These parameters are set to \( \Delta\theta = 10K, \delta_1 = 100km, \delta_2 = 1km, x_f = 1000km, \) and \( H_f = 5km \). This temperature perturbation is geostrophically balanced with an along-front jet \( \nu_j \), which is obtained numerically by downward integration of the thermal wind equation from domain top, where \( \nu_j = 0 \). In this case, thermal wind balance produces a northerly jet of \( \sim 20 m s^{-1} \) within the cold air to the west of the front. The \( \theta_f \) and \( \nu_j \) fields are shown in Fig. 9 of BRK. In this study, as in BRK, no large-scale forcing of the front is applied in order to isolate the effects of topography and surface friction on a nearly steady front. In essence, we have assumed that larger-scale forces had acted previously to intensify the front, but that these forces had ceased by the time the front was approaching the coast.

Figure 1 indicates some characteristics of the frontal development associated with surface friction effects (case OCF1). Figure 1a shows a very rapid deceleration of the near-surface northerly postfrontal jet in the first several hours, followed by a more gradual decrease thereafter. The deceleration of the frontal jet produces a thermal wind imbalance that causes pressure forces to accelerate a wide region of the zonal flow in the cold air toward the front (Fig. 1b). The resultant convergence
produces frontogenesis and upward motion (Fig. 1c) along the leading edge of the frontal zone. After about 20 h, the initial adjustment is nearly complete, the zone of maximum $u$ in the cold air near the surface narrows, and the front enters a stage of gradual decay (decreasing $\partial \theta / \partial x$). Dunst and Rhodin (1990) found a similar evolution of the maximum temperature gradient and vertical velocity.

For the simulations to be described in later sections, it was desired that the front be in this decaying stage as it approached the coast so that the effects of topography and the $z_0$ change were clearly distinguishable from the initial adjustment of the front to the boundary layer forcing. Initial conditions for these frontal simulations were obtained from the OCF1 simulation in Fig. 1 at $t = 26.7$ h, at which time the front was located near $x = 2050$ km. The initial frontal perturbations were obtained by subtracting the steady ocean boundary layer (OC1) temperature and velocity profiles from the frontal simulation (OCF1) and shifting the location of the front westward toward $x = 1200$ km.

The resultant initial perturbations for subsequent frontal simulations are shown in Fig. 2. The potential temperature perturbation field (Fig. 2a) shows a westward-sloping pool of cooler air with near-neutral lapse rates in the boundary layer. These temperature perturbations are qualitatively similar to the perturbations used by Dunst and Rhodin (1990) and to the frontal structure simulated by Keyser and Anthes (1982). An updraft of magnitude $>3 \text{ cm s}^{-1}$ occurs at the leading edge of the
front, with a weak gravity wave extending above the surface front. The boundary layer height shows a jump across the frontal updraft as a result of the change in stability across the front. The cross-front perturbation velocity field (Fig. 2b) indicates flow in the cold air toward the front within the boundary layer and weak flow toward the colder air above the boundary layer ($z = 1 \text{ km}$, $x = 1050\pm1250 \text{ km}$). The convergence of the cross-front perturbation flow at the leading edge of the front contributes to the narrow updraft there. The along-front velocity field (Fig. 2c) shows a northerly postfrontal jet centered near the top of the boundary layer and relatively strong cyclonic vorticity at the leading edge of the frontal zone. These perturbations were added to the steady solutions of OC1, CST1, and CSTP1 well upstream of the coast to produce simulations OCF2, CSTF1, and CSTPF1. Superposition of the perturbations onto the steady solutions produced no significant initial imbalances.

3. Flow across a coast: No mountain

a. Basic flow structure: No front

The structure of the flow crossing the coast in the absence of topography (case CST1) is shown in Fig. 3 at $t = 60 \text{ h}$. The cross-coast flow (Fig. 3a) reflects the increase in the boundary layer depth that occurs across the coast (Fig. 3c). Near the surface (Fig. 4), the cross-coast flow decreases rapidly downstream of the coast and reaches a new equilibrium approximately 300 km downstream. Most of the change in $u$, about 2 m s$^{-1}$, occurs within 40 km of the coast. A weak deceleration of the flow occurs upstream of the coast in a manner similar to Claussen (1987).

The along-coast flow (Figs. 3b, 4) also shows an upstream effect and indicates a weak relative maximum at the coast. Downstream of the coast, the strong deceleration of $u$ leads to an increase in $v$ by the Coriolis force, which subsequently accelerates $u$ in the positive
Fig. 3. Steady velocity perturbations associated with flow past a surface-roughness change (case CST1). (a) Cross-coast and (b) along-coast velocities, contour interval of 0.5 m s\(^{-1}\); (c) vertical velocity, contour interval of 1 cm s\(^{-1}\). Thick, lighter lines are potential temperature at 2-K intervals. (c) The thick, dashed line indicates the boundary layer height.

The vertical velocity field (Fig. 3c) indicates a weak vertically propagating gravity wave above the coast. The gravity wave is forced by a narrow zone of upward motion at the top of the boundary layer that results from the flow adjustment to the steplike jump in surface roughness. Along with the horizontal velocity components above the boundary layer, the vertical velocities indicate a flow pattern qualitatively similar to that associated with flow over a plateau (BRK; see also Fig. 8 in section 4a). In the latter case, the wave pattern is forced by a narrow zone of upward motion associated with flow up the windward slope, \( w = U\partial z_0/\partial x \). Numerical experiments (not shown) suggest that the strength of the gravity wave in the case of flow over a \( z_0 \) jump increases as the velocity \( U \) and magnitude of the \( z_0 \) change increase, and as the horizontal scale over which the \( z_0 \) change occurs decreases.

b. Frontal evolution

The flow perturbations associated with the change in surface roughness can be expected to modify the evolution of landfalling fronts. Figures 5 and 6 illustrate the structure of the frontal jet and vertical velocity from case CSTF1 prior to (Fig. 5) and following (Fig. 6) frontal passage at the coast. Prior to frontal passage, the northerly frontal jet is slightly weaker than at the initial time (cf. Fig. 2c). A weak southerly jet (>2 m s\(^{-1}\)) has formed above the front. Vertical velocities at the front are slightly weaker than those associated with the gravity wave at the coast. Following frontal passage, the northerly frontal jet weakens substantially as a result of...
the large increase in $z_0$, while the weak southerly jet intensifies slightly because of an apparent superposition of the jet with the inertia-gravity wave downstream of the coast (Fig. 3b). The frontal temperature gradient weakens as a result of weak divergence that follows the stronger, but less horizontally extensive, convergence near the coast. The frontal updraft deepens with the increase in boundary layer depth. At the coast, a slight decrease in the vertical wavelength of the gravity wave occurs because of the higher stability of the postfrontal air mass.

The temporal variation of the horizontal potential temperature gradient, $\partial \theta / \partial x$, at the lowest grid level and the maximum vertical velocity in a grid column, $w_{\text{max}}$, provide measures of the effects of the coastal $z_0$ change on the front. The strength of the frontal temperature gradient (Fig. 7a) decreases slightly as it approaches the coast (similar to the no-coast case in Fig. 1c). At the coast, there is a sudden increase in $\partial \theta / \partial x$ and a weak retardation of the frontal motion (indicated by the bending of the contours toward the top of the figure). Subsequently, the frontal gradient weakens and the frontal motion increases slightly to a value in equilibrium with the rougher surface.

The $w_{\text{max}}$ field (Fig. 7b) shows lifting at the coast of about 2–4 cm s$^{-1}$ prior to and after frontal passage. The lifting associated with the front moves from left to right with time. It is relatively constant prior to passage at the coast, increases sharply at the coast, but then weakens in the region within 200 km downstream. Farther downstream, $w_{\text{max}}$ gradually increases and the updraft deepens (Fig. 6b) as the flow adjusts to the rougher surface.

Not only does the coast have an effect on the vertical motions at the front, but the front has an effect on the vertical motions at the coast. At the coast, the vertical motion is maximum at the time of frontal passage. After frontal passage, the vertical motion remains more intense than its prefrontal value for about 4 h and is then weaker than the prefrontal value for roughly 6 h. In the postfrontal air, $w_{\text{max}}$ is the result of two competing ef-
fects: higher stability, which weakens the vertical motions; and stronger boundary layer–induced westerly flow, which strengthens the vertical motions through stronger boundary layer convergence at the coast. The enhanced vertical motion at the coast following frontal passage is associated with the horizontal cross-frontal velocity perturbations associated with the front \( u_f \), which are determined by subtracting the coastal flow in the absence of the front from that when the front is present \( u_f = u_{CSTF1} - u_{CST1} \). These velocity perturbations (Fig. 7c) indicate a narrow zone of enhanced westerly flow near the leading edge of the front within a broader region of westerly perturbations in the postfrontal air. The stronger the horizontal flow across the coast, the stronger the convergence forced by the surface-roughness change. Hence, when the narrow zone of enhanced westerlies reaches the coast, the strongest boundary layer convergence and largest vertical motions occur. As the strength of the postfrontal westerlies weakens, the vertical motions weaken. By about 25 h, the effect of the increased stability is greater than that of the enhanced westerlies, so the vertical motions are less than the prefrontal values.

The cross-coast circulation \( (u, w) \) in Fig. 3 qualitatively resembles that shown in Roeloffzen et al. (1986) for westerly onshore flow in that the flow is entirely in the onshore direction with a jump in the streamlines near the coast. Roeloffzen et al. (1986), however, showed that the streamline pattern can change significantly depending on the background wind direction. For example, with northwesterly background flow, the streamlines in their model results were all directed onshore with a weak jump near the coast, but with significant subsidence occurring farther downstream. For south-southwesterly flow (i.e., a weak onshore component), the streamlines showed a strong jump near the coast and the near-surface streamlines over land were directed offshore. Such a circulation would produce stronger lifting near the coast and promote greater frontogenesis than in the case with westerly onshore flow.

In our simulations, there is no along-shore component to the background flow, which may place some restrictions on the generality of the results. Assuming that our model would exhibit a similar dependence on the background wind direction, the results of Roeloffzen et al. suggest that the addition of a background southerly flow component to our simulations might enhance the frontogenesis and upward motion near the coast.

4. Flow across coastal terrain

a. Modification of the mountain flow by surface friction

In this section, we investigate how surface friction modifies the flow structure and dynamics associated with uniform flow over a plateau (BRK). Figure 8 shows the velocities associated with 10 m s\(^{-1}\) flow over a plateau (case P1, \( H = 1 \) km, \( L_s = 78 \) km) in the absence of boundary layer effects at \( t = 60 \) h. The cross-mountain velocity (Fig. 8a) indicates flow deceleration above the windward slope and a downstream inertia–gravity wave pattern. A barrier jet of about 13 m s\(^{-1}\) is located near the top of the windward slope in the along-mountain velocities (Fig. 8b). The vertical velocity pattern

![Fig. 6. Same as Fig. 5 but for frontal structure downstream of the coast at \( t = 40 \) h.]
Fig. 7. Time–space distributions of (a) $\partial \theta / \partial x$, (b) $w_{max}$, and (c) $u_f$ for a subset of the time–space domain for case CSTF1. The contour intervals are (a) 2 K (100 km)$^{-1}$, (b) 2 cm$^{-1}$, with an additional contour at 1 m s$^{-1}$, and (c) 0.5 m s$^{-1}$. (a), (c) The solid vertical lines indicate the coast.

(Fig. 8c) shows the maximum vertical motions near the surface above the windward slope.

The velocity components for the case with surface friction and a roughness change at the coast (CSTP1) are shown in Fig. 9. The cross-mountain velocity (Fig. 9a) is more strongly decelerated along the windward slope compared to the inviscid case (Fig. 8a). However, complete blocking of the flow does not occur. The downstream inertia–gravity wave pattern, partially evident above the boundary layer, is significantly damped by the effects of friction.

The barrier jet in Fig. 9b is slightly weaker than the jet in case P1 and is centered in the upper portion of the boundary layer. The location of the jet core above the upper portion of the slope differs from several observations of jets along the Sierra Nevada range in California (Parish 1982; Marwitz 1987). In most of the cases discussed by Parish (1982) and Marwitz (1987), aircraft in situ data indicated blocking of the low-level flow and a jet located near the base of the windward slope, in contradiction to our simulation results. However, Marwitz (1987) presented one case for which Doppler radar winds were analyzed (see his Fig. 8). In that case, the Doppler winds indicated no blocking of the cross-mountain flow and a barrier jet that was maximum over the middle-to-upper portion of the windward slope. Marwitz’s results, therefore, seem to suggest that, in the absence of complete blocking of the onshore flow,
the jet core location may indeed be located closer to the top of the windward slope.

The maximum vertical velocity is also displaced from the surface toward the top of the boundary layer. It occurs at the coast as a result of the combined effects of the lifting by the plateau and the change in \( z_0 \) at the coast. A secondary maximum of \( w \) occurs above the windward slope, where the slope is approximately steepest. While the vertical velocity associated with flow over a sharp surface-roughness change (Fig. 3c) is comparable to that for flow over a plateau (Fig. 9c), the total vertical displacement of air parcels is much smaller in the former case because of the very narrow character of the roughness change–induced gravity wave. Thus, one can anticipate that precipitation produced by flow over a coastal plateau is dominated by the effects of the elevation change rather than the roughness change. This supposition is validated by the simulations of Doyle (1997), which showed that coastal effects in the absence of topography produced only minor precipitation amounts compared to that produced by ascent over the coastal mountains. It is also evident from a comparison of the vertical displacement of the isentropes near the coast in Figs. 3 and 9.

A major focus of BRK was the generation of strong barrier jets by flow over plateaus. They showed from linear theory that the steady strength of the barrier jet relative to the background flow \( U \) varied approximately with the nondimensional mountain height \( h_m = NH/U \) and total nondimensional width \( L_2 = \overline{Ro}_2^{-1} fL_2/U \); \( \overline{Ro}_2 \) is the Rossby number corresponding to the total width of the plateau and the overbar indicates a nondimensional variable as \( v/U \approx h_m (1 + \ln L_2)/\pi \). Below, we examine how the relative strength of the barrier jet is modified by the effects of surface friction.

Figure 10 shows time series of the maximum barrier-jet strength \( (v/U) \) for six different cases. For each case, \( H = 1 \) km and \( L_2 = 2000 \) km (i.e., twice the plateau...

**Fig. 8.** Steady velocity perturbations associated with mean flow over a plateau in the absence of boundary layer effects (case P1). (a) Cross-mountain and (b) along-mountain velocities, contour intervals of 1 and 2 m s\(^{-1}\), respectively; (c) vertical velocity, contour interval of 1 cm s\(^{-1}\). Thick, lighter lines are potential temperature at 2-K intervals.
The mean velocity \( U \) was varied between 5, 10, and 20 m s\(^{-1} \), and for each velocity, one case was simulated with \( z_0 \) fixed at \( 10 \) m (cases OCP1, 2, 3, corresponding to \( U = 10, 5, \) and \( 20 \) m s\(^{-1} \)), while another case included the change in \( z_0 \) at the coast (cases CSTP1, 2, 3). Also shown along the right vertical axis are the jet amplitudes obtained from the inviscid numerical simulations (cases P1, 2, 3). Comparison of the curves associated with the \( z_0 \)-jump and fixed-\( z_0 \) cases indicates that the change in surface roughness produces an enhancement of the jet on long timescales. Over shorter timescales, the difference is small. For comparison to the inviscid simulations, consider only the cases with fixed \( z_0 \) (cases OCP1, 2, 3; thick lines). For weak jets (small \( h_m \) and \( L_z \), OCP3), the jet strength in the case with friction is only slightly less than that in the inviscid simulation. However, as the jet strength increases (\( h_m \) and \( L_z \) increase), the jet magnitude in the case with friction becomes increasingly weaker than that in the inviscid case. Therefore, it appears that as the relative jet strength increases, the effect of surface friction increasingly erodes the strength of the jet as compared to the predicted strength from the inviscid cases.

The increasing erosion of the jet strength by surface friction can be understood from the steady linear equation of motion for the along-mountain component:

\[
U\nu_1 = -fu + \tau, \tag{11}
\]

where \( \tau \) is the boundary layer stress and subscripts indicate derivatives. Let \( x = L_x x, t = U/L_x t, u = \bar{U} u, v = \bar{V} v, z = h\bar{z}, \) and \( \tau = (KU/h)\bar{\tau}, \) where the overbar indicates nondimensional variables and \( K \) is a characteristic scale for the exchange coefficient. The nondimensionalized form of (11) averaged over the boundary layer (\( \bar{z} = 0 \) to 1) is

\[
\{v_1\} = \frac{-1}{Ro}\{\bar{u}\} + KL_1\int_0^1 \bar{\tau}_z d\bar{z}, \tag{12}
\]

where \( \{\} \) indicates a boundary layer average. The nondimensional stress can be written in terms of dimensional variables as
surface friction in proportion to the square of the relative jet strength. Since \( v \) far upstream is small, a large \( \omega_0 \) implies a stronger jet along the windward slope. Typical magnitudes of the scaling parameters are \( h \sim 10^3 \) m, \( L_1 \sim 10^4 \) m, \( U \sim 10^2 \) m s\(^{-1}\), \( f \sim 10^{-4} \) s\(^{-1}\), and \( c^2 \sim 10^{-3} \) (for \( z_0 = 10^{-3} \) m). Thus,

\[
\{\tau_x\} = -\{\tau\} - 0.1 \left( \frac{v}{U} \right)^2 .
\tag{15}
\]

For \( h_m < 1 \), \( v/U \leq 1 \) and the surface friction term is small. Consequently, as indicated in Fig. 10, the strength of the jet in the case with surface friction is nearly the same as in the inviscid case. As the effects of orography act to increase \( v/U \), the dissipative effect of surface friction increases rapidly, in agreement with the orographic lifting. In this section, we examine the influence of surface friction on the above processes.

\footnote{As the magnitude of \( u_1 \) approaches \( U \), the applicability of linear theory diminishes. However, the analysis remains useful for qualitatively explaining the impact of friction on the jet magnitude, \( v/U \).}

b. Frontal evolution

Braun et al. (1999) showed with an inviscid model that the strong orographically induced flow deceleration upstream of the windward slope led to significant retardation of frontal motion, upstream frontogenesis, and subsequent strong frontolysis over the windward slope. Similar but weaker effects have been documented in cases of frontal motion over less steep orography (Zehnder and Bannon 1988; Keuler et al. 1992; Williams et al. 1992). BRK also demonstrated how frontal jets interact with the barrier jet along the windward slope. Winds within the coastal zone were approximately a superposition of the frontal and barrier jets. The result of this superposition effect was that a southerly prefrontal jet could combine with the barrier jet (also southerly) to produce very strong winds, while a northerly postfrontal jet could largely cancel the barrier jet to produce weak winds. Vertical motions were dominated by the orographic lifting. The front, which had no initial vertical motion and was not subject to large-scale forcing, did not develop significant vertical motions during its passage across the orography. In other words, the mountain forcing of frontogenesis did not result in a secondary circulation that was apparent in the presence of the strong mountain circulation. Zehnder and Bannon (1988) also found that the mountain circulation generally dominated that of the front, even in the presence of large-scale forcing. However, BRK found that the stability perturbations associated with the front modified the mountain circulation. For example, the higher stability behind the front reduced the magnitude of the orographic lifting. In this section, we examine the influence of surface friction on the above processes.

Figure 11 shows the evolution of the vertical motions during passage of the front across the coastal orography (case CSTPF1). At \( t = 13.3 \) h (Fig. 11a), the front is within 300 km of the coast and the vertical motion at

![Fig. 10. Time series of the barrier-jet strength for selected cases (see text). The thick line segments along the right vertical axis indicate values from the inviscid simulations. Relevant dimensional and non-dimensional variables are given, where \( \ell_t = \ell_L/U \) and the other parameters are defined in the text.](image)
the front is similar to the initial value in Fig. 2a. By \( t = 20 \) h (Fig. 11b), the leading edge of the front is right at the coast. The frontal updraft has merged with that associated with the lifting at the coast. At \( t = 26.7 \) h (Fig. 11c), the leading edge of the front is near the top of the windward slope, the frontal updraft is weak (about 2 cm s\(^{-1}\) near \( x = 2100 \) km), and the frontal temperature gradient is rapidly weakening. Subsequently (Figs. 11d,e), the front progresses downstream, the frontal updraft gradually deepens, and the temperature gradient continues to weaken slightly. The higher stability behind the front decreases the vertical wavelength of the mountain- and coast-induced gravity waves.

The temporal variations of \( \partial \theta/\partial x \) at the lowest grid level for case CSTPF1 are shown in Fig. 12a. A corresponding figure for the case without surface friction (PF1) is shown in Fig. 12d. In both cases, the gradient associated with the orography has been subtracted in order to remove the effects of the superposition of the mountain gradient with the frontal gradient (Zehnder and Bannon 1988; Keuler et al. 1992). In the inviscid case, the mountain component of \( \partial \theta/\partial x \) is approximately
Fig. 12. Time–space distributions of (a) $\partial \theta / \partial x$, (b) $w_{\text{max}}$, and (c) $u_f$ for a subset of the time–space domain for case CSTPF1. The contour intervals are (a) $2 \text{ K (100 km)}^{-1}$, (b) $1 \text{ cm s}^{-1}$, and (c) $0.5 \text{ m s}^{-1}$; (d) $\partial \theta / \partial x$ for the case without boundary layer effects (case PF1), contour interval of $2 \text{ K (100 km)}^{-1}$. (a), (c), and (d) The solid vertical lines indicate the coast and the location where the terrain height reaches 0.9 $H$ (in other words, they enclose the windward slope).

In the absence of surface friction effects, the frontal-gradient zone upstream of the coast is broader [e.g., cf. the 4 K (100 km)⁻¹ contour], with the maximum gradient $[-6.5 \text{ K (100 km)}^{-1}]$ near the center of the frontal zone. In contrast, in case CSTPF1, the frontal zone is narrow and the maximum gradient $[-8 \text{ K (100 km)}^{-1}]$ is located near the leading edge of the frontal zone. Just upstream of the coast, in case PF1, $\partial \theta / \partial x$ increases slightly as a result of the upstream flow deceleration, but then decreases rapidly inland as a result of frontolytic tilting and stretching by the orography. In case CSTPF1, stronger frontogenesis occurs near the coast,
with an intensity change of 2 K (100 km)$^{-1}$ in CSTPF1 versus 0.5 K (100 km)$^{-1}$ in case PF1, and the maximum intensity occurs just inland of the coast. This latter result occurs because the boundary layer mixing, which produces vertically oriented isentropes in the boundary layer, prevents significant tilting of the isentropes by the orography. Hence, the frontolysis over the windward slope is dominated by the effects of horizontal divergence.

In both cases CSTPF1 and PF1, after the strong frontolysis over the windward slope, the fronts incidentally reach nearly the same weak strength near $x = 2400$ km. Farther downstream (not shown), the front in case PF1 increases somewhat in strength as it moves through the inertia-gravity wave pattern (Fig. 8a). In case CSTPF1, the front remains weak as a result of the dampened inertia-gravity waves in the boundary layer (Fig. 9a).

The temporal variations of $w_{\text{max}}$ (Fig. 12b) show the combined effects of the lifting by the orography, the coastal $z_0$ change, and the front. Prior to the times included in the figure, the frontal updraft intensity increases only slightly. As the front approaches to within 300 km of the coast, its updraft gradually intensifies. Maximum vertical motions occur when the frontal updraft reaches the coast and combines with the coastal forcing. The frontal updraft subsequently weakens but remains relatively intense above the lower portion of the windward slope. Further progression of the front over the orography leads to rapid weakening of the frontal updraft. Beyond $x = 2250$ km, the frontal updraft begins to redevelop as the flow comes into equilibrium with the rougher surface (see also Figs. 11d,e).

As in the case with the $z_0$ change but no plateau (case CSTF1, Fig. 7b), enhanced lifting occurs at the coast immediately following frontal passage ($t = 21-25$ h; Fig. 12b) followed by a period of reduced lifting. Figure 12c shows the cross-frontal velocity perturbation, $u_f = u_{\text{CSTPF1}} - u_{\text{CST1}}$. Maximum westerly perturbations occur immediately behind the front and decrease rapidly upon landfall of the front. A tongue of strong westerlies extends inland with the frontal updraft, but it is rapidly reduced by the increased surface roughness. Behind the front, westerly velocity perturbations continue along and upstream of the coast for some time, contributing to the enhanced lifting at the coast just behind the front up to 25 h, after which time stability effects dominate and the vertical motions are reduced. The $w_{\text{max}}$ field implies possible impacts upon precipitation within the coastal zone, including 1) enhancement of frontal precipitation just prior to landfall and continuing partway up the windward slope, 2) reduced frontal precipitation downstream of the windward slope, and 3) enhanced precipitation along the slope and particularly at the coast for some time following frontal passage, provided that adequate moisture is available in the postfrontal air.

An important question to address is how surface friction modifies the superposition behavior of the frontal and barrier jets described in BRK. The interaction of these jets in the case with friction (CSTPF1) is depicted in Fig. 13. At $t = 13.3$ h (Fig. 13a), the northerly postfrontal jet weakens as it moves into the region of upstream orographic influence. In Fig. 9b, a southerly jet is located above the front and by $t = 20$ h (Fig. 13b) it is merging with the barrier jet. The strongest southerly flow occurs at $t = 26.7$ h (Fig. 13c), with peak velocities exceeding 12 m s$^{-1}$. The northerly frontal jet has a peak magnitude less than 2 m s$^{-1}$ at this time. At $t = 33.3$ h (Fig. 13d), the barrier jet and northerly frontal jet have combined to produce very weak winds in the coastal zone. Finally, by $t = 46.7$ h (Fig. 13e), the frontal and barrier jets reform. The northerly frontal jet remains weak because of the increased surface roughness.

The evolution of the jets in Fig. 13 suggests that superposition is valid at least qualitatively. In Fig. 14, we demonstrate the validity of superposition quantitatively. Figure 14 shows the strength of the northerly postfrontal jet at the lowest grid level as a function of the location of the jet within the domain. Case OCF2 shows the gradual weakening (decreasing negative values) of the jet associated with movement over an ocean surface. Case CSTF1 indicates the rapid weakening of the frontal jet as it moves past the surface-roughness change at the coast, while case CSTPF1 shows the rapid weakening and the reversal of wind direction (northerly to southerly) as the jet moves over the coastal plateau (including the effects of the $z_0$ change). A superposition estimate is obtained by isolating the plateau’s contribution to the flow and combining it with the frontal jet evolution in CSTF1. The plateau’s contribution to the flow is estimated by subtracting the near-surface flow in the case of the $z_0$ change but no orography from the flow in the case with the orography, $u_f = u_{\text{CSTPF1}} - u_{\text{CST1}}$. This plateau perturbation velocity is then added to the frontal-jet strength associated with a front moving across a flat coast (CSTF1; dashed line in Fig. 14). The superposition estimate for the frontal strength is shown as the dot–dashed line in Fig. 14, from which it can be seen that superposition is approximately valid quantitatively near the surface.

Above the surface, superposition is more difficult to verify, but it is perhaps most clearly evident in the interaction of the southerly jet above the front (Fig. 13a) and the barrier jet. In Fig. 13a, the southerly frontal jet has a magnitude of approximately 7 m s$^{-1}$. This magnitude appears to derive from a superposition of velocities associated with the orography ($\sim 5$ m s$^{-1}$; Fig. 9b) and velocities associated with the front in the absence of the orography ($\sim 2$ m s$^{-1}$; Fig. 5a). Superposition of the 2 m s$^{-1}$ frontal component with the approximately 10 m s$^{-1}$ velocities over the windward slope at about 2 km altitude gives a 12 m s$^{-1}$ combined strength. In Fig. 13c, the jet strength above the windward slope after the merger of the frontal and barrier jets is slightly in excess of 12 m s$^{-1}$. Thus, superposition appears to be valid above the surface as well.

Figures 11–13 depict the interaction of a front with
Many fronts are observed to have strong southerly flow ahead of, and weak along-front flow behind, the surface front. For this type of event, will the general character of the interaction of the front with the orographic circulation change? This case has not been examined numerically within this study, but we may speculate about whether significant changes may occur. In the case of a front with a southerly prefrontal jet, surface friction will reduce the magnitude of the jet near the surface in a manner similar to the case with a northerly postfrontal jet. Surface stresses will induce prefrontal flow in the cross-front component toward lower pressure (at the front) and result in convergence and frontogenesis within the frontal zone. Thus, quasi-steady frontal perturbations (analogous to those in Fig. 2) would likely be characterized by 1) vertical motion at the leading edge of the front, 2) negative cross-frontal velocity perturbations ahead of the front, and 3) a southerly jet ahead of the front. As long as the negative cross-frontal flow perturbations do not combine with the coastal and orographic perturbations to produce complete blocking of the flow, we anticipate a similar evolution of the front as depicted in Figs. 11–13, that is, superposition of the
barrier jet and frontal jet, with the following changes: 1) in the case with the prefrontal southerly jet, the frontal jet and barrier jet will combine to produce very strong southerly flow prior to frontal passage; and 2) since the cross-front velocity perturbation $u_f$ will be negative ahead of the front, we expect reduced convergence and lifting at the coast prior to frontal passage.

5. Conclusions

A previous study by Braun et al. (1999) investigated the effects of steep plateau-shaped orography on landfalling cold fronts using an inviscid numerical model. The present paper extends this previous work by incorporating a simple parameterization of surface friction into the model and describing the effects of surface friction on flow past a sharp change in surface roughness and flow over a plateau. Further, the effects of these frictionally modified flows on landfalling fronts is examined.

Flow over a sharp roughness change leads to rapid deceleration of the cross-coast flow and gradual acceleration of the along-coast flow. Vertically propagating gravity waves are present above the change in surface roughness and result from the temporary force imbalance that occurs as the flow adjusts to the steplike jump in surface roughness. The gravity wave pattern resembles that associated with flow over a shallow plateau, since the forcing of vertical motion at the top of the boundary layer resembles that produced along the windward slope of a plateau. The strength of the gravity waves increases with increasing onshore velocity, increasing magnitude of the surface-roughness change, and decreasing horizontal scale of the roughness change.

For the case of a front with a northerly postfrontal jet, the effects of the surface-roughness change at the coast on landfalling fronts are weak frontal retardation and rapid deceleration of the postfrontal jet. The roughness change produces weak frontogenesis at the coast and somewhat stronger frontolysis farther inland. The frontal updraft is substantially, but only briefly, enhanced upon passage at the coast. In addition, vertical motions at the coast are enhanced for a period after frontal passage as a result of westerly velocity perturbations associated with the front. (In the event of a southerly prefrontal jet, vertical motions are anticipated to be reduced somewhat prior to frontal passage because of easterly velocity perturbations ahead of the front.) The enhanced lifting probably produces negligible rainfall enhancement, however, since the horizontal scale of the updraft is very narrow, thereby producing small vertical displacements of air parcels.

Friction modifies the flow over the plateau by increasing the upstream deceleration, decreasing the barrier-jet strength relative to the basic-state flow, and damping the downstream inertia-gravity waves. The relationship between the barrier-jet strength ($u_f/U$) and the nondimensional mountain height ($h_m$) and plateau width ($L_2$) is essentially the same as in BRK, except that as $h_m$ and $L_2$ increase, thereby forcing a stronger barrier jet, the effects of friction become increasingly important in reducing the strength of the jet relative to the inviscid cases. Vertical motion is maximum at the coast because of the combined effects of the orography and the roughness change, but large parcel displacements are due mainly to the orographic forcing.

Compared to the case without friction, the effects of friction on a front moving over coastal orography are stronger frontal retardation and upstream frontogenesis, slightly stronger frontolysis over the orography, and elimination of downstream oscillations in frontal strength that, in the inviscid case, were associated with the interaction of the front with the downstream inertia-gravity waves. The frontal updraft is strongly enhanced upon reaching the coast and remains so until diminishing rapidly midway up the slope. The updraft is weakest near the top of the windward slope but gradually strengthens farther downstream as the frontal flow adjusts to the rougher underlying surface.

The superposition behavior of frontal and barrier jets described in BRK is not significantly changed by surface friction. The winds in the coastal zone are still well represented by the superposition of the landfalling frontal jets and the barrier jet. Maximum along-coast winds occur when a prefrontal southerly jet combines with the southerly barrier jet. The along-coast winds are weakest after frontal passage when a northerly postfrontal jet and the barrier jet at least partially cancel each other. As mentioned in BRK, this sequence of events varies depending on the presence of the prefrontal and post-
frontal jets as well as on the occurrence and duration of onshore flow prior to frontal passage. It is expected that variations in the basic state stability may also impact the flow evolution.

The numerical model used in this study differs from that of BRK only in the inclusion of surface friction effects. The two studies in tandem provide an important understanding of the effects of the broad, yet steep, orography of the western United States and Canada on landfalling frontal systems. The previous study described the basic dynamics of the problem, while this study provides an incremental advancement in our understanding of this significant coastal meteorological situation. Additional insights can be achieved through the examination of the certainly important effects of moisture and cloud microphysics, as well as the inclusion of more realistic initial frontal structures, large-scale forcing, and three-dimensionality.

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