The Use of Snow-Level Observations Derived from Vertically Profiling Radars to Assess Hydrometeorological Characteristics and Forecasts over Washington’s Green River Basin

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ABSTRACT

Two vertically pointing S-band radars (coastal and inland) were operated in western Washington during two winters to monitor brightband snow-level altitudes. Similar snow-level characteristics existed at both sites, although the inland site exhibited lower snow levels by ~70 m because of proximity to cold continental air, and snow-level altitude changes were delayed there by several hours owing to onshore translation of weather systems. The largest precipitation accumulations and rates occurred when the snow level was largely higher than the adjacent terrain. A comparison of these observations with long-term operational radiosonde data reveals that the radar snow levels mirrored climatological conditions. The inland radar data were used to assess the performance of nearby operational freezing-level forecasts. The forecasts possessed a lower-than-observed bias of 100–250 m because of a combination of forecast error and imperfect representativeness between the forecast and observing points. These forecast discrepancies increased in magnitude with higher observed freezing levels, thus representing the hydrologically impactful situations where a greater fraction of mountain basins receive rain rather than snow and generate more runoff than anticipated. Vertical directional wind shear calculations derived from wind-profiler data, and concurrent surface temperature data, reveal that most snow-level forecast discrepancies occurred with warm advection aloft and low-level cold advection through the Stampede Gap. With warm advection, forecasts were too high (low) for observed snow levels below (above) 1.25 km MSL. An analysis of sea level pressure differences across the Cascades indicated that mean forecasts were too high (low) for observed snow levels below (above) 1.25 km MSL when higher pressure was west (east) of the range.

1. Introduction

Hydrometeorological prediction in mountainous regions is particularly challenging because of the existence of a fluctuating rain–snow boundary. The altitude in the atmosphere where snow changes into rain (i.e., the snow level) during a storm is a product of storm characteristics and the interaction of the atmosphere with complex terrain (e.g., Marwitz 1987; Medina et al. 2005; Minder et al. 2011; Minder and Kingsmill 2013). The snow level, which should not be confused with the slightly higher free-atmosphere freezing level marking the 0°C isotherm, is an essential variable in prediction, used to estimate the fraction of river basin area that will receive rain and possibly produce flooding, versus the portion of the basin that will receive snow that may be stored for future runoff. For example, White et al. (2002) used the operational River Forecast System (Smith and Page 1993) at the National Weather Service (NWS) California Nevada River Forecast Center (CNRFC) to show that for several watersheds in northern California, increasing the snow level by 600 m could more than double the runoff in...
the watershed. A more recent study by White et al. (2010) further quantified the relationship between snow level and runoff in mountainous California watersheds. In contrast to the adverse effects of flooding, runoff that is generated by the melting of the winter snowpack during the dry warmer months and subsequently captured in reservoirs is vital to water supply there. Significantly, climate change may alter both the availability of snowpack as runoff, especially for lower elevation watersheds, and the timing of runoff (e.g., Knowles et al. 2006; Mote 2006; Minder 2010). Ultimately, it is important to monitor snow levels in storms during the winter, both for determining short-term flood-related impacts (e.g., White et al. 2002, 2010; Lundquist et al. 2008; Neiman et al. 2008a, 2011, 2013) and for assessing longer-term seasonal runoff (e.g., Knowles et al. 2006; Mote 2006; Lundquist et al. 2008).

One region of the United States especially prone to flood-producing winter storms with high snow levels is the Pacific Northwest, including western Washington. Many of these high-impact storms are accompanied by an atmospheric river (AR; Zhu and Newell 1998; Ralph et al. 2004)—a long (>~2000 km) and narrow (<~1000 km) corridor of enhanced water vapor transport in the warm sector of an extratropical cyclone where significant orographic precipitation enhancement is favored upon landfall (e.g., Ralph et al. 2006, 2011; Neiman et al. 2008a,b, 2011). The climatological precipitation distribution across western Washington is dominated by the effects of orography (e.g., Colle et al. 2000; Minder et al. 2008).

A strong AR, with an anomalously high snow level, generated record flooding across parts of western Washington in January 2009 (Mastin et al. 2010). It resulted in water seepage issues at the Howard Hanson Dam on the Green River, which drains a portion of the western slope of the Cascade Mountains (Fig. 1). Below the dam is where the greatest flood-related hazards in the state are found, that is, in the densely developed Kent Valley south of Seattle. In response to the seepage, the National Oceanic and Atmospheric Administration (NOAA) joined forces with the U.S. Army Corps of Engineers (USACE) and local and state agencies to enhance services to the communities at risk. As part of this coordinated effort, NOAA’s Earth System Research Laboratory (ESRL) deployed instrumentation to monitor in real time key hydrometeorological conditions in and near the Green River basin during subsequent winters (White et al. 2012).

Because of the sensitivity of mountain-basin runoff to snow-level altitude, two vertically pointing S-band precipitation-profiling Doppler radars (White et al. 2000) were operated in western Washington (Fig. 1, Table 1) during the 2009/10 and 2010/11 winter seasons to monitor snow-level altitudes during landfalling storms. One of these radars was situated along the coast at Westport (WPT) and the other inland along the Green River at Ravensdale (RVD). To demonstrate how the uniqueness of these radar datasets may be used to enhance the understanding and assessments of hydrometeorological processes within the watershed, diagnostics are performed on the radar data to address the following questions:
1) Do spatial and temporal snow-level relationships between the coastal and inland radar sites provide insights into how the snow level responds to the local and regional meteorology, and do they reveal the utility of using a single coastal site to infer snow-level information over the inland watersheds?

2) Can inland radar data be used as performance and representativeness measures of nearby NWS snow-level forecasts over the upper Green River basin using the methodology established in White et al. (2010) for northern California?

3) Can radar snow-level and wind data, when combined with precipitation, streamflow, surface temperature, and surface pressure measurements, identify potential hydrometeorological impacts related to snow-level misrepresentations that result from using single-point, snow-level forecasts during regionally complex meteorological conditions?

2. Datasets

The S-band radars at WPT and RVD simultaneously collected data from 3 December 2009 to 18 May 2010 and from 15 October 2010 to 27 April 2011. When precipitation was detected by these radars, 30-s-resolution vertical profiles of signal-to-noise ratio and vertical radial velocity were processed using the ESRL/Physical Sciences Division (PSD) snow-level algorithm (White et al. 2002) to produce 15-min snow-level observations at the bright band—a radar signature resulting from melting precipitation (e.g., Battan 1973; Fabry and Zawadzki 1995). In rare situations where multiple bright bands existed in a single profile (Martner et al. 2007), the algorithm performed suboptimally. The algorithm outputs were manually quality controlled to remove random outlier data and to occasionally add missing snow levels. During the period of simultaneous operations, 6997 and 5761 snow-level observations were produced for WPT and RVD, respectively. In the WPT dataset, 154 snow-level observations were flagged as outliers and removed, while 141 snow-level observations were added (2.2% and 2.0% of the total number of WPT snow levels, respectively). In the RVD dataset, 209 outlier observations were removed and 117 observations were added (3.6% and 2.0% of the total number of RVD snow levels, respectively). The snow-level additions resulted largely from the occurrence of double bright bands. In these situations, the altitude of the lower bright band was added. At WPT, the double bright band occurred most frequently during the passage of transient fronts aloft, which likely possessed a nearly isothermal layer, as in Martner et al. (2007). At RVD, the lower bright band occurred most often during the westward flow of cold continental air across the Cascade crest, thus likely locally lowering the snow level relative to what the large-scale conditions would dictate, as in Steenburgh et al. (1997).

The radar data were augmented with a variety of interagency datasets across the region (Fig. 1, Table 1). Operational freezing-level forecasts at Stampede Pass (SMP) were archived in real time and validated against the radar data to assess forecast performance. Historical records of freezing level (i.e., the altitude of 0°C) observed at the nearest operational radiosonde (raob) site at Quillayute (UIL) were used to independently verify coastal–inland spatial relationships in snow level and were also used to produce climatological vertical distributions.
of snow level. The same raob dataset was used in conjunction with long-term records of gauge-measured precipitation at Landsburg (LBG) to produce climatological rain-versus-snow distributions. When combined with 15-min river-flow data on the Green River at Auburn (ABN) and Howard Hanson Dam (HHD), the raob dataset described the hydrometeorological characteristics of historical flooding events on this river. Two 915-MHz radar wind profilers (e.g., Carter et al. 1995), one at WPT and another at the Seattle–Tacoma International Airport (SEA), provided hourly averaged vertical profiles of horizontal wind velocity from ~0.1 to 4.0 km above ground with ~100-m vertical resolution and ~1 m s\(^{-1}\) accuracy in all weather conditions. The profiler winds were objectively edited using the vertical–temporal continuity method of Weber et al. (1993). Sea level pressure data were used from the NWS operational surface meteorological sites at SEA and Ellensburg (ELN). An adjacent pair of U.S. Department of Agriculture (USDA) automatic snow monitoring stations [referred to as Snow Telemetry (SNOTEL) sites; e.g., Trabant and Clagett (1990)] at SMP and Rex River (RXR) provided hourly surface temperatures.

3. Observational diagnostics
   a. Spatial relationships

   The vertical distribution of all the radar snow-level observations over the two-winter period at each site is shown in Fig. 2, where the data are binned in 100-m increments. The two profiles were constructed independently of one another. The peak in the frequency-of-occurrence (FOC) distribution is 0.85 and 1.25 km MSL at WPT and RVD, respectively, with similar peak magnitudes. The altitude of the peaks differ because of a combination of meteorological conditions and data issues: 1) post-cold-frontal convective showers with low snow levels were less common in the Cascade foothills at RVD than on the coast at WPT (perhaps due, in part, to downslope warming and/or drying downwind of the Olympic Mountains in post-cold-frontal west-to-northwest flow), 2) data outages at RVD occurred during periods when low snow levels were detected at WPT, and 3) intermittent radar hardware issues at RVD occasionally prevented the detection of low snow levels. Below 0.6 and above 1.6 km MSL, the distributions are more similar. Section 3b will utilize common periods to compare the radar data between WPT and RVD. Figure 2 also shows the cumulative fraction of the basin area in the Green River watershed (GRB) above HHD as a function of basin elevation (from Neiman et al. 2011). These data indicate that the most frequently occurring snow levels over the Green River watershed at RVD coincided with rain (rather than snow) falling in a large majority (~90%) of the basin area.

   The coastal and inland radar snow-level analysis was extended to include comparisons with the NWS raob data from UIL to assess site representativeness over a larger region. This comparison uses a subset of radar data corresponding only to times when the balloons were launched twice daily (0000 and 1200 UTC) and when at least one radar snow-level observation was present within a 2-h window centered on the balloon launch time. Averaging was used when multiple snow-level observations were present within the 2-h window. The snow levels in the raob data were estimated by linearly interpolating these data to the 0°C freezing level and then subtracting a 233-m constant offset, as specified in White et al. (2010). This offset is discussed in more detail in section 4. The results from this comparison (Fig. 3) once again show good agreement in both the spread and magnitude of the distributions. The FOC peak magnitude at UIL is nearly equal to that in the radar dataset at both RVD and WPT. However, the height of the UIL peak magnitude is 300 m lower than at RVD and 100–200 m lower than at WPT. These
differences may reflect the fact that UIL is farther north than both radars.

To assess the historical representativeness of these dual-winter season snow-level results, the analysis was expanded to include the 45-yr record, from 1 January 1966 to 27 April 2011, of raob data from UIL (Fig. 4). Here, snow level was calculated in the same manner as for the dual-winter season comparison, except only during times when measurable precipitation occurred at the surface site at LBG, within the 2-h window centered on the balloon launch times. The most frequently occurring snow level over the 45-yr period at UIL occurred at 1.15 km MSL. This identically matches the height of the higher peak FOC at WPT and is 100 m lower than the peak FOC at RVD. A secondary raob peak of lesser magnitude also exists at 750 m MSL, 100 m lower than the lower peak at WPT.

Both the dual-season and historical raob comparison results support the notion that 1) the snow-level altitudes during the 2009/10 and 2010/11 winter seasons at RVD were characteristic of typical climatological conditions and 2) snow-level height distributions are quite similar across large spatial distances (~225 km) from the northwest coast of the Olympic Peninsula to the western slope of the central Cascade Range (CR) of Washington. This latter condition suggests that larger synoptic and/or mesoalpha scales play a major role in controlling the snow level across the region, although smaller-scale, topographically modulated impacts on snow-level altitude do exist and will be discussed in the following subsection.

b. Lead-time relationships

While the previous time-inclusive composite analysis demonstrated spatial consistency in snow-level height across western Washington, large variations in snow level can occur across the domain at any instance in time. To investigate this hypothesis, snow-level events were defined and identified in the WPT and RVD radar datasets as follows: 1) a large coherent cloud or precipitation radar signature, 5 km deep and 3 h duration, must mark the beginning of the event in the time–height series at each site within a 24-h period; 2) a minimum of a 6-h precipitation-free period (at the surface and aloft) must precede each event; and 3) the maximum time between detected contiguous snow levels must not exceed 6 h within the event. These criteria were chosen based on cloud and precipitation signatures in the radar data that most commonly characterized weather systems affecting both the coastal and inland sites. Using these criteria, 99 events were identified and matched in the WPT and RVD datasets.

Four hydrometeorologically meaningful categories were defined and tracked temporally within each event in both datasets. These categories included the first detected snow level of the event, the largest 2-h change
in snow level >200 m, and the lowest and highest snow levels of the event (Figs. 5–7). Figure 5a shows that, on average, the first detectable snow level occurred at RVD 4.47 h later than at WPT, with a standard deviation of 3.74 h. These data suggest that vertically pointing radars along the coast can provide useful lead-time information for the altitude of the snow level over the west slope of the Cascades, consistent with the fact that most winter storms affecting Washington move onshore. Occurrences of snow-level height changes exceeding 200 m per 2-h period (Fig. 5b) were observed to a lesser degree (i.e., 38 data points) but exhibited a similar average lag time of 3.95 h between the two sites. However, the standard deviation was more than triple that of the first-snow-level data, indicating that perhaps not all of these changes resulted from synoptic structures translating across the spatial domain.

Identifying coherent features at both sites for the lowest and highest snow-level categories posed the greatest challenge, as local variability often produced the highest and lowest snow levels. These instances were often offset by short time periods between the minimum and maximum values of the coherent features. To help minimize this source of noise, the lowest and highest snow levels were extracted from the lowest and highest triplets of snow-level observations that occurred within 1 h of each other at each site. Figure 6 shows mean-event WPT-to-RVD lag times of 4.94 and 4.74 h and standard deviations of 11.42 and 8.10 h for the lowest and highest snow levels, respectively. Like the 2-h height changes, local-scale processes may have produced significant variance. Despite the increased noise in these latter three categories, the mean statistics verify the frequent presence of coherent features at both sites with phase lags between WPT and RVD most commonly occurring in the 2–10-h range.

Under certain synoptic conditions, cold continental air on the east side of the Cascade crest can flow westward through mountain gaps into basins on the west side of the crest, thus producing snow levels that are lower there than in nongap locations west of the Cascades (e.g., Steenburgh et al. 1997). Such variability, if it occurs frequently enough, would produce a low bias in the snow level at RVD relative to WPT for the same events. Using the data in Fig. 6, the height differences between the two sites for the lowest and highest snow levels were calculated and plotted in Fig. 7 to identify systematic biases. These results demonstrate that, on average, both the lowest and highest snow levels were lower at RVD, with a mean bias of −70.1 and −73.5 m for the lowest and highest snow levels, respectively. These biases are opposite that in Fig. 2, where the distribution of the snow-level occupancies from the two sites was assessed independently of one another. In contrast, the event-matched analyses in Fig. 7 more accurately depict spatial differences in snow level on a storm-by-storm basis. In addition to the westward advection of cold air across mountain passes, other mechanisms may also account for the low inland bias, including terrain-induced adiabatic cooling, latent cooling from melting precipitation, and variations in the vertical melting distance of frozen hydrometeors (Minder et al. 2011).

4. Forecast performance and representativeness

Vertically pointing precipitation radars have been used previously to quantify the baseline performance associated with freezing-level forecasts made in northern California by the CNRFC (White et al. 2010). A similar, but more comprehensive, methodology is used here to assess the skill of freezing-level forecasts produced by the Northwest River Forecast Center (NWRFC) for the upper elevations of the Green River watershed along the windward slope of Washington’s Cascade Mountains.
The NWRFC systematically produces 48-h tabular operational forecasts of freezing level at 6-h intervals at 1200 UTC every day (www.nwrfc.noaa.gov/weather/temp_fcst.cgi) for 24 locations (Fig. 8). The freezing-level forecasts use numerical weather prediction models, the climatology-based Parameter–Elevation Regressions on Independent Slopes Model (PRISM; Daly et al. 1994), and NWS Weather Forecast Office Intersite Coordination (ISC) forecast grids. In addition, Hydrometeorological Analysis and Support (HAS) forecasters provide

**Fig. 5.** Time difference (h) between WPT and RVD observations during large-scale precipitation events for (a) the first detected snow level and (b) the max 2-h height change in snow level >200 m.

**Fig. 6.** As in Fig. 5, but for the (a) lowest and (b) highest snow levels.
crucial input, especially during stormy periods. Mean areal time series are generated using basin boundaries and the ISC grids, or by extracting forecast point values from those grids and employing station weights developed during hydrologic calibration.

Stampede Pass, located on the eastern border of the Green River watershed and on the Cascade crest, is one of the 24 freezing-level forecast points (Fig. 8). Forecasts for this site are compared with radar observations of the snow level at nearby RVD. The RVD 15-min snow-level data were used to verify the SMP forecasts for times when at least one radar-derived snow level occurred within a 2-h period centered on the forecast time. When multiple radar observations existed during the comparison period, the observation closest to the forecast time was used. The radar-derived snow levels were converted to freezing level by adding the average snow-level displacement height of 233 m found in White et al. (2010). Their study documented offsets ranging between 122 and 427 m, with the most frequently occurring offsets between 180 and 300 m. An earlier study by White et al. (2002) obtained similar results, with an average offset of 192 m and a root-mean-square error of 108 m. Comparable displacement heights were also reported in Stewart et al. (1984) and Fabry and Zawadzki (1995). The uncertainty in snow-level displacement heights, which can arise because of multiple factors, including variations in precipitation intensity and static stability, can introduce uncertainties in assessing the snow-level forecast performance. Nevertheless, given the general agreement between the independent studies cited above, we believe that the 233-m offset represents a valid first-order estimate.

Figure 9 highlights the freezing-level discrepancies between forecasts and observations for all forecast times and freezing levels. Two results are noteworthy: 1) the distribution is skewed toward forecasts being lower (i.e., colder) than observations, particularly for discrepancies exceeding ±1 km, and 2) the peak of the distribution is offset slightly from zero toward lower-than-observed forecasts. The first result can be viewed in greater detail by plotting the distribution of freezing-level forecast discrepancy as a function of observed freezing level and forecast verification time (Fig. 10). Here, regardless of forecast verification time, freezing-level forecasts that are lower than observations are far more common for discrepancies <0.5 km than those forecasts that are higher than observations with discrepancies >0.5 km. In addition, like the forecast-discrepancy behaviors presented in White et al. (2010), there is a tendency for these low forecast discrepancies to increase in magnitude with increasingly higher observed freezing levels. These types of discrepancies have the greatest hydro-meteorological consequences, as a large fraction of the basin area would receive precipitation in the form of rain that was anticipated to receive snow.

Figure 11 presents the mean and standard deviation of the freezing-level forecast discrepancies for each 6-h forecast verification time. A low forecast bias of 100–250 m is evident across all forecast times. Both the mean
and variance are consistent, that is, without deteriorating forecast performance, across the 48-h forecast period. Within the narrower range of forecast discrepancies between ±0.5 km, the mean biases are much smaller (i.e., 0–50 m) and random in sign for the different forecast periods, and the standard deviations are ~60% smaller than for all forecast discrepancies (not shown). Hence, the large standard deviations for all forecast discrepancies across all forecast periods (Fig. 11) arise because of the relatively small number of large low forecast discrepancies that skew the distribution in Fig. 10. As such, the cause of the low forecast bias for all forecast periods is related partly to systematic biases in the forecast process during those episodes when the altitude of the observed freezing level is significantly underforecast, that is, the forecast process may underrepresent the magnitude of warm-air advection aloft (as in California; White et al. 2010) and/or overrepresent the westward low-level cold-air intrusion through the SMP gap to RVD.

To explore the representativeness of the SMP snow-level forecasts at RVD, we compared hourly surface temperatures at SMP and RXR. RXR is located ~20 km west of SMP and 18 km east of RVD (see Fig. 1), and it is only 11 m lower than SMP. Figure 12a shows the observed surface temperature differences between SMP and RXR versus the discrepancy in freezing level between the forecasts at SMP and the observations at RVD. For 89% of the 714 freezing-level comparisons, surface temperatures are warmer at RXR, which suggests that colder air east of the Cascades frequently advects westward through the gap but with only a limited westward extent. This likely contributes to discrepancies between the observed and forecasted freezing levels.

Given the potentially impactful consequences of the largest negative forecast discrepancies (i.e., those ≤−1 km), that subset of 61 is analyzed in Fig. 12b. A large number of the freezing-level discrepancies are associated with observed freezing levels at RVD above the elevation of SMP and forecasted freezing levels at SMP below the elevation of SMP. Those cases where subfreezing surface temperatures were also observed at SMP confirm that the low forecast tendencies have validity (i.e., they are not necessarily errors). At the same time, almost all of those cases are also associated with warmer surface temperatures to the west of SMP at the same elevation (e.g., the first five cases on the left-most side of the plot). These concurrent observations
suggest an overlap of the following: 1) cold continental air often resides in the SMP gap with low freezing levels; 2) the cold continental air mass is either modified significantly as it advects westward or remains at low levels within the Green River basin, and/or never makes it as far west as RXR and RVD; and 3) a warm air mass exists aloft with high freezing levels. It is also evident that there are times when an inaccurate forecast contributes to these large freezing-level discrepancies. Cases in the middle of the distribution demonstrate this. Here, when the SMP surface temperature and the SMP/RXR surface temperature difference are close to 0°C, the observed freezing levels at RVD are also very close to the elevation of SMP. These observations suggest that a more homogeneous air mass exists between RVD, RXR, and SMP. However, the forecast freezing level at SMP is often erroneously low by 1.0–1.5 km.

Of the 61 largest negative freezing-level forecast discrepancies, 41 of them have unique comparison dates/times between the SMP forecasts and the RVD observations. Based on these 41 dates/times, composite analyses were generated from the 32-km-resolution North American Regional Reanalysis (NARR; Mesinger et al. 2006) dataset to provide synoptic-scale context at 900 and 700 hPa (Fig. 13). A prominent, deep-layer cyclone is situated west of British Columbia, with southwesterly flow impacting the Pacific Northwest. At 900 hPa, the NARR captures cold-air pooling east of the Cascade crest. At that level, strong warm-air advection is focused along coastal Washington and British Columbia. Aloft at 700 hPa, strong warm-air advection covers all of Washington. This scenario bolsters our assertion that the largest negative forecast discrepancies occur during strong warm-air advection aloft. Also, the cold-air pooling likely results in low freezing-level forecasts at SMP relative to the warmer location immediately to the west at RVD.

![Fig. 9. FOC of freezing-level forecast discrepancy (km) for all heights and all forecast verification times between 0 and 48 h. The observations are from RVD and the forecasts are valid at SMP.](image9)

![Fig. 10. Freezing-level forecast discrepancy (km) as a function of radar-derived freezing level (y-axis) and verification time (see key). The observations are from RVD and the forecasts are valid at SMP.](image10)

![Fig. 11. Mean discrepancy between NWRFC and radar-derived freezing levels (m) for different forecast lead times, and the std dev of those discrepancies. The observations are from RVD and the forecasts are valid at SMP.](image11)
5. Hydrometeorological implications

a. Historical rain-versus-snow distributions

Snow level, precipitation amount, and precipitation rate are the meteorological variables that determine the potential runoff in a watershed. Using historical data, the partitioning of observed surface precipitation accumulations and average precipitation rates for radar-derived and raob-inferred snow levels (Fig. 14) reveals the climatological distributions of runoff potential across the elevation profile of the Green River watershed above HHD. The radar data from the 2009/10 and 2010/11 winter seasons and the 45-yr raob record from UIL were used in conjunction with surface rain gauge data from those radar sites and from LBG to characterize the seasonal and climatological rain-versus-snow distributions across the Green River basin. Precipitation data at the radars (at LBG) were summed over 15-min (2 h) periods centered on the time of the snow-level observations at those radars (at LBG). Results for WPT and RVD (Fig. 14a) show peak accumulations of precipitation occurring with snow levels near 1.25 km MSL at both radars. While the peak precipitation at RVD was nearly 1.5 times larger than the next largest peak at 1.85 km, there was only a 2% difference in magnitude between the primary and secondary precipitation peaks at WPT (also at 1.85 km). Roughly 47% of the total seasonal precipitation at RVD fell as rain below 1.25 km. When combined with the elevation profile of basin-area cumulative fraction, ~90% of the basin area in the Green River watershed above HHD received precipitation in the form of rain for 47% of the total precipitation (734 of 1559 mm). A comparison of this analysis with the 45-yr raob snow levels from UIL (inferred by subtracting 233-m from the freezing level) reveals close agreement with the height of the primary precipitation peaks at both RVD and WPT (Fig. 14b).

The accompanying average hourly precipitation rates with sample sizes ≥35 (Figs. 14c,d) show increasing rates with increasing snow-level heights. This relationship supports the notion that an increase (decrease) in moisture associated with warmer (colder) air masses, as indicated by higher (lower) snow levels, is conducive to producing greater (lesser) rainfall rates. In addition, the largest rainfall rates at RVD occur with snow-level heights that are above the Green River watershed (>1.6 km MSL). This produces a compound effect on potential runoff, as the greatest precipitation rates are most likely to occur when the entire basin is receiving rain.

b. Hydrometeorological characteristics of Green River flood events

To assess the impact that the radar-verified, snow-level forecasts may have on flood forecasting in the
Green River watershed, it is first useful to present the hydrometeorological characteristics of historical flooding events. The NWS has the responsibility of alerting the public when flooding threats arise along the Green River. This alert system is based on the flow rate along the lower Green River at ABN (see Fig. 1), as it enters the populated, low-lying Green River Valley, and uses the criteria outlined in Table 2. This flow rate is determined by current gauge readings at ABN and/or the expected outflow from HHD located upriver (see Fig. 1).

Between October 1987 and April 2011, all 23 cool seasons of 15-min river gauge data at ABN and corresponding precipitation data at LBG were used to relate the meteorology to the hydrology in the context of the NWS flood-alert system. While the 15-min river gauge dataset is a shorter record than the daily dataset by 50+ years, it more accurately resolves the magnitude of instantaneous peak flows, which is more relevant to flood-forecasting applications. In Fig. 15, the daily maximum flow rates (i.e., the peak 15-min flow rates per day) at ABN are plotted against the 290 5-day periods with precipitation accumulation greater than zero at LBG. A 5-day accumulation period, ending on the day of the daily maximum flow rate, was chosen to smooth out flow-release variability related to HHD dam operations. A runoff relationship is evident, with flow rates increasing with increasing precipitation for accumulations >50 mm. The events of particular interest are the 28 (9.7% of the total) with daily maximum flow rates greater than 5000 ft$^3$ s$^{-1}$ ($142$ m$^3$ s$^{-1}$), which trigger an NWS internal alert flood-phase-1 (or greater) designation.

The raob data at UIL were used to investigate meteorological characteristics associated with these flood-alert events. The mean 5-day-inferred raob snow levels (Fig. 16a) show that 61% of all NWS flood-alert events occurred with mean snow levels above 1.25 km MSL, which is the same height as the peak in snow-level FOC (Fig. 2) and accumulated precipitation (Fig. 14a) at RVD. In the context of basin runoff, 61% of the alert events occurred when at least 90% of the upper Green River basin received rain during an event. All of the alert events occurred when $\geq$50% of the upper basin area received rain (i.e., snow-level heights exceeded 0.9 km MSL). The raob-derived integrated water vapor (IWV) plot (Fig. 16b) shows that 90% of the events were associated with AR conditions sometime during the event, based on a minimum IWV threshold of 2 cm first reported in Ralph et al. (2004). These relationships corroborate the findings in Neiman et al. (2011), who show that each of the top-10 annual peak daily flows on the Green River during a 30-yr period occurred with landfalling AR. The connection between flooding and ARs for multiple years was also reported in Ralph et al. (2006) on the Russian River in coastal northern California and in Dettinger et al. (2011) from California to Washington. In these mountainous regions, ARs are predisposed to generating orographically enhanced heavy precipitation and flooding upon landfall, because they are often characterized by strong, warm, and moist low-level onshore flow with weak stratification and high snow levels (e.g., Ralph et al. 2005, 2011; Neiman et al. 2008a, 2011, 2013). Baroclinic forcing in the AR environment can further enhance the precipitation intensity (e.g., Cordeira et al. 2013; Neiman et al. 2013).

c. Snow-level forecast-discrepancy impacts on rain-versus-snow distributions

The combination of the forecast-discrepancy trends revealed in section 4 coupled with the snow-level characteristics during flood-alert conditions described above
carry significant hydrometeorological implications. Specifically, many of the forecast discrepancies were associated with snow levels that would have produced frozen precipitation over a large fraction of the basin area that otherwise received rain.

When the Green River basin elevation profile above HHD is used in conjunction with the snow-level forecast discrepancies, it is possible to quantify the fractional changes in the basin rain-versus-snow distributions that result from these discrepancies (Fig. 17). All changes in a cumulative fraction of basin area that were associated with snow-level forecast discrepancies $>|\pm 1.0 \text{ km}|$ occurred for forecasts that were lower than the RVD radar observations. These changes, and the other nonzero
fractional changes that lie in the upper left-hand quadrant, are associated with increased rain coverage across the basin and potentially more runoff than forecast. In addition, these occurrences outnumber the discrepancy count associated with a decrease in basin-area rain coverage by 399–221 samples and comprise 64% of the total. Finally, 53% of the discrepancies in the upper left-hand quadrant occurred with observed snow levels above 1.25 km MSL. Within the climatological framework presented in section 5b, these are potentially impactful, since the majority of NWS flood-alert conditions historically occur with snow levels above the 1.25 km threshold. It is also noteworthy that all of the changes less than \( \frac{10}{20} \) in the cumulative fraction of basin area (lower right quadrant in Fig. 17) occurred with observed snow levels below 1.25 km. All of these fractional changes were associated with snow levels that were forecasted higher than what were observed at RVD.

The cumulative frequency of occurrence of low snow-level forecast discrepancies that lead to unanticipated increases in basin-area rain coverage, as depicted in the upper-left quadrant of Fig. 17, is shown in Fig. 18. Here, for example, 24% of the discrepancies resulted in a 40% or greater increase in basin-area exposure to rain. For these cases, 97% were associated with radar-derived snow levels above 0.9 km MSL (not shown), where rainfall accumulations and rainfall rates were the greatest (see Fig. 14) and where all historical NWS flood alerts occurred (see Fig. 16a). This underscores the potentially large impact that snow-level forecast discrepancies can have on runoff expectations, particularly during times when the hydrologic consequences are high. In the forthcoming subsection, vertical wind profile and sea level pressure relationships are investigated in an effort to identify meteorological conditions that may contribute to these large and hydrologically significant forecast discrepancies.

d. Snow-level forecast-discrepancy dependencies on meteorological conditions

The 915-MHz radar wind profilers at WPT and SEA were used to assess the relationship between flow structure and snow-level forecast discrepancy. Specifically, the geostrophic vertical directional wind shear (VDS) at the wind profilers was calculated using the vector-averaged wind velocity in the adjacent 600-m-deep layers above and below the snow-level height at RVD. Under conditions dominated by geostrophic flow, this thermal-wind-derived shear diagnostic can be used as a proxy for horizontal temperature advection (e.g., Neiman and Shapiro 1989).

The VDS calculations were performed for all observed snow-level heights at RVD using hourly averaged profiler winds and snow-level heights. Of the 1970
hourly averaged snow-level heights that were derived from the 15-min RVD dataset, VDS could not be calculated for 640 (32%) and 299 (15%) heights at SEA and WPT, respectively, because of missing wind observations. Figure 19 shows a preference for RVD snow levels to occur during positive directional shear (i.e., geostrophic warm advection) both inland at SEA and at the coast at WPT. The mean positive shear is larger at SEA than at WPT (18.3 versus 11.3 m s$^{-2}$), likely because of increased terrain-blocked, low-level, southerly component flow inland from the coast in the pre-cold-frontal environment. Data from SEA likely better represent the wind structure over RVD, owing to its closer proximity and more similar terrain characteristics than at WPT.

Using the SEA VDS, the snow-level forecast discrepancies at SMP were stratified by forecast verification time and by snow-level height observed at RVD above and below the 1.25-km-MSL peak snow-level FOC at RVD (which also corresponds roughly to the altitude of the SMP gap east of RVD). Figure 20 shows that 69% of all forecast discrepancies occurred with warm-advective conditions, reflecting the fact that ascent and precipitation are typically most prevalent in pre-cold-frontal conditions where warm advection occurs. For all forecast times, 76% of the forecast discrepancies below 1.25 km (Figs. 20a,c) were associated with higher-than-observed snow-level forecasts with warm advection. One possible physical mechanism for this bias may be the underrepresentation or poor modeling of cold easterly gap flow underlying the stable, warm-sector air mass. For discrepancies >1.25 km MSL (Figs. 20b,d), the high forecasts during warm advection exist with a similar population density to those below...
1.25 km, although a second population of discrepancies was associated with low forecasts with warm advection, and this latter population contained members with considerably larger discrepancy magnitudes. These low-bias forecast discrepancies have the greatest adverse impact on flood forecasts, when observed snow levels are 1–2 km higher than forecast and expose an increased, unforecasted fraction of the watershed to rain and subsequent enhanced runoff. These low-bias discrepancies occurred almost exclusively when snow levels were observed above 1.25 km MSL, that is, above 90% of the terrain in the upper Green River watershed.

Additional analyses were conducted to compare snow-level forecast discrepancies with observed west–east mean sea level pressure (MSLP) differences across the CR between SEA and ELN (Fig. 1, Table 1). Two significant results emerge. First, for observed snow levels <1.25 km MSL (Figs. 21b,d), most have high-bias discrepancies, and 79% of those are associated with higher MSLP west of the CR. This MSLP distribution suggests the presence of cold low-level air west of the CR. One possible mechanism for the cold pool is radiational cooling on clear nights preceding the landfall of extratropical cyclones, resulting in cold-air generation in the valley (similar to that observed in California’s Central Valley; e.g., Neiman et al. 2006). Another mechanism, occurring during cyclone landfall, is down-valley drainage caused by diabatic cooling from evaporating rain and melting snow over the high terrain where orographic precipitation enhancement is favored (e.g., Steiner et al. 2003). A third mechanism is blocking of stably stratified, low-level, onshore flow by the Cascades, resulting in an adiabatically cooled pressure ridge damned against the windward slope (e.g., Braun et al. 1997). A fourth mechanism involves the onshore advection of cold, post-cold-frontal anticyclonic flow. It is unclear which mechanism(s) contributed to the observed MSLP versus snow-level forecast-discrepancy behavior. The second significant MSLP result is that for observed snow levels >1.25 km (Figs. 21b,d), 77% of the low-bias forecast discrepancies are associated with lower MSLP west of the CR. The largest of those discrepancies, which are the hydrologically most impactful, are linked to the greatest MSLP gradients >0. This MSLP distribution is in phase with synoptic pressure distributions typically associated with pre-cold-frontal conditions ahead of landfalling extratropical cyclones. The positive cross-barrier MSLP differences can be enhanced by cold continental air masses.
east of the CR and/or by a strong cyclone west of the CR (e.g., Fig. 13). Figure 21 demonstrates that the forecast process may misrepresent key physical processes that cause enhanced cross-barrier pressure gradients of either sign, which, in turn, adversely impact the snow-level forecasts.

6. Conclusions

During the 2009/10 and 2010/11 winter seasons, NOAA ESRL deployed two S-band precipitation-profiling radars in Washington: one along the coast at WPT and another on the western slope of the Cascade Range at RVD. These data, combined with historical datasets composed of raob, precipitation, and surface temperature and pressure observations, NWRFC forecasts, U.S. Geological Survey river flows, and 915-MHz wind-profiler data, were used to characterize the local and regional snow-level behaviors and their influence on the local hydrology. These datasets were also used to assess operational snow-level forecast biases and potential hydrological forecast implications. The results of this study support the following conclusions.

Vertical distributions of nontemporally matched radar and raob snow-level observations confirm that larger atmospheric scales dominate the composite snow-level behavior similarly from the coast to the base of the Cascade Range. This implies that a single precipitation profiler sited along the coast could provide high temporal snow-level information that could be used to track snow-level behaviors over the Green River basin for composite-dependent applications. In addition, lead-time analyses using event-matched cases observed by the coastal and inland radars suggest that coastal radars can provide meaningful lead-time information for inland snow-level characteristics during coherent precipitation events. Assessing the timing of post-cold-frontal snow-level drops is important, as it provides confidence that the flood threat in the Cascades might be abating because of an anticipated decrease in areal coverage of rain across the upper Green River basin. Similarly, assessing the timing of warm-frontal snow-level rises provides observational evidence that the flood threat is increasing.

Comparing radar-derived snow-level observations at RVD to corresponding NWRFC forecasts at SMP.
revealed altitude discrepancies resulting from both natural variability and forecast errors. A low forecast bias of 100–250 m was evident for all 6-h forecast times out to 48 h. The cause of this bias was related to 1) observed variations in snow level between the comparison sites, when maritime and continental air masses were dominant in different parts of the basin; 2) a forecast underrepresentation of the magnitude of warm-air advection aloft, as deduced by NARR composite analyses and thermal wind-derived temperature advection diagnostics from wind-profiler observations; and/or 3) a forecast overrepresentation of the westward intrusion of low-level cold air through the SMP gap to RVD, as inferred from observed surface temperature differences over short east–west distances in the upper Green River basin. There was a tendency for low-bias discrepancies between the forecasts and observations to increase in magnitude with increasingly higher observed snow levels, with the magnitude of the largest negative discrepancies exceeding 1 km. These discrepancies were often associated with coexisting cold continental air masses east of the Cascade Range and strong cyclones west of the Cascade Range, which acted to enhance the cross-barrier pressure gradient and the resulting cold, low-level easterly gap flow. The dynamical interactions between these maritime and continental air masses, in the presence of the complex topography of the Green River basin, pose significant challenges for numerical forecast models. Hydrologic-forecasting applications for this region would likely benefit from future research that targets improved understanding and model representation of these complex processes.

Given the often complex snow-level behaviors across the Green River basin, the use of a single-point, snow-level forecast has the potential to mispredict the hydrologic response within the basin. In this study, the largest precipitation accumulations and rates were most likely to occur when most or all of the Green River basin was receiving rain, which, in the context of historical streamflows, were the times when the largest hydrologic responses and greatest flood risk occurred. These results confirm the hydrology-based study of Neiman et al. (2011), which demonstrated that significant flooding (i.e., >5-yr return rate) did not occur in western

FIG. 21. Snow-level forecast discrepancy (km) as a function of MSLP differences (mb) across the CR, stratified by radar-derived snow-level height at RVD [(a),(c) snow levels ≤1.25 km MSL; (b),(d) snow levels >1.25 km MSL] and forecast verification time at SMP [(top) 0–24-h forecasts; (bottom) 30–48-h forecasts].

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Washington unless the snow level exceeded ~2 km MSL, which almost always occurred during AR landfalls. In short, one can assess flood threats almost exclusively from the snow-level altitude. The largest snow-level forecast discrepancies in this study demonstrate that the current use of the single-point, snow-level forecast at SMP will often misrepresent the snow-level across the remainder of the Green River basin, particularly when cold continental air flows westward through the pass. During these times, snow-level forecasts at SMP (subjectively verified by surface temperature) were up to 1 km below pass level, while snow levels at locations ~20–38 km to the west were observed at or above the altitude of SMP. These types of discrepancies were associated with forecasted snow levels that would have produced frozen precipitation over a large fraction of the basin area that actually received rain, thus resulting in potentially large hydrometeorological consequences.

In conclusion, precipitation-profiling radars could be further utilized to provide real-time, high-temporal-resolution information about rain-versus-snow distributions for flood monitoring, operational runoff forecasts, and water resource management. Siting of these radars higher in the Green River basin could allow for near-real-time bias corrections in model output. For applications such as distributed hydrologic modeling, radars within the modeled watershed may also be utilized to capture this potentially important variability for accurate runoff calculations. Also, the high temporal resolution of the radar snow-level data, which provides multiple samples per hour as opposed to two samples per day from the raob data, is ideally suited to meet the rapid refresh initialization demands of these higher-resolution models. In fact, California has already deployed much less expensive but similarly accurate snow level–detecting radars in 10 major watersheds in the coastal and Cascade ranges in Oregon and Washington. In the long term, these radars offer an opportunity to detect rain-versus-snow distribution changes, within individual watersheds, associated with climate change. These opportunities to improve our understanding of the physical processes that modulate the hydrologically important snow level in western Washington would significantly benefit emergency managers, water managers, and ultimately the regional population.

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