

## 2.5 Bright-band radar

The transition elevation where hydrometeors turn from frozen to liquid, or freezing level, was determined from analysis of hourly Doppler-radar data from wind profilers located upstream of the LiDAR-acquisition area. Radar reflectance is greatest, or brightest, in the altitude range where precipitation changes from snow to rain, owing to a difference in the dielectric factor for water and ice and the aggregation of hydrometeors (White et al., 2009; Ryzhkov and Zrnic, 1998). We analyzed bright-band altitudes and thus identified freezing levels from observations collected over the 2010 water year snow-accumulation period (12/3-3/23) from the three nearest upwind locations, i.e. Punta Piedras Blancas, Lost Hills, and Chowchilla, California (data available at: <http://www.esrl.noaa.gov/psd/data/>, 2014) (Fig. 1).

## 4 Discussion

The overall increase in precipitation with elevation observed with airborne LiDAR is consistent with the orographic effect of mountains on precipitation (Roe, 2005; Roe and Baker, 2006). Variability of the snow accumulation along the elevation gradient and deviations from a regular increase with elevation can be attributed to the interactions of topography, wind and storm tracks. Deviations from a linear increase are apparent at the lower rain-snow-transition elevations, at higher elevations near the ridge, and at intermediate elevations that have a variety of aspect and steepness characteristics.

### 4.1. Variability of orographic trends

Elevations over 3300 m showed the greatest negative departure from the overall orographic trend, likely due to the southwest-to-northeast trending terrain flattening out and no longer providing the necessary lift for the same rate of adiabatic cooling (Figs. 2 and 4). Above 3300 m, the reduced lift over flatter terrain and an exhaustion of precipitable water as storms rise less steeply result in declining snow depths at the higher elevations (Houze Jr., 2012). Individual storms can give different orographic precipitation patterns, particularly near the upper elevations in a mountain basin, depending on the direction, size and overall moisture content of the air masses (Alpert, 1986).

As other researchers have noted, it is also difficult to identify the effects of specific storms on snowpack ablation due to the variability of atmospheric conditions close to the earth's surface (Lundquist et al., 2008). The extent to which high-altitude Sierra Nevada catchments receive more precipitation than adjacent low-altitude areas varies from storm to storm, and from year to year, from occasions during which nearly equal amounts of precipitation fall at high and low altitudes to occasions when 10 or more times as much precipitation falls at the higher altitudes (Dettinger et al., 2004). In the northern Sierra Nevada, the blocking and associated terrain-parallel southerly flow of air masses, referred to as the Sierra barrier jet, can influence enhanced lower versus higher-elevation precipitation (Neiman et al., 2008).

While the particular orographic patterns reported here could be unique to the 2010 water year, previous works have shown some consistency in the interannual spatial patterns of snow

accumulation (Sturm and Wagner, 2010; Deems et al., 2008). However, in the Central Sierra Nevada, it has been observed that seasonally, the ratio between higher- versus lower-elevation annual winter-season total precipitation averages about 3, but in some years, the ratio drops to as low as 1 (as in 1991) or rises to as much as 4 or 5 (Dettinger, et al., 2004).

## 4.2. Wind redistribution and radiation effects

While wind affects snow accumulation during a storm, the combined effects of wind and radiation are apparent in post-depositional changes in snow depth. As the same topographic variables influence both wind and radiation, separating the effects based on an analysis of the snow-depth data is challenging.

The high-spatial-resolution of LiDAR snow-depth measurements point to two possible controls of wind redistribution on snow. While wind patterns from a single station may be a poor indicator of the wind fields influencing snow redistribution across the entire domain, we expect snow transport by wind to be coarsely defined by the consensus of the local station's wind direction when temperatures are below zero within 24 h of a snowstorm (Figs. 1 and 3). However, the Topaz Lake station, located in smooth terrain with limited upwind influence, may best represent the wind patterns of the free atmosphere and predominant southwest storm winds. We attribute the inconsistent wind direction of other stations to the terrain induced turbulence of the free atmosphere upwind of the stations. The M3 and Emerald Lake sites have upwind obstacles, and the Wolverton and Panther stations have low wind speeds, reflecting the muting effect of tall forest cover on wind speed and consequently snow redistribution (Fig. 3).

Consistent with prevailing winds from the southwest, we observed more accumulation on the northeast slopes and less on the southwest; however, in our domain northeast has the least total area of all aspect quadrants and hence these areas may be underrepresented in the analysis (Fig. 2c).

The aspect intensity variable (*IA*) combines the influences of slope and aspect, and serves as a proxy for several processes affecting snow depth, e.g. radiation, upslope orographic deposition and potentially wind and gravitational redistribution. As a result some local anomalies, such as deep-snow-patch development, are likely masked when considering topography and snow depth as elevation-band means.

Examining residuals from a linear orographic trend by *IA* suggests that the steeper, northwest-facing slopes at the mid elevations and northerly slopes at the lowest elevations show the greatest snow depths, likely due to the combined effects of wind deposition and lower radiation influx (Fig. 4c and d). Conversely, low- to mid-elevation slopes prone to the combined effects of ablation and wind erosion have the least snow. These findings suggest that departures from the overall orographic trend can be observed in the elevation profile using *IA*; but there are limitations to the approach as used here.

It is also possible that there is limited utility in extrapolating prevailing winds from meteorological stations to predict effects of wind on snow redistribution because of the turbulence from local terrain. Research into the relationship between slope, aspect and wind has

revealed that small-scale slope breaks and surface roughness have the most-significant effects on where snow accumulates locally (Li and Pomeroy, 1997b; Winstral et al., 2002; Fang and Pomeroy, 2009; Pomeroy and Li, 2000). While not part of this analysis, classification of downwind terrain has been effective for identifying snow-patch development and persistence of localized wind deposition, offering a deterministic explanation for the spatial stationarity of snow (Winstral et al., 2002; Schirmer et al., 2011). The *IA* variable may also be effective for classifying locations where these processes are likely to occur.

### **4.3. Sublimation**

Wind-driven sublimation may also play a role in the departure from the linear increase in snow depth at the higher elevations, where the highest wind velocities and thus greatest suspension of snow occur (Figs. 3).

In dry intercontinental locations, sublimation rates can be in excess of 50 %, but are much lower in the maritime climate of the Sierra Nevada and lowest during the accumulation period (Ellis et al., 2010; Essery and Pomeroy, 2001). Studies conducted at 2800 and 3100 m in the Emerald Lake basin, located in the center of our measurement domain, found net losses due to evaporation and sublimation of <10% for the period between 1 December and 1 April (Marks and Dozier, 1992; Marks et al., 1992). Consequently, we consider the 2010 water year cumulative loss due to sublimation and snowmelt to be limited (< 10 %) prior to the March 23<sup>rd</sup> LiDAR acquisition at all elevation bands, with more melt occurring at the lowest elevations and on the southeast-facing slopes, as indicated by the loss of SWE measured at the low-elevation snow-pillow sites and reduced snow depths on the southeast mid-elevation slopes (Fig. 2c, 6, and 7).

### **4.4 Rain–snow transition**

At lower elevations, e.g. below 2050 m, a mix of rain and snow precipitation appears to influence the amount of seasonal snow accumulation. Local SWE measurements are only available at one location below 2050 m (GNF); and this station does show a very small loss of SWE in mid-February as a result of a rain-on-snow event (Fig. 6). Nevertheless, given the expected large storm-to-storm variation in freezing level, the relatively sharp transition in slope of LiDAR-measured snow accumulation at about 2050 m suggests most precipitation above this elevation fell as snow in the winter 2010.

In addition, seasonal snow at the lowest elevations and on south-facing slopes has greater positive net energy exchange (from radiation or condensation), and is most susceptible to melt during the accumulation period. LiDAR snow-depth results show lower depths on south-facing versus greater depths on north-facing slopes (Fig. 2c).

### **4.5 Snow density**

Our March 23<sup>rd</sup>, calculations of snow density based on snow depth and snow pillow measurements are uncorrelated with depth or elevation and varied < 11 % from the mean, which is within the combined uncertainty of the sensors used to calculate them (Figs.1 and 8). Thus we

consider the mean of our density calculations to be a reasonable estimate of the densities across the elevation gradient of our study.

Elder et al. (1998), Anderton et al. (2004) and Anderson et al. (2014) also found the variability of spring snow density to be insignificantly correlated with elevation in their studies, while Zhong et al. (2014) found negative correlations with elevation in the former USSR. A range of results has also been reported for the snow density correlation with depth showing both positive and negative correlations depending on the age of the snow and season (McCrieight and Small 2014). The likely explanation for these seemingly contradictory findings is due to the seasonal and climatic interactions between snow depth, the snowpack energy balance and snow density. Snow-depth is positively correlated with elevation and the energy balance of the snowpack can be negatively correlated with elevation, and each varies in magnitude and affect on snow density with season and climate (Jonas et al., 2009; Sturm et al., 2010b). For example in winter, when there are low levels of solar influx on low albedo snowpacks, snow depth, which is positively correlated with elevation, has a greater effect on density. Conversely, in springtime, or in a warmer climate, a warming snowpack may reverse any previous correlation or be uncorrelated with elevation. Thus our assumption of uniform density may not be accurate for early winter but is a reasonable estimate for spring a snowpack near peak accumulation and we expect these LiDAR snow depth measurements and estimates of cumulative seasonal precipitation to approximate each other.

#### **4.6 Other measures or orographic trends**

Although orographic precipitation is a well-documented first-order process, in the Sierra Nevada it is not well described at the watershed to basin scale owing to the very limited availability of ground-based precipitation measurements. Each set of comparative measurements used in this study provides a different index of orographic response: (i) LiDAR is a one-time snapshot of snow depth; (ii) point SWE data are small samples from highly variable spatial values, (iii) reconstructed snowmelt, or retrospective gridded SWE, reflect precipitation minus evaporation and sublimation; and (iv) PRISM is a retrospective precipitation estimate, based largely on lower-elevation stations. Nevertheless these complementary data offer spatially relevant indices of seasonally accumulated precipitation.

As Fig. 8 shows, snow depths from snow-pillow sites fail to capture the elevation patterns apparent in the LiDAR data. This pattern is also apparent in the SWE values from the same sites (Fig. 7b). While the shallowest depth is registered at the lowest elevation site (GNF, 2027 m), where a greater percentage of precipitation falls as rain, the other sites do not show a consistent increase in depth with elevation. Thus current operational measurements in the Sierra Nevada are insufficient to capture orographic trends in snow depth and precipitation.

The less-steep increase in precipitation with elevation seen in the two PRISM profiles versus the LiDAR results are thought to be primarily due to the limited number of mountain stations used to calculate the PRISM trends. SWE loss from ablation and rain versus snowfall are important components of the observed LiDAR lapse rates at lower elevations, particularly below 2050m; these processes should have only a small influence above that elevation. Evidence for this can be

seen in three locations of coincident SWE and cumulative precipitation measurements (Fig. 6). The accumulated SWE and total precipitation at the two higher-elevation stations, CRL and QUA, are in close agreement; and the lowest station, GNF, shows 21cm more total precipitation and slight loss of SWE on 18<sup>th</sup> - 23<sup>rd</sup> March, prior to the date of LiDAR acquisition, demonstrating that measurable rain and melt occurred at the site. In addition, precipitation station in the Kaweah basin near the LiDAR footprint (LDG, 2053 m) had an accumulation-period total of 72 cm, higher than the LiDAR SWE estimate and lower than both PRISM estimates for the same time period (Fig. 9). The difference in annual precipitation at these sites versus annual SWE accumulation reflects in part the contribution of both rain and snow and mid-winter melt at this elevation. Thus, divergence of the PRISM and reconstructed SWE at elevations below 2200m is expected. Temperature records in the area suggest only a small amount of winter melt at 2100m, with very little winter melt and precipitation as rain above 2400m (Rice and Bales, 2013).

The general pattern of SWE reconstructed from snowmelt by Guan et al. (2013) compares well with the LiDAR data, being somewhat higher at the highest elevations, lower in the mid elevations, and similar at the lower elevations. Even though the reconstruction was based on energy-balance modeling, the good match is somewhat surprising given the coarseness of the reconstruction model relative to the complex topography of the basin.

The Rice and Bales (2013) reconstruction, in which snowmelt was indexed to amounts and rates at the snow-pillow sites, has less SWE, particularly at the mid to lower elevations. This offset may stem in part from the higher 106% of average 2010 seasonal precipitation versus 90% of average precipitation in the 2000–2009 snowmelt reconstruction period. Further, the reconstructed SWE estimates by Rice and Bales (2013) are based on a temperature-index calculation, versus a full energy-balance approach by Guan et al. (2013).

Also, some offset in both reconstructed SWE estimates may reflect a bias in snow-covered-area estimates, which have a 500-m spatial resolution and are heavily influenced by canopy. That is, the LiDAR data represent open areas, and the reconstructed SWE values represent the full domain, but are empirically corrected for canopy.