WESTERN REGION TECHNICAL ATTACHMENT
NO. 01-02
March 20, 2001

WSR-88D PERFORMANCE IN NORTHERN UTAH DURING THE WINTER OF 1998-1999. PART-I: ADJUSTMENTS TO PRECIPITATION ESTIMATES

Steve Vasiloff - NSSL - WRH-SSD, Salt Lake City, UT

[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 <u>http://www.wrh.noaa.gov</u> under Technical Attachments.

1. Introduction

The National Weather Service (NWS) WSR-88D is a primary tool for remote estimates of precipitation. The WSR-88D Precipitation Processing Subsystem (PPS; Fulton et al. 1998) performs many functions to provide precipitation estimates. Many functions are related to data filtering and quality control. A key component is the equation used to estimate precipitation rates (R; mm h^{-1}) from the radar reflectivity (Z; mm^6 m^{-3}) values. This equation is widely referred to as the "Z-R relation" which is of the form $Z = aR^b$. The WSR-88D PPS has a default relation $Z = 300R^{1.4}$ for convective storms. The relation $Z = 250R^{1.2}$ is recommended for tropical storms.

The wide variety of microphysical processes responsible for precipitation means that no single Z-R relation works all of the time. For this reason, the WSR-88D Radar Operations Center (ROC; formerly called the Operational Support Facility) recently authorized WSR-88D sites to use three new Z-R relationships for cool season stratiform rain events with the relation $Z = 75R^2$ recommended for sites west of the continental divide (OSF memorandum 1999; Super and Holroyd 2000). The cool season includes snow events and the variable "S" can be substituted for "R" with the provision that "S" is in snow water equivalent (SWE) rate, also in units of mm h^{-1}. Vasiloff (1997) found that the coefficient of 75 worked well for an event at Snowbasin ski area in Utah. Barker et al. (2000) used the same relation in a test of the U. S. Bureau of Reclamation snow accumulation algorithm (USBR SAA).

This Technical Attachment describes an effort to understand and provide gross corrections to WSR-88D precipitation estimates for shallow cool season precipitation (CSP) storms in northern Utah. The SWEs estimates from the WSR-88D on Promontory Point are compared to accumulations from a network of eight snow gauges (Snownet) in and along the northern Wasatch mountains. The radar-based estimates were derived from the

U. S. Bureau of Reclamation snow accumulation algorithm (SAA) using the relation $Z = 75R^2$. Data from 23 cases during the winter of 1998-1999 are examined. Results are related to gauge location, radar scanning characteristics, and various storm parameters.

Section 2 provides the setting for the study, including local topography and the Snownet layout. Potential error sources are also discussed. Data analysis methodology is described in Section 3 and results are given in Section 4. Finally, recommendations for forecasters are discussed in Section 5 and a summary is provided in Section 6.

2. Setting

Figure 1 shows the layout of the Snownet and the KMTX WSR-88D. The gauges used in this study are situated in and along the Wasatch mountains to the east of the Provo-Salt Lake City-Ogden corridor. Note that the majority of the Snownet is southeast of the Great Salt Lake (GSL) which experiences lake-effect snow under certain conditions - mainly northwest flow and a strong vertical temperature gradient. The KMTX radar is located on Promontory point, 2002 m above sea level (ASL). Snownet altitudes and ranges from the radar are given in Table 1. Heights of the center of the 0.5 deg elevation angle radar beam as well as the beam's diameter above each gauge are also given. Because of beam blockage, the 1.4 deg elevation angle is used for Snowbasin and Deer Valley. In the absence of complex terrain, greater distances between the radar and gauge mean greater beam heights and diameters relative to the gauge. In complex terrain, the beam can be closer to the ground with increased distances. An example is shown in the schematic diagram shown in Fig. 2 with the 0.5 deg elevation angle beam pattern shown over Bountiful and Park City. Even though Park City is farther away from the radar than is Bountiful, Park City is actually closer to the beam. However, the beam is wider over Park City.

The biggest challenge for the radar in CSP is related to where the beams are sampling the precipitation. Obviously, if the beam is over-shooting the storm top, nothing is measured. Preferred is a narrow beam sampling just above the ground. The trick is to understand where the radar beam is and interpret the data accordingly. For example, radar samples of melting precipitation, i.e. in the bright band, have little or no correlation with gauge accumulations. Samples in the upper part of a storm may also have little correlation to surface measurements, especially if there is melting/rain occurring below the beam. Beam blockage by terrain also affects radar-gauge comparisons. If the beam at 0.5 deg elevation angle is partially or fully blocked, then the next highest elevation angle is used, in which case over-shooting is even a bigger problem.

Comparison between radar and gauge precipitation estimates is further complicated by the vast difference in the respective sampling volumes. For example, the volume of the radar beam for reflectivity sampling over Salt Lake City is a disk of dimensions 1133 m wide by 1000 m deep (velocity range bins are 250 m apart). The gauge is essentially sampling a point in space. Snow falling from the radar beam volume may travel some distance horizontally while falling toward the ground. Assuming a 1 m s^{-1} fall speed and a 10 m s^{-1}

uniform horizontal wind, a snow flake starting at 2 km above the ground (AGL) will move 20 km horizontally and take 33 min. For slower fall speed, the problem is worse. Besides melting, snow flake aggregation, break-up, riming, and other processes can change the character of the radar echoes and snowfall beneath the beam. The gauge itself is especially prone to under-catch in strong winds. A 4 m s^{-1} average wind can cause a 20 percent under-catch, even with the standard Alter shield (Fig. 3). Snow may stick to the side walls and certain types of gauges can become completely capped by snow. Side-wall snow will fall into the gauge at a later time not getting measured when it actually fell.

A special field experiment called the Intermountain Precipitation Experiment (IPEX) was conducted during February 2000. Among the special sensors deployed during IPEX was a vertically-pointing S-Band Doppler radar (10 cm wavelength) located at the base of Snowbasin ski area. Data from the S-Band on 12 February 2000 provide an example of how KMTX samples storms at Snow Basin (Fig. 4). The location and diameter of the 1.4 deg elevation angle beam is shown since the 0.5 deg beam is blocked by Mt. Ogden. In this case, the 1.4 deg sweep is sampling the upper part of the storm and the KMTX reflectivity values are qualitatively similar to the S-Band data. Furthermore, the S-Band shows that the echoes extend from the KMTX beam to the ground. Thus, a good match between the WSR-88D (numbers along the dashed line) and surface measurements is expected. However, there is some offset between reflectivity data and gauge accumulations. The lesson here is that data should be integrated over longer time periods to reduce uncertainties. Eventually, nearly all of the snow will be collected in the gauge and a reasonable comparison between radar and gauge storm totals can be made.

3. Data and Analysis Methodology

Data collected by the KMTX WSR-88D and the Snownet on 23 days during the winter of '98-'99 are used in this study. A variety of storms are in the data base and include mountain snow/ valley rain cases, as well as all snow cases. For most of the time, the radar beam was above the melting level.

The USBR SAA was run in the WSR-88D Algorithm Testing and Analysis System using the relation $Z=75S^2$. The SAA range correction was NOT applied to data used in this study to see if range effects could be identified. One-hour SWE amounts from the SAA were taken directly over the gauges. The SWE estimates are derived from reflectivity data from the 0.5 deg elevation angle sweep except where the beam is blocked. If more than 60 percent of the beam is blocked, data from the 1.45 deg sweep are used (the PPS uses a minimum of 50 percent blockage). The SAA reads a data file containing information on which elevation angle to use at each azimuth-range bin. The resultant reflectivity field may then contain data from different sweeps. This is referred to as a hybrid scan. The SAA hybrid scans are different from the PPS hybrid scans in that the PPS creates a constant-elevation reflectivity field from different sweeps and then, like the SAA, uses a topographic look-up table to determine where to use different tilts. Other details of the PPS processing can be found in Fulton et al. 1998.

Each volume scan the reflectivity data are converted into SWE rates. This is done by solving for S as follows:

$$Z = aS^b,$$

$$\log Z = \log a + b\log S,$$

$$10\log Z = dBZ = 10\log a + 10b\log S,$$

finally,

$$S=10^{[(dBZ-10\log a)/10b]}.$$

With "a" and "b" equal to 75 and 2, respectively, the resultant expression is

$$S=10^{[(dBZ-18.8)/20]}. \tag{1}$$

For example, a dBZ value of 25 results in an S of 2 mm h^{-1} (~ 0.08 in h^{-1}). The rate is multiplied by the volume scan time (5 or 6 min) and each volume scan SWE is added up over an hour. Gauge accumulations were recorded at 5-min intervals and were also integrated over an hour interval.

4. Results

A plot of totaled radar and gauge accumulations for each site is shown in Fig. 5. The SNW and SNX amounts are low because of missing data from those sites. The important feature in the plot is the comparison between the radar and gauge estimates. Overall, the radar underestimated precipitation at the mountain sites (SNB, SNC, and SNV), was fairly close to actual precipitation near the mountains (SNZ, SNH, and WSU), and overestimated precipitation west of the mountains (SNX and SNL). The data can be used to make a "seasonal" adjustment to output from the SAA and could be done site-by-site or according to proximity to the mountains.

Figure 6 shows storm total radar estimates versus gauge measurements for all sites and cases. In a scatter plot like this, if the relationship between the radar and gauge estimates was perfect, i.e., using $Z=75S^2$, all of the points would lie along a straight line. Here the linear fit is nearly one-to-one, indicating that $Z=75S^2$ is valid **overall**. However, there is a lot of scatter indicating quite a bit of error for individual cases.

Scatter plots of 1 hour precipitation estimates for each site are shown in Figs. 7a-14a. The amount of scatter varies considerably from site to site. Snow Basin (SNI) and Park City (SNC) show the best correlation between the radar and gauge, while Weber St. (SNW) shows the poorest skill. The plot for Antelope Island (SNX) indicates a great deal of radar over-estimation. The linear fit is good but there is concern about the exposure of the site to strong winds as there are no trees around the gauge, thus making the gauge values questionable.

Plots of storm totals for each site show much less scatter (Figs. 7b-14b), although the linear fits are not much different than for the 1-hr amounts. Again, plots for Snow Basin and Park City indicated the best results, while the Weber St. plot showed the poorest results. Even though the Antelope Island regression seems to show skill, the results are not used owing to suspect data. Overall, these results indicate that there tends to be more uncertainty in the 1-hr measurements compared to storm-total measurements. In fact, the radar-gauge storm total comparison for only the mountain sites shows the best correlation (Fig. 15).

5. Recommendations

What do these results mean for NWS operations? Recall that the WSR-88D ROC has allowed forecast offices to use various Z-S relations. It appears that respectable WSR-88D estimates are possible by setting the current PPS Z-S to $Z=75S^2$ and using a rough correction factor provided by the regression equations derived from the scatter plots. First, it is useful to establish expected SWE rates from a given reflectivity. Table 2 shows selected reflectivities and SWE rates using $Z=75S^2$. Differences among precipitation rates are small for reflectivity less than 20 dBZ. As reflectivity increases, the change in rates gets larger. It is left up to the reader to determine what a significant rate is since official definitions of snowfall rate is related to visibility. Table 3 shows corrections and offsets to SWEs from the regression equations for the different sites. Overall, radar SWE estimates for the mountain sites need increases by a factor of ~50 percent. Radar estimates for the airport and Sandy gauges need little correction. The large difference between corrections for Bountiful and the airport is interesting since the sites are only 10 miles apart. However, reflectivity gradients are extreme near the mountains and in different snowfall regimes and thus, different corrections are expected.

How should the different corrections be used? I suggest, after applying the $Z=75S^2$ relation, multiplying radar SWE estimates by 1.5 in the mountains and 1.0 along the base of the mountains. A factor less than one upstream of the mountains MAY be warranted but the current data do not support a specific number at this time. In between the 1 and 1.5 multiplicative factors, there could be a linear transition. Additional data must be examined to validate these corrections.

The corrections mentioned above apply to stratiform snow events that may have partial melting near the ground at valley locations. Often there are different mesoscale regimes embedded within a synoptic-scale system. Frequently, a synoptic-scale low/trough will have a sector with warm-air advection, frontal passage, and cold-air advection behind the front. Even in the heart of winter in Utah, there is often convective precipitation in the warm sector. Thus, I recommend use of the default $Z=300R^{1.4}$ relation ahead of the surface front and $Z=75S^2$ behind the front. Currently, there is no way to operationally apply two different Z-S relations so forecasters will have to choose one depending on the forecast emphasis. However, the National Severe Storms Laboratory has developed an algorithm that identifies convective vs. stratiform elements for the application of different Z-S relations and will hopefully work its way into the WSR-88D (see the web site www.nssl.noaa.gov/teams/western/qpe).

A tool that would have immediate operational impact is a real-time image showing the height of the hybrid scan relative to the bright band or melting level. The application would require input of at least the 0 C height. An example of such an application is shown in Fig. 16. The top schematic shows the centers of the 0.5 deg and 1.4 deg beams on the right. The height of the 0 C level is dashed. The difference between the beam height and the 0 C height is then color-coded. For example, the red ring indicates that the beam height is within 0 and 500 ft of the 0 C level. Note that the 0.5 deg is blocked on the right and the height of the 1.4 deg beam over that location is between 1500 and 2000 ft above the 0 C level. There is also a sliver of purple where the beam is within 1000 and 1500 ft of the 0 C level. The green ring indicates that the beam is 0 - 500 ft **below** the 0 C level. Note that looking at the colored rings one cannot tell which elevation angle is being used (recall that the PPS uses the lowest 4 tilts). The important thing is that it shows which range bins are above/below the 0 C level, as well as the height difference. Remember that these range bins are the ones from which the precipitation estimates are derived. Thus, the forecaster would instantly know where the precipitation estimates are suspect due to radar beam proximity to the bright band.

Finally, recipients of a "nowcast" are often concerned with how much it will snow. Thus, the conversion of SWE values to snow depth is an important issue. The USBR refers to the ratio of snow depth-to-SWE parameter as "FLUFF" in their SAA. (Fluffy/rare powder is an attraction for skiers and snowboarders and Utah is world-renowned for its powder.) Figure 17 shows 24-hr snow depth and SWE measurements at Alta, Utah, for 31 days during the winter of 1998-1999. Alta is at an elevation of 8800 ft ASL and is ~10 miles east of Sandy. Measurements are made by snow safety personnel twice daily at Alta as well as other ski areas. Snowfall amounts less than 4 in were not considered in this study. The chart represents a total of 348 in of snowfall. The maximum snowfall was 24 in with nearly half (15) of the events measuring 10 in or greater. A ratio of depth-to-SWE is 10-to-1 if the depth and SWE bars are of equal length. Note that most of the deeper snow events have ratios well over 10-to-1, indicating fluffier snow. Figure 18 shows 24-hr snow depth and the corresponding depth-to-liquid ratios. Ratios varied between 5 and 26, with a mean of 12.3 with a median value of 11.4. Depth-to-SWE ratios for each 24-hr period were also plotted as a function of mean temperature derived from the previous 12 hr maximum and minimum (Fig. 19). There is a tendency for smaller ratios (denser snow) to occur at warmer temperatures and higher ratios to occur at colder temperatures. However, a significant amount of events occurred at temperatures between 20 and 25 deg F where there does not seem to be much of a correlation between snow depth-to-SWE ratios and temperature. This is reasonable since the surface temperature does not necessarily reflect the actual crystal growth temperature, super-cooled water supply, vertical velocity, and wind regimes that dictate snowflake density.

6. Summary

Radar and snow gauge data for 23 cases during the winter of 1998-1999 in northern Utah were examined to understand radar performance and potential corrections for winter storms. Gauge sites were located in and along the Wasatch mountains. The USBR snow accumulation algorithm was used with the Z-S relation $Z=75S^2$. A range correction was not used.

Results indicated a good match between radar and gauge estimates for the mountain sites given an additional correction of ~50 percent. Radar estimates were fairly close to gauge estimates along the base of the mountains with little correction needed. One-hour estimates had more scatter than multiple-hour totals indicating possible problems with instantaneous gauge measurements.

It is important to understand that the results presented herein are simply guides based on limited data. Examination of data on a case-by-case basis showed a lot of variability in the radar estimates. Also, it is believed that much of the noise in the data set is from mixing data from all-snow cases with data from "snow in the mountains, rain in the valley" cases. A follow-on study is being conducted that incorporates additional data and segregates data sets by meteorological variables.

Table 1. KMTX beam characteristics over Snownet sites.

Site (ID)	Range (km)	Altitude (m MSL)	Beam center height (m AGL)	Beam width (m)
Weber St. (SNW)	43	1464	1030	716
Bountiful (SNZ)	65	1452	1345	1083
Antelope Isl. (SNX)	31	1287	1020	516
SLC (SNL)	68	1284	1570	1133
Sandy (SNH)	93	1452	1890	1550
Snowbasin (SNI)**	50	2257	1180*	833
Park City (SNC)**	103	2104	1425	1716
Deer Valley (SNV)**	107	2501	2920*	1783

*1.4 deg

** Ski area/Olympic venue

Table 2. SWE rates (in h^{-1}) associated with select reflectivity values using $Z=75S^2$.					
dBZ	15	20	25	30	35
SWE	.03	.04	.08	.14	.25

Table 3. Coefficients and offsets for the "storm-total" regression lines shown in Figs. 7b-14b.								
	SNI	SNW	SNX	SNZ	SNL	SNH	SNC	SNV
Coeff.	1.57	.67	.38	1.88	1.03	1.08	1.42	1.63
Offset	.01	.02	0	-.13	-.06	-.04	-.02	0

References

Barker, T., P. Felsch, T. Mathewson, C. Sullivan, and M. Zenner, 2000: Test of the WSR-88D snow accumulation algorithm at WFO Missoula. NWS Western Region Technical Attachment No. 00-13.

Fulton, R. A., J. P. Breidenbach, D.-J. Seo, D. A. Miller, and T. O'Bannon, 1998: The WSR-88D rainfall algorithm. *Wea. Forecasting*, **13**, 388-395.

Super, A. B., and E. W. Holroyd, 1998: Snow accumulation algorithm for the WSR-88D: Final report. Bureau of Reclamation Report R-98-05, Denver, CO, July, 75 pp.

WSR-88D Radar Operations Center (formerly the Operational Support Facility) memorandum, December 15, 1999: Recommended changes to improve WSR-88D precipitation estimates during cool season stratiform rain events.

Vasiloff, S.V., 1997: Interpretation of radar data during snow events in mountainous terrain. WR Tech. Attachment TA97-35.

Vasiloff, S. V., 2001: WSR-88D performance in northern Utah during the winter of 1998-1999. Part-II: Examples of error sources. Western Region Technical Attachment No. 01-02.