An Observational and Modeling Study of the Interaction of Low-Level Southwesterly Flow with the Olympic Mountains during COAST IOP 4

BRIAN A. COLLE AND CLIFFORD F. MASS

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

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ABSTRACT

This paper presents an observational and modeling study of the three-dimensional flow and precipitation around the Olympic Mountains of northwest Washington State for a case of relatively steady south-southwesterly flow. This study utilized data from COAST IOP 4 (the fourth intensive observing period during the Coastal Observation and Simulation with Topography field experiment) that took place during 9–10 December 1993. One of the most important data sources was the NOAA P-3 aircraft, which provided flight-level data, reflectivity, and Doppler winds as it flew around the Olympics. Features that were mapped successfully with the NOAA P-3 included flow splitting around the windward pressure ridge, strong southeasterlies (to approximately 25 m s⁻¹) northeast of the Olympics caused by downgradient acceleration into the lee trough, the rapid transition from strong southeasterlies to lighter easterlies (5–10 m s⁻¹) in the Strait of Juan de Fuca, strong downslope flow and a hydraulic jump–like transition over the lee slope, the Olympic rain shadow, and the enhancement of precipitation on the windward (southern) slopes of the Olympics.

This event was simulated at resolutions down to 3 km using the nonhydrostatic version of the Penn State–NCAR mesoscale model (MM5). Overall, the model realistically simulated the observed features.

The model simulation indicated significant asymmetry in the pressure field around the Olympics (the lee troughing was 5–7 mb deeper than the windward ridging). This asymmetry remained when the Olympics were replaced by a symmetrical bell-shaped barrier. A simulation without latent heating shows that this asymmetry was not the result of latent heating reducing the strength of the windward ridging. Instead, in an environment with high stability near crest level and negative vertical wind shear above the barrier, a mountain wave forced strong downslope flow on the lee side and the resulting pressure asymmetry. Momentum diagnostics along the western Washington coast indicate an approximate geostrophic balance normal to the coast and an antitropical balance parallel to the coast. To further demonstrate the three-dimensional flow around the Olympics, a series of trajectories launched upstream of the barrier are shown.

1. Introduction

The interaction of synoptic-scale flow with topography produces a spectrum of mesoscale phenomena that greatly influences the weather of the western United States. In the Pacific Northwest, low-level flow off the Pacific Ocean interacts with the Olympic and Coastal Mountains of Washington and Oregon (Fig. 1), producing a variety of local phenomena including convergence zones, lee troughs, rain shadows, and gap winds. The Olympics, an isolated terrain feature that is nearly bell-shaped in structure (80 km in diameter with peaks generally less than 1800 m), has a major impact on western Washington weather and is well suited for the study of airflow around a three-dimensional barrier. The coastal mountains of Oregon, with peaks generally less than 1000 m, have a lesser but still significant impact on the flow in the coastal zone.

The COAST (Coastal Observation and Simulation with Topography) field experiment occurred between 29 November and 13 December 1993. Of the six missions of COAST, four followed Pacific fronts to landfall, while the others investigated the interactions of relatively steady southwesterly flow with the coastal terrain. The key data collection platform of the intensive observational periods (IOPs) of COAST was the NOAA P-3 aircraft. In addition to providing flight-level data, the P-3 has a tail X-band (3.22 cm) radar that retrieves reflectivity and Doppler winds for a narrow volume in the vertical on either side of the plane, and a lower-fuselage C-band (5.59 cm) radar that provides reflectivity in the horizontal. In addition, profiler and radiosonde units were installed at Astoria, Oregon (AST on Fig. 1), and additional soundings were launched from the National Weather Service radiosonde site at Quillayute, Washington (UIL on Fig. 1).

Several IOPs of COAST have been simulated using the Penn State–National Center for Atmospheric Research (NCAR) mesoscale model (MM5). The pur-

Corresponding author address: Mr. Brian A. Colle, Dept. of Atmospheric Sciences, University of Washington, Box 351640, Seattle, WA 98195-1640.
E-mail: colle@atmos.washington.edu

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of the winds and precipitation around the Olympics and along the Oregon coast. Since there was not enough precipitation along the Oregon coast to obtain useful Doppler winds from the tail radar, the results shown in this paper will focus solely on the region around the Olympics. Overall, the observational and modeling efforts of this case provide the most detailed analysis to date of the three-dimensional airflow and precipitation around the Olympic barrier. In addition, this is probably the first time that aircraft-based Doppler winds have been retrieved over an orographic barrier. Before discussing this case in detail, a review of previous theoretical studies of flow around three-dimensional isolated topography is presented, followed by a review of past observational studies of atmospheric flow and precipitation around realistic mesoscale barriers.

2. Previous observational and theoretical studies

a. Theoretical studies

Most previous studies of flow around three-dimensional barriers have focused on strongly stratified regimes in which the Froude number \( Fr = U/NH \), where \( U \) is the up-slope wind component, \( N \) is the Brunt–Väisälä frequency, and \( H \) is the height of the obstacle) is much less than 1. Drazin (1961) showed that for this regime, inviscid, incompressible flow tends to pass horizontally around rather than over a three-dimensional barrier. This was verified with laboratory experiments by Riley et al. (1976), Brighton (1978), Baines (1979), and Hunt and Synyer (1980). When the Froude number is less than approximately 0.4, the flow may stagnate on the windward slope and split laterally around the barrier (Smith 1980). Using a nonhydrostatic numerical model to simulate the flow around the island of Hawaii, Smolarkiewicz et al. (1988) and Smolarkiewicz and Rotunno (1990) showed that low-Froude number flow \( (Fr < 0.5) \) leads to nonlinear effects such as flow reversal on the windward side and a pair of vertically orientated vortices on the lee side of the barrier.

When the Froude number is greater than one, linear theory becomes more accurate (Smith 1980). Using a linear, hydrostatic, analytic model of flow around an isolated bell-shaped mountain, Smith (1980) showed that the windward ridge and leeward trough have identical amplitude and shape. On the windward side, pressure ridging forces most air parcels around the barrier, while on the lee side there is low-level divergence and the sinking of potentially warmer air from aloft. The distance upstream of the barrier at which a parcel is decelerated and deflected for the large–Froude number regime is given by the radius of deformation \( \left( \frac{\nu}{\alpha} \right) \), where \( \alpha \) is the height of the barrier (Pierrehumbert and Wyman 1985).

Thorsteinsson (1988) used a hydrostatic, \( f \)-plane, primitive equation model to study stratified flow past...
an isolated bell-shaped topographic barrier for a variety of Froude and Rossby numbers.\(^1\) As noted by Mass and Ferber (1990), Thorsteinsson’s experiment using \(R_m = 4\), \(Fr = 1\), and \(L_m = 25\) km was the closest analog to the parameters characteristic of the Olympics during winter (i.e., southwesterly, slightly stable flow impinging on the Olympics with \(U \sim 15\) m s\(^{-1}\), \(N \sim 0.01\) s\(^{-1}\), and \(h_m \sim 1800\) m). The resulting shape and orientation of the windward pressure ridge and leeward trough were very similar to those shown by the linear model; however, Thorsteinsson’s pressure amplitude was asymmetric, with the lee trough of greater magnitude than the windward ridge.

Smith and Grønås (1993) applied a hydrostatic, sigma coordinate model on a Gaussian hill and showed that linear theory does well when the nondimensional mountain height (inverse Froude number, \(h^*\)) is small (\(h^* < 0.5\)). However, they demonstrated that there are considerable nonlinear effects as \(h^*\) increases. When \(h^*\) increases to about 1.1 (Froude number decreases to 0.9), stagnation first occurs at a point above the lee slope. This contrasts to linear theory, which overpredicts the magnitude of the windward ridge, thus resulting in stagnation occurring first on the windward slope.

Using scale analysis, Overland (1984) and Overland and Bond (1993, 1995) described the dynamical balances associated with flow adjacent to coastal barriers. Overland and Bond (1993) showed that when \(U/L\) is small (the cross-barrier length scale/along-barrier length scale is much less than 1) and the cross-barrier Rossby number, \(V/\Omega L\) (where \(V\) is the magnitude of the along-barrier flow), is approximately equal to or greater than one, ageostrophic down-gradient flow develops adjacent to the barrier, with the along-barrier pressure gradient primarily balancing the along-barrier acceleration and friction, and the mass field adjusting so that the pressure gradient normal to the barrier balances the Coriolis force associated with the terrain-parallel flow. Overland and Bond (1993, 1995) noted that these balances extend approximately a Rossby radius of deformation, \(L_r = N h_m/\Omega\), from the barrier. However, if the height of the disturbance \((h)\) is less than the height of the barrier \((h_m)\), the Rossby radius of deformation \(L_r\) is instead given by \(N h/\Omega\) (Overland and Bond 1995).

b. Past observational and modeling studies of flow impinging on realistic mesoscale barriers

Several observational studies have documented mesoscale effects induced by the Olympic Mountains.

On 13 February 1979, strong southwesterly flow, associated with an intense midlatitude cyclone, impinged on the Olympics, leading to the development of an intense mesoscale front southeast of the low and the subsequent destruction of the Hood Canal Bridge (Reed 1980). Smith (1981) noted that linear theory predicts the large pressure gradient observed to the south of the mesoscale during the Hood Canal Storm; however, he suggested that the magnitude of the observed lee low was greater than the windward pressure ridge because of nonlinear effects and the asymmetry of the Olympics. Walter and Overland (1982) found that light winds and weak troughing were present in the lee of the Olympics for Froude numbers of approximately 1, and that the intense lee low associated with the Hood Canal Storm was associated with a Froude number of 4.6. Mass (1981) discussed the Puget Sound convergence zone, which occurs primarily when west-northwesterly flow along the Washington coast splits around the Olympics and converges over Puget Sound. Mass and Dempsey (1985) described the topographic convergence west of the Olympic Peninsula that is associated with low-level northeastly flow impinging on the Olympics. Overland and Walter (1981) studied the accelerating gap flow through the Strait of Juan de Fuca (cf. Fig. 1).

Using conventional synoptic data and a microbarograph network around the Olympics, Mass and Ferber (1990) found that the lee trough was generally stronger than the windward ridge. This asymmetry diminished as the Froude number increased to approximately 1. It was also noted that the amplitude of the pressure perturbations was correlated better with wind speed at 850 mb than with Froude number or stability.

Steenburgh and Mass (1996) used both mesoscale analyses and the Penn State–NCAR mesoscale model (MM5) to document some of the time-dependent aspects of the pressure perturbations around the Olympics for a case in which an intense extratropical cyclone interacted with the coastal orography of Washington and Oregon (Inauguration Day storm 1993). They showed that the development of the lee trough to the east of the Olympics occurred within 2–3 h after frontal passage, and did not explain the peak winds observed over Puget Sound. They further showed the asymmetries in the pressure perturbation around the Olympics (the lee trough was twice as strong as the windward ridge) occurred when the Froude number was relatively high (Fr = 1–2).

There have been a number of observational studies of the flow around other mesoscale three-dimensional barriers. The “Denver cyclone,” a mesoscale vortex that forms in the lee of the Palmer Divide (an east–west ridge to the south of Denver), has been shown to form when the ambient flow is primarily southerly and

\(^1\) The Rossby number, \(R_m = U/\Omega L_m\), where \(U\) is the upstream velocity, \(\Omega\) is the Coriolis force, and \(L_m\) is the half-width of the mountain, is the ratio of the inertial and Coriolis accelerations. The Coriolis acceleration is small compared to inertial accelerations when \(R_m \gg 1\).
the Froude number is low ($Fr < 0.75$) (Crook et al. 1990). Enhanced northerly low-level flow impinging on the coastal mountains of the southern California bight results in lee troughing and the generation of the Catalina eddy (Mass and Albright 1989). Smolarkiewicz et al. (1988) and Smith and Grubisic (1993) studied the terrain-altered wind field induced by the island of Hawaii.

The distribution of precipitation is greatly affected by the Olympic Mountains. Mass and Ferber (1990) showed the climatological mean annual precipitation for northwestern Washington (their Fig. 4), with windward enhancement (greater than 4 m) over the southwestern side of the Olympics and a leeward rain shadow (less than 0.4 m) northeast of the barrier. Parsons and Hobbs (1983) described the evolution of warm-frontal and warm-sector rain bands that moved across the Olympics during the CYCLES (Cyclonic Extratropical Storms) project. They noted that warm-frontal rainbands entering the central Puget Sound (Fig. 1) attenuated more when the flow at approximately 850 mb was west-southwesterly rather than southwesterly. This was attributed to the greater subsidence off the Olympics for those cases that had more of a westerly component.

c. Objectives

Prior to the COAST project, a detailed dataset describing the three-dimensional flow around the Olympics did not exist. By utilizing the Doppler winds and reflectivity from the NOAA P-3 aircraft as well as a high-resolution simulation of COAST IOP 4 using the Penn State–NCAR MM5, this paper addresses several questions concerning the flow and precipitation around three-dimensional barriers in general, and the Olympics in particular, including the following:

- What is the three-dimensional airflow around the Olympics during relatively steady southwesterly flow at low-levels?
- What is the observed precipitation distribution around the barrier?
- Can one successfully simulate the flow and precipitation around the Olympics using a high-resolution mesoscale model?
- What are the mechanisms responsible for the asymmetries in the pressure field around the Olympics?
- What is the offshore length scale of orographic influence and the momentum balances around the barrier?

3. Brief synoptic overview

At 1800 UTC 9 December 1993 (about the time the NOAA P-3 took off), a deep occluded cyclone with a minimum pressure of 950 mb was situated southwest of British Columbia (Fig. 2a). A well-defined occluded front extended eastward along 50°N, while a weak warm front was moving northeastward toward the Washington and Vancouver Island coasts. Ahead of this warm front, the surface winds along the Washington coast were mainly southeasterly.

The 1800 UTC sounding at Quillayute (UIL on Fig. 1) on the northwest Washington coast (Fig. 3a)
shows that the low-level flow below 900 mb was southeasterly at 20 m s\(^{-1}\), with strong veering immediately above. Above the weak frontal inversion at 900 mb, the atmosphere was moist adiabatic. At 1900 UTC, surface observations indicated the warm front was located just north of Astoria, Oregon (AST on Fig. 1), where the surface winds had changed from easterly (5 m s\(^{-1}\)) at 1800 UTC to southerly (10 m s\(^{-1}\)) by 1900 UTC (not shown). The 1830 UTC sounding at AST indicates the warm front near the surface, with south-southwesterly flow of 25 m s\(^{-1}\) from 950 mb upward (Fig. 3b).

By 0000 UTC 10 December 1993, the surface low had moved northeastward and maintained its central pressure (Fig. 2b). The winds along the Washington and northern Oregon coasts had veered to the south and strengthened as the warm front moved northward along the Washington coast. At 850 mb, the tight height gradient over western Washington and Vancouver Island led to south-southwesterly winds reaching 35–40 m s\(^{-1}\) upstream of the Olympics (Fig. 4). UIL’s sounding at 0000 UTC 10 December (Fig. 3a) shows that the winds between 900 and 950 mb veered to a more southerly direction (at about 20 m s\(^{-1}\)) between 1800 and 0000 UTC with the approach of the warm front. The sounding upstream of the Olympics at AST (Fig. 3b) at 0000 UTC 10 December shows that strong south-southwesterlies (approaching 40 m s\(^{-1}\)) were present throughout the lower troposphere.
winds were recomputed using this vertical motion field, the estimated fall speeds, and the observed radial velocities; this process was iterated until the three-dimensional wind field converged.

a. Horizontal and vertical structures

At 1800 UTC 9 December, the southerly flow ahead of the warm front led to the development of pressure perturbations on the windward (southern) and leeward (northern) sides (Fig. 5). Near the center of the lee trough the winds were nearly calm. The lee troughing produced an enhanced pressure gradient on the northeast side of the Olympics, resulting in winds of approximately 20 m s\(^{-1}\) southeast of the San Juan Islands. Along the Washington coast the winds were mainly southeasterly at 15–20 m s\(^{-1}\).

The computed horizontal winds from the NOAA P-3 at \(\sim 800\) m above sea level (Fig. 6a) provides some evidence of flow deflection around the south side of the Olympics at 1800 UTC. The winds were mainly south-southeasterly immediately to the north of Hoquiam (HQM) and became more east-southeasterly and weakened as the flow approached the windward slopes the Olympics (\(\sim 20–30\) m s\(^{-1}\) near the coast versus 10–20 m s\(^{-1}\) farther inland). A vertical cross section along the south flank of the Olympics (AA’) shows the sloping ascent of the strong southerlies (greater than 30 m s\(^{-1}\)) over the barrier and the weakening of the southerly flow as the air was deflected by the barrier (Fig. 6b).

A weak stratiform rainband (less than 30 dBZ) was located over the southern foothills of the Olympics around 1800 UTC (Fig. 6a), with the corresponding bright band approximately 2 km above mean sea level (Fig. 6b). This quasi-stationary rainband likely was forced by upslope flow and low-level convergence associated with the flow deflecting around the Olympics.

By 2100 UTC 9 December (Fig. 7), the weak warm front had passed the southern coastal stations as suggested by the transition to strong southerlies at Hoquiam (HQM). As shown previously with theUIL soundings (Fig. 3), between 1800 and 0000 UTC the low-level winds at 1–2 km increased 3–7 m s\(^{-1}\) and became southwesterly. As a result of this stronger cross-barrier flow, the lee troughing near the northeast corner of the Olympics had intensified between 1800 and 2100 UTC to 994 mb (7-mb perturbation from the synoptic pressure field). Strong southeasterly flow (25 m s\(^{-1}\)) continued south of the San Juan Islands in the region of enhanced pressure gradient.

At 2100 UTC the NOAA P-3 aircraft was flying at 150–200 m above the surface along the Strait of Juan de Fuca. Flight-level winds (Fig. 7) show the rapid transition from strong southeasterly flow (25 m s\(^{-1}\)) over the Pacific to easterly flow (15 m s\(^{-1}\)) within the western entrance of the strait. These easterlies weakened to less than 5 m s\(^{-1}\) north of the Olympics, where the temperature reached a local maximum. The rapid transitions down the strait (between points B and B’ in Fig. 7) can be seen more clearly by looking at 5-s running means of temperature, sea level pressure, and horizontal winds as well as 1-s means of vertical velocity (Fig. 8). As the aircraft entered the strait from the west at 2034 UTC, the winds shifted rapidly from south-southeasterly (160°) to east-southeasterly (120°) and weakened from 25 to 13 m s\(^{-1}\) within 1 min (5 km). At point C, the temperature spiked to 11.5°C, the winds became light and variable, and there was a local increase in turbulence (larger fluctuations in vertical velocity). The lighter winds around point C can be attributed to the relatively flat pressure gradient across that part of the strait. Such a weak pressure gradient resulted from the superposition of the eastward-increasing synoptic-scale pressure field and the oppositely directed mesoscale pressure gradient associated with the lee trough.\(^2\) In contrast, in the eastern strait the superposition of the lee trough and synoptic-scale pressure field resulted in the strong southeasterlies to the northeast of the Olympics.

Figures 9, 11, 12, and 14 show the wind and precipitation structures at various levels around the Olympics (i.e., 300, 800, 1300, and 1800 m ASL) using the tail Doppler radar of the NOAA P-3 aircraft as it flew around the barrier between 1900 and 2200 UTC 9 De-

\(^2\) As observed from the P-3 aircraft, many ships were grouped around this location in order to take advantage of the calm seas.
November (cf. Fig. 1). At 300 m (Fig. 9), a warm-frontal rainband was located just offshore of the Washington coast. The winds near the coast were primarily southeasterly at 25–30 m s\(^{-1}\), but veered to more southerly offshore within the rainband. South of Vancouver Island at 300 m, the winds veered from south-southeasternly offshore to southeasterly within 30 km of the Vancouver Island coast. Within this coastal transition zone, enhanced precipitation of 30 dBZ extended offshore approximately 20 km. Cross section \(DD'\) shows strong southerly flow (greater than 30 m s\(^{-1}\)) rising over the barrier (Fig. 10). Orographically induced low-level convergence and resulting upward motion (\(-1\) m s\(^{-1}\)) is situated at the leading edge of the precipitation enhancement along the coastal zone of Vancouver Island. The northward advection of precipitation particles generated by this enhanced vertical motion resulted in the precipitation enhancement immediately southwest of Vancouver Island. Another area of enhanced upward motion (\(~0.8\) m s\(^{-1}\)) was collocated with the edge of the sloping terrain of Vancouver Island. To the east of the Strait of Juan de Fuca the NOAA P-3 flew through the heart of the Olympic rain shadow where the reflectivities dropped dramatically (Fig. 9). The winds at 300 m over the eastern Strait of Juan de Fuca were primarily southeasterly at less than 10 m s\(^{-1}\). In contrast, over Puget Sound, reflectivity was high with some echoes exceeding 40 dBZ. The winds at this level were primarily southwesterly at 5–10 m s\(^{-1}\) over central Puget Sound, but became more southeasterly to the north as the flow was deflected around the northeast corner of the Olympics. Meanwhile, the winds were light and variable near the southeast slopes of the Olympics, which corresponds to the weak pressure gradients.
within the windward pressure ridging (Fig. 7). The few available wind vectors to the south of the Olympics show westward deflection of the incoming flow by the windward pressure ridging.

At 800 m (Fig. 11), orographic precipitation enhancement and the coastal wind transition to southeasterly were still evident within 25 km of southwestern Vancouver Island. A large contrast in reflectivity existed between the rainshadow on the northeast side and precipitation enhancement on the windward sides of the Olympics. Even in areas of strong downslope flow immediately northeast of the Olympics, weak reflectivities (less than 20 dBZ) were still present. This light precipitation hugging the lee was probably the result of precipitation particles being blown over the Olympics and surviving the descent down the lee slopes. Hobbs et al. (1973) described this process in a study of precipitation fallout in the lee of the Washington Cascades. Southerly flow to the south of the Olympics was deflected around the barrier, with southeasterlies on the western side and southwesterlies to the east (Fig. 11). Near the pressure ridge on the southeast side of the Olympics, the winds were light and variable. The nature of the precipitation on the southern slopes of the Olympics at 2200 UTC 9 December was considerably different than taken by the NOAA P-3 approximately 4 h earlier (Fig. 6). Because of the advection of warmer and less stable air into the region after 2100 UTC (increasing low-level $\theta_e$), the reflectivities over the windward foothills of the Olympics had intensified (30–40 dBZ) and had become more convective in nature.\(^3\)

Flow splitting on the southern slopes of the Olympics is still evident at 1300 m (Fig. 12). Southerlies in excess of 30 m s\(^{-1}\) were observed along the west slopes of the Olympics while the winds veered to southwesterly offshore. Deflection of the winds near Vancouver Island was still evident at this level. Bands of enhanced windward precipitation greater than 30 dBZ were situated over and immediately downwind of the ridges.

\(^3\) The convective nature of the precipitation was observed better in the raw data where obvious convective towers were hugging the sloping terrain.
along the western side of the Olympics as a result of forced ascent over the terrain and the subsequent advection of precipitation particles.

The winds at 1300 m were less than 10–20 m s$^{-1}$ over the eastern Strait of Juan de Fuca, and stronger downslope flow (greater than 30 m s$^{-1}$) was occurring on the north side of the Olympics. The contrast in winds between the strait and the northern foothills of the Olympics can be seen in a cross section of P-3 tail radar radial velocities along section EE' (Fig. 13). Since this section is pointed primarily toward the southwesterly low-level flow, it gives a good estimate of the magnitude of the downslope winds. The cross section shows the acceleration of the downslope flow to 40–45 m s$^{-1}$ over the lee slope approximately 20 km south of the coast. There is a subsequent rapid deceleration of the flow to about 30 m s$^{-1}$ to the north and an ascent of the wind maximum to 2.5 km ASL over the strait. This wind transition has a structure resembling a hydraulic jump and is similar to that observed during downslope windstorms near Boulder, Colorado, using Doppler lidar (Neiman et al. 1988). Radial velocities over the Strait indicate flow away from the radar below 1 km, thus indicating a component of the easterly flow at low levels. As a result of the strong vertical wind shear at low levels over the strait, turbulence was present as evinced by the large horizontal variations in radial wind speeds and the previously mentioned vertical velocities fluctuations shown in the flight-level data (point C on Fig. 8). The reflectivities are shallow (less than 5 km deep) over the lee slope of the Olympics as a result of the downslope flow, while there is a recovery to higher reflectivities (7.5 km) over the Strait.

At 1800 m (Fig. 14), the flow was primarily southwesterly away from the immediate vicinity of the Olympics. In contrast, the flow over the central Olympics was primarily southerly at this level (the mean crest level of the Olympics is approximately 1800 m). At this level the wind transition near the southwestern side of Vancouver
Island had disappeared. Orographic enhancement of precipitation was still evident along the windward side of Vancouver Island and the Olympics. The Olympic rain shadow to the northeast of the barrier had approximately the same spatial coverage and was nearly vertically aligned with the rain shadowing observed at lower levels.

b. Observed precipitation measurements around the Olympics

The precipitation data for two time periods, 1800–2100 UTC 9 December and 2100 UTC 9 December to 0000 UTC 10 December (Figs. 15a and 15b, respec-
Fig. 11. NOAA P-3 Doppler winds and reflectivity (shaded) at 800 m above mean sea level between 1900 and 2200 UTC 9 December 1993.

southern foothills of the Olympics ranged from 2.0 to 4.0 cm in 3 h. This enhanced variability can be attributed to the more convective nature of the precipitation in the post-warm-frontal regime.

5. Model simulation of COAST IOP 4

a. Model description

The Penn State–National Center for Atmospheric Research (PSU–NCAR) MM5 was used in nonhydrostatic mode to simulate the COAST IOP 4 case and to provide additional data for diagnosing the three-dimensional flow and precipitation around the Olympics. The simulation used the explicit moisture scheme of Hsie et al. (1984), with improvements to allow for mixed liquid–ice phase below 0°C (Grell et al. 1994), and the Grell cumulus parameterization (Grell 1993) was applied except for the inner domain where the precipitation could be explicitly resolved. The planetary boundary layer (PBL) was parameterized using the Zhang and Anthes (1982) scheme. In order to prevent gravity waves from being reflected off the model top, Durrant and Klemp’s (1983b) upper radiative boundary condition was applied.
For this simulation, stationary 9- and 3-km domains were nested within a 27-km domain using a one-way interface (Fig. 16a). Thirty-one unevenly spaced full-sigma levels were used in the vertical, with the maximum resolution in the boundary layer.\(^6\) Five-minute-averaged terrain data were interpolated to the 27- and 9-km model grids using a Cressman analysis scheme and filtered by a two pass smoother/desmoother. For the 3-km domain, the Cressman scheme and filtering was applied to a 30-s topography dataset to better resolve the Olympic Mountains (Fig. 16b). To generate the initial and boundary conditions, the 12-h National Centers for Environmental Prediction (NCEP) global analyses (2.5° latitude–longitude resolution) were interpolated to the model grid to serve as the first-guess fields. Surface and upper-air observations were incorporated into the analysis using a Cressman-type analysis scheme (Benjamin and Seaman 1985). The gridded analyses were subsequently linearly interpolated in time in order to provide the lateral boundary conditions for the 27-km domain. An 18-h simulation initialized at 1200 UTC 9 December 1993 was completed for the 27- and 9-km domains. The 1-h output from the 9-km domain provided the boundary conditions for a separate 3-km simulation for the same time period.

\(b.\) High-resolution simulation of the flow around the Olympics

Figure 16a shows sea level pressure, surface winds,\(^7\) and 950-mb temperatures for the 27-km domain 6 h into the simulation at 1800 UTC 9 December 1993. In this outer domain, the positions of the fronts and cyclone center were very close to the observed (Fig. 2a) and the strength of the cyclone (954 mb) was within 4

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\(^6\) The 31 full-sigma levels were: \(\sigma = 1.0, 0.99, 0.98, 0.96, 0.93, 0.90, 0.87, 0.84, 0.81, 0.78, 0.75, 0.72, 0.69, 0.66, 0.63, 0.60, 0.57, 0.53, 0.49, 0.45, 0.41, 0.37, 0.33, 0.29, 0.25, 0.21, 0.17, 0.13, 0.09, 0.05, 0.0\).

\(^7\) The surface winds in Fig. 16a, as well as in subsequent figures, are for the model’s lowest half-sigma level (\(\sigma = 0.995\)), which is about 40 m above the surface.
mb of the analyzed central pressure. In agreement with the observations, the surface winds were south-south-easterly upstream of the Olympics.

Figure 17 shows the model sea level pressure and surface winds for the 3-km domain at 2100 UTC 9 December 1993. Compared with the observations (Fig.

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**Fig. 13.** Cross section $EE'$ showing radial velocities (shaded) from an aft scan of the NOAA P-3 tail radar. Negative and positive radial velocities indicate the component of flow toward and away from the direction of the radar beam, respectively.

**Fig. 14.** NOAA P-3 Doppler winds and reflectivity (shaded) at 1800 m above mean sea level between 1900 and 2200 UTC 9 December 1993.
7) and the 300-m radar data (Fig. 9), the model realistically simulated the surface flow and pressure patterns around the Olympics. Over the southeast slopes of the Olympics the winds were nearly calm and realistic flow splitting was evident. West of the Olympics, strong southerly winds (\(\sim 22 \text{ m s}^{-1}\)) over the Pacific Ocean became more southeasterly within about 30 km of the Washington coast.

The model realistically simulated the transitions within the Strait of Juan de Fuca (Fig. 17). Strong offshore south-southeasterlies (maximum of 22 m s\(^{-1}\)) switch rapidly to easterlies within the strait. In agreement with observations, the winds decrease dramatically to less than 10 m s\(^{-1}\) midway through the strait where the down-strait pressure gradient weakened. Further to the east, the model simulated realistically the

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Fig. 15. (a) Three-hour observed precipitation contoured every 0.3 cm starting at 0.1 cm for 1800–2100 UTC 9 December 1993. (b) Same as (a) except from 2100 UTC 9 December to 0000 UTC 10 December 1993.

Fig. 16. (a) Sea level pressure (solid) every 4 mbar, 950-mb temperature (dashed) every 2°C, and surface wind vectors for the 27-km domain at 1800 UTC 9 December 1993 (hour 6 of the simulation). The boxes show the locations of the 9- and 3-km nested domains. (b) The high-resolution terrain used for the 3-km domain contoured every 100 m.
magnitude (~23 m s⁻¹) and areal extent of the strong southeasterlies around the San Juan Islands.

A minor difference between the model and the observations is that the simulated lee trough was 2–4 mb deeper than analyzed at this time (cf. Fig. 7); however, no surface pressure observations exist on the northern slopes of the Olympics to verify the model pressures directly. Cross section FF' (cf. Fig. 17 for location) shows that the model developed a strong mountain wave with high momentum air (speeds greater than 30 m s⁻¹) thrusting over the barrier and then descending and accelerating rapidly on the lee side of the Olympics (Fig. 18). A hydraulic jump-like feature was present over the lee slope, analogous to observed structure seen in cross section EE' (Fig. 13). The mountain wave did not contain a self-induced critical layer, only an area of weaker horizontal winds at about 500 mb. Looking at a model sounding near the crest of the Olympics (not shown), the atmosphere was nearly isothermal from the surface up to 700 mb.

To facilitate the comparison of the simulation with the NOAA P-3 observations (Figs. 9, 11, 12, and 14), the model winds around the barrier were interpolated to 300, 800, 1300, and 1800 m ASL (Fig. 19). At 300 m (Fig. 19a), as observed, strong southerlies (~24 m s⁻¹) over the Pacific Ocean veered to more southeasterly near the coast. Lighter easterlies (~10 m s⁻¹) were present in the central Strait of Juan de Fuca, and stronger southeasterlies of 25 m s⁻¹, about the same strength as the NOAA P-3 observations (Fig. 7), were located southwest of the San Juan Islands. Also in agreement with the NOAA P-3 observations (Fig. 9), weak flow existed around the southeast foothills of the Olympics, and southerlies over the central Puget Sound became more southeasterly to the north.

At 800 m (Fig. 19b), the winds around the San Juan Islands were still strong (~30 m s⁻¹) but more southerly than at lower levels. In contrast, weak southerlies (~15 m s⁻¹) were present to the north of the Olympics over the strait. As observed (Fig. 11), flow splitting on the southeast side of the Olympics is apparent at 800 m. The model winds at this level and 300 m were not as southeasterly over the central and southern Puget Sound as observed by the NOAA P-3 (cf. Figs. 9 and 11). Meanwhile, south-southeasterly flow impinging on the western Washington coast rotated to southeasterly along the western slopes of the Olympics (as observed).

At 1300 m (Fig. 19c), the transition from southerlies to southeasterlies over southern Vancouver Island is absent. In agreement with the NOAA P-3 radar data at 1300 m (cf. Fig. 12), southerlies were present over the northwest slopes of the Olympics, southwesterlies were located offshore over the Pacific, and weaker southeasterlies were situated over the southwestern slopes of the Olympics. Between 800 and 1300 m (Figs. 19b and 19c), the area of lighter southerlies (10–20 m s⁻¹) over the central strait shifted northeastward toward the San Juan Islands. As in the observations (cf. Fig. 14), southeasterly flow existed around the Olympics at 1800 m except for weaker southerlies over of the Olympics (Fig. 19d).

c. Model precipitation

The model precipitation for 1800–2100 UTC 9 December and 2100–0000 UTC 9–10 December 1993

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Fig. 19. Model horizontal winds at (a) 300 m, (b) 800 m, (c) 1300 m, and (d) 1800 m above mean sea level for 2100 UTC 9 December 1993 (9 h).

(Figs. 20a and 20b) was compared quantitatively with the observations shown earlier (Figs. 15a and 15b, respectively). For the 1800–2100 UTC period, the model realistically simulated the rain shadows to the northeast of the Olympics and on the west side of the Cascades, as well as the maximum precipitation on the windward (south) slopes of the Olympics. The model also indicated precipitation enhancement extending approximately 30 km offshore of central Vancouver Island, consistent with the NOAA P-3 reflectivity data (Fig. 9). Along the western Washington and Vancouver Island coasts, and throughout Puget Sound the model did very well in simulating the 3-h rainfall totals (generally within 0.1 cm of the observations). Only on the southern side of the Olympics were the model amounts somewhat less than observed (by about 0.3 cm).

Between 2100 UTC 9 December and 0000 UTC 10 December (Fig. 20b), the model, as in the observations (Fig. 15b), maintained the Olympic and Cascade mountain rain shadows. As the flow veered to more southwesterly with the approaching warm front, the maximum precipitation amounts increased along the western side of the Olympics and Vancouver Island.
coast. The limited observations available suggest that precipitation increased in these areas. The offshore extent of the terrain-enhanced precipitation decreased along Vancouver Island because the terrain blocking was reduced as a result of the stronger cross-barrier (southwesterly) flow and decrease in low-level stability. On the southern foothills of the Olympics the model precipitation amounts were about 1.5 cm less than observed. The simulated rainfall was also too low over central and southern Puget Sound (between 0.2 and 0.5 cm below the observed).

6. Discussion

a. Pressure perturbations around the Olympics

Based on a microbarograph network and conventional observations, Mass and Ferber (1990) found that the asymmetry between the windward high and the lee low decreased as the Froude number increased. This agrees with linear theory, which suggests that pressure perturbations should be symmetric for large Froude numbers (Smith 1980). Using a representative wind and moist static stability upwind of the Olympics (at \(\sim 900 \text{ mb} \)) during COAST IOP 4 (\(U \sim 34 \text{ m s}^{-1}, N \sim 0.01 \text{ s}^{-1}, \) and \(H \sim 1800 \text{ m} \)), the internal Froude number, \(U/NH\), was about 1.9. Even for this relatively high Froude number, a well-defined pressure asymmetry existed around the Olympics in the simulation, with the lee low being 5–8 mb deeper than the windward high (Fig. 17). There are a variety of mechanisms that might explain this pressure asymmetry. For example, latent heating (condensation) associated with the precipitation enhancement over the windward slopes of the Olympics may hydrostatically reduce the strength of the windward pressure ridge, thus contributing to the observed pressure asymmetry. On the other hand, the asymmetry of the Olympic barrier with its relatively steep northern slope favors enhanced downslope flow, leeside warming, and troughing along the northern slope (Smith 1981). Miller and Durrell (1991) used a two-dimensional numerical model (Durrell and Klemp 1983b) with surface friction to show that the downslope flow increases as the lee slope steepens. In addition, the presence of an inversion near crest level (Durrell 1986) or negative vertical wind shear (Durrell and Klemp 1983) may have amplified the vertically propagating mountain waves.

To determine the impact of latent heating on the pressure asymmetry, a separate MM5 simulation (NOLH) was completed for the 3-km domain in which the diabatic effects associated with precipitation were turned off. Rainfall was allowed to prevent unrealistic moisture values from developing. Figure 21 shows the sea level pressure and surface winds around the Olympics for the NOLH simulation at 2100 UTC 9 December 1993. Without latent heating the pressure asymmetry remained essentially the same (the lee troughing was approximately 6 mb greater than the windward pressure ridge), which suggests that latent heating does not play an important role in the asymmetry observed around the Olympics.

To determine the role of the asymmetry of the Olympic barrier (steeper northern slope, cf. Fig. 16b) on the pressure asymmetry, the Olympics were replaced with an axisymmetric Gaussian mountain of similar height (1700 m) to the observed orography (Fig. 22). The pressure perturbations at 2100 UTC 9 December 1993.
Strait of Juan de Fuca inhibit the high momentum air from reaching the bottom of the slope.

**b. Flow splitting and deceleration over the windward side of the Olympics**

As seen from surface data (Fig. 7), Doppler wind data from the NOAA P-3 (Figs. 6 and 11), and model data (Fig. 17), flow splitting and deceleration occurred around the southern (windward) side of the Olympics. This low-level flow came close to stagnating over the southeastern slopes of the Olympics, with winds less than 2 m s⁻¹ below 300 m (Figs. 9, 17, and 19). The 900-mb Froude number of 1.9 seems too high for such light winds to develop. However, this Froude number may not be representative for the flow approaching the barrier closer to the surface given the strong positive wind shear from the surface to 900 mb (cf. Fig. 18). Near the surface (~975 mb), the Froude number was approximately 0.83 (where \( U \sim 15 \text{ m s}^{-1} \), \( N \sim 0.01 \), and \( H \sim 1800 \text{ m} \)), which is closer to the predicted threshold value of \( Fr \sim 0.76 \) necessary for stagnation to occur for weak forward shear (Smith and Gronás 1993).

The average Froude number for IOP 4 in this baroclinic environment was much higher than for those studies focusing on the flow around the island of Hawaii (Smolarkiewicz et al. 1988; Smith and Grubišić 1993), where Froude numbers are typically less than 0.5. Unlike Hawaii, there was no upwind flow reversal or lee-side vortices around the Olympics during IOP 4. To further prove that it was the relatively high—Froude number flow (\( Fr > 0.8 \)) that inhibited the development

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**Fig. 22.** Surface analysis at 2100 UTC 9 December 1993 for the Gaussian mountain simulation showing sea level pressure (solid) every 2 mb, terrain (dashed) every 100 m starting at 50 m, and wind vectors in meters per second.
of lee vortices and not the complex topography surrounding the Olympics (i.e., Vancouver Island), a separate 3-km simulation without Vancouver Island was completed (not shown). It showed no lee vortices, only weak low-level east-southeasterlies (5–10 m s$^{-1}$) in the wake to the north of the Olympics.

c. Momentum budgets around the Olympics

A variety of interesting low-level wind patterns were present around the Olympics during COAST IOP 4 (cf. Figs. 7 and 17): the strong low-level southeasterlies northeast of the Olympics, the transition from southeasterlies to easterlies near the western entrance to the Strait of Juan de Fuca, and the transition from southeasterlies to south-southeasterlies near the western Washington coast. Associated with these transitions, there were rapid changes in momentum balances. To study these balances, various terms of the momentum equation were diagnosed using output from the 3-km simulation. Friction, which represents both surface drag and momentum mixing from aloft, was estimated from the residual of the momentum equations; that is,

\[
F_x = \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} - fu + g \frac{\partial z}{\partial x},
\]

(1)

\[
F_y = \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + fv + g \frac{\partial z}{\partial y}.
\]

(2)

Since most of the calculations were not done over sloping terrain, the vertical advection terms were small and therefore neglected, and the local changes were approximated using centered time differences ($\Delta t = 30$ min).

Figure 23 shows directions and magnitudes for the terms in the momentum equation at the surface for the transition region between the strong southeasterlies (south of the San Juans Islands) to weaker easterlies (within the strait). To the northeast of the Olympics the pressure gradient is enhanced by the superposition of the synoptic-scale and the terrain-induced (meso-scale) pressure gradients. This large pressure gradient, combined with an over-water trajectory (with relatively weak drag), led to an acceleration of the southeasterly surface flow south of the San Juan Islands (point A). This acceleration continued until increasing frictional and Coriolis forces balanced the decreasing pressure gradient at point B in the area of maximum winds ($\sim 23$ m s$^{-1}$). The flow decelerated approaching Vancouver Island and the Strait of Juan de Fuca as the pressure gradients weakened further and became more perpendicular to the wind. Strong deceleration can also be seen near the north shore of the Olympic Peninsula (point D) as the strong pressure gradient force is pointed opposite the wind. Even though the flow in the Strait had a large component of the pressure gradient force orientated to the left of the wind, the flow was not in geostrophic balance (point C), since frictional
forcing (the residual) also had a large component in
the cross-strait direction (to the right of the wind). If
one considers only surface drag, the residual vector
should be aligned directly opposite the wind vector.
However, with strong mixing, as noted by Bell and
Bosart (1988), the orientation of the friction vector can
deviate substantially from what might be expected.
Around point C, the large northward component of
friction suggests that downward mixing of southerly
momentum from aloft was appreciable in the strait.8
The Richardson numbers were below the threshold
value for turbulence (0.25) at low levels in the strait
(not shown), and the potential for vertical mixing can
be seen by noting the large vertical wind shear from
the surface to 800 m (Figs. 17, 19). Similarly, Lack-
mann and Overland (1989) showed the importance of
downward entrainment of momentum into the PBL in
their study of the Shelikof Strait in Alaska.

The vertical mixing (northward frictional forcing)
was stronger in the western strait (point E on Fig. 24)
because the south-southeasterly flow above the low-
level easterlies was greater over the western strait than
over the eastern and central portions (cf. Fig. 19). The
combination of this northward frictional component
and the rotating pressure gradient force resulted in a
northward acceleration of those parcels from easterly
to east-northeasterly around point E. Meanwhile, near
the northwest tip of the Olympic Peninsula (point F)
air parcels experienced a more southwesterly compo-
nent to the pressure gradient force, which rotated the
winds to a more easterly direction near the entrance of
the strait.

Mass and Ferber (1990) documented the develop-
ment of pressure ridging along the coast of western
Washington associated with southwesterly flow at low
levels. The superposition of the topographically in-
duced windward pressure ridging and the synoptic-
cale scale gradients can lead to strong surface winds di-
rected approximately parallel the coast. During
COAST IOP 4 the pressure ridging on the west side of
the Olympics was relatively weak (1-mb amplitude)
because the low-level winds (south-southwesterly at
900 mb) did not have a large terrain-normal compo-
nent. There was a transition from southerly to south-
southeasterly flow 30 km offshore of the western coast
of the Olympic Peninsula (cf. Fig. 17). This wind tran-
sition can also be seen in the model fields within cross
section GG' taken across the Washington coast (Fig.
25). The effect of the weak pressure ridging on the
windward side of the Olympics can be seen within 30
km of the coast as the isentropes begin to bend upward
over the barrier. This offshore scale is also reflected in
the u-component of the wind; within 30 km of the coast
the easterly component of the wind becomes larger
since the flow tends to parallel the northwest–southeast
orientated terrain.

Figure 26a shows the 950-mb momentum balances
normal to the coast for a portion of cross section GG'.
The flow is nearly geostrophic over the Pacific, with
friction and Lagrangian accelerations increasing within
30 km of the coast. The pressure gradient force peaks
at 20–30 km west of the coast in response to the ter-
rain-induced pressure ridging (Fig. 26a). In the terrain-
parallel direction (Fig. 26b), there was an approximate
balance between the pressure gradient force, on one
hand, and Lagrangian acceleration, friction and the Cori-
olis force, on the other, over the offshore waters.

The horizontal width of the wind transition near the
western Washington coast fits the length scales deter-
mined by the scale analysis of Overland (1984) and
Overland and Bond (1993 and 1995). The calculated
Rossby radius, \( R_L = \frac{R}{f h_m} \) (where \( h_m \) is assumed to be the
height of the western 700-m peak in cross section GG',
\( N \sim 0.01 \text{ s}^{-1} \), and \( f \sim 1.07 \times 10^{-5} \text{ s}^{-1} \)), was 65 km,
which puts the calculated location of where the distur-
bance has decayed to 1/e of its maximum amplitude 25
km offshore and close to the observed position. Simi-
larly, the transition to more southeasterly low-level
flow south of Vancouver Island both in the model
(Fig. 17) and the observations (Fig. 9) occurs within an
approximate Rossby radius (~35 km) of the Vancouver
coastal mountains given \( N \sim 0.01 \text{ s}^{-1}, h_m \sim 400 \text{ m}, \) and
\( f \sim 1.12 \times 10^{-4} \text{ s}^{-1} \).

d. Trajectory analysis

Using 15-min model output (interpolated to 5 min),
forward trajectories starting at 1800 UTC 9 December
1993 were calculated for an array of points south of the
Olympics (Figs. 27a–c). At the lowest level, about
40 m above the surface (Fig. 27a), trajectory 1 shows
no deflection. Trajectories 2–8 show increasing west-
ward deflection west of the Olympics but with very
little vertical displacement (about the same arrow
width) near the barrier. Trajectories 1–8 begin to rise
from near the surface to 900 mb approximately 30 km
south of Vancouver Island, which verifies well with the
offshore precipitation enhancement seen both in the
model (Fig. 20) and the NOAA P-3 reflectivity (Figs.
9 and 11). Trajectory 9 moves toward the center of the
windward pressure ridge (Fig. 17) of the Olympics and
rapidly ascends the barrier, which matches well with
the area of precipitation enhancement seen both in the
observations (Fig. 15a) and the model (Fig. 20a). Traj-
ecory 9 then descends rapidly (from about 850 mb to
near the surface) over the Strait of Juan de Fuca but
soon rises again to 900 mb. Trajectories 10 and 11 de-
scent the western foothills of the Cascades (not com-
pletely shown) resulting in the rain shadow in that area
(Figs. 15a, 20a). As these trajectories approach the
Olympics they are deflected to the right by the wind-

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8 The large horizontal variations in radial velocities shown in the
NOAA P-3 radar cross section (Fig. 12) and the fluctuations in ver-
tical velocity found in the flight-level data (Fig. 8) are also indicative
of the turbulence associated with this mixing.
ward pressure ridging, and then turn northwestward around the northeast corner of the Olympics.

Trajectories launched approximately 50 mb (~400 m) above the surface experience a smaller horizontal deflection around the Olympics (Fig. 27b). Trajectories 16–19 still show some westward deflection around the pressure ridging. Trajectory 20 rises to 800 mb as it ascends the Olympics and then descends near the strait. Trajectories 21 and 22 pass over the northeast corner of the Olympic Peninsula as they are deflected around the barrier.

Trajectories launched approximately 120 mb (~1100 m) above the surface experience substantially less horizontal deflection (Fig. 27c). Trajectories 26–30 descend on the north side of the Olympics and therefore get caught up in the more southeasterly flow at low levels. Trajectories 29–32 get channelled northward at 850 mb over the San Juan Islands and the Strait of Georgia.

7. Summary

This paper documents the three-dimensional flow and precipitation around the Olympic Mountains for a case of relatively steady south-southwesterly flow during COAST IOP 4. This was achieved using both high-resolution data from the NOAA P-3 aircraft, conventional observations, and output from a nonhydrostatic mesoscale model simulation. The observations and
model simulation showed a hydraulic jump–like feature associated with a mountain wave over the northern slopes of the Olympics. Strong downslope flow over the northern slopes of the Olympics resulted in enhanced (by 4°C) low-level temperatures over the central strait. Flow splitting was evident in the simulation and observations over the southern slopes of the Olympics. Precipitation enhancement was also observed on the southern and western slopes of the Olympics, both in the radar reflectivity and the precipitation totals. To the northeast of the Olympics, the NOAA P-3 mapped the Olympic rain shadow as well as the strong south-easterly low-level flow (~25 m s⁻¹), the latter resulting from a superposition of the synoptic and orographically induced pressure gradients. In contrast, over the central Strait of Juan de Fuca the pressure gradient associated with the lee troughing partially cancelled the eastward-directed synoptic-scale pressure gradient, resulting in weak easterly flow. Near the western entrance of the strait, there was a sharp transition from easterly to southeasterly flow. A transition from southerly to southeasterly (terrain parallel) flow and precipitation enhancement occurred within 20 km of Vancouver Island.

Overall, the Penn State–NCAR mesoscale model (MM5) realistically simulated the three-dimensional flow around the Olympics. The model precipitation distribution was close to the observed, both in areas of topographic enhancement and the Olympic and Cascade rain shadows. The MM5 model simulation indicated a large pressure asymmetry around the Olympics with the lee troughing 5–7 mb deeper than the wind-
ward pressure ridging. It was shown that this asymmetry was not the result of the asymmetry of the Olympics or latent heating on the windward (southern) slopes of Olympics reducing the strength of the windward ridge. Based on previous theoretical studies, it is hypothesized that the mountain wave steepened as a result of higher stability near crest level and reverse shear above the crest, which resulted in strong downslope flow and the pressure asymmetry. Although strong downslope flow was observed over the upper slopes of the steeper peaks of the northern Olympics, high momentum air did not reach the surface at the bottom of the slope. This may be attributed to the three-dimensionality of the Olympics in which low-level flow can pass around the barrier, thus preventing the strong downslope flow from reaching near sea level.

The various terms in the momentum equation were calculated around the Olympics using model output. It was shown that geostrophic balance could not be attained in the Strait of Juan de Fuca partially because of the large component of vertical mixing in the cross-barrier direction. The southerly flow approaching the western entrance of the strait became east-southeasterly in the strait because of a more southward-orientated pressure gradient. Within 20–30 km of the coast, the weak ridging along the western Olympic Peninsula resulted in the pressure gradient normal to the barrier being primarily balanced by the Coriolis force, while pressure gradient and frictional forces approximately balanced in the terrain-parallel direction.

Overall, this study has shown the great potential of current mesoscale models for realistically simulating the airflow and precipitation around mesoscale orographic barriers such as the Olympics. This capability promises improved forecasting in the Pacific Northwest and other locations where the interaction of the synoptic-scale flow and terrain is important. In addition, because conventional observations are sparse in areas of high terrain, this paper also shows the value of using a dual-Doppler research aircraft for studying the flow and precipitation around orography.

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Fig. 27. Forward trajectories (1–33) starting at 1800 UTC 9 December originating (a) on the lowest half-sigma level (40 m above the surface), (b) 50 mb above the surface, and (c) 120 mb above the surface. The width of the trajectory is inversely proportional to pressure as shown in the inset scale. The length between lines inside a trajectory represents the distance the air parcel traveled in 1 h.
possible by the Microscale and Mesoscale Meteorological Division of the National Center for Atmospheric Research (NCAR). The mesoscale model was run at the Scientific Computing Division of NCAR. Editing and interpolation of NOAA P-3 radar data was performed using the RDSS and REORDER programs, respectively, developed by the Atmospheric Technology Division of NCAR. We appreciate Scott Braun’s assistance in showing us how to edit and synthesize the radar data.

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