Boundary Layer Turbulence and Orographic Precipitation Growth in Cold Clouds: Evidence from Profiling Airborne Radar Data

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ABSTRACT

Airborne vertically pointing Doppler radar data collected in 10 winter storms over the Medicine Bow Range in Wyoming are used to examine the importance of boundary layer (BL) turbulence for orographic precipitation growth. In all 10 cases, the cloud-base temperature was below 0°C and the bulk Froude number was more than 1.0, implying little or no blocking of the flow by the mountain barrier. Seven of the 10 storms sampled were postfrontal, with weak static stability and relatively shallow cloud tops. Doppler vertical velocity transects depict an approximately 1-km-deep turbulent layer draped over the terrain, sometimes clearly distinct from the stratified flow in the free troposphere aloft, where vertical motion is largely controlled by gravity wave dynamics. Spectral analysis of near-surface Doppler vertical velocity data in terrain-following coordinates reveals an inertial subrange with decreasing power with height toward the BL top. The composite of radar data profiles from the 10 flights is analyzed in frequency-by-altitude diagrams, with altitude expressed above ground level. These diagrams indicate a wide range of vertical velocities in the BL, and rapid snow growth within the BL as air rises through the cloud base, especially when BL turbulence is more intense. This snow growth is concentrated on the windward side of mountains, above the terrain–cloud base intersection. The dominant snow growth mechanism in the BL (i.e., by accretion or vapor deposition) cannot be established because of restrictions in aircraft flight level over complex terrain. Snow aggregation may have contributed to the observed rapid increase in reflectivity in the BL along the windward slope.

1. Introduction

Orographic clouds, in particular shallow orographic clouds, are remarkably efficient at generating precipitation. Processes other than the condensation of water by large-scale ascent over a mountain barrier must be at play to enable the rapid precipitation formation and fallout, in fact so rapid that most precipitation generally falls on the windward slope rather than the lee side of the crest (e.g., Smith 1979; Frei and Schär 1998; Roe 2005). The detailed cross-mountain distribution of precipitation depends on several factors; the main factors studied are the terrain height and width combined with the cross-mountain wind speed (e.g., Smith and Barstad 2004; Colle 2004), the static stability (e.g., Kirshbaum and Durran 2004), and cloud-active aerosol concentrations (e.g., Saleeby et al. 2009). A complete explanation requires both dynamical and cloud microphysical processes and, regarding the latter, both warm and cold cloud processes. The present study examines cases with surface snowfall and cold cloud bases (at most −5°C). Thus, only cold cloud processes are addressed herein. At sufficiently low temperatures even relatively shallow clouds can produce persistent snowfall over mountains (e.g., Steenburgh 2003), although such shallow clouds usually produce light snowfall only (e.g., Grant et al. 1974).

The main hydrometeor growth mechanisms in cold clouds are ice initiation, vapor deposition, and riming.
The main difference between collisional (riming) and diffusional (the Bergeron process) growth is that the former yields hydrometeors with a higher fall speed (i.e., terminal velocity), and thus a higher probably for the hydrometeors to reach the ground upwind of the crest, rather than downwind of the crest where some of the ice water mass sublimates on its way toward the ground (e.g., Hobbs et al. 1973), reducing the mountain-scale precipitation efficiency. Diffusional growth tends to dominate under weak updrafts, while riming tends to dominate under strong updrafts (e.g., Houze 1993, 197–199). The cloud condensation nuclei (CCN) concentration also matters. A high CCN concentration in the upstream flow leads to numerous but small cloud droplets, which suppresses riming because riming efficiency decreases sharply toward zero as the droplet diameter decreases from 20 to 5 \( \mu \text{m} \) (Pruppacher and Klett 1997; Wang and Ji 2000). Thus, high CCN concentrations tend to suppress precipitation on the windward side, as can be inferred observationally (Borys et al. 2003) and numerically (Lynn et al. 2007; Muhlbauer and Lohmann 2009; Saleeby et al. 2009).

Several dynamical processes affect orographic precipitation and its distribution across the mountain range. These include the release of potential instability by ascent over the mountain range (Banta 1990; Kirshbaum and Durran 2004), stratified or potentially unstable ascent over a stable air mass trapped by the mountain barrier (Rotunno and Houze 2007), orographic modification of advected convection (Colle et al. 2008), upstream tilt of the leading mountain wave (Colle 2004), gravity waves due to secondary terrain features perpendicular to the main crest (Garvert et al. 2007), lee-side stratification (Zängl, 2005), and turbulence generated by vertical wind shear (Houze and Medina 2005).

One dynamical process that has received little or no attention in the study of orographic precipitation is mechanical or thermal turbulence in the planetary boundary layer [referred to simply as the boundary layer (BL)] over the mountain, at elevations above cloud base. Turbulent eddies can produce transient yet strong updrafts, and since most water vapor resides in the BL, the potential importance of BL turbulence for orographic precipitation growth seems intuitive. The reason why this process has received little attention probably is that the BL over complex terrain is difficult to document, be it with instrumented aircraft (for safety reasons), ground-based scanning radars (because of beam blockage and lack of vertical resolution), or ground-based profiling radars (because of the close-range radar blind zone or lack of temporal resolution). The role of BL turbulence, especially shear-driven turbulence, in orographic precipitation growth is also a challenging numerical problem because of the requirement to resolve a broad spectrum of scales ranging from hundreds of kilometers to tens of meters.

The objective of this paper is to demonstrate that shear-driven or convectively driven turbulence in the BL can be an important precipitation growth mechanism over mountains whose elevation is above the upstream cloud base. The main source of evidence is an airborne Doppler cloud radar. The cloud radar affords a resolution high enough to capture a significant fraction of the spectrum of turbulent eddies, and its nadir view provides data within about 30 m of the terrain.

The data source and processing are described in section 2. The atmospheric environment of the 10 cases is described in section 3. Radar transects are illustrated in section 4. These are analyzed in terms of velocity power spectra in section 5, and in a composite sense in section 6. Flight-level data are examined in section 7. Section 8 then assembles the evidence for the significance of BL turbulence in snow growth.

2. Experimental design and data sources

A series of straight and level legs was flown repeatedly in winter storms over the Medicine Bow Mountains of Wyoming on a total of 10 flight days in early 2006, 2008, and 2009. This mountain range is about 30 km across and 1500 m high above the surrounding plains (Fig. 1). Only along-wind legs were flown on the three flights in 2006 (Table 1). During the seven flights in 2008 and 2009, most flight legs followed geographically fixed tracks, oriented roughly normal to the prevailing wind (Fig. 1). These slightly shorter fixed tracks were designed to maximize data collection near and upwind of the mountain crest.

The aircraft, the Wyoming King Air (WKA), carried in situ cloud microphysics probes and a sensitive 94-GHz (3 mm) Doppler radar with zenith- and nadir-pointing antennas, the Wyoming Cloud Radar (WCR; see http://www.atmos.uwyo.edu/wcr/). The synergy between the radar profiles above and below the aircraft, and the in situ data sandwiched in between, is quite powerful. The WCR data are especially useful (a) close to flight level (~120 m above and below flight level), since this allows the study of various relationships between flight-level and nearly coincident radar data [e.g., particle fall speed estimates and reflectivity–snow rate \((Z–S)\) relationships], and (b) in very close proximity to the terrain. At 94 GHz (W band) the radar echo in a mixed-phase cloud is dominated by ice crystals scattering in the Mie regime. This scattering process is complex and highly dependent on crystal shape, orientation, and size distribution. Matrosov (2007) computed \(Z–S\) relationships for this radar frequency based on Mie scattering theory,
assuming different ice size distributions, shapes, and densities. His relationship $S = 0.11Z^{1.25}$ ($S$ in mm h$^{-1}$, $Z$ in mm$^3$ m$^{-3}$) appears to apply for the storms analyzed herein (Geerts et al. 2010).

The WCR zenith- and nadir-pointing antenna Doppler velocities are processed to remove the motion and 3D attitude variations of the aircraft. When an antenna is tilted slightly off-vertical, the observed Doppler velocity contains an uncertain horizontal wind contribution. This contribution is removed by means of the wind profile measured upwind of the mountain (Fig. 1). The resulting field, referred to as WCR vertical velocity, approximates the hydrometeor vertical motion with an accuracy of $\pm 1$ m s$^{-1}$ (Damiani and Haimov 2006). We do not remove an estimated particle fall speed (terminal velocity) to obtain the air vertical motion because of fall speed uncertainty. In addition to the nadir and zenith antennas, the WCR operated a slant-nadir antenna 30° forward of nadir. The nadir and slant-nadir antenna Doppler data are synthesized to obtain 2D (horizontal along-track and vertical) velocity data below flight level (Leon et al. 2006; Damiani and Haimov 2006).

3. Description of the cases

a. Synoptic situation

The synoptic situation during each of the 10 flights is summarized in Fig. 2, using 3-hourly North American Regional Reanalysis (NARR) data, whose horizontal resolution is 32 km. On 7 days (Figs. 2c–f, h–j) an upper-level trough was overhead or had passed through already. This typically implies deep-tropospheric subsidence, a deep cold air mass, reduced low-level stability, and shallow orographic clouds.

On the first day (Fig. 2a) a strong zonal upper-level jet stream was present, with significant subsidence warming in the lee of the Rockies. A deep trough in a rather weak 300-hPa jet was present over Utah on the second day (Fig. 2b). In both cases the cold front from the Pacific remained well to the west of the Medicine Bow Range during the flight, as evident in the low-level wind pattern. An upper-level trough and cold front approached from the north on 20 February 2009 (Fig. 2g). The first part of the flight on this day was prefrontal. Surface station and radiosonde (Fig. 3g) data indicate that the cold air mass had just reached Saratoga at 2325 UTC, 2 h into the WKA flight. The other seven flights occurred in postfrontal conditions with little baroclinicity in the area of interest. One day was exceptionally cold across Wyoming (Fig. 2h).

The low-level flow was highly ageostrophic in the area of interest. Strongly downgradient westerly flow occurred on 9 out of 10 days. [The one exception, the second day (Fig. 2b), features a very weak 800-hPa height gradient in the area of interest, but the flow still has a westerly component.] Strong downgradient flow is typical across south central Wyoming in winter, on account of the terrain (Martner and Marwitz 1982). This flow can be quite vigorous under steep west-to-east height falls (e.g., Fig. 2d).

b. Vertical structure

On each of the 10 days, sounding data were collected just upstream of the Medicine Bow Mountains to document the low-level stability and wind profiles (Table 1). No radiosondes were available in 2006, so we resorted to aircraft soundings for the first three flights. The drawback of aircraft soundings is an increased probability of incorporation of horizontal variability (e.g., due to gravity waves) into the profiles, as apparent on 28 January 2006 near the lifting condensation level (LCL) (Fig. 3b).

The LCL was determined from near-surface data in the Saratoga radiosonde or aircraft sounding. This estimate is important because it determines the “LCL–terrain intersection” point, one of the two boundaries we will use in the composite analysis (section 6). The LCL is a good estimate for the height of the base of the cloud.
### Table 1. Summary of the 10 flights used in this study. The first two rows indicate the number of flight legs (illustrated in Fig. 1) used. Rows 3–5 represent averages for all these flight legs. Row 3 is the average $-20$-dBZ echo top height. The mean fall speed of hydrometeors (row 4) is based on a comparison between the air vertical velocity measured by the gust probe and the mean WCR particle vertical motion measured at the nearest radar gate above and below the aircraft, at a range of $\sim 120$ m. The BL depth (row 5) is subjectively estimated from transects (such as Fig. 4b) and corresponding velocity power spectra as function of height AGL (such as Fig. 9). The data in the remaining rows are based on soundings. The data for the first 3 days were derived from multiple aircraft soundings just upwind of the mountain (sounding type A). Radiosondes released at Saratoga Airport in the middle of the $\sim 4$-h flights were used for the last 7 days (sounding type R). The numbers shown in rows 7–12 represent averages between the near-surface (practically, 100–200 m above the Saratoga airport; i.e., the lowest flight level for the A soundings) and the elevation of Medicine Bow Peak (3661 m). The wind direction offset is positive if the wind direction is clockwise relative to the flight direction. $Fr$ is calculated as the wind speed divided by $N$ and by the height of the mountain peak above the lowest data level. $Ri$ is the ratio of $N^2$ over the square of the shear vector between the lowest data level and the mountain top level. The LCL (a cloud base estimate) is derived from the near-surface sounding data. If more than one sounding was available (last row), the data are linear averages for all soundings. The 5 days with the lowest static stability are in boldface.

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FIG. 2. Maps of NARR 800-hPa height (thick black contours, 30-m interval), temperature (color, key on the right), and winds (barbs, full barb equals 5 m s⁻¹) plus 300-hPa height (thick white contours, 60-m interval) for the 10 flight days, at a time close to the middle of the flight. The date/time format is yymmdd hhZ. State boundaries and latitude and longitude lines (dashed lines, 5° interval) are shown as well. The purple square in southern Wyoming outlines the map in Fig. 1. A cold front is analyzed in (g).
FIG. 3. Stability and wind profiles near Saratoga, upstream of the Medicine Bow Range (Fig. 1) during the 10 flights. (a)–(c) Aircraft soundings, starting at about 100 m above the Saratoga airport. (d)–(j) From radiosondes released near the middle of each flight period. In each panel, the profiles of (equivalent) potential temperature ($\theta$ and $\theta_e$, solid and dashed lines) are shown on the left and the wind profile on the right (full barb equals 5 m s$^{-1}$). An isothermal lapse rate is added for reference.
hugging the mountain since all days were rather windy and lacked a decoupled stable layer in the valley. This LCL was usually slightly higher than the lowest level with cloud liquid water for days with aircraft sounding data (the first three flights), and slightly higher than the lowest ceilometer cloud base, for the 2008 and 2009 flights, during which ceilometer data were collected just upwind of the mountain (Fig. 1). The cloud-base temperature was below \(-5.0^\circ\text{C}\) during all flights (Table 1).

The average wind speed below mountain top level varied between 11 and 21 m s\(^{-1}\) (Table 1). The mean Brunt–Väisälä (BV) frequency\(^1\) below mountain top level was below \(10^{-2}\) s\(^{-1}\) during all but one flight (i.e., the low-level upstream stability was rather weak). The first day was the most stable (Fig. 3a): the lowest 1 km was nearly isothermal in the first sounding, but the stability decreased later during the flight. On 3 days (Figs. 3c,e,i), the lower troposphere was potentially unstable (i.e., \(\theta_e\) decreases with height between the surface and mountain top level). On these days, the BV frequency was close to zero (Table 1) and the clouds over the Medicine Bow Range were rather shallow, with cumuliform tops. Static stability was low also on two other days, with minor potential instability (Figs. 3d,h). The five low-stability days are highlighted in boldface in Table 1.

The mean wind shear below mountain top varied between 3 and 9 meters per second per vertical kilometer, and this shear generally was distributed well (i.e., not concentrated in a thin layer). A low-level jet below mountain top level was present on 7 out of 10 days (e.g., Figs. 3c,f,i). If this jet were a barrier jet (Parish 1982) or channeled flow (Wippermann 1984), the flow would be mainly southerly, aligned with the North Platte valley (Fig. 1). Instead, this jet was generally westerly. A westerly low-level jet is not uncommon in winter, especially about 50 km north of Saratoga, in the middle of a large gap in the Rocky Mountains just north of the Medicine Bow Range. This jet is due to the often large zonal pressure gradient across southern Wyoming, pushing air through the gap, while flow to the south (in western Colorado) and the north (in northwestern Wyoming) is often blocked (Marwitz and Dawson 1984).

The upstream below-mountain Froude number Fr, calculated from the bulk BV frequency and mean wind speed, exceeded 1.0 during all flights (Table 1). This implies that the upstream air mass was not blocked but was rather readily advected over the mountain. During the five flights highlighted in boldface in Table 1, Fr was particularly low, Fr high (Fr \(\geq 1.9\)), and the vertical shear of the wind rather strong, such that the bulk Richardson number Ri (also calculated between the surface and mountain top level) was less than one (Ri < 1). Ri was even lower below the low-level jet, for 3 out of the 7 days with a low-level jet. A low local Ri value implies little resistance to vertical exchange by turbulence, and \(\text{Ri} < 0.25\) is a necessary condition for local shear instability.

None of the profiles indicates a frontal boundary aloft, except the 20 February 2009 profile (Fig. 3g). The sharp backing of wind with height and the high static stability in the lowest 0.5 km are consistent with a shallow cold front. The early arrival of the front took us by surprise. Even the NARR data 30 min after the sounding suggest the cold front to still be north of the area of interest (Fig. 2g). The frontal snowband was rather deep, and clouds became shallower following cold-frontal passage (Geerts et al. 2010).

WCR data indicate that snow continuously fell over the mountain during each flight. The average WCR echo tops ranged from 4.1 to 6.6 km MSL (Table 1). The average echo top height was only 5.3 km MSL for the seven postfrontal days. Given a terrain height of 2–3 km MSL, this implies relatively shallow storm systems. The dominance of postfrontal shallow orographic cloud situations was, in effect, by design. The 2008 and 2009 flights were conducted under northwesterly low-level flow (Fig. 1). Such flow was desired because the 2006 flights had shown that under southwesterly to westerly flow, gravity waves were common aloft, presumably triggered by an upwind range (the Sierra Madre, whose foothills can be seen in the lower-left corner of Fig. 1). Under west to northwesterly flow, there are no significant terrain obstructions upwind of the Medicine Bow Range over at least 300 km. Northwesterly flow conditions tend to be postfrontal. The bias toward shallower, less stable postfrontal orographic clouds will be revisited later, as it is likely to affect the dominant hydrometeor growth process (e.g., Cooper and Saunders 1980).

### 4. Example radar transects

#### a. Separating BL processes from deeper tropospheric processes

The WCR transects reveal orographic precipitation structures in unprecedented detail, such as in a transect aligned with the wind on 18 January 2006 (Fig. 4). Some convective cells are embedded aloft (5–7 km MSL) over the mountain, growing on the upwind side, with small updrafts up to 4 m s\(^{-1}\) (Fig. 4b), and decaying in the lee.
This elevated convection is consistent with a potentially unstable layer around 4 km MSL (Fig. 3a), lifted isentropically from the windward valley over the mountain: widespread stable ascent breaks up into several small-scale updrafts aloft right over the mountain. Because this convection is elevated and the wind strong (Fig. 3a), most snow falls in the lee of the crest in this transect. Deep subsidence in the lee (Fig. 4b) undoubtedly evaporates the cloud droplets but snow persists. The thinning wedge of snow near the right end of Fig. 4a is an indication of strong downslope winds.

Finescale turbulent vertical motion is apparent in a roughly 1.0-km-deep layer draped over the terrain, especially on the upwind side (Fig. 4d). Some near-surface eddies can be attributed to airflow over local terrain features, but most eddies appear random in location and strength. More detailed closeups (not shown) reveal eddies at the scale of individual WCR gates, about $30 \times 30$ m$^2$. The updrafts in many eddies exceed peak values of 2 m s$^{-1}$ [i.e., stronger than the mean mountain-scale ascent rate$^2$ (0.5–1.0 m s$^{-1}$)].

Note that the observed turbulent vertical motion cannot be attributed to Doppler velocity measurement uncertainty or processing errors. The signal is strong and the eddies are coherent, so the Doppler velocities are not random noise. As will be shown later, the power spectrum follows the $-5/3$ power law down to the smallest scale resolved by the radar. An error in the corrections for aircraft motions would be continuous along vertical radar beams (i.e., it would be evident as vertical stacks of alternating errors).

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$^2$ The mountain-scale ascent rate scales as $\frac{U}{H}$, where $U$ is the mean cross-mountain wind speed ($10–20$ m s$^{-1}$, Table 1) and $H$ the height change (1 km) over a distance $L$, the mountain half-width (20 km).
The layer of turbulent motions is remarkably well defined during the 18 January 2006 flight. We refer to this layer as the BL. (The BL top in Fig. 4d is drawn subjectively.) The turbulence is shear driven, as the static stability is high and the wind strong (Fig. 3a). Most flight legs used in this study (except two, including the one shown in Fig. 4) were flown at about 660 m above Medicine Bow Peak (i.e., the lowest level allowed under instrument flight rule). Thus, in situ data within the BL are limited. Therefore our description of the flow and microphysical processes in the BL mostly relies on WCR profile data below the aircraft. On the upwind side within the BL, echoes appear to strengthen in a Lagrangian (downwind) direction, suggesting snow growth, mainly in the range $2 < x < 9$ km (Fig. 4d). This growth, which starts above the LCL (Table 1), seems to be separate from the growth aloft that we earlier interpreted as the result of smooth isentropic ascent and subsequent potential instability release. In this case the snow growth within the BL appears insignificant compared to the growth aloft, but there is much variability between transects.

b. Surface-induced ice initiation?

A transect identical to Fig. 4 was flown 10–15 min earlier, but at the 14 000-ft (4267-m) flight level (Fig. 5). Convective cells aloft are weaker or absent, maybe because there is less ascent, but snow growth within the BL is similar to that in Fig. 4. In situ data are shown as well because some of the BL turbulence reached flight level. Particle concentrations measured by optical array probes

![Fig. 5](image-url)
(Fig. 5d) correlate well with the near-flight-level radar reflectivity. The gust probe vertical velocity (Fig. 5e) correlates well with the near-flight-level Doppler velocity and confirms the transition of kinetic energy from low to higher frequencies as the terrain rises, suggesting that the WKA samples some BL air over the mountain. In this case snow does not reach the ground at the upwind margin of this transect. The first BL echoes (near 13 < x < 17 km in Fig. 5a) are elevated. This suggests that these echoes are not simply due to lofting of freshly fallen snow. Blowing snow is quite common over this mountain range, but its concentration normally drops off rapidly with height, becoming insignificant above about 50 m AGL (Schmidt 1982).

The cloud base (LCL) is below the level of the first BL echoes (Fig. 5a). Blowing snow particles, lofted even at small concentrations by turbulence, could be the first ice crystals in a supercooled water cloud, starting snow growth at temperatures well above typical ice nucleus activation temperatures. Ice could also be initiated when droplets impact rimed vegetation: these droplets, upon freezing, may splinter into numerous ice crystals. This is the Hallett–Mossop mechanism (Hallett and Mossop 1974), but in this case droplets impact stationary rimed surfaces rather than rimed crystals falling through a water cloud. Thus, the impact velocity can be rather high. Secondary ice crystal generation by riming is most effective under restrictive: secondary ice crystal generation by riming is most effective under rapid with height, becoming insignificant above about 50 m AGL (Schmidt 1982).

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The Glacier Lake Ecosystem Experiment Site (GLEES) surface station located downwind of the crest (Fig. 1) recorded a 10-m wind speed of 7.3 m s$^{-1}$ at the time of the transect. Such wind speeds are higher than typical impact velocities of falling rimed particles, but even at such speeds ice splinter formation may occur (Mossop 1985), and weaker winds are likely to occur within the vegetation canopy and at the canopy edges during lulls between gusts. Second, particle temperature is more restrictive: secondary ice crystal generation by riming rapidly ceases at temperatures below $-8^\circ$C (e.g., Griggs and Choularton 1986). This eliminates 6 of the 10 days (i.e., those with LCL temperatures at or below $-8^\circ$C) and virtually eliminates three other days, those with LCL temperatures at $-7^\circ$C (Table 1). And third, on the remaining day, 18 January 2006, the flight-level FSSP data indicate a mean droplet diameter large enough for ice splinter formation (Fig. 5e).

In other, warmer climes, ice splinter formation on rimed surfaces such as tree needles may well be an important ice initiation mechanism. Both this mechanism and blowing snow may explain why clouds sampled on Elk Mountain (Fig. 1) have been found to contain 10–1000 more ice crystals than clouds in the free atmosphere upstream (Rogers and Vali 1987). Both surface-induced ice initiation mechanisms are speculative (Vali et al. 2008) and should be investigated further. The WCR data only show that on 18 January 2006 snow growth occurred within the upwind BL, above the LCL, without ice crystal seeding from aloft.

If the surface does induce small ice crystals, they should readily be lofted into the BL by the turbulent eddies. The dual-Doppler synthesized hydrometeor streamlines$^4$ for the transect in Fig. 4 indicate that snow falls down through the upwind BL in this case (Fig. 6a). But small particles with an insignificant fall speed can be carried up in the upwind BL: if we boost the dual-Doppler vertical velocity by 1.0 m s$^{-1}$ (an estimate of the typical hydrometeor fall speed in this case, Table 1), some streamlines rise from near the surface on the upwind side and cross the terrain crest (Fig. 6b). This rise is facilitated by the relatively weak cross-mountain wind in the upwind BL (Fig. 6b).

No fall speed assumption is needed for the snow streamlines (Fig. 6a). In this case hydrometeors entering the transect at the upwind edge at a height of at least 1.9 km AGL end up on the lee side. The introduction mentions a number of factors affecting hydrometeor transport across the mountain crest. Dual-Doppler analyses for select transects on each of the 10 flights (not shown) show that wind speed is the primary factor explaining the particle slope and where particles reach the ground relative to the crest. The steepening of the streamlines in the upwind BL in Fig. 6a may reflect an increased particle fall speed due to aggregation and riming in the BL. Yet the first-order explanation simply is the deceleration of the flow (Fig. 6b).

c. BL turbulence, stability, and wind speed

To further describe precipitation growth in the BL over the upwind terrain slope, we examine two transects collected under distinctly different stability and wind conditions. On 2 February 2006, the BV frequency was quite low (Table 1), and the cloud tops were somewhat cumuliform and shallow. Clouds were generally confined

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3 The Medicine Bow Range is mostly covered by needle trees between elevations of about 2300 (the sage brush boundary) and about 3200 m ASL (the tree line).

4 “Streamlines” are tangential to the instantaneous 2D vector field. They would also be trajectories if the flow was steady.
to the mountain. Turbulence extended to the cloud tops (Fig. 7b) and appeared to change scales from small, shear-driven eddies near the surface to more coherent, thermally driven vertical motions aloft. Unlike the 18 January 2006 case (Figs. 4 and 5), there were no clouds above the turbulent layer. The towers reaching flight level contained much liquid water (Fig. 7c), and snow was generated rapidly in all transects (Figs. 7a,d). It is not clear what initiated ice crystals on this rather cold day (Table 1).

BL turbulence was evident in all WCR transects on all flights. Generally the top of the BL turbulence was not as well defined as on 18 January 2006 (Fig. 4). The most shallow, least defined BL was encountered on 28 January 2006 (Fig. 8a), the least windy day (Table 1), on a flight leg over the Sierra Madre, an adjacent mountain range to the southwest (Fig. 1). Surface-induced ice initiation in the upwind BL theoretically is possible in this case but is unlikely to be important for snowfall production, for two reasons. First, there is a continuous layer of significant reflectivity (i.e., ice crystals) from the BL to the echo top, whose temperature is about $-30^\circ$C. Second, WCR dual-Doppler data indicate that hydrometeors gently fall down through this deep layer, notwithstanding widespread rising air currents along the entire upwind section (Figs. 8b,e). This broad updraft likely is part of a mountain-scale upwind-tilting gravity wave, as the stability and wind speed support such waves in this case. Vertically propagating upwind-tilting waves occur when the BV frequency $N$ exceeds the harmonic
frequency of the mountain $2\pi U/L$, where $U$ is the mean wind speed and $L$ the mountain wavelength (e.g., Lin 2007). In this case ($U \approx 11$ m s$^{-1}$, $L \approx 50$ km), the ratio $(NL/2\pi U)^2 = 7 \gg 1$. The lack of convectively or shear-generated rising eddies above the BL results in a smooth reflectivity field and very little liquid water at flight level, even close to the crest (cf. Figs. 8c and 7c). The droplets appear to be consumed effectively through depositional growth in the broad updraft region. Some water is present in a region of slightly stronger orographic ascent (near $x = 20$ km; Fig. 8c), but the droplets are very small (Fig. 8e).

5. Power spectra

Boundary layer turbulence essentially is due to continuous forcing of the mean flow toward a state in which shear or convective instabilities grow. These instabilities feed energy into eddies that scale with the depth of the unstable layer, and those eddies in turn transfer energy into smaller scales. A range of eddy scales (larger than the scale of viscous dissipation) normally exists at which buoyancy or shear of the mean flow is insignificant to the eddy statistics compared to the effects of other turbulent eddies. This range (the inertial subrange) is characterized by a spectrum of the turbulent kinetic energy (TKE) per unit wavenumber $k$ proportional to $k^{-5/3}$. WCR vertical velocity data in the BL indicate that this power law applies well for eddies with dimensions between about 30 m (the smallest resolvable scale) and several hundred meters (Fig. 9). At larger scales the variations in the power spectrum are largely controlled by local terrain variations (i.e., the near-surface vertical velocity is directly forced by details in the upwind terrain). For instance, the terrain spectrum features a spike near 0.35 Hz ($\sim 300$ m) in Fig. 9a, and that is evident in the near-surface WCR vertical velocity traces. At yet larger scales, mountain-scale gravity waves become apparent in some spectra [e.g., the spike at 0.007 Hz]
(~13 km) in the vertical velocity spectra at higher elevations in Fig. 9d].

The inertial subrange is evident in all Doppler vertical velocity transects within the BL on all 10 days. Generally, the TKE is highest near the ground and decays with height in the BL, consistent with observations above the surface layer in a stratified BL over flat terrain (e.g., Fig. 3.25b in Garratt 1992). The power spectra on 18 January 2006 confirm a rapid decrease of TKE with height near 1.0 km AGL, consistent with the transects (Figs. 4d and 5b). The BL depth, listed in Table 1, is estimated based on both the level of rapid TKE decrease in the inertial subrange and visual inspection of the vertical velocity transects (e.g., Fig. 4d). The definition of the BL top is exaggerated in Fig. 9b because there are a number of “bad” levels near the BL top (i.e., levels intersected by the radar blind zone). But data from a higher flight level (Fig. 9a) confirm the clear definition, with a rapid TKE dropoff between 1.0 and 1.1 km AGL for eddies around 100 m in wavelength. The power spectra also confirm a relatively shallow BL for the weak-wind transect in Fig. 8b (Fig. 9d) and high values of TKE extending to the echo tops on the day with shallow orographic convection in Fig. 7b (Fig. 9c). Thus on days without convection, a mean BL depth over the mountain can be inferred rather well from the WCR vertical velocity power spectra.

6. Frequency-by-altitude displays

a. Method and purpose

Several transects of high-resolution WCR data have been used to illustrate orographic precipitation growth and the impact of BL turbulence (section 4). These transects are snapshots, dominated by rather transient features. Thus, to examine profiles of hydrometeor growth
and decay across a mountain more generally, WCR data for a total of 53 along-wind and 132 across-wind flight legs (Fig. 1) from 10 flights (Table 1) are synthesized in the form of a frequency-by-altitude display (FAD) (Yuter and Houze 1995). These displays are derived as follows, both for radar reflectivity and vertical velocity (Fig. 10). All WCR profiles collected along straight and level flight sections were grouped into one of three classes, depending on location. We distinguish three regions, defined by two boundaries, the LCL–terrain intersection and the crest. The former is the terrain contour at the height of the LCL (Table 1). Profiles upwind of the crest where the terrain is below (above) the LCL are labeled “upwind below (above) LCL” and all other profiles are labeled “downwind of crest.” Some of this classification was done manually because the crest line actually depends on wind direction (Fig. 1). Profiles were discarded in the vicinity of an ill-defined crest line. The WCR profiles were remapped as a function of height above ground level, and occurrences of reflectivity (vertical velocity) values in any of the three classes were then counted in bins with dimensions of 0.5 dB (0.1 m s\(^{-1}\)) and 30 m, for all radar range gates and all profiles, the total number of which is listed at the top of Fig. 10, for the three classes. The vertical resolution (30 m) equals the WCR range resolution. The value shown in any of the six panels of Fig. 10 is this count in each bin, normalized by the total count of all occurrences in all bins in both dimensions. Thus, the summation of all bin values in any panel equals 1.0. In the absence of the mountain, the top panels (and the bottom panels) of Fig. 10 should have identical distributions, at least for a
sufficiently large sample size. Thus, the difference between the normalized distributions can be interpreted as an orographic effect. There are significant differences between the 10 days in terms of upwind water vapor content, LCL, echo top height (all listed in Table 1), and other factors that affect precipitation intensity. Yet the proportional amount of flight time in each of the three regions was rather constant for each of the 10 flights, namely 26%, 37%, and 37% on average for the three regions from the upwind side to the lee, respectively, with a variation of at most \( \pm 9\% \) for individual flights, mostly due to variations in the LCL. Therefore it is meaningful to examine differences between the three regions, defined in terms of the LCL–terrain intersection and the mountain crest, in order to isolate orographic precipitation growth.

The WCR does not measure vertical air motion but rather the vertical motion of echoes. The fall speed of hydrometeors can be estimated by comparing the WCR vertical velocity at the close-range gates above and below the aircraft to the vertical air motion measured by the gust probe on the aircraft. Average values for each flight are included in Table 1. The FAD of WCR vertical motion is analyzed (Fig. 10). The 10-day mean fall speed of 0.95 m s\(^{-1}\) (shown as a dashed vertical line in Fig. 10) gives a first-order separation between updrafts and downdraughts: air generally rises (sinks) for bins to the right (left) of this line. Clearly this is a generalization: the actual fall speed probably varies significantly around this average (e.g., Mitchell and Heymsfield 2005).

b. Reflectivity and vertical velocity changes across the mountain range

Measurable snowfall occurred near the ground for almost all upwind above LCL profiles,\(^5\) and also for almost all downwind of crest profiles, because the flight tracks did not extend far downwind of the crest (Fig. 1). Near-surface echoes were often absent upwind of the LCL–terrain intersection (Fig. 10a). The relative dearth of low-level echoes and the decrease of mean reflectivity from 1.5 km AGL toward the ground in the upwind-below-LCL group suggest low-level sublimation of snow in the Saratoga valley.

The most profound change along the upwind slope is the sudden abundance of strong low-level echoes above the LCL (Fig. 10b). The precipitation growth mostly is

\(\text{FIG. 10. Normalized FAD of (a),(c),(e), WCR reflectivity and (b),(d),(f) WCR vertical velocity for all flight legs on 10 flights, shown (a),(b) upwind of the LCL–terrain intersection, (c),(d) between this intersection and the crest, and (e),(f) on the lee side. Also shown in all panels are the mean profile (white line), the “data presence” (red line; i.e., the percentage of WCR range gates with radar echo as a function of height, with values ranging from 0% to 100%), and the mean hydrometeor fall speed [black dashed line, in (b),(d), and (f) only].}\)
confined to the lowest 1.0 km above the terrain (i.e., roughly the BL). The difference in normalized FADs for profiles above and below the LCL–terrain intersection highlights how this growth clearly is confined to low levels (blue area in Fig. 11a). Remarkably, this change cannot be explained by enhanced low-level rising motion over the ascending terrain. The mean low-level radar vertical velocity is the same over terrain above and below the LCL–terrain intersection (Fig. 12b). There is only a slight increase in vertical velocity variance below 1 km across the LCL–terrain intersection, as can be seen in Fig. 12b, and as is suggested by the slightly overlapping bell-shaped low-level blue “mountains” in Fig. 11b. The variance is only partly due to turbulence, the other cause being more coherent vertical motions associated with the terrain or convection. Good WCR velocity spectra in the region below the LCL–terrain intersection rarely can be derived because echoes are absent or discontinuous, but in all likelihood BL turbulence over lower terrain was not very different from that over higher terrain. In essence, rapid hydrometeor growth occurs when the turbulent BL is lifted above cloud base. The WCR data do not reveal the growth mechanism, nor do they prove that BL turbulence affects this growth. The growth mechanism will be explored in section 7. Evidence that turbulence matters is given in section 6c.

The horizontally stretched dipole of extreme values between 1 and 2 km AGL in the difference FADs (mainly in Figs. 11a, b, and to a lesser extent in Figs. 11c, d) is an artifact of the radar blind zone. Specifically, the different heights (AGL) of this blind zone in the three different regions (Fig. 10) yield artificial vertical gradients in the difference FADs. The data in the 1–2-km AGL height zone is still useful with the knowledge of this artifact, which does not affect horizontal variations in Fig. 11.

While the low-level reflectivity substantially increases as the flow approaches the mountain crest, it tends to decrease at higher levels (Fig. 12a). The strongest echoes aloft become less common as the terrain rises above the LCL (Fig. 11a). This may be due to the prevailing

![Figure 11](https://example.com/image11.png)

**Fig. 11.** Difference in normalized FAD of (a), (c) WCR reflectivity Z and (b), (d) WCR vertical velocity. (a), (b) The difference between [above LCL] and [below LCL] [i.e., (a) in this figure equals Fig. 10c minus Fig. 10a]. (c), (d) The difference between [downwind of crest] and [upwind, above LCL]. The precipitation rate shown in the upper abscissa of (c) and (d) is inferred from $S = 0.11Z^{1.25}$ (Matrosov 2007).
presence of an upwind-tilting gravity wave, with sinking motion just upwind of the crest (Figs. 12b and 11b). The highest echo top is found 6 km upwind of the crest, on average. Vertically propagating upwind-tilting waves are expected on the five most stable days, as $N = 0.77 \times 10^{-2}$ s$^{-1}$ on average, much larger than the mean terrain harmonic frequency ($0.18 \times 10^{-2}$ s$^{-1}$) (see section 4c). A wave tilt yielding upward motion over the lower upwind slope affects the distribution of surface snowfall, depending on wind speed and other factors (e.g., Colle 2004). No streamlines emerging from above the BL on the transect’s upwind edge reach the ground upwind of the LCL–terrain intersection in any of the hydrometeor streamline analyses for select along-wind flight legs on the 10 flights. (Only two are shown, in Figs. 6a and 8b.) Yet most if not all streamlines emerging at the upwind edge above the BL did reach the ground upwind of the crest on days with sufficiently deep echoes on the upwind edge, including 18 and 27 January 2006 (Figs. 6a and 8b, respectively). Thus the upwind gravity wave tilt may contribute to the apparent low-level growth across the LCL–terrain intersection (Fig. 10).

Subsidence prevails downwind of the crest (Fig. 12b), resulting in a clear dipole in the lee–luff difference FAD for vertical velocity over a substantial depth (Fig. 11d). BL turbulence is still present in the lee—in fact, it may be more intense as the flow accelerates downslope (Figs. 11d and 12b)—but this turbulence becomes immaterial for snow growth (except maybe aggregation) as cloud droplets rapidly evaporate and the environment becomes subsaturated with respect to ice. The vertical velocity change across the crest is greatest at low levels (as illustrated dramatically in Fig. 8b), and thus the decrease in snow rate and radar reflectivity values is best defined at low levels (Fig. 11c). Some snow growth still occurs aloft as the flow crosses the crest, but this is not significant for surface precipitation. The composite mean reflectivity profiles (Fig. 12a) confirm that the near-surface snow rate is higher upwind than downwind of the crest for these 10 cases and this particular mountain range, consistent with literature mentioned in section 1. Clearly this difference is very sensitive to the location of the profiles in the lee.

c. Does BL turbulence really matter?

To examine whether turbulence in the BL affects low-level snow growth along the upwind terrain slope, we compute BL turbulence intensity for each flight leg as the variance of the WCR vertical velocity for all gates upwind of the crest and below the BL top (listed in Table 1). All geographically fixed flight legs on the seven flights in 2008 and 2009 are used. The flight legs then are partitioned in roughly equal parts based on their BL turbulence intensity.

Such partitioning reveals that the low-level snow enhancement across the LCL–terrain intersection is systematically greater when turbulence is more intense in the BL (Figs. 13a,b). This suggests that the low-level precipitation enhancement along the upwind slope (Fig. 11a) is at least partly due to BL turbulence, not just an upwind-tilting gravity wave. The effect of BL turbulence is even more pronounced on the more stable days (not shown): under stronger BL turbulence (i.e., strong wind) and high-stability conditions, low-level snow growth
across the LCL is much enhanced compared to the weaker BL turbulence (i.e., weak-wind), low-stability cases. Here, static stability is determined based on the upwind soundings, as listed in Table 1.

Next, we contrast the FADs for the five flights on the more stable days to those for less stable days. The precipitation systems are shallower on the less stable days anywhere over the mountain, as evident from the mean echo top listed in Table 1. The WCR vertical velocity transects do not reveal upwind-tilting gravity waves on these days, and Geostationary Operational Environmental Satellite (GOES) imagery (not shown) further indicates that clouds tended to be more confined to the mountains on these days. The partitioning by stability shows that low-level growth across the LCL–terrain intersection is most evident on the more stable days (Figs. 13c,d). This is seen as evidence that the onset of turbulence enhances snow growth: on more stable days little growth occurs in the stratified cloud until BL turbulence is encountered. On less stable days deeper turbulence, sometimes convective turbulence up to cloud top, is present upstream of the LCL–terrain intersection, at least on those days with echoes at all in that region. On these days growth does occur as the flow ascends the terrain because more liquid water becomes available in a deeper turbulent cloud, but the change across the LCL–terrain intersection is less dramatic.

7. Snow growth mechanisms

Snow growth can be due to deposition, riming, or aggregation. Aggregation does not convert liquid water to ice, but it does increase the median diameter of snow flakes and thus the reflectivity. It also slightly increases the fall speed (Brandes et al. 2008). Aggregation enhanced by BL turbulence may be a significant factor in the observed low-level reflectivity increase across the LCL–terrain intersection (Fig. 11a).

The propensity for snow to grow by riming versus deposition depends on vertical velocity, liquid water content (LWC), and the drop size spectrum. These variables are measured at flight level, but for safety reasons the flight level mostly remained above the BL top. Close to the mountain peak the WKA probably did sample air rising...
from the BL in some cases (e.g., Figs. 5b and 7b), especially on the less stable days. Visual examination of 2D-C and 2D-P data collected in the vicinity of the mountain peak did not reveal distinctly higher riming amounts. Heavily rimed particles (graupel) were not encountered on any of the 10 flights. This may be because the median droplet diameter was rather small on most days (e.g., Fig. 8e); the largest median diameter values, observed on 18 January 2006 (Fig. 5e), were still less than 20 $\mu$m. Thus, according to Wang and Ji (2000), the riming efficiency probably was quite low for all but the largest droplets on most flights.

There is no correlation between Doppler vertical velocity and reflectivity in the BL (as is evident from the various WCR transects shown), nor do 2D-C or 2D-P ice crystal concentrations or ice mass correlate with gust probe vertical velocity at flight level where the flight track penetrates the BL (not shown). Individual BL eddies probably are too transient to detect the snow growth response. But cloud liquid water does correlate somewhat with vertical air velocity over the highest terrain where BL air is most likely to be sampled at flight level (Fig. 14). In the absence of snow growth, the LWC would be a function of height only (i.e., the adiabatic LWC). A positive correlation between LWC and vertical velocity indicates that some liquid water has been consumed by snow by the time the eddy descends. The correlation is stronger for the five less stable days, probably because the (convective) eddies are larger and deeper. It is stronger also on some individual days since LWC and cloud-base temperature vary much among the 10 cases (Table 1; Figs. 5c and 7c).

It is not clear whether this consumption of liquid water by snow is via the Bergeron process or by riming. The riming amount is difficult to measure; riming estimates based on particle images from optical array probes are subjective and cannot be readily composited for entire flight legs. Particle density and thus fall speed increase rapidly with increasing riming fraction (Locatelli and Hobbs 1974), so it is tempting to examine variations in fall speed among the 10 cases (Table 1; Figs. 5c and 7c).

A gust probe measures the air vertical motion at flight level. The hydrometeor fall speed near flight level can be estimated by removing the gust probe vertical velocity from the average close gate (~120 m up and down) radar vertical velocities above and below the aircraft (section 6a). This fall speed estimate is biased toward the largest snowflakes because they dominate the received power and thus the Doppler velocity estimate. It is also rather uncertain, given the lack of spatial coincidence of radar and gust probe data and the limited accuracy of the instantaneous (1 Hz) radar and gust probe vertical

![Figure 14](https://example.com/figure14.png)

**Fig. 14.** Scatterplot of PVM LWC against gust probe vertical velocity for (a) the five most stable days and (b) the five least stable days in boldface in Table 1. The 1-Hz data sampling conditions are LWC > 0.05 g m$^{-3}$, and the aircraft is within the BL, according to the BL depth estimates listed in Table 1. The regression line is shown.
velocity measurements. Thus we average the 1-Hz fall speed estimate for both the in-BL and out-of-BL sections for each flight leg.

This fall speed estimate is plotted against the forward scattering spectrometer probe (FSSP) median droplet diameter averaged over corresponding sections, but only where the FSSP droplet concentration exceeds 20 cm\(^{-3}\), for all flight legs where at least part of the track penetrated the BL according to the BL top estimate listed in Table 1. Both the fall speed and the median diameter are averaged for the sections within and above the BL, respectively, for each leg. Also shown are the linear best-fit lines for the in-BL (solid line) and above-BL (dashed line) data.

8. Discussion: Snow growth in the boundary layer

Turbulent vertical motion does not imply any net lifting; it simply tends to mix, leading to a uniform distribution of conserved quantities. Yet above the cloud base turbulent motions can affect hydrometeor growth. Houze and Medina (2005) suggest that turbulent rising eddies create pockets of higher LWC and larger droplets, resulting in more riming as well as more aggregation, and thus snow particles falling out more rapidly. Their observations were made in the Oregon Cascades, and the turbulence they examine was shear induced near the top of the blocked flow layer (i.e., above the BL). We believe that BL turbulence, possibly combined with shallow convection, may be important also in warm clouds, through accelerated growth by collision and coalescence. For instance, BL turbulence may contribute to the generation of occasionally heavy rain from shallow non-brightband systems rising over the California coastal mountains (White et al. 2003; Neiman et al. 2005). Rapid precipitation growth in the saturated BL may contribute to extremely tight precipitation gradients across ridges (e.g., Anders et al. 2007; Kirshbaum and Smith 2009). Minder et al. (2008) document observed climatological precipitation totals about 50% higher on top of an 800-m-high ridge, pointing into the prevailing wind, relative to valleys just 10 km on either side. We believe that the ridge enhancement is at least partly due to BL turbulence above cloud base [which according to soundings in Minder et al. (2008) generally is below 800 m], resulting in rapid growth mainly by collision–coalescence.

This study has focused on airborne profiling radar data, a powerful resource to examine vertical velocity patterns and precipitation growth over complex terrain. Radar reflectivity relates reasonably well to precipitation mass and precipitation rate, but it says little about precipitation growth mechanisms. To study these mechanisms, LWC, drop size distributions, and riming amounts must be measured within the BL. This is difficult to do over complex terrain because of flight restrictions. Further investigations with large eddy simulation models of sufficient resolution to capture both a sufficiently large domain and a significant fraction of the TKE spectrum in the BL are needed. Such work can go further than observational studies in proving the significance of BL turbulence in orographic precipitation growth. Such work may also indicate the need to parameterize the impact of BL turbulence on hydrometeor growth over mountains in models in which the eddy transfer across the BL is parameterized.

9. Conclusions

This study points to the potential significance of BL turbulence to orographic precipitation, specifically but not exclusively in mixed-phase clouds. The main source of evidence is high-resolution vertically pointing
airborne Doppler radar data, collected in 10 winter storms over a mountain in the high-elevation, continental environment of Wyoming. The main findings are as follows:

- All transects depict a turbulent layer draped over the terrain, sometimes clearly distinct from the stratiform flow aloft. This layer, the BL, varies in depth from about 0.4 to 1–2 km, the latter on the least stable days when the BL top corresponds with the cumulus echo top.
- Sometimes ice crystals appear to initiate within the BL. Possible mechanisms include blowing snow and ice multiplication (splintering) near the ground. The latter involves supercooled droplets colliding with rimed vegetation. These surface-based ice initiation mechanisms remain unproven.
- Spectral analysis of Doppler vertical velocity data reveals an inertial subrange with the highest power near the surface, and decreasing power with height toward the BL top. Larger-scale eddies are directly forced by terrain undulations.
- Comprehensive frequency-by-altitude diagrams indicate a broad range of vertical velocities in BL and rapid snow growth within the BL as the BL air rises through the cloud base. The location of the snow growth, upwind of the crest, can be explained by an upwind-tilting gravity wave in some cases, but in all cases BL turbulence appears to contribute to the growth, as it is more marked when the BL turbulence is more intense and in more stratified flow cases experiencing a rapid transition toward turbulent flow near the mountain.
- Little is known about the microphysical pathways of this low-level snow growth, since the flight level generally remains above the BL. The radar-documented change may at least partly be due to aggregation. Aside from aggregation, it is not clear whether relatively more growth is due to depositional or accretional growth in the BL. Limited flight-level data within the BL indicate that supercooled liquid water is effectively consumed by snow in the rising eddies.

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