Airborne Doppler Observations of a Cold Front in the Vicinity of Vancouver Island*

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Abstract

This study uses airborne Doppler radar and flight-level measurements from the Coastal Observations and Simulations with Topography experiment to examine the detailed mesoscale structure of an oceanic cold front upstream of Vancouver Island on 13 December 1993. These aircraft observations show that there were dramatic differences in frontal structure and movement between nearshore and offshore regions, presumably due to the effects of blocking by the terrain. The aircraft observations are considered in two parts, since the behavior of the front evolved over the flight period. During the early and middle portion of the flight, the low-level flow east of the front was out of the south-southeast in the nearshore region, rather than southerly as found farther offshore. The nearshore segment of the front was oriented south-north and appeared to be stationary. The zone of orographic influence was found to extend to a region ~20 km offshore and confined primarily to the lowest 1.5 km (MSL). In distinct contrast to the nearshore segment, the offshore segment of the front was oriented southwest-northeast and retrograded slowly northwestward. It exhibited more deep inflow and its associated slope was less steep. During the latter part of the flight, the offshore portion of the front moved eastward as is more typical of a cold front. Meanwhile, the nearshore segment of the front remained virtually stationary, and thus a significant distortion developed in the front. This distortion featured a local minimum in low-level convergence along the front and, hence, also reduced precipitation rates.

The structure and evolution of this front was related to the interactions between synoptic, orographic, and boundary layer effects. The low-level portion of the front did not resemble a classic gravity current in its structure or propagation. The frontal updraft at low levels appeared to be less due to the cooler air behind the front undercutting the warmer air, but rather more due to Ekman pumping, that is, frictional convergence, associated with the cyclonic vorticity concentrated at the front.

1. Introduction

The weather along the west coast of the United States during the cool season is dominated by the landfall of extratropical storm systems originating over the Pacific Ocean. These synoptic-scale disturbances are subject to mesoscale modifications in their winds and precipitation due to the coastal orography in this region. Our understanding of these orographic effects is limited by the scarcity of observations within the coastal zone. This problem was addressed in the Coastal Observations and Simulations with Topography (COAST) field experiment along the coast of the Pacific Northwest (Bond et al. 1997). The centerpiece of COAST was a National Oceanic and Atmospheric Administration (NOAA) P-3 research aircraft equipped with a tail-mounted scanning Doppler radar, which was tasked to document the detailed structures of frontal systems near the coastal zone. In this paper we present results from an airborne Doppler analysis of a cold front on the windward side of Vancouver Island on 13 December 1993. We focus on aspects concerning the interaction between the cold front and coastal orography.

Incident flow is generally blocked by a mountain barrier when the upstream Froude number (Fr = U/NH, where U is the cross-barrier flow component, H is the mountain height, and N is the Brunt–Väisälä frequency) is less than unity (Smith 1979). A theoretical study by Pierrehumbert and Wyman (1985) showed that the Burger number (B = Hf/L, where f is the Coriolis parameter and L is the mountain half-width), not just the Froude number, is an important parameter in determining the nature of blocking. When B exceeds unity, the

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deflection of the flow extends well upstream of the barrier and the mountain is said to be “hydrodynamically steep” (Overland and Bond 1995). Pierrehumbert and Wyman (1985) also showed that, when considering the effect of the earth’s rotation, the Coriolis force acts to constrain the upstream extent of blocking that attains a maximum on the order of the Rossby radius of deformation \( L_R = NH/f \). For a relatively shallow blocked flow (compared to the mountain height), Overland and Bond (1995) further suggested that the seaward extent of flow blocking scales as \( UL/f \).

In contrast to the simple conditions considered in studies such as Pierrehumbert and Wyman (1985), orographic effects in conditions of unsteady, nonuniform incident flow (such as that found in the vicinity of fronts and cyclones) tend to be more complicated and are not as well understood. Recent numerical studies have attempted to examine processes and dynamics relevant to the fronts approaching the steep coastal terrain along the coast of the Pacific Northwest. For instance, Doyle (1997) used high-resolution simulations of the Coupled Ocean–Atmospheric Mesoscale Prediction System (COAMPS) model to investigate an event of a prefrontal low-level wind maximum occurring adjacent to steep terrain along the central California coast. His results indicate that the southerly coastal jet developed as a result of low-level blocking of prefrontal southwesterly flow. This blocking also served to slow the onshore propagation of the front. Braun et al. (1999) used a dry, two-dimensional, Boussinesq numerical model in conjunction with an idealized coastal terrain typical of the western United States and Canada (i.e., a broad mountain range with a steep windward slope), to examine the formation of coastal barrier jets and their impacts on cold fronts. Their results indicate that the strength of the simulated barrier jet in the coastal zone is more related to the longwave characteristics of the orography (i.e., the total width, instead of the width of the windward slope) and are approximately determined by a superposition of the southerly barrier wind and the southerly prefrontal jet. Moreover, the strong upstream deceleration in association with the occurrence of low-level blocking is found to produce a significant retardation of the frontal motion and frontogenesis ahead of the windward slope.

Despite a large number of past studies dealing with the finescale structure of fronts, relatively few have investigated the modulation of frontal precipitation by orography. Some radar observations of the mesoscale structure of frontal precipitation along the Pacific Northwest coast were collected as part of the Cyclonic Extratropical Storms (CYCLES) project (e.g., Hobbs et al. 1975; Parsons and Hobbs 1983a), but these earlier studies lacked dual-Doppler radar observations. As such, their ability to address specific processes relevant to the orographically modified precipitation near fronts was hampered. COAST essentially built upon CYCLES to more completely document the mesoscale structure and evolution of fronts adjacent to the steep coastal terrain. Several recent articles from COAST have presented various observational aspects on this topic. Braun et al. (1997) studied an intense cold frontal system oriented parallel to the mountain barrier as it advanced toward the southern Oregon coast. Their observations suggested the importance of the irregular coastline near Cape Blanco on contributing to the dissipation of a precipitation core at the front as it approached shore. Colle et al. (1999) investigated a cold front as it interacted with the Olympic Mountains, a relatively isolated, quasi-circular barrier. Their analysis of the aircraft observations showed weakening of frontal precipitation as the system began to ascend the western slopes of the Olympics; the front subsequently deformed as it passed around this barrier. Yu and Smull (2000) used airborne Doppler radar observations to document the mesoscale structure and evolution of a cold front as it made landfall on the mountainous coast of Oregon and northern California. Their results show that upstream blocking by the steep coastal terrain evidently led to a significant increase in coastal surface winds and modifications of precipitation in the vicinity of the front including the rapid formation of a narrow cold-frontal rainband (NCFR) and nearshore enhancement of two prefrontal precipitation bands. In a corresponding numerical study, Colle et al. (2002) further suggested that the rapid evolution of frontal precipitation was also related to the enhanced deformation frontogenesis associated with the prefrontal blocked flow. Chien et al. (2001) utilized observations and fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) simulations in a study of an intense cold front making landfall along the northwest coast of the United States. Based upon their simulations, the formation of a prefrontal rainband was attributed to the low-level convergence between strong southwesterly winds ahead of the front and nearshore southerlies. Doyle and Bond (2001) used the aircraft observations and the numerical model to investigate a warm front making landfall on Vancouver Island. They observed changes in precipitation patterns near the front and a more concentrated wind shift zone as it neared shore.

The primary objective of this study is to use airborne Doppler radar and flight-level measurements on 13 December 1993 from COAST Intensive Observation Period 6 (IOP 6) to document the mesoscale structure of a cold front in the vicinity of Vancouver Island. Vancouver Island is a quasi-two-dimensional barrier and oriented roughly northwest–southeast, with a high elevation of \( \sim 1200 \) m MSL and a mountain width of \( \sim 50 \) km (Fig. 1). In the case presented here, the synoptic-scale cold front was oriented approximately perpendicular to the axis of Vancouver Island. As suggested by Yu and Smull (2000), such a configuration, in which the terrain effect is not uniform along the front, would favor a stronger and more complicated interaction between the front and the coastal orography. In contrast
to other COAST events (e.g., Yu and Smull 2000), whose aircraft observations were primarily confined to the nearshore region, the aircraft observations for the present case were collected in two zones: a nearshore region within 50 km of the coast and an offshore region ~80–120 km from the coast (Fig. 1). These data permit a unique description of three-dimensional kinematic and precipitation fields in the vicinity of the front in both nearshore and offshore regions. This case is also different from previous COAST studies of fronts proximal to coastal terrain (e.g., Braun et al. 1997; Yu and Smull 2000) in that it involves a cold front that is slow moving and does not resemble a gravity current. Nevertheless, this front did feature strong gradients in the low-level winds coincident with a well-defined band of enhanced precipitation. Our analysis documents these structures, in particular how they relate to boundary layer processes, and how they differ in the offshore and nearshore regions of Vancouver Island.

2. Synoptic overview and aircraft observation

Intensive observations by the NOAA P-3 for this case were conducted in the coastal zone south of Vancouver Island between 0230 and 0800 UTC 13 December 1993. The geographic region for the P-3 flight track is shown in Fig. 1. Synoptic conditions accompanying this event at 850 mb at 0000 UTC 13 December (Fig. 2) were dominated by a cyclone centered over the Gulf of Alaska near 53°N, 140°W. Strong south-southwesterly geostrophic flow was present to the west of the Washington coast and immediately upstream of the southern coast of Vancouver Island. The National Centers for Environmental Prediction (NCEP) surface analysis at 0000 UTC 13 December indicated a front (heavy thick line in Fig. 2) extending from near the low center analyzed at 850 mb in the Gulf of Alaska to ~36°N, 138°W. The portion of the front north of 51°N was designated as occluded, while the southern portion was designated as a cold front. A pressure trough (indicated by the heavy dashed line in Fig. 2) was present along 139°W, ~600 km behind the cold front.

The cold front generally advanced eastward, as indicated by a sequence of NCEP surface analyses (Fig. 3). The propagation of this front was not uniform, however. Notably, the portion of the front between roughly 46° and 50°N was virtually stationary between about 0000 and 0600 UTC 13 December. Note that this portion extends from Vancouver Island to a region at least ~400–600 km offshore, well upstream of coastal terrain. This lack of propagation therefore does not seem to be simply related to orographic blocking, but rather due to the absence of rear-to-front low-level flow in the cool air west of the front, on the synoptic scale. The
The NCEP surface cold front isochrones with 3-h intervals from 1500 UTC 12 Dec to 1200 UTC 13 Dec 1993. The aircraft observations of frontal structure are considered for two phases: a phase in the early and middle part of the flight mission in which the front was quasi-stationary, and a phase in the latter part of the flight when the front was advancing eastward. Based on this scenario, the aircraft observations of frontal structure are summarized using a composite of the flight-level measurements taken before 0600 UTC (Figs. 4 and 5). The frontal zone featured a distinct cyclonic change in wind speed from 25–30 m s\(^{-1}\) east of the front to 10–15 m s\(^{-1}\) west of the front. The change in wind direction was much less prominent, although there was some cyclonic turning near the coast. Within the warm air mass east of the front, there was a noticeable change in direction from southerly in the offshore region to southeasterly near the coast. Based on the sea level pressure distribution deduced from the aircraft measurements (which was consistent with the synoptic-scale analyses by NCEP), the winds near the coast included a strong ageostrophic component, in contrast to more geostrophic southerlies farther offshore (Fig. 4). There was some suggestion of pressure ridging occurring along the coast of Vancouver Island. Such a signal in the low-level pressure field is consistent with the air temperatures observed by the aircraft, which indicate that the nearshore region was generally colder than offshore (Fig. 5). The warmest air was found in a narrow tongue immediately ahead of the front coincident with enhanced southerly winds. Relatively cold air was flowing out of the mouth of the Strait of Juan de Fuca (location shown in Fig. 1) in the eastern portion of the analysis domain. Unlike the warm-frontal situation studied by Doyle and Bond (2001), it appears that this cold air out of the strait was not reaching the frontal zone, or at least it was highly diluted by the prefrontal flow originating over the open ocean.

The basic thermodynamic characteristics upstream of Vancouver Island were obtained from a P-3 ascent sounding. This sounding was ~90 km offshore (see Fig. 4) and thus indicates the vertical structure of the incident flow upstream of the coastal terrain. The vertical thermodynamic profile reveals saturated conditions and stable-to-neutral convective stability between 1 and 3 km (MSL) (Fig. 6). Stronger stratification was found in the
lowest 1 km except a shallow mixed layer near the surface. Winds veered with height from south-southeasterly near the surface to southwesterly at 850 mb, and exhibited a maximum of $\sim 35 \text{ m s}^{-1}$ at a height of $\sim 1000$ m. For the case studied here, the representative terrain height ($H$) of Vancouver Island in the region of aircraft observation is $\sim 1000$ m (cf. Fig. 4), and the latitude of $50^\circ$ gives a value of $f = 1.1 \times 10^{-4}$ s$^{-1}$. Since the air at low levels was saturated as shown in Fig. 6, static stability is approximated using the saturated Brunt–Väisälä frequency ($N_s$) derived by Durran and Klemp (1982) and is equal to $8.2 \times 10^{-3}$ s$^{-1}$. The mountain half-width ($L$) is assumed to be $\sim 50$ km and the cross-barrier flow averaged below 1000 m (MSL) is equal to $23 \text{ m s}^{-1}$. With these values, we obtain a Froude number ($Fr$), Burger number ($B$), and the Rossby radius of deformation ($L_R$) equal to 2.8, 1.5, and 74 km, respectively. The relatively large value for $Fr$ implies that the low-level flow should not be fully blocked by the terrain and indeed a significant onshore component was observed in the nearshore region just ahead of the front. Even though the deflection of the prefrontal flow just ahead of the front was modest, given the large value of $B$ (i.e., $>1$), the effects of the coastal terrain are expected to extend upstream of the windward slope as suggested by the flight-level winds (Fig. 4). Due to the coarse resolution of flight-level measurements shown in Fig. 4, the nature and extent of orographic effects on the flow in the coastal zone are not evident. We pursue this issue in section 3.

It is interesting that the flow in the nearshore region farther to the southeast (i.e., $\sim 60–120$ km well ahead of the nearshore front) was oriented to be nearly parallel to the coastal terrain. This enhanced blocking would partially be attributed to a relatively drier oncoming flow as revealed by flight-level data and Doppler radar observations showing generally unsaturated conditions at low levels and lack of observed radar echoes over the region upstream of this mountain-parallel flow. In this condition, the dry (instead of saturated) Brunt–Väisälä frequency ($N = 1.4 \times 10^{-2}$ s$^{-1}$) would be more suitable to represent upstream static stability, which gives a $Fr$ of $\sim 1.6$, about one-half of the foregoing estimated value. Additionally, the presence of cool air flowing out of the Strait of Juan de Fuca (cf. Fig. 5) would presumably further enhance the blocking effect through an increase in static stability (and a decrease in the onshore component of the flow) resulting in a $Fr$ of approximately unity or less.

3. Structure of the front during its quasi-stationary phase

In this section we compare the frontal structure in the nearshore and offshore regions during the front’s quasi-stationary phase. The detailed three-dimensional airflow and precipitation in the frontal region were obtained by use of a pseudo-dual-Doppler synthesis technique described by Jorgensen and Smull (1993). The airborne Doppler radar data editing, interpolation, and synthesis applied in this study are following the procedures outlined by Yu and Smull (2000). Analyses for two segments along the front are presented: one from the coast to about 40 km offshore (hereafter referred to as the nearshore segment/domain) and the other located $\sim 80–120$ km farther offshore (hereafter referred to as the offshore segment/domain). The locations of the dual-Doppler analysis domains for these two segments are indicated in Fig. 4. The horizontal and vertical analysis grid spacing were set to 1.5 km and 0.25 km, respectively, over multiple volumes encompassing $81 \times 81$ km$^2$ in the horizontal and 8.5 km in the vertical, with.
the lowest analysis level located at 0.25 km MSL. A complementary perspective on the low-level frontal structure, including kinematics and thermodynamics, was provided by in situ flight-level measurements as the aircraft penetrated the nearshore and offshore segments of the front at multiple altitudes below about 2 km (see Fig. 1 for the locations).

Dual-Doppler synthesis results derived from a series of flight legs confirm that the nearshore segment of the front was stationary. A representative view of the winds and radar reflectivity in the nearshore domain at 0.5 km (all heights MSL) is shown in Fig. 7. The frontal zone was oriented approximately south–north and was marked by a change in wind speed from about 30 m s\(^{-1}\) east of the front to about 10 m s\(^{-1}\) west of the front over a horizontal scale of 10 km. This zone of cyclonic wind shear coincided with a continuous, elongated band of enhanced precipitation (a reflectivity maximum of \(\sim 25-30\) dBZ). As shown in the cross-front vertical section (AB indicated in Fig. 7a), this low-level enhanced frontal precipitation, embedded within widespread stratiform precipitation, was associated with a deceleration of the low-level cross-front flow component (Fig. 8).

Stronger updrafts (maximum analyzed velocities of \(\sim 1.5\) m s\(^{-1}\)) were found at the leading edge of the front, with a more downshear tilt (toward the warm air mass) above \(\sim 1\) km, in good agreement with the environmental shear profile. As revealed by the vertical profile of the wind taken offshore and shown in Fig. 6, a maximum of the along-front flow component (38 m s\(^{-1}\)) was also observed immediately ahead of the front at a height of \(\sim 1\) km (Fig. 8c). The elevated maximum at this altitude represents the combined effects of two mechanisms. A warm tongue was present at the leading edge of the front (Fig. 5); such low-level baroclinity typically occurs just ahead of cold fronts and induces a low-level jet (Browning and Pardoe 1973). Maximum wind speeds also occur above the surface (commonly near 1 km as in this case) because of friction.

The partial blocking of the flow nearshore within the warm air mass, as seen in the analysis of in situ data (Fig. 4), was also evident in Fig. 7a. The low-level southerly flow was deflected by the coastal terrain to become more south-southeasterly near the coast and hence more parallel to the coastal barrier. While the low-level strongest wind speed (\(\sim 34\) m s\(^{-1}\)) was found \(\sim 25\) km offshore rather than within the partially blocked zone near the coast (Fig. 7b), the alongshore component of the flow was actually stronger near the coast. This is shown in Fig. 9, which is a vertical section of cross-and along-barrier velocities (CD indicated in Fig. 7a) in a frame of reference perpendicular to the orientation of Vancouver Island [\(\sim 43^\circ\) from the north (counterclockwise)]. The cross-barrier flow component at about 0.5 km decreased from \(\sim 20-22\) m s\(^{-1}\) at a position 20 km offshore to \(\sim 14-16\) m s\(^{-1}\) at the coast, with a concomitant increase in along-barrier flow (Fig. 9). The maximum of the enhanced along-barrier flow was pre-
Fig. 8. Vertical cross section of dual-Doppler-derived ground-relative winds along A–B in Fig. 7a. (a) Radar reflectivity (dBZ, shading key at upper left), (b) the component of horizontal velocity parallel to the section (i.e., cross-front flow) with a contour interval of 3 m s\(^{-1}\), and (c) the component of horizontal velocity normal to the section (i.e., alongfront flow) with a contour interval of 3 m s\(^{-1}\). The shading in (b) and (c) indicates regions of vertical velocity greater than 1 m s\(^{-1}\).

Fig. 9. Vertical cross section along C–D in Fig. 7a showing radar reflectivity (dBZ, shading key at upper left) in conjunction with the (a) cross-barrier and (b) along-barrier flow component (2 m s\(^{-1}\) contour interval) at 0635 UTC 13 Dec 1993. Heavy solid line in lower-right portion of each panel indicates height of coastal topography along the section.

Sent inland over the sloping terrain surface, a feature similar to that of barrier jets observed along other approximately two-dimensional mountain barriers (e.g., Marwitz 1987). The results suggest that the orographic influence on the low-level winds extended to a region \(\sim 70\) km from the mountain crest of Vancouver Island (i.e., \(\sim 20\) km from the coastline). This observed offshore extent is broadly consistent with the horizontal scale of the upstream influence of topography (i.e., the Rossby radius of deformation calculated in section 2) as predicted by idealized theory (e.g., Pierrehumbert 1984; Pierrehumbert and Wyman 1985). Above the height of 1.5–2 km MSL, the vertical section indicates a relatively unperturbed wind pattern.

Dual-Doppler synthesis results for the offshore analysis domain at 0503 UTC (Fig. 10) show that the low-level flow was virtually uniformly from the south. The northeast-southwest-oriented front was characterized by a marked discontinuity in wind speed and, hence, a zone of cyclonic shear and convergence. As in the nearshore region, this change in the wind was accompanied by a localized, enhanced region of precipitation with maximum reflectivity greater than 30 dBZ (near \(X = 40\) km, \(Y = 49\) km). A sequence of airborne Doppler observations from three consecutive dual-Doppler volumes during this period show that this segment of the front moved northwestward with a mean speed of \(\sim 4\) m s\(^{-1}\). The cross-front vertical section (Fig. 11) indicates several structural features distinct from those observed near the coast. A deep layer of low-level inflow of 12–18 m s\(^{-1}\) from the warm air...
mass toward the cold air mass was present offshore. In contrast, a relatively shallow and weak inflow was confined to the lowest 750 m (Fig. 8) in the nearshore domain. This difference in inflow is due to the difference in the local orientation of the front in the two regions. The front was oriented north–south in the nearshore region, and northeast–southwest in the offshore region. Since the prefrontal flow was from the south in both regions, it was directed much more toward the front in the offshore region. The slope of the frontal zone (identified as the enhanced shear of the alongfront flow component in Fig. 11c) below 1.5 km was less steep offshore, and its associated sloping updrafts were spread over a greater width (the region of vertical velocities greater than 1 m s\(^{-1}\) spanned ~18 km in the horizontal). The precipitation was generally stratiform except at the surface front, where radar reflectivities greater than 30 dBZ extended to ~500 m from the surface. As in the nearshore domain, a maximum of the alongfront flow component (~33 m s\(^{-1}\)) was found immediately east of the front at a height of ~1 km.

While the coverage from the flight-level data is much less complete than that from the Doppler radar, it is

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**Fig. 10.** As in Fig. 7 except showing the offshore domain at 0.75 km MSL at 0503 UTC 13 Dec 1993. Thick line segment E–F in (a) marks location of vertical cross section shown in Fig. 11. Heavy dashed line in (b) marks approximate position of flight legs used to generate vertical cross sections shown in Fig. 13.

**Fig. 11.** As in Fig. 8 except along E–F in Fig. 10a.
valuable in that it can portray the thermodynamic structures accompanying the kinematic and reflectivity fields presented above. Vertical cross sections were prepared for the nearshore segment of the front using observations collected during 0635–0731 UTC (Fig. 12), and for the offshore segment of the front using observations collected during 0503–0612 UTC (Fig. 13). The locations of flight legs used to construct the nearshore and offshore vertical sections are indicated in Figs. 1, 7b, and 10b. Because the flight legs were not necessarily oriented perpendicular to the front, the flight-level data were projected onto sections normal to the front. Because the front was quasi-stationary during the analysis period, the frontal propagation speed was assumed to be zero in the time–space conversion. The contour analyses presented in Figs. 12 and 13 are performed objectively after projection of these data onto a grid in a vertical cross section normal to the front. Kinematic structures for both nearshore and offshore vertical sections were highly analogous to those obtained from the dual-Doppler analyses shown in Figs. 8 and 11. The nearshore segment of the front exhibited a steeper slope below 1.5 km (1:5 and 1:16 for the nearshore segment and offshore segment, respectively) and its inflow was weaker and confined to the lowest 1 km. The alongfront component of the near-surface winds was about 6 m s⁻¹ stronger in the nearshore section than in the offshore section. The frontal updrafts were of greater peak intensity (~2–3 m s⁻¹) and of relatively smaller horizontal scale (~3–5 km for the nearshore segment and ~8–10 km for the offshore segment) in the flight-level data than from the dual-Doppler synthesis (cf. Figs. 8 and 11), due to the inherent smoothing of the latter.

The sections of potential temperature and equivalent potential temperature include similarities and differences between the nature of the front in the nearshore and off-
Fig. 13. As in Fig. 12 except for the offshore segment of the front during 0503–0612 UTC 13 Dec 1993.

Both sets of sections indicate weaker convective stability ahead of rather than behind the front. Each also shows that the frontal updrafts occurred on the warm side of the equivalent potential temperature gradients; as expected the flow in the cross-frontal plane approximately follows surfaces of equivalent potential temperature. Regions of localized updrafts (downdrafts) tended to be collocated with local minima (maxima) in potential temperature and equivalent potential temperature along individual transects. The horizontal variability in these thermodynamic properties and the vertical velocity was greater in the nearshore than in the offshore section. The frontal zone as defined by the region of horizontal potential temperature and the equivalent potential temperature gradient was much steeper in the nearshore than in the offshore section; this difference is consistent with, but more evident than, the differences seen in frontal slope as defined by kinematic properties. While steeper in the nearshore section, the strength of the front here, as gauged by the differences in potential temperature and equivalent potential temperature across the front at low levels, was less in the nearshore than the offshore section. This result is due to the prefrontal flow below about 500 m being about 0.5 and 1.5 K cooler in potential temperature and equivalent potential temperature, respectively, in the nearshore than in the offshore section. A stronger degree of convective stability was found below the elevated frontal zone (i.e., within the cold air mass) for both nearshore and offshore segments of the front. Even though the offshore segment of the front was retreating during this portion of the flight (i.e., with a warm-frontal-like movement toward the cold air mass), its frontal character, in terms of temperature and wind discontinuities across the front, remained more like a cold front. As shown by these flight-level measurements, the offshore frontal zone included an abrupt change in temperature (~1.8 K) and wind speed (~20 m s⁻¹) over a horizontal scale of only ~6 km. The oceanic warm fronts, as investigated by recent observational studies using high-resolution aircraft measurements, are
generally characterized by a quite gradual transition of temperature and wind over a broad horizontal distance and a relatively flat frontal slope (e.g., Doyle and Bond 2001; Wakimoto and Bosart 2001).

It bears noting that there was a local maximum of potential temperature close to the leading edge of the nearshore segment of the front, immediately ahead of the frontal updrafts (Fig. 12c). While it is therefore possible to conclude that this portion of the front was occluded, we prefer to label it as a cold front. This temperature maximum was limited in its magnitude and its vertical and horizontal extents, and given that the winds in this location were representative of those of the prefrontal region, it does not appear that this air mass had a different recent history than the prefrontal flow, in general. Given the spatial scale of the local temperature maximum, we suspect it was caused by compensating subsidence accompanying the primary frontal updraft.

4. Distortion of the frontal zone

As the offshore segment of the front moved eastward again during the latter part of the flight, distortion of the front occurred near the coast. A schematic diagram (based on the series of dual-Doppler radar analyses) illustrates the movement of the front and its distortion over the period of the flight (Fig. 14). The synoptic perspective provided by the NCEP analyses (Fig. 3) indicated that the entire length of the front tended to move simultaneously across Vancouver Island near and after the end of the flight mission (i.e., between 0600 and 0900 UTC), but the aircraft observations show that the nearshore segment of the front actually remained stationary throughout the flight. The pronounced differences in frontal propagation between the nearshore and offshore regions caused a substantial distortion of the front in the coastal zone.

Evidence for this distortion is provided by a composite of low-level radar reflectivities obtained from three selected dual-Doppler syntheses (0722, 0732, and 0742 UTC) as shown in Fig. 15. The relative locations of these dual-Doppler analysis domains and corresponding flight-level winds are also superimposed on the figure. The front is characterized by a sharp discontinuity in wind speed and a narrow zone of enhanced radar reflectivities (25–30 dBZ). The discrete cores of heavy precipitation and intervening regions of lighter precipitation along the leading edge of the front are similar to those of narrow cold-frontal rainbands (NCFRs) observed previously in midlatitude cyclones (Hobbs and Biswas 1979; James and Browning 1979; Hobbs and Persson 1982; Wakimoto and Bosart 2000).

One of the most striking features in Fig. 15 was a region of distinctly weak reflectivities (less than 15 dBZ) found along a segment of the front near and approximately parallel to the coast. A sequence of dual-Doppler observations in the nearshore region (not shown) indicated this precipitation gap did not occur until the appearance of the distortion of the nearshore segment of the front. A detailed view of kinematic and precipitation structure in the vicinity of the precipitation gap is provided by the dual-Doppler synthesis at 0722 UTC within the north domain in Fig. 15, as shown in Fig. 16. While the low-level airflow has some similarities with that observed ~50 min earlier at 0635 UTC (cf. Fig. 7), the frontal zone clearly became deformed over the intervening period. Note the prominent curve of the isolach contours as well as the axis of maximum cyclonic vorticity, with an apex located ~25 km offshore (Fig. 16b). Within the warm, prefrontal air mass the flow in the immediate vicinity of the coast included slightly more of an easterly component than that farther offshore, presumably due to terrain effects (Fig. 16a).

The distribution of horizontal divergence at 0.5 km indicates that there was a locally enhanced convergence immediately to the north of the distortion point with a maximum of ~2.8 × 10⁻³ s⁻¹ (Fig. 17b), which is much greater than that of the strongest convergence (~1.6 × 10⁻³ s⁻¹) observed earlier along the front at 0635 UTC (Fig. 17a). This intensified convergence was accompanied by heavier frontal precipitation and a stronger frontal character (in terms of the horizontal gradient of wind speed across the front as indicated in Fig. 16b). In contrast, a region of weak divergence (~0.4 × 10⁻⁴ s⁻¹), associated with weaker radar echo (minimum radar reflectivities less than 15 dBZ), was found along the front near and to the south of the distortion point. This variation in low-level convergence along the front is
consistent with the observed variation in precipitation intensity.

The differences in divergence patterns between 0635 and 0722 UTC are clearly related to the change in the orientation of the front. Since the low-level southerlies to the east of the front were stronger than the southerlies on its west side, once a segment of the front developed a northwest–southeast orientation (i.e., south of the distortion point), there was a greater cross-front flow component away from the front on its warm side than toward the front on its cold side. In other words, it was the cross-front component of the flow that was causing the low-level divergence along this segment of the front. Similar geometrical considerations explain the occurrence of enhanced convergence along the northeast–southwest segment of the front to the north of the distortion point.

In this case, the development of distortion in the front (and its associated precipitation features) is largely the

Fig. 15. The horizontal composite radar reflectivity (dBZ shading key at top) at 0.5-km height derived from three Doppler analysis domains (inset top, middle, and bottom squares), corresponding to the synthesis times 0722, 0732, and 0742 UTC 13 Dec 1993, respectively. Flight-level winds (flags) at heights of −380 m (south of the letter H) and −1100 m (north of the letter H) measured during the composite period are superimposed on the reflectivity field and the dotted line indicates the flight track. The thick dashed line approximately marks the surface front position.

Fig. 16. As in Fig. 7 except for showing dual-Doppler analysis time of 0722 UTC 13 Dec 1993. Heavy dashed line in (b) indicates axis of maximum cyclonic vorticity, marking the surface front position.
result of differential frontal propagation between the nearshore and offshore regions. In particular, the stationary character of the front near the coast is likely caused by deflection of the low-level southerlies upstream of the coastal terrain. Note that the abrupt change in frontal orientation occurred in a position collocated with the seaward extent of the zone of deflected low-level flow (~20 km offshore from the coast, denoted by an L in Fig. 14) (cf. Fig. 16). The effect of the deflection by the terrain is to cause backing of the winds and hence a greater component toward the front in the warm air, and a lesser component toward the front in the cold air. In turn, this creates a tendency for reduced (or in this case, virtually zero) frontal propagation speed in the coastal zone. We consider this conjecture to be tentative, because of complications related to variable frontal movement farther offshore (this issue will be further addressed in the next section). Our results are consistent with the recent study by Doyle and Bond (2001), which found that the progress of a warm front slowed markedly in the vicinity of Vancouver Island, but there are important distinctions between the two situations. In the case of Doyle and Bond (2001), the slowing of the warm front (which was oriented approximately parallel to Vancouver Island) was due to the front encountering a trapped mass of cool air that had flowed out of the Strait of Juan de Fuca. As mentioned earlier, in the present case cool air was flowing out of the mouth of the Strait of Juan de Fuca, but it does not appear that it was reaching the front. If not, it was orographic effects acting more on the flow inherent to the front, and less interactions between the front and a preexisting, trapped air mass, that were responsible for the evolution of the front in the coastal zone.

5. Discussion

a. Movement of the front

Many previous observational studies have noted similarities between the leading edge of atmospheric cold fronts and laboratory gravity (density) currents, in terms of their kinematic structure and propagation speed (e.g., Carbone 1982; Parsons and Hobbs 1983b; Shapiro et al. 1985; Roux et al. 1993; Wakimoto and Bosart 2000; Yu and Smull 2000). Yet some studies have also questioned the overall relevance of gravity-current dynamics to atmospheric fronts (e.g., Smith and Reeder 1988). For the present case, both the movement of the observed front (i.e., the retreating/quasi-stationary characteristics) and its kinematic structures did not resemble those of laboratory gravity currents (e.g., Simpson and Britter 1980). This suggests little applicability of gravity-current dynamics to this particular front. Instead, the movement of this front appears to be largely determined by the large-scale, low-level winds on its cold side. These winds evolved as the offshore segment of the front underwent its transition from a quasi-stationary retrograding stage to an eastward-propagating stage. During the early stage, the winds on the cool (west) side of the front were from the south, and hence directed slightly away from the front (Fig. 10a). Over the next 2–3 h, these winds veered from out of the southwest to south-southwest (Fig. 18). While this represented a modest
shift in wind direction, it was sufficient to bring about a cross-front component of 8–10 m s$^{-1}$ toward the warm air. With regard to the nearshore portion of the front during the period of about 0600–0800 UTC, the cross-front component of the winds on its cool side was close to zero (Fig. 7) and then directed away from the front (Fig. 16). This segment of the front actually appears to have retrograded slightly, and rotated from a north–south to a northeast–southwest orientation during the interval between the Doppler syntheses. As in previously observed low-level cold fronts, the frontal zone in both the nearshore and offshore domains appears to have been a semimaterial surface whose movement was controlled principally by the cross-front component of the flow on its cold side.

Latent cooling produced by evaporating hydrometeors can produce a more intense postfrontal rear-to-front flow and a faster movement of fronts (e.g., Oliver and Holzworth 1953; Wakimoto and Bosart 2000). As shown in Fig. 15, widespread stratiform precipitation was occurring behind the present front. Conceivably, diabatic effects helped cause the observed changes in the postfrontal flow. The thermodynamic and radar reflectivity observations, however, do not support such a process as being important for the present case. In particular, the flight-level measurements revealed that the troposphere was generally saturated at low levels behind the front (not shown). The low-level temperatures and precipitation intensities within the postfrontal region remained essentially constant with time. It appears that the changes in the postfrontal region can be best ascribed to subtle evolution in the synoptic-scale flow. As described in section 2, a pressure trough existed well behind the surface front (Fig. 2) and generally progressed eastward with time over the flight period. The approach of this trough brought about a transition in the low-level postfrontal geostrophic winds from southerly to southwesterly and, hence, greater rear-to-front flow.

b. Role of frictional convergence

Unlike classical cold fronts, the present cold front included a minimal low-level directional wind shift and propagation speed, particularly early in the flight (cf. Figs. 7 and 10). Nevertheless, the leading edge of the front did feature enhanced precipitation coincident with a localized updraft, similar to many previously documented cold fronts. Since this region also included substantial cyclonic shear, it is plausible that the surface friction was responsible for the observed frontal updraft. Following Bond and Fleagle (1985), the contribution of frictional convergence to the vertical velocity at the top of the boundary layer is approximated by the expression

$$W_i = \frac{1}{\rho f} \nabla \times \tau \cdot \hat{k},$$

(1)

where $\tau$ is the surface stress, $\rho$ is the air density, $f$ is the Coriolis parameter, and $\hat{k}$ is the unit vertical vector. This expression is based on a vorticity equation for the boundary layer (e.g., Fleagle et al. 1988); in synoptic situations this mechanism is often referred to as Ekman pumping (Fleagle and Nuss 1985). The full expression for estimating the vertical velocity at the top of the boundary layer includes terms related to the local rate of change of vorticity and the horizontal advection of vorticity (Fleagle and Nuss 1985). These terms are expected to be relatively insignificant in the present case, compared to most fronts, because the propagation speed
of the front was relatively small. The surface stress is estimated by

$$
\tau = C_d \rho (V - V_c)(V - V_s) \quad \text{and} \quad (2)
$$

where $C_d$ is the drag coefficient using the formulation of Large and Pond (1982), $V$ is the horizontal velocity, and $V_c$ is the horizontal velocity of the ocean surface (assumed to be zero). Three dual-Doppler synthesis periods (0503, 0635, and 0742 UTC) have been selected in evaluation of (1). The winds at 500 m (for 0635 and 0742 UTC) and 750 m (for 0503 UTC) were used to estimate with the assumptions of zero vertical wind shear between these levels and 100 m (the nominal top of the surface layer), and a log profile in wind speed based on neutral static stability between 100 and 10 m. To obtain representative values, the frictional vertical velocities estimated by computing (1) have been averaged over an elongated zone (~6 km in width) parallel to the front and centered at its leading edge. Calculated results are listed in Table 1. The frictional vertical velocities estimated from each of the dual-Doppler periods were approximately 1.3–1.6 m s$^{-1}$. The magnitudes are close to their corresponding in situ vertical velocities measured at the leading edge of the front. This result suggests that indeed surface friction, through its forcing of boundary layer convergence, was crucial to the low-level structure of the front. This finding is consistent with previous observational (e.g., Browning and Harrold 1970; Bond and Fleagle 1985) and numerical modeling (Keyser and Anthes 1982) studies of cold fronts.

The result that frictional convergence accounts for much of the low-level convergence at fronts with substantial cyclonic shear has interesting implications. First, as implied by (1), the vertical mass flux at such a front depends solely on the difference in alongfront velocity across the frontal zone, assuming that variations in the winds along the front are negligible. The magnitude of the vertical velocity is inversely proportional to the width of the frontal zone; it is therefore reasonable to expect that conditions of lesser (greater) static stability would be accompanied by stronger (weaker) updrafts and hence narrower (wider) frontal zones. Both the vertical mass flux and vertical velocity are independent of boundary layer depth in (1), which implies that deeper (shallower) boundary layers are accompanied by weaker (stronger) low-level convergence, all other factors being equal. The upshot is that orographic effects can be manifested at fronts not just directly through modulation of the cross-front winds, but also indirectly. As discussed above, these indirect effects can be both through modulation of alongfront winds and hence Ekman pumping at the front, and through impacts on boundary layer thermodynamic properties that ultimately help determine the magnitude and spatial scales of the low-level circulations associated with the Ekman pumping.

### Table 1. Estimated frictional vertical velocities averaged along the front from three selected dual-Doppler synthesis periods. For comparison, observed vertical velocities (provided from the flight-level data) averaged within the leading edge of the front (~4–6 km in width) at the altitude of ~360–500 m (MSL) are also indicated.

<table>
<thead>
<tr>
<th>Time UTC</th>
<th>Frictional velocity (m s$^{-1}$)</th>
<th>Observed velocity (m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0503</td>
<td>1.5</td>
<td>1.5</td>
</tr>
<tr>
<td>0635</td>
<td>1.6</td>
<td>1.3</td>
</tr>
<tr>
<td>0742</td>
<td>1.3</td>
<td>1.4</td>
</tr>
</tbody>
</table>

6. Conclusions

The detailed mesoscale structure of a cold front in the vicinity of Vancouver Island on 13 December 1993 has been documented using Doppler radar and flight-level measurements from a NOAA P-3 research aircraft as part of the COAST experiment. Comparisons have been made between two distinct segments of the front that were well sampled by the aircraft. One segment extended from the coast to about 40 km offshore and the other was located ~80–120 km offshore. The analysis shows substantial differences in frontal structure and movement between the nearshore and offshore regions. During the early and middle portion of the aircraft observations, the nearshore segment of the front was oriented south–north and was stationary. The low-level flow within the warm air mass ahead of the front appears to have been deflected by the topography. This deflection consisted of a transition in wind direction from southerlies offshore to south-southeasterlies near the coast. The zone of orographic influence was found to extend to a region ~20 km offshore and confined mainly to the lowest 1.5 km MSL. In distinct contrast to the nearshore segment, the offshore segment of the front was oriented southwest–northeast and retrograded slowly northwesterly. The coastal mountains appear to have had little or no direct influence on this segment of the front. It exhibited deeper inflow (from the warm side) and its associated slope was less steep. While the offshore front exhibited a warm frontal-like movement (i.e., toward the cold air mass) during its westward retreat, its frontal character, in terms of the nature of the temperature and wind discontinuities across the front, remained more like a cold front.

During the later portion of the aircraft observations, the offshore segment of the front moved eastward at 7–10 m s$^{-1}$, while the nearshore segment remained stationary. This difference in frontal propagation between the nearshore and offshore regions resulted in a pronounced distortion of the nearshore segment of the front. A region of weak radar echoes along the front (i.e., a precipitation gap) was evident near the distortion point. Our analyses suggest that divergence perturbations associated with the occurrence of the frontal distortion led to this modification of precipitation along the nearshore portion of the front. Observations of the relationships between frontal distortion and mesoscale
precipitation patterns have been reported previously (Trier et al. 1990; Braun et al. 1997; Wakimoto and Bosart 2000). The present study illustrates how orographic effects can cause frontal distortion. Such a process is expected to be particularly complex for those frontal systems approaching an irregular coastline and/or intersecting a coastal barrier at a significant angle because the effects of coastal orography are not uniform along the front.

Our results indicate gravity-current dynamics are not relevant to the observed front, in terms of its propagation speed and kinematic structure. Instead, the propagation of the front seems to have been due to a combination of orographic and synoptic/mesoscale processes. In particular, orographic effects appear to have been responsible for the deflection of the low-level flow in the nearshore region (due largely to upstream effects of the coastal terrain of Vancouver Island, with the outflow of cool air from the Strait of Juan de Fuca playing a secondary role). This deflection of the flow served to increase (decrease) the component of the winds toward the front on its warm (cool) side and, hence, delay its southeastward progress in the immediate vicinity (within ~20 km) of the coast. In contrast, movement of the front in the offshore region appeared to mostly be related to the changes in synoptic flow. Of particular importance here was the approach of a synoptic-scale pressure trough from the west. This trough served to change the low-level postfrontal winds from southerly to southsouthwesterly, that is, from a direction virtually parallel to the front to a direction with a rear-to-front component.

This cold front was characterized by a relatively small (<20°) change in wind direction, unlike most previously documented cold fronts. As mentioned above, it was essentially stationary until late in the flight, when the offshore portion began making significant propagation toward the warmer air. Yet the leading edge of this front was marked by enhanced precipitation and moderate updrafts (~2 m s⁻¹) over most of its length throughout the period of observation, as is typical of propagating cold fronts with prominent wind shifts. The low-level convergence creating the primary frontal updraft in the present case appears to have been induced by surface friction. Friction was effective at forcing updrafts at the front because even though the directional shift was modest, the cyclonic shear was substantial (10–15 m s⁻¹ over 5–10 km). Estimates of the vertical velocity at the top of the boundary layer based on a vorticity balance (i.e., the basis for Ekman pumping) yielded values similar to those observed. Similar conclusions have been drawn in previous observational (Browning and Harrold 1970; Bond and Fleagle 1985) and numerical modeling (Keyser and Anthes 1982) studies of cold fronts.

This study has revealed that orographic effects can be manifested on fronts in complicated ways. These front–terrain interactions are also mediated through boundary layer effects. Our understanding of these dynamical processes is still incomplete. Further progress on this issue requires future detailed observations and numerical modeling of cold fronts encountering coastal barriers in a variety of terrain configurations and synoptic situations.

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