Vertical Structures of Precipitation in Cyclones Crossing the Oregon Cascades

SOCORRO MEDINA, ELLEN SUKOVICH,* AND ROBERT A. HOUZE JR.

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

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ABSTRACT

The vertical structure of radar echoes in extratropical cyclones moving over the Oregon Cascade Mountains from the Pacific Ocean indicates characteristic precipitation processes in three basic storm sectors. In the early sector of a cyclone, a leading edge echo (LEE) appears aloft and descends toward the surface. Updraft cells inferred from the vertically pointing Doppler radial velocity are often absent or weak. In the middle sector the radar echo consists of a thick, vertically continuous layer extending from the mountainside up to a height of approximately 5–6 km that lasts for several hours. When the middle sector passes over the windward slope of the Cascades, the vertical structure of the precipitation exhibits a double maximum echo (DME). One maximum is associated with the radar reflectivity bright band. The second reflectivity maximum is located approximately 1–2.5 km above the bright band. The secondary reflectivity maximum aloft does not appear until the middle sector passes over the windward slope of the Cascades, suggesting that this feature results from or is enhanced by the interaction of the baroclinic system with the terrain. In the intervening region between the two reflectivity maxima there is a turbulent layer with updraft cells ($>0.5$ m s$^{-1}$), spaced 1–3 km apart. This turbulent layer is thought to be crucial for enhancing the growth of precipitation particles and thus speeding up their fallout over the windward slope of the Cascades. In the late sector of the storm, the precipitation consists of generally isolated shallow convection echoes (SCEs), with low echo tops and, in some cases, upward motion near the tops of the cells. The SCEs become broader upon interacting with the windward slope of the Cascade Range, suggesting that orographic uplift enhances the convective cells. In the SCE period the precipitation decreases very sharply on the lee slope of the Cascades.

1. Introduction

Extratropical cyclones moving from the Pacific Ocean into the mountainous western United States are the main source of precipitation in the populous west coastal region, especially during the nonsummer months. Furthermore, the runoff from these storms constitutes one of the main weather hazards of the region. Understandably, the precipitation processes in Pacific storms moving into the mountainous western United States have been the focus of a long history of mesoscale studies and field projects (e.g., Nagle and Serebrenyi 1962; Elliott and Hovind 1964; Farber 1972; Hobbs 1975; Hobbs et al. 1975; Houze et al. 1976; Matejka et al. 1980; Carbone 1982; Marwitz 1987; Braun et al. 1997; Yu and Smull 2000; Neiman et al. 2004; James and Houze 2005; Houze and Medina 2005; Evans et al. 2005; Garvert et al. 2005; Woods et al. 2005; Ralf et al. 2005; Kingsmill et al. 2006). In this paper, we describe an observational study aimed at understanding precipitation processes in midlatitude Pacific cyclones as they move over an inland mountain range.

The precipitation in extratropical cyclones is controlled by a hierarchy of dynamical and physical processes. Upward air motion occurs on the scale of the baroclinic trough in which the cyclone occurs. Frontogenetic processes associated with warm, cold, and occluded fronts within the storm tend to concentrate the precipitation into elongated mesoscale zones. The frontal precipitation is in turn superposed with still smaller-scale maxima in the form of rainbands and convective-scale features (e.g., Fig. 1). The rainbands in extratropical cyclones have been categorized relative to their location within the synoptic-scale storm (Browning and...

* Current affiliation: University of Colorado, Cooperative Institute for Research in Environmental Sciences, and National Oceanic and Atmospheric Administration/Earth System Research Laboratory, Boulder, Colorado.

Corresponding author address: Socorro Medina, Atmospheric Sciences, University of Washington, Box 351640, Seattle, WA 98195-1640.
E-mail: socorro@atmos.washington.edu

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When an extratropical cyclone moves over a mountainous region, the precipitation pattern is further modified by the interaction of the airflow with the terrain. The hierarchical horizontal patterns of precipitation associated with fronts, rainbands, and smaller features then become difficult to recognize and separate from orographic effects. It is then unwieldy if not impossible to categorize and synthesize the precipitation processes in different sectors of the storm based on their horizontal structures. We suggest that to understand the variation of precipitation mechanisms in a storm as it crosses a mountain range it is more fruitful to focus on the sequence of detailed vertical structures observed at a particular site.

White et al. (2003) used an S-band vertically pointing radar (S-Prof) to analyze precipitation processes in a portion of a frontal system crossing the mountains of coastal California. From this analysis, they deduced precipitation growth mechanisms above and below the melting level as the storm passed over the coastal terrain. In this study, we use S-Prof radar data collected in the Oregon Cascade Mountains as part of the second phase of the Improvement of Microphysical Parameterization through Observational Verification Experiment (IMPROVE-2; Stoelinga et al. 2003) in November–December 2001 (Fig. 2). These observations were available continuously through the entire project whenever precipitation was over the radar. The time section of S-Prof data, available approximately every 30 s with 0.1-km vertical resolution, exhibited certain repeatable characteristic reflectivity and Doppler velocity profiles.

In this study, we link those characteristic vertical profiles of radar data to storm sector.

Figure 1 shows the precipitation pattern found by Nagle and Serebreny (1962) to be typical of maritime cyclones moving from the eastern Pacific Ocean toward the west coast of North America. We have added three line segments to indicate three simple and basic sectors. The early sector is the first to pass over a surface site and is identified with the warm advection region of the cyclone. It often contains a warm front that slopes gradually down toward the center of the storm, and it is characterized by relatively weak precipitation, which is often aloft and does not reach the surface. The middle sector is the portion of the storm where the warm advection transitions to cold advection. In this sector the warm front often reaches the surface, and cold and occluded frontal structures may occur. In principle, it is possible that the middle sectors of different storms contain widely diverse dynamical conditions; however, the middle sectors observed during IMPROVE-2 were very similar, since they came through the experimental domain at similar stages of their life cycles, during a very uniform westerly regime (section 2). Precipitation in the middle sector was predominantly stratiform and moderate in intensity with embedded smaller-scale features. The late sector is in the cold, unstable zone behind the cold and/or occluded front and is often called...
the postfrontal period. It is typically characterized by cellular convective precipitation, which may be locally quite strong but transient. Farber (1972) also subdivided midlatitude cyclones moving from the Pacific Ocean to the Washington Cascade Mountains into three stages: prefrontal, transitional, and postfrontal, which correspond to what we call here early, middle, and late sectors, respectively.

By associating the echo structures in the S-Prof data with these storm sectors, we aim to characterize the precipitation in these three basic periods in a way that will indicate the variation of precipitation processes owing to the passage of each storm sector over the mountain range. We will show that high-resolution radar reflectivity and Doppler velocity structures change sharply from one storm sector to the next and that the unique echo structures in the early, middle, and late phases are repeatable from storm to storm. Since the radar echo structures are closely related to precipitation mechanisms, these results indicate the variability of precipitation processes across the storm.

2. Synoptic setting

During the IMPROVE-2 project, the large-scale flow was remarkably persistent. Figure 3 shows a longitude–time plot of the 850-hPa zonal wind averaged between 40° and 50°N. The location and time period of the IMPROVE-2 project are encompassed by the black box. Within this region, the flow was particularly steady, with westerly winds persisting throughout the field experiment. A total of 16 baroclinic shortwave troughs formed in this environment and passed over the study area during the project.

3. Data and method of analysis

A key instrument used in this study is the S-Prof radar, which was deployed on the windward slope of the Oregon Cascades at an elevation\(^1\) of 0.512 km (Fig. 2). The S-Prof is a vertically pointing Doppler S-band radar operated by the National Oceanic and Atmospheric Administration/Earth System Research Laboratory (NOAA/ESRL). As baroclinic systems moved from the Pacific Ocean into the study region, they interacted with the terrain of the Cascade Mountains. From its location, the S-Prof radar surveyed each weather system that moved over the Oregon Cascades throughout IMPROVE-2. During IMPROVE-2, the S-Prof radar emitted rays alternately at pulse lengths of 300 ns (range resolution of 45 m, maximum range of \(\sim 3.8 \text{ km} \)) or 700 ns (range resolution of 105 m, maximum range of \(\sim 9.8 \text{ km} \)). For this study, the 700-ns data is utilized\(^2\) to focus on the processes occurring in the region extending above the 0°C level. Instrument specifications are given in more detail by White et al. (2000, 2003).

A network of scanning radars was also part of the IMPROVE-2 field program. The National Center for Atmospheric Research (NCAR) S-band dual-polarized radar (S-Pol) was located approximately 80 km west of the Cascade crest (Fig. 2). It focused on mapping the vertical structure of the precipitation over the lower

\[^{1}\] Unless otherwise stated, all altitudes are above mean sea level (MSL).

\[^{2}\] The S-Prof data presented in this study have been corrected for the signal saturation that was often observed below intense bright bands (A. White 2006, personal communication). This procedure decreased the mean temporal resolution of the data from 3–8 to 30 s. Additionally, all of the S-Prof data have been thresholded at 0 dBZ.
windward slopes of the Cascades by primarily collecting data over an eastward-looking sector. Broader Doppler radar coverage was provided by the Portland, Oregon, National Weather Service (NWS) Weather Surveillance Radar-1988 Doppler (WSR-88D; Fig. 2).

We also employed a full range of synoptic and mesoscale data from soundings, model analysis fields, and geosynchronous satellite. Rawinsonde data were from Salem, Oregon (plus symbol in Fig. 2). Infrared satellite images were from Geostationary Operational Environmental Satellite-10. Synoptic fields from the fifth-generation Pennsylvania State University–NCAR Mesoscale Model (MM5) 36-km horizontal resolution operational runs conducted at the University of Washington (UW) were used to place the radar data into a large-scale synoptic context. The temperature and geopotential height fields from the MM5 simulations used in this study were validated against observed analysis at 0000 and 1200 UTC. The main features of the fields, such as the location and intensity of the maxima and minima compared well with the observations.

In a study similar to ours, Kingsmill et al. (2006) analyzed midlatitude cyclones moving over the California Coastal Range. They compiled composites of the S-Prof radar data by first defining synoptic situations and then averaging the S-Prof data in each regime. The vertical profiles of radar data, however, differed only slightly from one regime to another. Perhaps this lack of distinctive differences arose because it is hard to a priori define characteristic synoptic regimes, which consist of so many large-scale variables, in a way that relates distinctly to radar structure. Moreover, the precise boundaries of regimes are not always obvious from synoptic data. We found a more clear result by reversing the procedure: first identifying characteristics echo types within the S-Prof data and then inquiring about the associated synoptic situations with which they were associated. This approach led to a distinct 1:1 association between the characteristic radar structures and early, middle, and late cyclone sectors.


One of the most intense storms passing over the IMPROVE-2 data network occurred near the beginning of the project on 28–29 November 2001. Before examining the detailed vertical structure of the precipitation, we describe the synoptic-scale situation of the storm with the aid of Fig. 4.

a. Large-scale conditions

During the early stage of the passage of the 28–29 November storm, a 500-hPa low pressure system was located west of the IMPROVE-2 study area, and westerly winds (inferred from the geopotential height contours) were advecting warmer air toward the Oregon coast (Fig. 4a). At 850 hPa, temperature gradients were clearly defined and warm advection was occurring near the Oregon coast (Fig. 4b). A well-defined cloud shield with a sharp back edge was beginning to move over the region; however, the coldest (highest) cloud tops were still far west of the experimental area (indicated by the X in Fig. 4c). By the time the middle sector was passing over the experimental area, the 500-hPa trough had deepened and progressed eastward (Fig. 4d), and the previously strong 850-hPa temperature gradients had moved eastward and weakened (Fig. 4e). The coldest cloud tops were nearly directly over the region of study (Fig. 4f). During late sector passage, westerly winds were advecting slightly cooler air toward the Oregon coast (Fig. 4g), and at 850 hPa the horizontal temperature gradients were weak in the Pacific Northwest and adjacent seas (Fig. 4h). Small cellular cloud tops were observed over the experimental area and the main cloud shield was located east of the domain (Fig. 4i).

b. Static stability and winds

Time cross sections of wind, potential temperature ($\theta$, solid lines) and equivalent potential temperature ($\theta_e$, shaded contours), measured upstream of the terrain, at Salem (plus symbol in Fig. 2), are shown in Fig. 5a. During the early sector of the storm (~0800 UTC 28 November), $\theta$ and $\theta_e$ increased with height, indicating stability (Fig. 5a). The maximum $\theta_e$ values at nearly all levels occurred when the middle sector was passing over (~1500 UTC 28 November). During the late sector (~0300 UTC 28 November), the $\theta_e$ contours were oriented vertically from approximately 0.5 to 6 km, indicating moist static neutrality, with a shallow potentially unstable layer ($d\theta_e/dz < 0$) within 1 km of the surface and strong stability above ~6 km.

Figure 5b displays the MM5-simulated fields corresponding to the sounding analysis in Fig. 5a. During the early sector, winds were south-southeasterly near the surface. They intensified and veered with height to westerly near the top of the domain. In the middle sector, they were strong at all levels and slightly veered from south-southwesterlies near the surface to west-
southwesterlies near the top of the domain. The low-level easterly component was no longer present.

The late sector on 28–29 November was interrupted by a secondary trough passage, evident in the low-level southwesterly flow between 0000 and 1000 UTC 29 November (Fig. 5b). Between 1200 and 1600 UTC 29 November, the low-level flow sharply shifted to westerly and the $\theta_e$ values dropped further, indicating that the secondary trough had passed over the site (Fig. 5b).

c. Mesoscale precipitation features seen in horizontal displays of radar data

As observed in previous studies by Houze et al. (1976), and Matejka et al. (1980), the mesoscale substructure of a storm is often evident in the radar PPI scans. Figure 6 contains 0.5° elevation PPI scans of the Portland WSR-88D. During the early sector of the storm, echoes first appeared aloft (not shown) and gradually descended toward the surface until a widespread layer of echo nearly covered the radar domain (Fig. 6a). The radial velocity (Fig. 6b) showed south-easterly low-level winds and a pronounced veering of the winds, consistent with the large-scale maps (Figs. 4a,b) and the Salem sounding (Fig. 5). In the middle sector, where both the winds speeds and the $\theta_e$ values are maximum (Fig. 5), mesoscale bands of high reflec-

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**Fig. 4.** Progression of the 28–29 Nov 2001 storm over the IMPROVE-2 area. (a), (d), (g) Model 500-hPa geopotential height (60-m intervals in black lines) and temperature (shaded); (b), (e), (h) as in (a), but for 850 hPa; and (c), (f), (i) infrared satellite imagery. Model output is from the UW MM5 real-time 36-km simulation. The X indicates the location of the S-Prof radar (discussed in the text). Panels (a)–(c) correspond to 0800 UTC 28 Nov. The model output in (a), (b) is from an 8-h forecast initialized at 0000 UTC 28 Nov. Panels (d)–(f) correspond to 1500 UTC 28 Nov. The model output in (d), (e) is from a 3-h forecast initialized at 1200 UTC 28 Nov. Panels (g)–(i) correspond to 0300 UTC 29 Nov. The model output in (g), (h) is from a 3-h forecast initialized at 0000 UTC 29 Nov.

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4 In Fig. 6, distances from the radar of 100 (inner ring) and 200 km (outer ring) correspond to altitudes of ~1.9 and ~4.6 km, respectively.

5 When the zero radial velocity contour exhibits an S shape in the PPI, the winds are veering with height (Baynton et al. 1977), which is often associated with warm advection.
activity were embedded in the widespread echo (Fig. 6c). The radial velocity indicated southerly winds at low levels, with southwesterly to west-southwesterly flow at higher levels (Fig. 6d), consistent with the large-scale flow (Figs. 4d,e, and 5). In the late storm sector, as low-level instability passed over the study area (Fig. 5a), the echoes became more scattered and cellular (Fig. 6e). Radial velocity indicated that southwesterly flow dominated at all levels (Fig. 6f), consistent with Figs. 4g,h and 5.

**d. High-resolution vertical structure of radar echoes**

The S-Prof radar depiction of the vertical structure of the precipitation in the 28–29 November storm is shown in Fig. 7. We used time sections like these to identify early, middle, and late sectors of each storm in IMPROVE-2. Figure 7a shows the characteristic reflectivity sequence. In the early sector of the storm (approximately 0000–1300 UTC 28 November), the echo appeared first aloft with a top at 7–8 km and a lower echo boundary that descended toward the surface with time. The middle sector of the storm (approximately 1500 UTC 28 November–0000 UTC 29 November) consisted of a vertically continuous echo layer containing a radar bright band slightly below 2 km, just below the 0°C level shown by the Salem sounding (crosses in Fig. 7). The late sector of the storm (approximately 0100–1200 UTC 29 November) consisted of relatively isolated echoes with tops at approximately 2.5–5 km. The storm sectors are discussed in more detail below.

Figure 7b shows the vertical radial velocity corresponding to the reflectivity profiles in Fig. 7a. No correction was applied to the radial velocity to subtract the particle fall speed. Thus, positive radial velocities are a lower limit of updraft speed. Regions shaded in red indicate particles moving upward at speeds of at least 0.5 m s⁻¹ and show embedded updraft cells. The transition from white to blue shows the sudden increase in fall velocity at the melting level (White et al. 2003).

**1) Early storm sector**

Figures 4a–c and 6a,b show the relative position of the S-Prof site (indicated by the X) with respect to the early sector synoptic fields, satellite-observed cloud shield, and general radar echo pattern. The vertical echo structure observed in the early sector of the storm is shown with more detail in Figs. 8a,b. We call this structure the leading edge echo (LEE). The descent of the base of the LEE is consistent with the classic cyclone model (Bjerknes and Solberg 1926; Carlson 1998) and with radar observations in the early stages of a storm passing over the Sierra Nevada Range in California (Rauber 1992).

The LEE reflectivity (Fig. 8a) was stratiform. However, a bright band could not be distinguished during the first 10 h of the LEE period. Since the echo was very deep the precipitation at higher levels was most likely ice or snow. The upstream Salem sounding data suggest that the 0°C level was ~1 km during this inter-
Fig. 6. Portland WSR-88D radar 0.5° PPI of (a), (c), (e) reflectivity and (b), (d), (f) radial velocity for (a), (b) 0800 UTC 28 Nov 2001; (c), (d) 1530 UTC 28 Nov 2001; and (e), (f) 0300 UTC 29 Nov 2001. The radar was operating in clear-air mode at 0800 UTC 28 Nov 2001; therefore, the color scale in (a) is different from that in (c) and (e). The range ring spacing is 100 km. Negative (positive) velocities are toward (away from) the radar.
val (indicated by the crosses in Figs. 8a,b). In the Cascades, cold easterly flow typically comes through the passes during the early stages of storm passage, preventing melting from occurring at low levels. Farber (1972) pointed out that in early (“prefrontal”) sectors of Pacific Northwest storms, characterized by deep and high cloud tops, cold-type ($T < -20^\circ C$) ice crystals dominated over the Cascades. The S-Prof reflectivity from 0000 to 1000 UTC 28 November decreased with decreasing height near the lower boundary of the echo, suggesting that the ice crystals were small and were probably undergoing sublimation. According to Farber (1972), ice crystals in the early sector of storms moving from the Pacific Ocean toward western Washington (just north of the IMPROVE-2 area) tend to be small, rarely exceeding ~1000 $\mu$m in length. Melting was only suggested after 1000 UTC 28 November, where a subtle bright band is seen in the LEE period (Fig. 8a). The bright band coincides with the lower $0^\circ C$ level seen at 1200 UTC 28 November.

During this event, Ikeda et al. (2005) analyzed S-Pol radar data and noted a double bright band at ~1009 UTC 28 November that lasted for ~30 min. This feature is consistent with the height and occurrence of a double $0^\circ C$ level observed in Salem at 1200 UTC 28 November (Fig. 8a). According to Ikeda et al. (2005), the double bright band was produced as a maritime and relatively warm air mass advancing from the southwest overran the cold continental air mass originating east of the Cascades.

The Doppler velocity in the LEE (Fig. 8b) showed little significant upward motion of precipitation particles (i.e., radial velocities $<0.5$ m s$^{-1}$). The absence of upward motions strong enough to generate liquid cloud water, which subsequently could be accumulated on the ice particles by riming, is further consistent with the idea that the ice particles were relatively small. Closer to the surface, the radial velocities were only slightly negative, indicating low fall speeds, consistent with the precipitation being composed of small particles.
2) MIDDLE STORM SECTOR

Figures 4d–f and 6c,d show the relative position of the S-Prof site (indicated by the X) with respect to the middle sector synoptic fields, satellite-observed cloud shield, and general radar echo pattern. The radar echo structure in the middle sector of the storm (shown in detail in Figs. 8c,d) is of the type described by Houze and Medina (2005). We call it the double maximum echo (DME) region because it had reflectivity maxima at two altitudes: a bright band was located just below 2 km, while a secondary maximum of reflectivity occurred at an altitude of ~4 km (Fig. 8c). In between the two reflectivity maxima was a layer of cellular overturning (Fig. 8d). Houze and Medina (2005) and Rotunno and Houze (2007) have discussed how this turbulent layer likely plays an important role in the enhancement of precipitation as a cyclonic storm passes over the windward slope of a mountain range. In section 6, we present evidence that the echo structure observed by S-Prof in the middle sector was either produced or enhanced over the windward side of the mountain range.

The lower-level reflectivity maximum was clearly associated with melting. The S-Prof radial velocity field (Fig. 8b) shows strong negative values below the bright band, where the particles have melted into rapidly falling drops. The origin of the upper-level reflectivity maximum is less clear and will be discussed in section 6. Within the DME period and near the surface, the S-Prof radial velocity field shows strong negative values associated with large raindrop fall speeds (Fig. 8d). Soon after the back edge of the contiguous frontal cloud shield passed over the S-Prof site, the turbulent layer was no longer apparent on the S-Prof radar, nor was the upper-level reflectivity maximum.

3) LATE STORM SECTOR

Figures 4g–i and 6e,f show the relative position of the S-Prof site (indicated by the X) with respect to the late sector synoptic fields, satellite-observed cloud shield, and general radar echo pattern. At this stage, the main cloud shield had already passed over the S-Prof site (Fig. 4i). The radar echo observed by the S-Prof in the late sector of the storm (approximately 0100–1200 UTC 29 November in Fig. 7) is called the shallow convection echo (SCE) period. The reflectivity pattern of the SCE consisted of relatively isolated and shallow cellular con-
vective echoes (Fig. 8e). A bright band is subtly suggested in the reflectivity field. The melting layer is well defined in the radial velocity cross section slightly above 1 km (Fig. 8f), which is consistent with the 0°C level indicated by the Salem soundings (crosses in Figs. 7e,f). Large upward motions (i.e., radial velocities > 0.5 m s⁻¹) were observed during the SCE period, particularly at the top edges of the cellular echoes (Fig. 8f).

5. Comparison to other storms

The LEE, DME, and SCE radar structures, just described for the 28–29 November storm were recurrent features during IMPROVE-2. Although every storm did not necessarily exhibit all three echo structures, these signals were repeatedly observed. In this section we present examples from other storms to illustrate their repeatability.

a. LEE period

Figure 9 shows the synoptic situations in which LEE structures were observed in three additional storms. Comparing Figs. 4a–c with Fig. 9 shows that the S-Prof radar (indicated by the X) was in a similar location relative to the early sector of each storm. As the leading portion of the upper-level cloud shield for each case was passing over S-Prof (Fig. 4c and Figs. 9c,f,i), the 850-hPa geopotential height and temperature fields showed warm advection over the site (Figs. 4b and 9b,e,h) and the main baroclinic trough was located up-stream (Fig. 4a and Figs. 9a,d,g). To facilitate comparison with these three additional cases, the LEE period of 28–29 November is displayed again in Figs. 10a,b. The reflectivity panels (Figs. 10a,e,g) show that the overall shape of the echoes was similar in all four cases. This shape was also seen with an 8.6-mm wavelength vertically pointing radar by Locatelli and Hobbs (1987) in
the early sector of a storm moving over the Washington coast (their Fig. 7). Each case showed a particularly steep segment in the lower reflectivity boundary just before the echo reached the surface (e.g., 0630–0700 UTC 10 December in Fig. 10c). The back edge of the LEE tended to drop abruptly to the ground. This back edge of the echo may have marked the effective edge of the warm frontal lifting zone, as suggested by Locatelli and Hobbs (1987). Just prior to the conclusion of the LEE period, most of the cases showed the passage of a relatively deep and short-lived feature suggestive of a rainband (e.g., 1100 UTC 28 November in Fig. 10a).

The LEE period also exhibited some variability in its substructure from case to case. In particular, the LEE period of 18 December consisted of areas of higher reflectivity and more embedded cellular upward motion than the other cases (Figs. 10g,h). The cloud system on 18 December was closer to the cold trough aloft and the 500-hPa temperatures were colder than in the other cases (cf. Figs. 9g–i with 9a–c, d–f, and 4a–c). This position of the LEE relative to the cold upper trough evidently favored convective upward motions within the LEE period. Interestingly, the overall LEE structure was maintained even though it contained more unstable air and embedded convection.

b. DME period

Figure 11 shows the synoptic situations for three additional times when DME structures were observed.
In each of these additional cases, the DME occurred in the middle sector of the storm (Fig. 11), similar to 28–29 November case (Figs. 4d–f). In each case, the 500-hPa trough was near the IMPROVE-2 domain, and the cloud tops of the upper-level cloud shield were passing over the experimental area. Figure 12 compares the time sections of the S-Prof reflectivity and radial velocity of the four observed DME structures. In each case, a layer of updraft cells comprising a turbulent layer of the type described by Houze and Medina (2005) was located between two layers of enhanced reflectivity: the melting layer at lower levels and an upper-level layer approximately 1 to 2.5 km above the lower maximum. A notable variation of the radial velocity structure occurred on 18 December, when the updrafts tended to extend in height to the tops of the echo (Fig. 12h). As noted in the discussion of the LEE period (section 5a), this date was characterized by upper-level instability, associated with a cold trough aloft, located farther south and east than in other cases. Overall, the similarities of the DME structures in the four storms illustrated in Fig. 12 is striking. Houze and Medina (2005) noted this structure in IMPROVE-2 and also in Alpine storms.

c. SCE period

Figure 13 shows the large-scale environments of three additional IMPROVE-2 periods when the S-Prof observed a SCE signature. In each case, the late sector

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6 The middle sectors shown in Fig. 11 do not correspond to the storms whose early sectors are shown in Fig. 9, except for the bottom rows.

7 The late sectors shown in Fig. 13 do not correspond to the storms whose middle sectors are shown in Fig. 11, except for the top row of Fig. 11 and the bottom row of Fig. 13, which correspond to the same storm.
was moving over the S-Prof (indicated by the X). The satellite imagery for each case showed shallow clouds (with warm tops) over the experimental area. Figure 14 compares the S-Prof radar reflectivity and radial velocity structures for these three additional SCE cases with the 28–29 November event. All cases differed significantly from the LEE and the DME periods. Echo-top heights were lower than in the LEE and DME regimes. Even when the cells were in mesoscale clusters or otherwise connected by weak echo, the embedded regions of higher reflectivity within the clusters formed distinct narrow and vertically aligned cells. The S-Prof radial velocity data indicated that some cells within the SCE period had positive (upward) radial velocity, usually located near the tops of the echoes. In many cells, significant updrafts were not observed, suggesting that the radar echo was most likely fallout from previously, but no longer, active updrafts. Alternatively, the updrafts could have been located in the clouds above the precipitation showers, at levels where particles had not yet reached radar-detectable size. Finally, it is also possible that the falling precipitation within the cells masked the updrafts.

There was some variability in the SCE substructure. The 28–29 November SCE had cells that were deeper and had stronger updrafts (Figs. 14a,b) than in the other cases (Figs. 14c–h). This variability was probably associated with the large-scale environment. The SCE in

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**Fig. 12.** As in Fig. 7 but during the passage of DME periods: (a), (b) 1600–1900 UTC 28 Nov 2001; (c), (d) 2000 UTC 13 Dec–0200 UTC 14 Dec 2001; (e), (f) 0000–0500 UTC 17 Dec 2001; and (g), (h) 1720–2000 UTC 18 Dec 2001. Note that the color scale in (a) is different than those in (c), (e), and (g).
Figs. 14a,b occurred during the passage of a secondary trough (section 4b) and under low-level southwesterly winds (Figs. 4g,h), as opposed to the other three cases, where the low-level flow was north-northwesterly, and the 500-hPa trough was already over or just downstream of the S-Prof (Fig. 13).

6. Orographic effects

The S-Prof radar observations presented in the previous section were collected over the windward slopes of the Cascade Mountain Range (Fig. 2). Therefore, it is not known whether the features described in section 5 were a part of the baroclinic system before it reached the barrier or if they resulted from the interaction between the baroclinic system and the Cascade terrain. Ideally, concurrent data from vertically pointing radar located near the Oregon coast and/or in the Willamette Valley would have provided us with critical information to address this issue; however, such a dataset was not collected during IMPROVE-2. In this section, we instead use S-Pol radar (Fig. 2) data collected on 28–29 November to investigate orographic effects on midlatitude cyclones. The S-Pol radar collected an RHI scan (rotation of the antenna in elevation at constant azimuth) over roughly the same azimuth angle about every 11 min. We have analyzed data from two azimuth angles: one extending over the windward slopes of the Oregon Cascades and another pointing upstream of the Cascades. The S-Pol RHI scans used in this section were collected exactly due east, over the windward slope of the Cascades and toward the west, upstream of the radar. The upstream azimuth varied with low-level wind direction as follows: 250° azimuth from 0000 to 1734 UTC 28 November; 270° azimuth from 1743 to 1929 UTC 28 November; and 225° azimuth from 1938 UTC 28 November to 1449 UTC 29 November.
of the vertical structure of radar echo as a function of distance from the radar in the cross-barrier direction.

a. Cross-barrier distribution of reflectivity aloft as a function of time

Figure 15 shows the maximum S-Pol reflectivity as a function of time and distance from the radar in the cross-barrier direction on 28–29 November. For each RHI scan, we calculated the maximum reflectivity in the 3–10-km layer. Echoes below 3 km were ignored to avoid contamination by high reflectivity values in the melting layer and because the radar’s view east of the Cascade crest was blocked below this level. The individual regions of the storm are distinct and clearly separated over the Pacific Ocean at −100 to −150 km from the radar (Fig. 15a). Where these features extend over the western slopes of the Cascades (0–90 km in Fig. 15b), the individual periods are still identifiable but they are located in closer proximity to each other. Where the LEE and DME move from the Coastal...

9 The observations at −67 km from the S-Pol radar in Fig. 15b roughly correspond to the range of the S-Prof radar. However, the S-Prof was located at an azimuth angle of −110° from the S-Pol (i.e., outside the 90° azimuth used to construct Fig. 15b).
Range (−100 to −40 km) to the Willamette Valley (−40 to 0 km in Fig. 15), the maximum reflectivity values decrease, probably associated with downslope drying on the lee side of the Coastal Range. Over the Pacific Ocean (−150 to −100 km in Fig. 15a), the SCE consists of a few isolated cells with nearly echo-free areas around them. As the SCE reaches the steepest western slopes of the Cascades, the reflectivity also consists of isolated cells but the area around the cells is no longer echo free (range of 35–90 km in Fig. 15b). Thus, the precipitation associated with the convective cells becomes more contiguous over the windward terrain, likely as a result of orographic lifting. This result is in agreement with the observations of postfrontal precipitation made by Woods et al. (2005) in the 13–14 December 2001 IMPROVE-2 storm and by Farber (1972) in the western Washington storms.

Over the western slopes of the Cascades, the DME consists of two broader portions separated by a thin reflectivity maximum, clearly defined at 1900 UTC 28 November. This feature is a narrow cold frontal rainband (NCFR; Kessler and Wexler 1960; Browning and Parrod 1973; Houze et al. 1976; Carbone 1982; Hobbs and Biswas 1979; Hobbs et al. 1980; Hobbs and Persson 1982; Houze 1993, 475–481; Braun et al. 1997), readily identifiable when PPI scans of reflectivity from the S-Pol radar are animated and also seen in Fig. 7a as the short-lived feature with echo tops near 7 km at 2000 UTC 28 November. Despite the passage of the NCFR, the upper-level echo maximum remained in evidence until about 0000 UTC 29 November (Fig. 7a). This behavior suggests that the upper-level maximum is associated with orographic effects tied to the mountain range rather than an inherent feature of the midlatitude cyclone.

Another orographic effect is evident in the SCE. Over the Willamette Valley (−40 to 0 km in Fig. 15a) the occurrence of echoes outside the cells decreased

Fig. 15. Maximum reflectivity (in the layer between 3 and 10 km) during the 28–29 Nov 2001 storm as a function of range from the S-Pol radar. The calculation used data from RHI scans collected at an azimuth facing (a) west and (b) east of the radar. (c), (d) The underlying orography as a function of range from the radar. The white horizontal line is a period of missing data. The orography in (c) is directly due west of the radar, while the azimuth direction of the data in (a) changed with time (see text).
from what was observed over the Pacific Ocean. This result suggests that convection may be suppressed by downward motion in the lee of the Coastal Range. The SCE precipitation completely collapsed just to the lee of the Cascade crest (ranges >90 km in Fig. 15b), where downslope motions were pronounced (Fig. 15d).

This result agrees with Farber’s (1972) finding of clearing on the eastern slopes of the Cascades in postfrontal conditions.

b. Cross-barrier distribution of precipitation and airflow as a function of height

Mean vertical cross sections of S-Pol reflectivity data have been constructed for each of the periods (Fig. 16). The reflectivity cross sections (Figs. 16a,c,e), extend from west to east of the radar site. Radial velocity extends only to the east of the radar (Figs. 16b,d,f). Since the radar did not scan exactly due west during most of this storm (see footnote 8), the westward-looking radial velocity is not readily comparable with the eastward-looking one.

1) LEE PERIOD

In the LEE period the S-Pol reflectivity (Fig. 16a) was generally lower on the western slopes of the Cascades than farther upstream, probably as a result of the low-level easterly flow (Fig. 16b), producing downslope winds on the western slopes of the Cascades and suppressing the precipitation. The easterly flow below ~1 km brought cold continental air over the mountains at low levels, consistent with the low bright band in Fig. 8a. Figure 16b shows the radial velocity changing to a strong westerly component aloft, consistent with the warm advection occurring over the region (Figs. 4b and 5b).
2) DME PERIOD

Comparison of the upstream and downstream cross sections of reflectivity in the DME period (Fig. 16c) suggests that the DME structure seen over the windward slope of the Cascade Range was orographically induced. First, the brightband structure was enhanced over the western slope compared to that over the Willamette Valley (cf. echo between −20 and 0 km with that between 0 and 20 km). Second, the upper-level secondary reflectivity maximum (at heights of ∼3.5 to 5.5 km) exhibits a well-defined structure over the windward slope of the Cascade Range but is not evident upstream of the Cascades (Fig. 16c). The secondary maximum of reflectivity thus appears to have been produced or enhanced by interaction of the baroclinic system with the terrain.

The physics of the secondary reflectivity maximum aloft are not entirely clear. Since being first noted by Houze and Medina (2005), enhanced reflectivity at altitudes ∼1.5–2.5 km above the 0°C level have been reported in midlatitude cyclones passing over California (Kingsmill et al. 2006). In all cases, the upper-level layer of enhanced reflectivity is well above the 0°C level, where ice particles dominate. The Salem sounding at 1800 UTC 28 November indicated that the temperature at the level of the secondary reflectivity maximum (∼4 km in Fig. 8c) was ∼8°C. Houze and Churchill (1987) found aggregates concentrated in a similar temperature regime in stratiform precipitation areas in the Asian monsoon. They found the aggregates to be associated with large dendritic crystals that had apparently grown by deposition at higher levels and drifted downward and aggregated at this level. These factors suggest that the upper-level reflectivity maximum is associated with aggregation producing large particles that have high reflectivity because of their large size. Kingsmill et al. (2006) arrived at a similar explanation in regard to California storms.

However, this purely microphysical explanation does not account for the occurrence of the enhanced reflectivity aloft only over the windward slope of the mountain range (Fig. 16c). Garvert et al. (2007) found that the upper maximum of reflectivity (in the 13–14 December case illustrated in Figs. 12c,d) coincided with enhanced upward velocity associated with a vertically propagating gravity wave anchored to the crest of the Cascades. Kingsmill et al. (2006) rejected an orographic dynamic explanation because they observe this enhancement at a site in the Central Valley. However, the Central Valley is unlikely to be free from orographic influences. Under several prevailing flow directions, the site is downstream of the Coastal Range and, perhaps more importantly, its nearness to the Sierra Nevada foothills suggests that it may also be under upstream effects from the larger mountain range. The fact that Fig. 16c shows the secondary maximum only over the windward slope of the mountain range suggests that a dynamic orographic enhancement may be needed to bring this feature out strongly.

Although the secondary reflectivity maximum aloft is interesting, it is not important to the orographic enhancement of precipitation on the west slopes of the Cascades. High values of cross-barrier flow were observed by the S-Pol near the upper reflectivity maximum (Fig. 16d), suggesting that the precipitation that formed here did not fallout over the windward slopes but was instead advected farther downstream, as discussed by Medina et al. (2005) and Garvert et al. (2007). The importance of the secondary maximum to this study is that it indicates that the radar echo over the windward slope had a characteristic structure that was unique to the zone of orographic lifting and absent upstream. Most of the precipitation enhancement on the windward slope of the barrier was associated with lower-altitude echo structure. Houze and Medina (2005) and Rotunno and Houze (2007) have discussed how the precipitation fallout as the storm passes over is evidently enhanced by a layer of small-scale updrafts (∼1–3 km wide) in a topographically enhanced shear layer separating weak low-level flow from stronger upper-level flow (Medina et al. 2005). During the DME period, shear was indeed concentrated in a thin layer near the surface (clearly seen between 0 and 20 km in Fig. 16d).11 The shear in this layer was much stronger than the moderate shear seen earlier during the LEE period (Fig. 16b). The layer of high shear is a key feature of the orographic dynamic modification of the middle sector of the storm. It provides a layer in which cellular overturning can occur and accelerate particle growth at the low levels over the windward slope.

3) SCE PERIOD

In the SCE period, the reflectivity had a transient character (Fig. 15a). A 9-h average of reflectivity data (Fig. 16c) exhibits a bright band at ∼1 km and mean echo tops at ∼4 km. The leeward side of the Cascades

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11 Medina et al. (2005) used airborne radar data to show that the shear layer is observed throughout the extent of the windward slope of the terrain. The shear layer is not seen at far ranges by ground-based radars, like the one used to produce Fig. 16, because of beam widening and blocking by the terrain.
is nearly echo free (Fig. 16e). At this stage the radial velocity is positive (i.e., had a westerly component) at all levels (Fig. 16f).

c. Low-level wind reversal

The three-dimensional structure of the airflow during the different sectors of the 28–29 November storm can be inferred by jointly analyzing the radial velocity signatures shown in Figs. 6 and 16. While Fig. 6 represents a snapshot, the data in Fig. 16 are robust time means. In the LEE period, the southeasterly low-level flow (Fig. 6b) has an easterly projection in the cross-barrier direction (Fig. 16b). The southerly low-level flow observed during the DME (Fig. 6d) has a weak westerly projection in the cross-barrier direction (Fig. 16d). The transition in the nature of the cross-barrier low-level flow from easterly during the early sector to westerly during the middle sector is often observed in Pacific Northwest cyclones reaching western Washington (e.g., Farber 1972; Hobbs 1975). While in the LEE period there was a moderate amount of shear (extending from the surface up to ~6 km), during the DME period the shear and radial velocity increased significantly (Figs. 16b,d). During the SCE period, the deep southwesterly winds’ (Fig. 6f) projection in the cross-barrier direction changed to moderate westerly flow (Fig. 16f).

The orographic effect on the precipitation was then different during each of these periods. During the LEE period, the low-level easterly flow (Fig. 16b) slightly suppressed the precipitation on the western Cascade slopes and most likely enhanced the precipitation on the eastern slopes, in agreement with the observations of Farber (1972). During the DME period, the neutral or statically stable low-level flow was strongly sheared over the western slopes (Fig. 16d), and the precipitation was enhanced by small-scale turbulent cells (Houze and Medina 2005; Rotunno and Houze 2007). In the SCE period, precipitation enhancement on the western slopes was produced as convection was released in westerly upslope low-level flow (Fig. 16f).

7. Conclusions

Previous studies have provided insight into how precipitation varies from one sector of an extratropical cyclone to another. These studies have focused on the horizontal precipitation patterns (especially rainbands) of midlatitude cyclones. However, when extratropical cyclones move over a mountain range, the horizontal precipitation patterns are strongly affected by the terrain and, in turn, become highly complex. To categorize and describe the precipitation processes in extratropical cyclones passing over the Cascade Mountains of western Oregon, we analyzed vertical structures observed by a vertically pointing (S-Prof) radar, deployed on the windward slope of the Cascades during IMPROVE-2. These structures were analyzed in relation to other datasets collected by the scanning S-Pol radar, also located on the windward slope of the Cascades, the Portland NWS WSR-88D radar, model output, soundings, and satellite data. Examination of 16 baroclinic systems moving from the Pacific Ocean into and over western Oregon in a rather constant westerly flow regime suggests a conceptual model of the typical echo structure characteristic of each sector of an extratropical cyclone as it passes over the windward slopes of the Cascade Range.

The early period of a storm’s passage over the windward slopes of a mountain range (Fig. 17a) is marked by an LEE in the warm advection region of the storm. The LEE describes a deep layer of precipitation that appears initially aloft and descends toward the surface, gradually at first, then more abruptly, until a deep stratiform echo period extends from the surface to approximately 6–7 km. Updraft cells inferred from the vertically pointing Doppler radial velocity may be embedded in the LEE at upper levels. When cold air comes in aloft in this sector, the cells in the LEE may be numerous, but often they are absent or weak. The LEE period is similar to the warm frontal radar echo seen by Locatelli and Hobbs (1987) in storms moving over the Pacific Northwest coast and therefore does not appear to be qualitatively altered by the orography. However, it appears to be weakened by the easterly lower-level flow.

In the middle sector of the storm (Fig. 17b), the precipitation over the windward slope is more intense and the radar echo comprises a thick vertically continuous layer extending from the mountainside up to a height of about 5–6 km. This echo region in the middle of the storm takes several hours to pass over a point on the windward slope, and it exhibits a characteristic vertical structure, which we call the DME. The lower echo-intensity maximum is the bright band associated with particle melting. A second region of high reflectivity is located approximately 1–2.5 km above the bright band. This secondary reflectivity maximum aloft is not present when the middle sector of the storm passes over the lowlands upstream of the Cascades but appears and becomes well defined over the windward slope of the range, apparently as a result of dynamic interaction of the baroclinic system with the terrain. Sandwiched between the two reflectivity maxima, the DME of all the cases analyzed contained a layer of turbulent overturning with small-scale updraft cells >0.5 m s$^{-1}$. This layer

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of cellular vertical motions coincided with a strong shear layer over the windward slope. Houze and Medina (2005) and Rotunno and Houze (2007) have suggested that the upward motion cells in the turbulent layer play a key role in the enhancement of precipitation on the windward slope of the mountain barrier.

In the late sector of the storm (Fig. 17c), the precipitation takes the form of generally isolated SCEs. Offshore from Oregon and Washington, the late sector of an extratropical cyclone is characterized by instability with convective cells forming and evolving through their life cycles and at times organizing upscale into lines or groups of cells (Houze et al. 1976). The S-Pol radar shows that as the maritime polar air moves over the windward slope of the Cascade Range, the cells become broader in their horizontal dimensions, probably in response to orographic uplift. The S-Prof data collected on the windward slope show that the SCE period is characterized by shallow cellular echoes with top heights lower than those observed when the LEE and DME moved over the S-Prof site. The radial velocity measured by the S-Prof indicated some upward motion near the tops of the shallow cells. However, updrafts were not always observed. The SCE cells disappeared abruptly on the lee side of the Cascades, in contrast to the precipitation in the LEE and DME periods, which decreased more gradually on the lee side.

Although the three types of echo periods (LEE, DME, and SCE in Fig. 17) do not appear in the S-Prof radar in every single event, they recurred in recognizable form during the IMPROVE-2 storms. The LEE, DME, and SCE echoes exhibited some variability from case to case. These case-by-case variations in echo structure may be related to the large-scale environment in which the cyclone exists, the position of the storm relative to the instrument site, and the proximity of the cyclonic system to the previous storm. However, the recurrence of these three basic echo structures from storm to storm suggests that they are directly associated with the basic dynamic structure of the parent cyclone, as modified by its passage over the mountain range.

The structures identified in this study should be tested observationally for their repeatability in other West Coast regions of complex terrain, such as the California Sierra Nevada Mountains, the midlatitude Andes, the New Zealand Alps, and the mountains of Scandinavia. To fully understand the precipitation mechanisms giving rise to the echo structures repeatedly observed in storms crossing an orographic barrier (e.g., the secondary maximum in the DME) will require high-resolution numerical models that include ice-phase cloud microphysical processes correctly. The recurring structures seen in this study provide a goal for the modeling efforts. A model correctly reproducing the multiscale dynamical and physical mechanisms of the precipitation processes over the mountains should produce the echo structures in Fig. 17. Then the numerical models can help separate the precipitation processes intrinsic to the cyclone from those arising from the interaction of the flow with the orography.

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