The 5–9 February 1996 Flooding Event over the Pacific Northwest: Sensitivity Studies and Evaluation of the MM5 Precipitation Forecasts

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(Manuscript received 28 December 1998, in final form 23 March 1999)

ABSTRACT

This paper describes the flooding event of 5–9 February 1996 in which a series of landfalling Pacific storms brought 30–70 cm of rain to many mountain sites over southwest Washington and northwest Oregon. This event was simulated at 36-, 12-, 4-, and 1.33-km horizontal resolution using the Pennsylvania State University–National Center for Atmospheric Research mesoscale model (MM5). The model precipitation was verified with over 300 rain gauges in western Washington and Oregon as well as WSR-88D radar data from Portland, Oregon. There was a significant improvement in the precipitation forecast skill as the grid spacing was decreased from 36 to 4 km; however, the 12- and 4-km resolutions had excessive precipitation shadowing in the lee of barriers. Although increasing resolution from 4 to 1.33 km did not produce a significant improvement in precipitation skill for the entire domain, the 1.33-km domain had more precipitation in the immediate lee of the Cascades, and thus verified better in those regions.

Additional simulations explored the effects of changing vertical resolution. Enhancing the number of levels from 29 to 38 increased the model precipitation over the windward slopes of the Cascades by 10%–30%; further increasing the number of levels to 57 decreased windward precipitation to the 29-level amounts. The leeside precipitation varied by 20%–80% for the different vertical resolutions as a result of variations in mountain wave structure over the Cascades.

Five different MM5 microphysical schemes were compared for a 24-h period during the February 1996 flooding event. The warm rain scheme dumped too much precipitation along the windward slopes, underlining the importance of ice microphysics during the cool season. The most sophisticated scheme (i.e., graupel scheme) also did not provide the best forecast. An additional microphysical sensitivity run, in which the snow fall speed was reduced by approximately 20%, improved model skill by advecting more precipitation advecting to the lee of the Cascades. Overall, these microphysical results suggest that further improvements to microphysical schemes are needed in order to more accurately predict precipitation.

1. Introduction

Heavy precipitation and flooding are undoubtedly some of the most significant weather problems along the mountainous coastal zone of western North America. As documented in the NOAA publication Storm Data, flooding along coastal California, Oregon, and Washington during the past decade has resulted in damage exceeding 6 billion dollars and the loss of dozens of lives.

Flooding over the Pacific Northwest is usually the result of warm, moist subtropical southwesterly flow interacting with the coastal and Cascade mountains of Washington and Oregon (Fig. 1). Since the flow typically originates near Hawaii, this weather pattern is sometimes dubbed the “Pineapple Express.” During the winter of 1995–96, exceptionally heavy seasonal precipitation (125%–175% of the mean) occurred over the Pacific Northwest (Halpert and Bell 1997). The most significant event of that winter took place on 5–9 February 1996, during which many rivers of northwestern Oregon and southwest Washington experienced the worst flooding in 30 years, with the Columbia and Willamette Rivers rising as much as 10–20 feet above flood stage. Eight deaths were directly attributed to the floods, over 30,000 residents were forced from their homes, and damage approached 500 million dollars (Storm Data 1996).

Prior to this event subfreezing temperatures and a
Significant snowpack were present at relatively low elevations (>500 m). A series of much warmer synoptic systems moved through between 5 and 9 February and were associated with an unusually strong upper-level jet over the central Pacific. The liquid precipitation for this 4-day event ranged from 10–25 cm over the lowlands of western Washington and Oregon to 35–75 cm over the coastal and Cascade mountains. Many stations across western Oregon set all-time records for the most precipitation observed during a 4-day period (G. Taylor 1997, personal communication). In addition to the wet conditions, temperatures were unusually mild. Many stations in the lowlands reported minimum temperatures that were 1–5°C greater than normal maximum values for early February. The warm temperatures augmented flooding since high freezing levels (above 2 km) resulted in 10–30 cm of water equivalent from melting snow at many mountain sites.

Quantitative precipitation forecasting has recently been labeled as a research priority by the U.S. Weather Research Program (Fritsch et al. 1998). Precipitation forecasting over mountainous regions, such as the U.S. west coast, is especially difficult since orographic precipitation is controlled by a number of dynamical and microphysical factors. For example, the amount of stable orographic uplift and precipitation is determined by the size and shape of the barrier as well as the distribution of moisture and static stability at low levels. Large barriers such as the Sierras and Cascades (half-width, $L > 100$ km) typically have the maximum precipitation 10–20 km upstream of the crest with significant precipitation shadowing in the lee (Rauber 1992; Westrick 1998), while narrow barriers such as the Southern Alps of New Zealand ($L < 30$ km) usually have the maxima closer to the crest (Sinclair et al. 1997). Sinclair et al. (1997) noted that the precipitation distribution is sensitive to how much of the low-level flow is blocked by the terrain as determined by the Froude number ($Fr = U/N\theta_h$, where $U$ is the upstream wind speed toward the barrier, $N$ is the static stability, and $\theta_h$ is the height of the barrier), where a $Fr > 1$ favors unblocked flow and more precipitation spilling over the crest. Sinclair et al. showed that the Froude number could explain 78% of the spillover variations across the Southern Alps. There are also precipitation sensitivities to the amplification and wavelength of mountain waves. For example, Bruins et al. (1994) showed the importance of mountain waves in the generation of significant cloud-liquid water over the narrow ridges of northern Arizona during passage of a winter storm.

The popular conceptual model for stratiform precipitation enhancement over orography is known as the
“seeder-feeder” effect (Bergeron 1949, 1960), where precipitation particles generated aloft grow within an orographically forced lower cloud. However, recent studies have revealed considerable complexity related to orographic microphysics. For example, observational studies such as Hobbs (1975) and Rauber (1992) have shown the presence of Hallet and Mossop (1974) ice multiplication over the coastal terrain of Washington and California, in which large numbers of ice fragments are formed when supercooled droplets contact ice crystals. Hobbs et al. (1973) showed that an enhancement in the number of small ice particles immediately over the Washington Cascades results in more precipitation carried downwind of the crest. Mountain waves associated with upstream orography may also alter seeder clouds, such as the California coastal range enhancing the seeder cloud upstream of the Sierras (Meyers and Cotton 1992).

Diabatic feedbacks can also occur between the orographic cloud and the flow across the barrier. For example, Durran and Klemp (1983) showed that the inclusion of moisture reduces effective stability and mountain wave amplitude, which, in turn, significantly alters the orographic cloud. Marwitz (1983) suggested that an isothermal melting layer near crest level in the Sierras increased low-level static stability and blocking upstream of the barrier, resulting in a displacement of maximum precipitation further upwind of the crest.

The detailed diagnosis and prediction of precipitation structures over terrain has been hindered by the lack of horizontal resolution in numerical models. However, recent high-resolution simulations of precipitation events over various locations of the western United States (Bruintjes et al. 1994; Colle and Mass 1996; Gaudet and Cotton 1998; Westrick 1998) have suggested that numerical models run at high resolution (i.e., down to ∼10 km) can realistically predict observed precipitation structures over complex terrain. For example, Colle and Mass (1996) showed that the Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) mesoscale model (MM5) run at 3-km resolution was not only able to realistically simulate the observed orographic precipitation structures (e.g., windward enhancement on Vancouver Island and Olympics, rain shadows to the lee of the Olympics and Cascades), but produced precipitation amounts within 30% of the storm totals for approximately 30 observing sites within the domain. More recently, Westrick (1998) simulated the 28 December 1996–3 January 1997 flooding event over the central Washington Cascades down to 4-km resolution using the MM5. Results showed that although the model precipitation accuracy improved as horizontal resolution was increased from 12- to 4-km resolution, precipitation was underpredicted by 30%–40% within the Stampede Gap region of the central Washington Cascades in the 4-km domain.

Mesoscale models are currently run daily at a number of sites (Mass and Kuo 1998), thus allowing precipitation forecasts to be evaluated over topography for longer time periods (Colle et al. 1999; McDonald and Horel 1998). For example, Colle et al. (1999) verified the MM5 using real-time forecasts down to 12-km resolution over Washington and Oregon during the 1996–97 cool season, and compared the results with the National Centers for Environmental Prediction’s 10-km resolution Eta model (Eta-10). They noted a noticeable improvement in bias and skill scores as the horizontal resolution was increased from 36 to 12 km in the MM5; however, statistics across the region suggested that the 12-km MM5 and Eta-10 generate too much precipitation along the steep windward slopes and not enough in the lee of major barriers.

By utilizing rain gauge and WSR-88D radar data over Washington and Oregon, as well as simulations using the PSU–NCAR MM5 down to 1.3-km resolution, precipitation was verified at higher resolutions and over a wider spatial area than previous studies in the region (Colle et al. 1999; Westrick 1998). This paper addresses several key questions concerning West Coast flooding events and the 5–9 February 1996 event in particular:

- What is the synoptic evolution accompanying the 5–9 February 1996 flooding event?
- How does the ability of the MM5 to simulate the mesoscale precipitation structures change as horizontal resolution increases from 36 to 1.3 km, and the vertical resolution is increased from 29 to 57 vertical levels?
- What is the spatial variability in model precipitation accuracy across the region?
- How does the model precipitation change as more sophisticated microphysics are applied?

Section 2 discusses the observational datasets and some of the verification methods. Section 3 describes some of the large-scale observations for this event. Section 4 shows MM5 verification of the winds and temperatures for a few selected sites as well as detailed precipitation verification as horizontal resolution was increased from 36 to 1.3 km. This section also presents some sensitivity studies using different vertical resolutions and microphysics. A summary and conclusions are presented in the final section.

2. Datasets and methods

Several different precipitation datasets were utilized in this study. Approximately 30 National Weather Service (SAO), 400 National Climatic Data Center cooperative observer (COOP), 100 snow telemetry (SNOTEL), and 10 Northwest Avalanche Center (NWAC) sites were used to verify the MM5 precipitation forecasts in Washington and Oregon (Fig. 1). Most of the SAO stations are shielded tipping-bucket rain gauges that record hourly at 0.254 mm (0.01 inch) resolution. Approximately two-thirds of the COOP stations use unshielded weighing gauges that record hourly at either
0.254-mm or 2.54-mm resolution, while the remainder record at daily intervals. The SNOTEL stations are maintained by the U.S. Department of Agriculture Natural Resources Conservation Service (McMillan 1981). These are automated snow-pillow sensors and weighing rain gauges located in remote mountainous regions, and measure total water-equivalent precipitation since the start of the water year, 1 October. The NWAC gauges are electrically heated tipping-bucket gauges with an 8-in. collection diameter, and are monitored by Campbell CR-10 data-loggers and accessed over phone lines.

Each type of rain gauge has its own inherent problems, so special care has to be taken when interpreting the verification results. For example, errors include excessive evaporation for heated gauges (Groisman and Legates 1994), snow capover for storage gauges, and snow bridging over snow-pillow sensors at SNOTEL sites (Ferguson et al. 1997). These problems, combined with the high winds and frozen precipitate often observed over mountainous terrain, can result in significant precipitation underestimation. For example, an unshielded gauge can fail to catch over 60% of snowfall when the winds are 10 m s<sup>-1</sup> (Dingman 1994). Because of the lack of colocated wind observations with these rain gauges, this study does not attempt to estimate undercatchment.

To verify the MM5 precipitation at the various observational locations, precipitation from each model was first interpolated to these stations using the Cressman (1959) method

\[ P = \frac{\sum_{n=1}^{4} W_n P_n}{\sum_{n=1}^{4} W_n}, \tag{1} \]

where \( P_n \) is the model precipitation at the four model grid points surrounding the observation. The weight \( W_n \) given to the surrounding gridpoint values is given by

\[ W_n = \frac{R^2 - D_n^2}{R^2 + D_n^2}, \tag{2} \]

where \( R \) is the model horizontal grid spacing and \( D \) is the horizontal distance from the model grid point to the observational station. It has been shown previously that this method produces acceptable results (Colle et al. 1999).

3. Brief synoptic overview

The large-scale evolution of this 4-day event is similar to Lackmann and Gyakum’s (1999) composite study of Pacific Northwest heavy precipitation events. Important aspects of this flooding case are the persistent southwesterly flow with a long fetch that started just north of Hawaii and the series of shortwave troughs and associated frontal systems that crossed the coast approximately once a day. During the beginning of this event (1200 UTC 5 February 1996), longwave upper-level ridges at 500 mb were situated over northeastern Siberia and western North America (Fig. 2a), while a longwave trough was located over the central Pacific. A series of shortwave troughs (A, B, C, and D on Fig. 2a) were rotating around the longwave trough. At 1200 UTC 6 February (Fig. 2b), large-scale confluence was occurring over the central Pacific, as flow originating north of Alaska merged with the more westerly–southwesterly flow to the south. The flow became more southwesterly over the West Coast as weak upper-level ridging developed along the coastline. By 1200 UTC 7 February 1996 (Fig. 2c), the 500-mb shortwave trough (E) over the central Pacific had deepened near 160°W. During the next day this trough moved to near 140°W and sharpened as downstream ridging occurred near the crest of the Rockies (Fig. 2d). The end of the heavy precipitation event occurred approximately 24 h later after this upper-level trough moved past the Pacific Northwest (not shown).

During this event a band of mid- and upper-level clouds stretched nearly continuously between 160°W and the coast (Fig. 3). At 1200 UTC 5 February (Fig. 3a), shortwave troughs A and B are evident as comma-shaped cloud patterns in the infrared imagery. At 1200 UTC 7 February 1996 (Fig. 3b), strong southwesterly flow extending from just north of Hawaii to the Pacific Northwest was transporting significant amounts of moisture toward the region. This so-called Pineapple Express was associated with a train of mid- and high-level clouds stretching from near Hawaii to the Pacific Northwest.

4. Simulations of the 5–9 February 1996 flooding event

a. Model description

Three forecasts from the PSU–NCAR MM5 initialized at 0000 UTC 5 February, 1200 UTC 6 February, and 0000 UTC 8 February 1996 were used to create the model storm-total precipitation. Since the SNOTEL precipitation is only reported every 6 h at local standard time (GMT −8 h), the 8–44-h forecast period was chosen in order to match the model with the SNOTEL observation times. The control simulations used the explicit moisture scheme of Hsie et al. (1984), with improvements to allow for mixed liquid-ice phase below 0°C and graupel microphysics (Grell et al. 1994; Reisner et al. 1998). The Kain–Fritsch cumulus parameterization (Kain and Fritsch 1990) was applied, except for the two innermost domains, where convective processes could be resolved explicitly. The planetary boundary layer (PBL) was parameterized with the Blackadar scheme (Zhang and Anthes 1982). Klett and Durran’s (1983) upper-radiative boundary condition was applied in order to prevent gravity waves from being reflected off the model top.

The simulations had stationary 12, 4, and 1.33 domains nested within the 36-km domain using one-way
FIG. 2. The 500-mb ECMWF analysis at (a) 1200 UTC 5 Feb, (b) 1200 UTC 6 Feb, (c) 1200 UTC 7 Feb, and (d) 1200 UTC 8 Feb 1996 showing geopotential heights (solid lines every 60 m), temperature (dashed every 5°C), and winds (one pennant, full barb, and half-barb denote 25, 5, and 2.5 m s⁻¹, respectively). The shortwave trough axes are dashed in bold and labeled A, B, C, D, and E.

interfaces (Fig. 4), and all four domains were run simultaneously. Thirty-eight unevenly spaced full-sigma levels were placed in the vertical,¹ with the maximum resolution in the boundary layer. To resolve the coastal and Cascade Mountains within the 4- and 1.33-km resolution domains (see Fig. 8 for the 4-km domain topography), the model topography was obtained by interpolating a 30-s topography dataset to the grid using the Cressman scheme and then was filtered using a two-pass smoother-desmoother. Initial atmospheric conditions and sea surface temperatures (SSTs) were generated by first interpolating the NCEP global analyses (2.5° lat × 2.5° long resolution) to the model grid.² These analyses were then improved by incorporating surface and upper-air observations using a Cressman-type analysis scheme (Benjamin and Seaman 1985). Additional analyses were generated every 12 h in the same manner and then linearly interpolated in time in order to provide lateral boundary conditions for the 36-km domain. Four-dimensional data assimilation was applied to the 36-km domain during the first 12 h of the integration of all the simulations in order to obtain a more

¹ The 38 full-sigma levels were: $\sigma = 1.0, 0.99, 0.98, 0.97, 0.95, 0.93, 0.91, 0.89, 0.87, 0.85, 0.83, 0.81, 0.79, 0.77, 0.75, 0.73, 0.71, 0.69, 0.67, 0.65, 0.62, 0.59, 0.56, 0.53, 0.50, 0.47, 0.44, 0.41, 0.37, 0.33, 0.29, 0.25, 0.21, 0.17, 0.13, 0.09, 0.05, 0.0.$

² Higher resolution (1.125°) European Centre for Medium-Range Weather Forecasts (ECMWF) grids were not available at the time the MM5 simulations were completed.
realistic simulation. Specifically, winds and moisture within the PBL and the temperature, winds, and moisture above the PBL were nudged toward the NCEP analyses.

3 The FDDA scheme applies Newtonian relaxation to nudge the model toward the NCEP 3-h surface and 12-h upper-air gridded analyses (Stauffer and Seaman 1990; Stauffer et al. 1991).

b. Model time series verification

The evolving surface wind, temperature, and precipitation over and to the west of the Cascades were verified for the 4-km domain using surface observations from Salem, Oregon, in the Willamette Valley (SLE in Fig. 1) and June Lake in the southwest Washington Cascades (JNL in Fig. 1) between 0800 UTC 5 February and 1400 UTC 9 February 1996 (Figs. 5a and 6a). Figures 5b and 6b show the model time series at the lowest half-sigma
level (~40 m above the ground) for the SLE and JNL surface stations, which were derived by interpolating the four model grid points surrounding these locations in the 4-km domain. In addition, lower-tropospheric winds from a 915-MHz wind profiler of Astoria, Oregon (AST on Fig. 1), were compared to model output between 0600 UTC 5 February and 0000 UTC 9 February 1996 (Fig. 7).

1) SURFACE TIME SERIES

The observed warm frontal passage at Salem occurred shortly after 1800 UTC 5 February 1996 and was associated with a dramatic increase in surface temperature (from 0° to 11°C) and strengthening southerly flow (WF1 on Fig. 5a). The temperatures continued to increase more gradually until an ocluded front (OF) moved through shortly before 0200 UTC 6 February. Another period of pressure rises, weak warming, and veering winds from south-southeasterly to southerly occurred with the passage of another weak warm front (WF2) at 1500 UTC 6 February. The surface temperatures increased to 16°C after the passage of the final warm front (WF3) at 0000 UTC 8 February. Cold frontal passage (CF) at 2000 UTC 8 February was marked by a rapid pressure rise, a gradual wind shift to northwesterly flow, and an 8°C temperature drop.

Many of the observed transitions at SLE are well duplicated by the model (Fig. 5b). There are three distinct warming periods associated with the three warm frontal passages (WF1–WF3). The primary problem was the timing of the various surface fronts, in which the model was 2–4 h early with the warm frontal transitions and a few hours late with the occluded and cold fronts. The model also had some difficulty capturing the magnitude of the postfrontal (OF) and nocturnal cooling observed on 6–7 February 1996.

The heaviest precipitation at SLE occurred during three separate periods, which were realistically simulated by the model. The first was associated with the OF and WF2 frontal passages, in which 20–30 mm of precipitation occurred every 6 h (Fig. 5a). The precipitation also intensified to 30 mm in 6 h with the passage of the WF3, but then became light in the warm sector between 0500 and 1700 UTC 8 February. Finally, a brief period of heavy precipitation (22 mm in 6 h) occurred shortly before CF. The 4-km simulation captured 87% of the observed storm-total precipitation. Most of the underprediction occurred between 1700 UTC 5 February and 0200 UTC 6 February, in which only 20%–30% of the observed precipitation was predicted (Fig. 5b). Many other sites, including JNL (Fig. 6), also had significant underprediction during this period, suggesting that the model had difficulty with prefrontal precipitation before the main plume of subtropical moisture arrived over the Pacific Northwest on 6 February 1996.

The observed temperatures at June Lake (JNL) at 1000 m above mean sea level (MSL) also increased significantly with the warm frontal passages around 1900 UTC 5 February, 1900 UTC 6 February, and 0700 UTC 8 February 1996 (Fig. 6a). Unlike SLE, the heaviest precipitation at JNL occurred during the warm episodes rather than during or shortly before frontal passages. JNL is situated along the southwest slopes of the Cascades, where orographic precipitation is favored during the warm southwesterly flow conditions that occur.

FIG. 4. Model terrain contoured every 200 m for the 36-km domain. Terrain heights of 500–1300 m and greater than 1300 m are shaded light and dark gray, respectively. The inner boxes shows the location of the 12-, 4-, and 1.33-km nested domains.
after warm frontal passage, while SLE is located to the lee of the coastal mountains. Overall, 726 mm of precipitation fell at JNL during this event compared to 210 mm at SLE.

For JNL (Fig. 6b), the model was 2–6 h too fast with warm and occluded frontal transitions, but the postfrontal temperatures were within $1^\circ$–$2^\circ$C of observed. The nocturnal cooling errors on 6–7 February 1996 were $1^\circ$–$3^\circ$C larger than at SLE. As observed, the 4-km simulation had heavy precipitation after warm frontal passages when there was strong (10–15 m s$^{-1}$) southwesterly (upslope) low-level flow (Fig. 6b). However, the precipitation during these periods was underpredicted, resulting in the simulated storm total (503 mm) being 69% of observed.
FIG. 7. (a) Wind speed and direction from the Astoria 915-MHz wind profiler (see Fig. 1 for location) displayed as a time–height section of hourly observations between 0600 UTC 5 Feb and 0000 UTC 9 Feb 1996. One full barb is equivalent to 5 m s⁻¹ and a flag represents 25 m s⁻¹. The various surface warm (WF1, WF2, and WF3), occluded (OC), and cold (CF) frontal passages are indicated by the arrows. (b) Same as (a) but for the 4-km domain interpolated to the Astoria profiler site using three separate MM5 simulations between 0600 UTC 5 Feb and 0000 UTC 9 Feb 1996.
2) Astoria profiler

Between 0600 UTC 5 February and 0000 UTC 9 February 1996 (Fig. 7a), the low- and midlevel flow at the Astoria profiler site was generally dominated by strong (20–25 m s\(^{-1}\)) southerlies or southwesterlies. WF1 occurred shortly before 1200 UTC 5 February, as the low-level flow veered from southeasterlies to strong southerlies at \(\sim 20 \text{ m s}^{-1}\). This was followed by OF at 0300 UTC 6 February and WF2 at 1700 UTC 6 February, both of which were marked by the low-level flow veering to more southwesterly. WF3 at 2100 UTC 7 February was also associated with a transition from relatively weak (5–10 m s\(^{-1}\)) southeasterlies to strong (15–25 m s\(^{-1}\)) south-southwesterlies at low levels. The low-level flow veered to west-northwesterly with the passage of the CF shortly before 2000 UTC 8 February.

Many of the observed low- and midlevel wind structures at the AST profiler site were well simulated by the model (Fig. 7b), with winds near crest-level (\(\sim 1500 \text{ m}\)) generally within 3 m s\(^{-1}\) of observed. The largest errors originated from modest timing errors, in which the model was around 2–3 h too fast with the CF and WF2 and WF3. In addition, the simulated CF was associated with a more rapid wind shift to northwesterlies than observed, and the simulated warm and occluded frontal wind shifts were too weak. Overall, the model captured the important wind, temperature, and precipitation transitions during this event.

c. Precipitation verification

1) Observed precipitation distribution

Using all available precipitation data, a storm-total map was constructed for western Washington and Oregon (Fig. 8). Between 5–9 February 1996 the heaviest precipitation fell over the coastal range of northwest Oregon and the Cascades near the Oregon–Washington border. In some of these regions, over 500 mm of liquid-equivalent precipitation fell, while in the lowlands of northern Oregon, around 200 mm of precipitation was observed. This heavy precipitation combined with the 200–300 mm of snowmelt at the mountain sites (not shown) resulted in significant flooding and damage in the region.
shown) resulted in severe flooding west of the Cascade Mountains. In contrast, precipitation shadowing is evident downwind (east) of the Cascades and to the northeast of the Olympics, where generally less than 50 mm fell.

The precipitation enhancement over the relatively low (500–700 m MSL) Oregon coastal range was nearly equal to that of the higher Cascade range (2000 m MSL), which illustrates that the seeder–feeder enhancement process was quite efficient for the modest topography. The downstream effect of the coastal range is quantified in section 4f.

The observations also suggest that the maximum precipitation (>400 mm) was closer to the crest of the Oregon coastal mountains than the Cascade crest. This is consistent with previous observational and modeling studies, which have shown that relatively narrow (<30-km half-width) and low barriers tend to have the maximum precipitation closer to the crest than wider and higher barriers. Thus, the orographic precipitation distribution for this event cannot be duplicated by a simple elevation-based enhancement scheme.

2) MODEL PRECIPITATION

Figure 9 shows scatterplots of observed versus model precipitation for the 36-, 12-, and 4-km domains using only verification sites within the 4-km domain. There is a noticeable and progressive improvement in model precipitation as the resolution is increased from 36- to 4-km resolution, with the bias increasing from 64% to 85% and the rms errors decreasing from 110 to 70 mm.4 The greatest reduction in errors with resolution is for the moderate to high precipitation amounts (>200 mm) over the high terrain.

Figure 10 shows the storm-total precipitation for the 4-km domain as well as the percentage of observed precipitation at each station (bold numbers with x). Overall, the best bias scores (80%–120%) are located over the Oregon coastal range and the Cascades, while a few areas of high bias (>130%) are situated over the central Oregon coast and some isolated volcanic peaks such as Mt. Rainier. In contrast, the model predicted less than 60% of the observed storm-total precipitation

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4 The bias is defined as forecast divided by observed ($F/O$) precipitation, and then multiplied by 100.
at many stations downwind of the coastal ranges, the Cascades, and within the Cascade gaps.

To show differences in model skill between a station located at the Cascade crest and a station downwind of orography, Fig. 11 presents observed and model precipitation at the Bumping Ridge and Meadows Pass SNOTEL sites (points A and B, respectively, on Fig. 10). At the crest-level Bumping Ridge site (Fig. 11a), the model simulates precipitation accurately throughout the event. Meanwhile, within Stampede Gap at Meadows Pass (Fig. 11b), which is frequently downwind of the southern Washington Cascades and Mt. Rainier, the model precipitation error grows with time as a result of excessive rainshadowing. Model precipitation biases of $<60\%$ were also evident in the lee of the Cascades in the 12-km resolution domain (not shown). Because of this excessive rainshadowing in the higher-resolution domains, the 36-km precipitation verified better than the 12- and 4-km resolutions downwind of the Cascades (not shown).

Using only observations within the 1.33-km domain (see Fig. 10 for domain location), the 4- and 1.33-km model precipitation fields were also compared (Fig. 12). The improvement from 4- to 1.33-km resolution is much less obvious than seen from 12 to 4 km (cf. Figs. 9b,c). The average 1.33-km bias score is closer to one and the least squares best-fit line (dashed) is slightly better at higher precipitation thresholds than the 4 km. However, the 1.33 km has more scatter around the one-to-one line, resulting in a larger average rms error than the 4-km resolution.

Over the lowland stations to the west of the Cascades
the model biases in the 1.33-km domain are nearly identical to those in the 4-km grid over the lowland stations (Fig. 13). Interestingly, 1.33-km resolution produced significantly more precipitation in the lee (east) of the Cascades, where the precipitation shadowing was overdone in the 4-km domain (cf. Fig. 10); however, there are still stations 10–20-km downwind of the Cascades at which the 1.33-km resolution precipitation fields are still too low by 25%–50%. Along the windward side of the Cascades the model bias at June Lake (see JNL in Fig. 1) improved from 68% to 75% going from 4- to 1.33-km resolution. The 1.33-km resolution tended to produce large “bull’s-eyes” of precipitation (>1000 mm) in the lee of major peaks. Model cross sections taken across these peaks suggest that these precipitation maxima were created by downward motion associated with mountain waves bringing hydrometeors down to the surface (not shown). Unfortunately, the magnitude of this leeside response cannot be evaluated by observations since there are no rain gauges in those areas and the radar beam from the Portland, Oregon, WSR-88D (see RTX on Fig. 1) is either blocked by terrain or too high at these locations.

To verify further the model precipitation over the Cascades, simulated and observed reflectivities were compared. Figure 14a shows reflectivities from the WSR-88D radar at Portland, Oregon, (see RTX on Fig. 1) for a cross section taken across the northern Oregon Cascades (see segment XX' on Fig. 13 for location) at 0300 UTC 7 February 1996. Figures 14b–e present simulated reflectivities and winds from the 1.33-, 4-, 12-, and 36-km domains for the same cross section and time (15 h into the simulation). The model reflectivities were calculated using empirical relations based on the model cloud and precipitation mixing ratios (Fovell and Ogura 1988). Comparing the 4-km model winds and temperatures immediately upstream of the barrier with the nearly coincident Salem, Oregon, (SLE on Fig. 1) sounding at 0000 UTC 7 February 1996 (Fig. 14), indicates that the ambient flow and low-level temperatures impinging on the Cascades were well simulated. At 1.33-km resolution (Fig. 14b), the precipitation enhancement over the windward side of the Cascades is generally in good agreement with the radar. Both the observed and modeled radar reflectivity suggest that orographic cloud over the western Cascades was modulated by mountain waves induced by the mesoscale ridges, with the heaviest precipitation falling downwind in the valleys. In contrast, over the crest and in the lee of the barrier the model reflectivities at around 4 km MSL were less than the radar observations. At 4-km resolution (Fig. 14c),

5 The cross section was obtained by bilinearly interpolating the WSR-88D scans to a Cartesian grid with a horizontal and vertical spacing of 2 and 0.5 km, respectively.
the structures over the windward slope are similar to the 1.33-km domain, but the lee wave to the east of the Cascades was significantly weaker. The stronger subsidence in the immediate lee of the Cascade crest at 1.33-km resolution brought down more hydrometeors to the surface than at 4-km resolution, which resulted in more storm-total precipitation in the 1.33-km domain in the lee of the barrier (cf. Figs. 10 and 13).

There is a large difference in precipitation and wind structures over the Cascades when the resolution is decreased from 4- to 12-km resolution (Figs. 14c,d). The relatively smooth 12-km topography results in a single vertically propagating gravity wave over the Cascades, in which the subsidence extends upwind of the crest at midlevels. The lack of mountain waves and associated upward motions with the mesoscale ridges of the windward side of the Cascades results in less precipitation approaching the crest than at 4- and 1.33-km resolutions. The 36-km domain produced less precipitation over the windward slopes. In addition, the 36-km simulation could not resolve the leeside subsidence associated with the Cascades (Fig. 14e), and therefore more precipitation existed at midlevels over the lee slope.

The excessive drying in the lee of the Cascades for higher resolutions could have resulted from several sources such as: 1) the mountain wave and resulting subsidence might have been overpredicted and/or 2) the microphysical scheme was not generating or maintaining enough ice aloft so that too little was being advected over the barrier. In section 4g of the paper we attempt to explore how the precipitation distribution across the Cascades changes using the other microphysical schemes of varying sophistication in the MM5.

d. Precipitation distribution over the Cascades versus horizontal resolution

The precipitation distribution over the Cascades was sensitive to horizontal resolution. Figures 15a,b shows the average east–west terrain and precipitation profiles for the boxed region in Fig. 13. As the resolution increased from 36 to 1.33 km, the terrain of the Willamette Valley and Cascade foothills was reduced while the height of the Cascades increased. The terrain differences are most dramatic between the 36 and 12 km resolutions, while from 4 to 1.33 km the orographic slopes become steeper and terrain variability increases dramatically near the Cascade crest. An increase in horizontal resolution resulted in a distinct downwind (eastward) shift in model precipitation toward the crest (Fig. 15b). The coarse 36-km grid has the maximum precipitation displaced toward the lowlands, where the greatest synoptic-scale forcing and moisture influx occurred. At higher resolution (12 and 4 km) the orographic contribution to the total precipitation increases dramatically and the peak is displaced eastward. There is a significant increase in average precipitation near the crest and on the lee slopes going from 4- to 1.33-km resolution, which suggests the steeper orographic slopes at 1.33 km make an important contribution to the orographic precipitation rates.

Another question deals with the relative importance of decreasing the grid spacing versus better resolving the model topography. To address this issue, the 4-day event was simulated at 4-km resolution, but with the topography from the 12-km grid interpolated bilinearly to the 4-km domain. Comparing this simulation with the control 12-km domain, there was a small shift in precipitation from the lower to upper windward slopes.
Fig. 13. Storm-total model precipitation (shaded in mm) for the 1.33-km MM5 domain from 0800 UTC 5 Feb 1996 through 1400 UTC 9 Feb 1996. The numbers at the × locations indicate the percent of observed. Terrain is contoured every 200 m for reference. Line XX' is the location of the cross section shown in Fig. 14. The dashed-boxed region denotes the area in which the average terrain and precipitation cross sections were derived for Fig. 15. The location of the 1.33-km domain is shown in Fig. 10.

of the Cascades, with slightly more precipitation shadowing in the lee. Overall, this suggests that 10%–30% of the precipitation change over the Cascades from 12- to 4-km resolution resulted from the decreased grid spacing, with the 4-km grid spacings better resolving the terrain circulations and microphysics.

e. Vertical resolution

It was shown above that the precipitation forecasts generally improved when the horizontal resolution was increased from 36 to 4 km, but what happens when the vertical resolution is varied? To address this issue, simulations were completed at 4-km horizontal resolution for a 24-h period between 1200 UTC 6 February and 1200 UTC 7 February using vertical resolutions ranging from 29 to 57 full-sigma levels (Table 1). The control run shown above used 38 vertical levels.

For the 8–24-h forecast period, the 29-level run generated 10%–30% (10–30 mm) less precipitation along the windward (west) slopes and crest of the Cascades and 20%–80% (5–10 mm) more immediately in the lee of the crest (Fig. 16a). The differences are even more dramatic (>200% in some areas) further downwind of the crest, where the precipitation is generally light (<5 mm) in this region. A representative cross section (YY') taken across a lower portion of the northern Oregon Cascades at hour 18 (0600 UTC 7 February) reveals that the 38-level run had greater concentrations of graupel near the crest than the 29-level (Figs. 17a,b), resulting in more precipitation near the crest and the immediate lee. The leeside mountain wave and subsidence
was also more significant in the 38-level run, which limited the production and advection of hydrometeors downwind of the Cascades.

An increase in the number of vertical levels from 38 to 57 resulted in 20%–40% (10–40 mm) less precipitation over the upper western slopes and crest of the Cascades (Fig. 16b), while 20%–80% (10–20 mm) more precipitation fell in the lee. The 57-level run had somewhat weaker upward motions immediately upstream of the crest between 500 and 700 mb and more subsidence over the crest at midlevels compared to the 38-level...
simulation (Figs. 17b,c). This resulted in a less snow and graupel over the barrier, and in turn less precipitation over the windward slope. The 57-level run also had a broader area of upward motion and more precipitation over the eastern portion of the domain compared to the 38-level run.

Overall, these experiments show that different vertical resolutions can yield subtle differences in mountain wave structure, which in turn creates significant changes to model precipitation. Along the windward slope of the Cascades, the 38-level run precipitation was closer to the observations than the other vertical resolution runs (not shown), while the 29- and 57-level runs verified better in the lee; however, more cases need to be simulated to confirm these results.

f. No coastal range

A potentially important issue is the modification of orographic precipitation by upstream topography. Meyers and Cotton (1992) showed that the California coastal range increased ice concentrations downwind over the Sierras between 5 and 6 km MSL. However, this additional ice had little impact on the precipitation distribution at the surface over the Sierras, since it appeared that the strong low-level winds advected the crystals over the orographic feeder cloud. Since the configuration of the Washington and Oregon terrain is somewhat similar to California with a low coastal range and higher terrain inland (cf. Fig. 1), one might expect a similar result. To explore this, the three MM5 simulations of the flooding event were rerun without coastal terrain (NOCST experiment) south of the Olympics (between the two arrows on Fig. 18).

Figure 18 shows the percentage change in 4-km storm-total precipitation between the NOCST and CTL simulations. The coastal terrain increased precipitation by 100%–200% immediately upwind and over the coastal range. A precipitation enhancement of 10%–20% also extended westward of the coastal mountains by 20–30 km, which is close to a Rossby radius given a moist adiabatic atmosphere ($R = Nh/f \sim 25$ km, where $h_w \sim 500$ m, $f \sim 1.03 \times 10^{-4}$ s^{-1}, and moist $N \sim 0.005$ s^{-1}). In contrast, subsidence in the lee of the coastal range resulted in 10%–50% less precipitation downwind over the Willamette Valley, and a 10%–20% reduction further eastward over the lower windward slopes of the Cascades. There was also some sensitivity in the lee of the Cascades, where the CTL run had approximately 10% more precipitation. The CTL run had slightly less ice and snow aloft over the Willamette valley as a result of mountain-wave-induced subsidence in the lee of the coastal terrain (not shown). Therefore, the increase in precipitation in the lee of the Cascades was not the direct effect of enhanced ice production by the coastal range falling out downstream.

g. Microphysical sensitivity studies

This paper has identified some model deficiencies in precipitation prediction over the Pacific Northwest for the 5–9 February flooding event, most notably excessive precipitation shadowing in the lee of barriers. Excessive precipitation shadowing was also found in the MM5 forecasts down to 12-km resolution over this region during the 1996–97 cool season (Colle et al. 1999) and in recent simulations down to 4-km resolution of another flooding event over western Washington in late December 1996 (Westrick 1998). Deficiencies with the bulk microphysical parameterizations might lead to problems. For example, not enough snow and cloud ice may be blowing over the crest of these barriers because the number concentration of these hydrometeors are too low and/or the fall speeds are too large.

Although detailed evaluation of model microphysics is beyond the scope of this study, the sensitivity of the orographic precipitation distribution to varying microphysical schemes was examined. Five separate schemes (the warm rain, simple ice, Reisner1, Reisner2, and Reis1VS) in the MM5 were compared for the 24-h period from 2000 UTC 6 February 1996 through 2000 UTC 7 February using the 38-vertical level forecast initialized on 1200 UTC 6 February 1996. The explicit warm rain scheme of Hsie et al. (1984) includes no ice microphysics while the simple ice scheme of Dudhia (1989) includes snow and cloud ice below 0°C. The Reisner1 scheme is similar to the simple ice scheme except that the Reisner1 allows for supercooled water below 0°C and ice does not immediately melt above 0°C. The Reisner2 scheme, which was used for the previous runs above, includes graupel and has prognostic equations for cloud-ice number concentration (Reisner

### Table 1. A listing of the full-sigma levels used in the vertical resolution experiments.

<table>
<thead>
<tr>
<th>Number of levels</th>
<th>Full sigma levels</th>
</tr>
</thead>
<tbody>
<tr>
<td>29</td>
<td>1, 0.99, 0.98, 0.97, 0.95, 0.91, 0.87, 0.83, 0.79, 0.75, 0.71, 0.67, 0.63, 0.59, 0.55, 0.51, 0.47, 0.43, 0.39, 0.35, 0.31, 0.27, 0.23, 0.19, 0.14, 0.11, 0.07, 0.03, 0.0</td>
</tr>
<tr>
<td>38</td>
<td>1, 0.99, 0.98, 0.97, 0.95, 0.93, 0.91, 0.89, 0.87, 0.85, 0.83, 0.81, 0.79, 0.77, 0.75, 0.73, 0.71, 0.69, 0.67, 0.65, 0.62, 0.59, 0.56, 0.53, 0.50, 0.47, 0.44, 0.41, 0.37, 0.33, 0.29, 0.25, 0.21, 0.17, 0.13, 0.09, 0.05, 0.0</td>
</tr>
<tr>
<td>57</td>
<td>1, 0.99, 0.98, 0.97, 0.96, 0.95, 0.94, 0.93, 0.92, 0.91, 0.8975, 0.885, 0.8725, 0.86, 0.8475, 0.835, 0.8225, 0.81, 0.795, 0.78, 0.765, 0.75, 0.735, 0.72, 0.705, 0.69, 0.675, 0.66, 0.645, 0.63, 0.615, 0.60, 0.585, 0.57, 0.555, 0.54, 0.525, 0.51, 0.495, 0.48, 0.46, 0.44, 0.42, 0.40, 0.38, 0.36, 0.33, 0.30, 0.27, 0.24, 0.21, 0.18, 0.15, 0.12, 0.08, 0.04, 0.0</td>
</tr>
</tbody>
</table>
et al. 1998). The Reisner and simple ice schemes noted above use a fixed slope intercept formulation for the Marshall–Palmer snow distribution (Grell et al. 1994). Reisner et al. (1998) implemented a variable slope intercept for the Reisner schemes in the MMS based on observations by Sekhon and Srivastava (1970). Thus, in addition to the above schemes, a variable slope intercept version of the Reisner1 scheme (denoted by Reis1VS) was also tried.

Table 2 presents some of the bias and rms error statistics for the 4-km domain using the various microphysical schemes. The rms errors are generally worse for the warm rain scheme, but for the highest threshold (>130 mm) the Reis1VS has larger rms scores. The bias and rms errors for the simple ice scheme are slightly worse than the Reisner schemes. Overall, Reisner1 with the fixed slope intercept for snow has the best bias and rms error scores for this case. This illustrates that using the most sophisticated microphysical schemes (i.e., Reis1VS and Reisner2) does not guarantee a better precipitation forecast, and that future research is needed to make improvements in these more sophisticated schemes.

Figure 19a shows the observed and model precipitation (in mm) from 2000 UTC 6 February 1996 through 2000 7 February for a portion of the 4-km domain using the Reisner2 (graupel) scheme. The observed precipitation during this 24-h period ranged from 80 to 200 mm over the southern Washington and northern Oregon Cascades, and 25 to 50 mm in the lowlands of western Washington and Oregon, to generally less than 20 mm east of the Cascades. The model verified best (within 20% of observed) over the western Washington and Oregon lowlands and near the Cascade crest, while it generally underpredicted (by more than a factor of two) in the lee of the Cascades and within major gaps such as Stampede Gap (not shown).

Figures 19b–e present the difference in 24-h precipitation between the Reisner2 run and the various other schemes at 4-km resolution. The differences are also apparent in cross section YY' taken across a lower portion of the northern Oregon Cascades at 0600 UTC 7 February (Fig. 20). The warm rain scheme predicts too much precipitation along the windward slope (50%–75% more than the Reisner2 scheme) and not enough over the crest (10%–30% less than Reisner2) (Fig. 20b). Without ice microphysics, excessive condensate resides as cloud water, which rapidly gets autoconverted to rainwater and precipitates over the windward slopes (Fig. 20a). The inclusion of ice in the simple ice scheme reduces and broadens the maximum precipitation around the crest (Figs. 19c, 20b); however, the simple ice scheme still produces 10%–20% more precipitation along the lower windward slopes and 5%–10% less over the crest than the Reisner2 (graupel) scheme.

The inclusion of super-cooled water in the Reisner1 scheme reduces the amount of precipitation along the lower windward slope while more falls near the crest and immediate lee compared to the simple ice scheme (Figs. 20b,c); however, Reisner1 still produces slightly more precipitation over the windward slope than Reisner2 (Fig. 19e). The inclusion of graupel in the Reisner2 increases the precipitation in the immediate lee of the crest of the lower Cascades (Figs. 20c,d), producing a double maximum; however, for many higher portions
Fig. 17. Cross section YY' showing the 4-km winds in the section, rainwater mixing ratio (solid black every 0.04 g kg$^{-1}$), snow mixing ratio (solid gray every 0.1 g kg$^{-1}$), and graupel (dashed every 0.04 g kg$^{-1}$) for the (a) 29-, (b) 38-, and (c) 57-vertical level simulations at 0600 UTC 7 Feb 1996 (18 h). The winds in the sections are not plotted at all vertical levels. The location of the section is shown in Fig. 16b.
of the Cascades, the Reisner1 had more precipitation in the immediate lee than the Reisner2 (Fig. 20c).

The variable snow intercept (Reis1 VS) also had a significant impact on the precipitation distribution compared to the fixed intercept (Reisner1) and graupel (Reisner2) schemes. The Reis1 VS scheme had 10%–30% less precipitation near the Cascade crest compared to the other Reisner schemes (Fig. 19e). Cross section YY’ indicates Reis1 VS had approximately 30% less snow aloft, the melting band was more diffuse and lower over the crest, and there was no secondary maximum of rainwater generated by snow being advected downward near the crest (Fig. 20e).

Figure 21 summarizes the microphysical results by showing the average precipitation profile for the 24-h period using the boxed region in Fig. 14 from all the experiments. The warm rain scheme produces significantly more precipitation than the other schemes over the windward slope and much less precipitation over the lee slope. The simple ice scheme reduces the precipitation over the windward slope compared to the warm rain and increases the precipitation in the lee. The Reisner schemes are similar along the windward slope, with somewhat less precipitation than the simple ice, and the Reisner2 has the most precipitation near the crest because of graupel processes.

Both the simple ice and the Reisner1 schemes placed more precipitation on the leeside of the Cascades than the Reisner2 scheme; however, neither scheme removed the underprediction downwind of orographic barriers (not shown). Colle et al. (1999) also noted an underprediction in the lee of the Cascades for the real-time MM5 forecasts, which used the simple ice scheme during the 1996–97 cool season. On the other hand, these sensitivity results suggest that the Reisner schemes in the MM5 may reduce some of the high biases that Colle et al. (1999) noted over the lower windward slopes at 12-km resolution (see their Fig. 15).
Table 2. (a) Bias scores and (b) rmse for the various microphysical schemes in the MM5 for the period 2000 UTC 6 Feb 1996 through 2000 UTC 7 Feb 1996. The bias score is defined here as the total model precipitation within each threshold range divided by the total observed within each threshold range. The bold numbers indicate the best score for a given threshold category.

<table>
<thead>
<tr>
<th>Threshold (mm)</th>
<th>Reisner2</th>
<th>Reisner1</th>
<th>Reis1VS</th>
<th>Simple ice</th>
<th>Warm rain</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(a) Bias</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>All</td>
<td>0.876</td>
<td>0.896</td>
<td>0.866</td>
<td>0.875</td>
<td>0.885</td>
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<tr>
<td>≤20</td>
<td>2.884</td>
<td>2.796</td>
<td>2.573</td>
<td>2.401</td>
<td><strong>1.851</strong></td>
</tr>
<tr>
<td>20–60</td>
<td>1.282</td>
<td>1.331</td>
<td>1.259</td>
<td>1.292</td>
<td><strong>1.134</strong></td>
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<tr>
<td>60–100</td>
<td>0.913</td>
<td><strong>0.934</strong></td>
<td>0.925</td>
<td>0.904</td>
<td>0.918</td>
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<tr>
<td>100–130</td>
<td>0.629</td>
<td>0.641</td>
<td>0.628</td>
<td>0.638</td>
<td><strong>0.677</strong></td>
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<tr>
<td>&gt;130</td>
<td>0.670</td>
<td>0.678</td>
<td>0.627</td>
<td>0.676</td>
<td><strong>0.796</strong></td>
</tr>
<tr>
<td>(b) Rmse error</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>All</td>
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<td>34.336</td>
<td>35.035</td>
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<tr>
<td>60–100</td>
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<td><strong>30.772</strong></td>
<td>31.224</td>
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<td>100–130</td>
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<td>45.883</td>
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<td>&gt;130</td>
<td>66.776</td>
<td>66.342</td>
<td>72.888</td>
<td>65.916</td>
<td>66.350</td>
</tr>
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</table>

Fig. 19. (a) Total model precipitation using the Reisner2 scheme (solid lines contoured every 25 mm) for a portion of the 4-km MM5 domain from 2000 UTC 6 Feb 1996 through 2000 UTC 7 Feb 1996. The numbers at the × locations are observed precipitation in mm. Topography is shaded for reference. (b) Precipitation differences (in mm) between the warm rain and the Reisner2 scheme (warm rain − Reisner2) for a selected region of the 4-km domain centered around Portland, Oregon (PDX). The contour interval is every 20 mm and positive values are shaded. Terrain is lightly contoured every 300 m for reference. (c) Same as (a) except for the simple ice minus Reisner2 and the contour interval is every 10 mm. (d) Same as (c) except for Reisner1 minus Reisner2. (e) Same as (c) except for Reis1VS minus Reisner2.
Fig. 20. Cross section YY’ showing the 4-km winds in the section, rainwater mixing ratio (solid black every 0.04 g kg⁻¹), snow mixing ratio (solid gray every 0.1 g kg⁻¹), and graupel (dashed every 0.04 g kg⁻¹) for the (a) warm rain, (b) simple ice, (c) Reisner1, (d) Reisner2, and Reis1VS simulations at 0600 UTC 7 Feb 1996 (18 h). The location of the section is shown in Fig. 16b.

At 12-km resolution, differences in precipitation between the various microphysical schemes were similar to those found in the 4-km domain (not shown). In contrast, the precipitation differences in the 36-km domain between the warm rain and Reisner2 was generally less than 15% (not shown) compared to 50%–75% at 4-km resolution. As a result, there is little benefit in using a sophisticated microphysical scheme in complex terrain when the model resolutions are greater than 30 km since much of the error originates from the lack of terrain resolution.

One of the crucial parameterizations in these microphysical schemes is the empirical fall speed relation for snow and ice particles. Since snow and ice particles are readily carried by the horizontal flow, small differences in fall speeds may substantially influence where the precipitation ultimately reaches the ground. For example, all of the above MM5 microphysical schemes use the Rutledge and Hobbs (1983) expression for the fall speed of snow, which is based on the Locatelli and Hobbs (1974) fall speed for unrimed radiating assemblages of plates, side planes, bullets, and columns; namely,

\[ V = 11.72D^{0.41}, \]  

(3)
Fig. 22. Fall speed vs snow diameter for unrimed aggregates of side planes, columns, and bullets (LH 74) and the modified fall speed expression by Cox (1988). The gray line shows an empirical relationship for unrimed radiating assemblages of dendrites from Zikmunda and Vali (1972).

where \( D \) is the particle diameter (in m) and \( V \) is the fall speed (in m s\(^{-1}\)).

Cox (1988) used the following fall speed expression for snow,

\[
V = 16.8D^{0.527}.
\]  (4)

Figure 22 shows the Cox (1988) snow fall speed changes as a function of particle diameter compared to the Rutledge and Hobbs (1983) relationship. The Cox expression yields snow fall speeds that are generally 20% smaller than Rutledge and Hobbs. Overall, the Cox expression is closer to Zikmunda and Vali's (1972) empirical expression for unrimed radiating assemblages of dendrites.

To determine how much impact the Cox snow fall speeds would have on precipitation at the surface, the 1200 UTC 6 February MM5 simulation using the simple ice scheme was run down to 4-km resolution using the Cox fall speed [Eq. (4)] through 36 h (COX experiment). Figure 23 shows the difference in 24-h precipitation between the COX simulation and the simple ice run. The reduction in snow fall speeds shifted precipitation from the windward (west) side of major barriers to the lee, with 10%–20% less precipitation on the windward side of the Cascades and 10%–60% percent more near the crest and immediate lee of the barrier. Figure 24 shows cross section XX' taken across the Cascades at

![Figure 23](image-url)  
**Fig. 23.** Precipitation difference (mm) between the COX experiment and the control (simple ice) run (COX-CTL) from 2000 UTC 6 Feb 1996 through 2000 UTC 7 Feb 1996 for the 4-km resolution. The contour interval is every 10 mm and positive values are shaded. Terrain is lightly contoured every 300 m for reference.

![Figure 24](image-url)  
**Fig. 24.** (a) Cross section XX' showing model reflectivities and winds for the 4-km simple ice simulation at 0300 UTC 7 Feb 1996 (15 h). The relative humidities with respect to ice are dashed every 20%. (b) Same as (a) except for the Cox experiment (altered snow fall speeds). The location of the section is shown in Fig. 13.
0300 UTC 7 February 1996 for the COX and control simulations. The larger model-calculated radar reflectivities over the windward and lee slopes in the COX experiment suggests that there is significantly more snow getting advected over the barrier. Furthermore, the COX simulation appears to verify better over the lee of the Cascades compared to the observations (Fig. 14a). Because more precipitation falls around the crest, there are some bias and rms error improvements using this new fall speed relation for higher thresholds (Table 2). However, the overall verification scores are not significantly better than the control since the bias and rms errors worsened for the lower thresholds. More cases need to be simulated with this COX fall speed relationship to determine whether there are significant benefits for a range of synoptic situations. Furthermore, the microphysical sensitivities noted above point toward the need for more detailed verification of the microphysics in these numerical models. To accomplish this, better and more comprehensive observations of both microphysics and basic-state variables are required.

5. Summary and conclusions

This paper describes the flooding event of 5–9 February 1996 in which a series of landfalling Pacific storms dumped 30–70 cm of rain at many mountain sites across southwest Washington and northwest Oregon. The heavy precipitation combined with the 200–300 mm of liquid precipitation from snowmelt in the mountains resulted in severe flooding west of the Cascade Mountains and approximately 500 million dollars in damage. The upper-level pattern during this event was characterized by strong southwesterly flow throughout the troposphere extending from near Hawaii to the Pacific Northwest. This subtropical connection brought copious amounts of moisture and anomalously warm temperatures over the Pacific Northwest.

Three 8–44-h forecasts from the PSU–NCAR MM5 were used to simulate the storm-total precipitation. To determine how model precipitation changes with increasing horizontal resolution, the simulations were run at 36-, 12-, 4-, and 1.33-km resolution and verified using over 300 rain gauges in western Washington and Oregon. Overall, there was a significant improvement in the precipitation forecast skill as the grid spacing was decreased from 36 to 4 km; however, the 12- and 4-km resolutions had too much precipitation shadowing in the lee of barriers. Going from 4- to 1.33-km resolution did not produce a significant improvement, although the 1.33-km resolution did produce more precipitation in the immediate lee of the Cascades and therefore verified better in those regions. However, the 1.33-km domain generated bull’s-eyes of heavy precipitation immediately downwind of the major peaks, which were not confirmed by observations.

Additional simulations were conducted to show the effects of changing vertical resolution from 29 to 57 sigma levels. The model precipitation increased over the windward slopes of the Cascades by 10%–30% going from 29 to 38 levels; further increasing the number of levels to 57 decreased windward precipitation to the 29-level amounts. The leeside precipitation also varied by 20%–80% for the different vertical resolutions as a result of subtle differences in mountain wave structures over the Cascades.

The simulated precipitation over the Cascades was also compared with the radar reflectivities obtained from the Portland, Oregon, WSR-88D radar. The cloud and precipitation water in the model was underpredicted over the Cascade crest and lee slopes, which suggests that the model may be overpredicting the leeside mountain wave and/or the microphysical scheme is not generating or maintaining enough ice aloft to get advected over the barrier. Unfortunately, there has been relatively little quantitative evaluation of the microphysical packages in mesoscale models, mainly due to the absence of simultaneous microphysical and basic-state data over orographic barriers. Therefore, future field experiments using research aircraft are needed to gather such datasets over complex terrain.

Sensitivity results using microphysics of varying sophistication were shown by comparing 24-h precipitation amounts from four different MM5 microphysical schemes for the February 1996 flooding event. The warm rain scheme dumped too much precipitation along the windward slopes, which underlines the importance of ice microphysics during the cool season. However, the most sophisticated schemes (i.e., schemes including graupel or a variable slope-intercept for snow) did not provide the best forecast. An additional microphysical sensitivity run was completed in which the snow fall speed in the model was reduced by approximately 20%, which resulted in substantially more precipitation advecting to the lee of the Cascades.

Finally, a simulation without the coastal range was presented in order to show how barrier effects precipitation downwind over the Cascades. Precipitation shadowing in the lee of the coastal range decreased precipitation over the lower windward slopes of the Cascades by 10%–20%. The coastal terrain also indirectly enhanced precipitation in the lee of the Cascades by around 10%.

Acknowledgments. This research was supported by the USWRP program (ATM-9612876) and the Office of Naval Research (Grant No. N00014-94-1-0098 and N00014-98-1-0193). We wish to thank Dr. Brad Colman for the Portland WSR-88D data; Scott Pattee for the SNOTEL observations; Dr. Brad Smull and the Environmental Technology Laboratory (ETL) in Boulder, Colorado, for the Astoria profiler data; Dr. David Schultz for programs to read the NCAR-archived ECMWF grids; Curtis James for software necessary to interpolate and display the WSR-88D data, and Ken Westrick for generating the topography/rain gauge map.
Comments and suggestions by Ken Westrick and the anonymous reviewers helped improve the manuscript. Use of the MM5 was made possible by the Microscale and Mesoscale Meteorological Division of the National Center for Atmospheric Research (NCAR), which is supported by the National Science Foundation.

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