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WSR-88D PERFORMANCE IN NORTHERN UTAH DURING THE WINTER OF 1998-1999. PART-II: EXAMPLES OF ERROR SOURCES

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[Note: Because of the large number of figures, only the text will be published in hard copy. The figures can be accessed on the Web version at http://www.wrh.noaa.gov under Technical Attachments.

1. Introduction

This is the second in a two-part Technical Attachment (TA) regarding the ability of the WSR-88D to estimate precipitation during winter storms. Part I of this study (Vasiloff 2001) discussed general corrections to radar snow water equivalent (SWE) estimates from 23 days of data during the winter of 1998-1999. Radar estimates were derived by the U.S. Bureau of Reclamation snow accumulation algorithm (SAA) using a Z-S relation $Z = 75S^2$. A network of snow gauges (Fig. 1) was used to determine the corrections. It was shown that overall, the radar underestimated snowfall in the mountains and overestimated snowfall upstream of the mountains. Radar estimates were fairly accurate along the foothills. However, there were significant variations among the individual cases. Reasons for the variations include evaporation, beam blockage, overshooting, and the bright band. This TA shows examples of the error sources that occurred during three cases. In addition, instances of good radar performance are described.

2. Case Examples

As mentioned in Part I, performance is largely tied to where in the storm the radar is scanning. A schematic of a radar beam scanning a storm with a melting level somewhere above the surface is shown in Fig. 2. The radar is sampling rain at near ranges and at farther ranges it is sampling snow. In between the rain and snow is the bright band where the radar reflectivity values are exaggerated because of melting precipitation. Note that, depending on the height of the 0 C level, the radar may be sampling all snow, all rain, or a combination of the two. For snow measurement, the ideal situation is for the entire atmospheric column to be sub-freezing so the radar is sampling snow everywhere. Obviously, precipitation estimates from the bright band are nearly useless.

2.1. Radar Over-Estimation

Two cases are used to illustrate radar over-estimation. In both cases, the melting level is somewhere above the ground. The radar is over-estimating precipitation because of evaporation and sampling in the bright band.

2.1.1. Evaporation Case of 15 January 1999

On 15 January 1999, radar storm total SWE estimates upstream (west) of the Wasatch mountains were much greater than gauge measurements (Fig. 3). The radar estimates were much closer to gauge measurements in the mountains. For example, the radar estimated .13 in over SNL but only .02 in were measured by the gauge. A similar discrepancy is seen at SNZ and SNH. The 1-hr SWE total (1600-1700 UTC) from the SAA is shown in Fig. 4. The 1-hr estimate for SNL was .06 in while the actual gauge catch was .02 in. A loop of reflectivity from the 0.5 deg sweep between 1600 UTC and 1700 UTC (Fig. 4) shows that echo intensities over SNL ranged from 22 to 25 dBZ. Recall from Part I that 25 dBZ is equivalent to ~.08 in h^{-1}, according to the relation $Z=75S^2$. The 1200 UTC 15 January and 0000 UTC 16 January soundings (Fig. 6) reveal the reason for radar over-estimation. Though the storm occurred between the two sounding times, both soundings were dry near the ground, indicating the potential for evaporation, especially the 1200 UTC sounding which was taken only 2 hours before the echoes formed. Surface data at SNL also indicated low relative humidity during the storm, ranging between 50 percent and 80 percent (Fig. 7). Note that the temperature remained well above freezing, indicating liquid precipitation. Thus, blowing snow can be eliminated as a potential gauge under-catch problem.

3.1.2 Bright Band Case of 3 May 1999

On this day, the radar over-estimated precipitation because of the bright band. The 0000 UTC 3 May 1999 sounding (Fig. 8) shows that the height of the 0 C level was just below 700 mb (~10000 ft MSL). The sounding also shows lower relative humidity near the ground, indicating potential evaporation. Figure 9 shows radar and gauge SWE estimates. The radar values are much greater than gauge measurements for sites upstream of the Wasatch mountains. Note that the radar estimates are fairly similar to gauge measurements at the mountain sites with the exception of SNI where the 1.4 deg beam is being used.

Reflectivity from the 0.5 deg sweep at 0027 UTC (Fig. 10) shows enhanced reflectivity (yellow and orange areas) in the bright band over the gauges that are within 60 km range and upstream of the Wasatch mountains. Note that the enhanced area does not extend over SNH where the radar estimates were similar to the gauge measurements. A vertical cross section along the line in Fig. 10 is shown in Fig. 11. The thin bright band is obvious, especially within 10 n mi range of the radar where the 0.5 deg sweep was beneath the bright band and indicated a decrease in reflectivity below the bright band. The height of the bottom of the 0.5 deg beam can be seen increasing with increasing range and the decrease in reflectivity below the bright band is cut off beyond 10 n mi range. More accurate radar estimates would be expected below the bright band.

2.2. Under-Estimation Case of 10 February 1999

This event was well-defined synoptically with a sharp frontal passage and heavy post-frontal snow. Figure 12 shows a time series of surface data for SNH. Note the wind shift and rapid temperature drop just after 0600 UTC. Temperatures at the surface and aloft (not shown) were at or below freezing during the time period for which the radar and gauge data are compared. Water-equivalent snowfall rates between .1 in h^{-1} and .2 in h^{-1} were common over a broad area with some valley locations getting 8-10 in of snow in a 5 to 6-hr period. Storm total radar estimates and gauge measurements were very similar at SNW and SNL (Fig. 13). However, the radar underestimated the gauge measurements by 30 percent - 50 percent at the other sites.

A 3-hr radar precipitation field (Fig. 14) shows the horizontal variation of snow between 0711 UTC and 1010 UTC. There is a band of precipitation that stretches from northeast of the radar to south of the radar. There is also a maximum in precipitation southeast of the radar between 40 n mi and 50 n mi range. Peak values of SWE estimates are between .25 in and .3 in. The 0.5 deg sweep at 0749 UTC (Fig. 15) shows the typical reflectivity structure. Highest reflectivity values are just east of the radar. A band of echo is aligned from northeast through south of the radar and corresponds to the band seen in Fig. 14. There is also a secondary reflectivity maximum over SNH.

It is difficult to determine the exact cause of under-estimation, but several possibilities are suggested. First is the problem of overshooting by the beam. A schematic of the 0.5 deg radar beam (Fig. 16) illustrates this effect. The horizontal line at 9000 ft MSL represents the upper limit of the peak reflectivity in the storm. It follows that the beam over SNH is missing most of the storm while well-sampling the part at closer ranges (e.g., at SNL). A vertical cross section at 0749 UTC (line AB; Fig. 17) shows that the reflectivity seems to lose definition at farther ranges. For example, note the detailed structure and strong vertical gradients within 30 n mi range and the rather flat echo appearance beyond 30 n mi. The widening of the beam with range may contribute to underestimation by smoothing out details in the echo structure.

Another possible contributor to under-estimation is horizontal advection. Note that in Figs. 14 and 15, SNH is on the southern edge of an area of enhanced precipitation. Even though reflectivity in the echo overhead was weak, northerly winds along the mountains suggest that snow could have fallen undetected below the 0.5 deg beam, and arriving at SNH having originated in the enhanced area to the north.

Reflectivity cross sections AC and AD (Figs. 18 and 19, respectively) show why precipitation was severely underestimated at SNI. Section AC cuts through echo that is not blocked at 0.5 deg elevation angle. Beyond 25 n mi range, the echo tops are lower than at closer ranges. Line AD is through the region over SNI that is blocked by Mount

Ogden and the reflectivity pattern ends abruptly just beyond 25 n mi range. As a result, the 1.4 deg sweep is used for precipitation estimation. However as seen in cross section AC, the echoes in the mountains do not extend high enough to be adequately sampled by the 1.4 deg beam. This has implications for the potential use of a vertical profile of reflectivity to adjust reflectivity at far ranges or in areas of beam blockage; the vertical structure of storms in complex terrain varies according to proximity to mountains.

3. Summary

This TA describes sources of errors in radar precipitation estimates during winter storms in Utah. The accuracy of the estimates is largely tied to where in the storm the radar is sampling. Elevated melting levels produce a radar bright band which results in over-estimates of precipitation. Evaporation below the radar beam can also cause over-estimation. On the other hand, beam blockage and overshooting the storm top results in under-estimation. Horizontal drift of snow in storms with strong reflectivity gradients can also cause problems. Ways to compensate for these effects were described in Part I of this study (Vasiloff 2001). Real-time assessment of how the radar is scanning the storm relative to the melting level and complex terrain is crucial to interpretation of radar data.

References

Vasiloff, S. V., 2001: WSR-88D performance in northern Utah during the winter of 1998-1999. Part-I: Adjustments to precipitation estimates. Western Region Technical Attachment No. 01-01.