Radar Measurement of Rainfall—A Summary

Abstract

Radar can produce detailed precipitation information for large areas from a single location in real time. Although radar has been used experimentally for nearly 30 years to measure rainfall, operational implementation has been slow. Today we find that data are underutilized and both confusion and misunderstanding exist about the inherent ability of radar to measure rainfall, about factors that contribute to errors, and about the importance of careful calibration and signal processing.

Areal and point rainfall estimates are often in error by a factor of two or more. Error sources reside in measurement of radar reflectivity factor, evaporation and advection of precipitation before reaching the ground, and variations in the drop-size distribution and vertical air motions. Nevertheless, radar can be of lifesaving usefulness by alerting forecasters to the potential for flash flooding.

The most successful technique for improving the radar rainfall estimates has been to “calibrate” the radar with rain gages. Simple techniques that combine sparse gage reports (one gage per 1000–2000 km²) with radar produce smaller measurement errors (10–30%) than either system alone. When high accuracy rainfall measurements are needed (average error less than about 10–20%) the advantage of radar is diminished, since the number of gages required for calibration is itself sufficient to provide the desired accuracy.

1. Introduction

For operational forecasting of river flow and flash floods, dense rain gage observations (telemetered) are desirable, but their installation has not been practical. Thus there has been considerable interest in utilizing weather radar, since it provides spatially and temporally continuous measurements that are immediately available at one location.

Both scatter and attenuation of microwaves by precipitation targets can be bases for estimation of precipitation. Currently, reflectivity data, either alone or in combination with rain gages, are considered most practical for operational measurement of rainfall over large areas. Figure 1 illustrates the detail in storm rainfall distribution that can be obtained by this method. Flooding from this storm devastated Enid, Okla., located near the 400 mm contour. Note that the spatial detail in the rainfall could not have been defined by the existing gage network. Unfortunately, radar does not measure rainfall rate directly, but rather the backscattered energy from precipitation particles in an elevated volume; and is prone to errors from 1) variations in the relationship between the backscattered energy and rainfall rate, 2) changes in the precipitation before reaching the ground, and 3) anomalous propagation of the beam. To circumvent the errors caused by the variations in the relationship between the backscattered energy and rainfall rate, a number of other possible precipitation measurement techniques have been suggested that utilize various properties of the radar signal.

Since microwave attenuation is related to rainfall rate, it has been proposed that the average rainfall along a path be determined by measuring signal attenuation (Rogers and Wexler, 1963; Atlas, 1964; Battan, 1973). Attenuation is also dependent on radar wavelength; thus Eccles (1978) has shown moderate success in obtaining rainfall rate from measurements of differential attenuation in tests of dual wavelength radars. Ulbrich and Atlas (1975) have suggested that improved measurements could be obtained by measuring both radar reflectivity factor and microwave attenuation. These radar values could then be used to determine two parameters of a drop-size distribution of assumed shape, from which rainfall rate could in turn be calculated. Similarly, Seliga and Bringi (1976, 1978) have proposed calculating rainfall from drop-size distribution parameters estimated from differential reflectivity between horizontally and vertically polarized waves. Only the reflectivity technique is compatible with the radar system in operational use today, and the others require considerable additional evaluation before they could be certified for operational use.

The potential for radar to improve warnings of floods and severe storms, together with the ever-increasing availability of inexpensive high speed data processing equipment, stimulated the National Weather
2. Theory and potential errors

a. Theory

The backscattered radar power from precipitation particles is proportional to the summation of the sixth power of particle diameters \(D_i^6\) in a unit volume illuminated by the radar beam; and hence the radar reflectivity factor \(Z\) (in units of \(\text{mm}^6/\text{m}^3\)) is defined as

\[
Z = \sum_i N_i D_i^6 = \int_0^\infty N(D) D^6 dD
\]  
(1)

where \(N_i\) is number of drops per unit volume of air with diameter \(D_i\) and \(N(D)\) is the number of drops with diameters between \(D\) and \(D + dD\) in a unit volume of air.

The desired parameter, rainfall rate \((R)\), is related to \(D\) through the following equation assuming vertical air motions are absent.

\[
R = \frac{\pi}{6} \int_0^\infty N(D) D^3 V_t(D) dD
\]  
(2)

where \(V_t(D)\) is the drop terminal velocity of a drop of diameter \(D\) that is approximated, in units of \(\text{cm/s}\), by \(V_t = 1400 \ D^{1/2}\) (Spilhaus, 1948). By substituting the Marshall-Palmer exponential drop-size distribution
sources; i.e., comparison with rain gages or disdrometers. It is
desirable that calibration include an independent
carefully electronic system calibration. Large
drop-size distribution is rarely known and it varies in time and space. In addition,
vertical air motions are frequently of the same magni-
tude as the terminal velocities (particularly in thunder-
storms). Thus the Z-R relationship is not unique and
we are forced to rely on average empirical relationships.
Battan (1973) presents a comprehensive list of Z-R
relationships derived by a number of investigators.
A widely used expression, based on the empirical study
of Marshall and Palmer (1948), is

\[ Z = 200R^{1.6} \]  

b. Sources of error
Numerous factors can cause errors in radar rainfall measurement. These sources can be categorized as
1) errors in estimating radar reflectivity factor, 2) variations in the Z-R relationship, and 3) gage and
radar sampling differences.

1) Measuring radar reflectivity factor
One source of reflectivity measurement error arises from
hardware calibration. Frequently, even after presumably
careful electronic system calibration, large unexplainable systematic errors in rainfall measure-
ments remain (Wilson, 1964; Harrold et al., 1974; Klazura, 1977; Saffle and Greene, 1978). Hence, it is
desirable that calibration include an independent source; i.e., comparison with rain gages or disdrometers.

Potentially serious sources of measurement error, not associated with hardware, are 1) beam blockage by
obstacles close to the radar site (Harrold and Kitchingman, 1975; Wilson 1975a), 2) anomalous propagation
(bending) of the radar beam (Battan, 1973; Brandes and Sirmans, 1976), 3) the build-up of precipitation
films on the radome (Cohen and Smolski, 1966; Wilson, 1978), and 4) attenuation by precipitation (Burrows
and Attwood, 1949; Wexler and Atlas, 1963), cloud (Ryde and Ryde, 1945), and atmospheric gases
(Blake, 1970). Attenuation most affects the shorter wavelengths; thus, it is frequently advocated that a
wavelength of 10 cm be used for quantitative rainfall measurement.

Random errors in signal estimates are associated with the independent motions of numerous target
precipitation particles. This error is generally reduced
to less than 1 dB by averaging a number of consecutive pulses in time and sometimes several pulses in range.

When logarithmic radar receivers are used, the averaging of pulses from different radar sampling volumes
will result in signal underestimates in regions of reflectivity gradients (Rogers, 1971; Sirmans and
Doviak, 1973). The bias becomes significant for large reflectivity gradients and large averaging intervals.

2) Variations in the Z-R relationships
Drop-size distributions have been measured in many
types of rains and characteristic Z-R relationships
have been reported (Jones, 1955; Imai, 1960; Fujiwara,
1965; Joss et al., 1970; Jatila and Puhakka, 1973a). Usually the coefficient (Eq. 4) increases and the
exponent decreases with increasing convective intensity. Even for similar type storms in the same geographical
area, wide variations in the reported Z-R relationship exist. Such variations are thought to reflect the predom-
ine of one or another physical process that influences the drop-size distribution.

Studies by Atlas and Plank (1953), Blanchard (1953),
Mason and Andrews (1960), and Joss and Gori (1978)
show narrow drop-size spectra that vary in time and space. Recent drop-size observations within Great Plains thunderstorms by Martner (1975) and Carbone and Nelson (1978) show high coefficients and low
exponents during growth (updraft) phases of the thunderstorm and low coefficients and high exponents
during declining (downdraft) phases. The large scatter in drop spectra over short time periods (5–10 min)
and the wide disparity among Z-R relationships reported by others prompted Twomey (1953) to
conclude that radar rainfall estimates derived with average Z-R relationship were only approximate at best (a factor of 2 high or low).

Physical mechanisms that may alter drop-size distributions are listed in Table 1 with an indication of
their probable influence on the Z-R relationship and the storm region where the effect is probably at a
maximum. Such processes act in combination to modify the drop-size distribution and produce a complex
net result.

Depending on size and number concentration, hail
may enhance reflectivity measurements or, if the hail-
stones are water-coated, a significant attenuation may
occur (Battan, 1973). Since the hail location, composi-
tion, shape, and size distributions are rarely known,
routine corrections to reflectivity measurements are
not possible.

3) Gage-radar sampling differences
Radar data ordinarily are obtained by scanning in
azimuth at a low elevation angle and making measure-
ments at discrete intervals of range and angle. Reflect-
vitvity factor values are converted to rainfall rate with
an appropriate Z-R relationship and accumulated in
time to yield rainfall depth (e.g., Fig. 1). Regardless of
the Z-R relationship utilized, this procedure results

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2 This paper discusses only rainfall measurement. Snowfall measurements are evaluated by Wilson (1975b), Collier and
Larke (1978), and Boucher (1978).
in time and space sampling errors that relate to rain rate histories (i.e., storm element size, intensity, duration, and motion characteristics).

Spatial precipitation features smaller than approximately twice the signal-averaging interval are not resolved in the radar measurements but are sometimes detected by rain gages and hence tend to decorrelate point gage-radar comparisons. For example, if a gage location coincides with a local rainfall maximum, the smoothed radar estimate tends to appear low.

Agreement between radar and surface (gage) rainfall estimates generally decreases with increasing radar range (Wilson, 1976; Brandes and Sirmans, 1976). Increasing radar sampling volume and height of the beam above the ground at far ranges leads to a higher probability that the precipitation observed aloft is different from that reaching the ground. The most important factors are probably evaporation, advection, and vertical air motions. With a larger radar sampling volume it becomes more likely that the beam is not uniformly filled by precipitation. Such problems are particularly important in stratiform precipitation where the radar beam intercepts the freezing level (bright band phenomena (Wilson, 1975a; Harrold and Kitchingman, 1975)).

Sampling and instrument error (both gage and radar) combine with error associated with the selection of the $Z-R$ relationship to reduce the agreement between gage and radar. Figure 2 illustrates the effect the radar time sampling interval can have on the ratio of gage and radar depth estimates, i.e., the parameter $G/R$. The lower curve gives the systematic error, while the upper curve gives the spatial variability. For the convective storm in Oklahoma on 7 April 1975, the dispersion among point gage-radar comparisons increases rapidly for sampling periods $>5$ min. Note that considerable residual error remains in the relative dispersion of $G/R$ ($\geq 22\%$) even for the shortest sampling interval (20 s). Radar bias (an underestimate in this example) increases slowly with sampling interval. Apparently reflectivity variations important for determining total rainfall deposition are inadequately sampled when measurements are made less frequently. We conclude that spatial smoothing in the basic radar

![Figure 2: Effect of radar sampling interval on gage/radar comparisons. The lower curve illustrates the effect on the average ratio of gage and radar depths ($G/R$) and the upper curve illustrates the effect on the relative dispersion of individual $G/R$ ratios. Curves are based on data from 190 gage/radar comparisons of total storm rainfall from Oklahoma thunderstorms observed on 7 April 1975 (radar data obtained from the NSSL WSR-57).](image)

For example, the lower curve indicates that for a radar sampling interval of 15 min, the average $G/R$ ratio of the 190 gages on 7 April was 1.3, and the upper curve shows that for the same sampling interval the relative dispersion of the 190 $G/R$ ratios was 36%.
instrument and integration of time series measurements introduces bias and scatter in radar-gage comparisons (see also Desautels and Gunn, 1970; and Zawadski, 1975; for further discussion).

Differences between point gage and radar observations can be minimized if the radar totals are spatially averaged about the gage site. Although an averaging radius that minimizes point gage/radar differences exists, the specific radius selected is not critical, since the average error changes only slowly near the optimum radius (Wilson, 1963; Brandes and Wilson, 1979).

4) ERRORS IN RAIN GAGE MEASUREMENTS

Since gages are used to adjust radar estimates and dense networks are used to verify radar areal rainfall estimates, it is important to examine errors in the gage measurements. The major cause of error in point measurements is from turbulence and increased wind flow about the gage. Larson and Peck (1974) summarized a number of independent studies and report about a 12% deficiency in gage catch for a wind of 5 m/s and 19% at 10 m/s. By extrapolating these data we estimate the undercatch in strong outflow (10-35 m/s) regions of thunderstorms to be as much as 20-40%.

Error in areal measurements of rainfall by gages arises from sampling a rainfall field with substantial variation. Sampling error decreases with increasing area size, increasing time period, increasing gage density, and increasing rainfall amount (Linsley and Kohler, 1951; McGuinness, 1963; Nicks, 1966; Huff, 1971; Woodley et al., 1975). For Illinois thunderstorms Huff (1970) reported an average error less than 5% in estimating average storm rainfall in a 1000 km² area with a network density of 65 km² per gage and rainfall amounts >1 mm. Comparable data from Woodley et al. (1975) for Florida air-mass thunderstorms indicate considerably larger errors (10-40% depending on rainfall amount). Typical gage verification networks used in radar rainfall studies are about 1 gage per 10–20 km². Thus, with the possible exception of small air-mass thunderstorms and regions of strong windflow, errors in gage measurements of areal rainfall are ordinarily less than 5%.

3. Radar-gage comparisons

Table 2 summarizes comparisons between radar and gage measurements for 14 Oklahoma storms observed by the NSSL WSR-57 radar. The gages were spread over an 8000 km² area at radar ranges between 45 and 100 km. The basic utility of a single relationship for converting radar-received power to rainfall rate is indicated by the variability in $G/R$. Ratios vary from 0.41 (radar overestimate) to 2.41 (radar underestimate). Note the large day-to-day variability with little indication of day-to-day correlation. Large storm-to-storm differences in $G/R$ exceeding a factor of two are fre-

<table>
<thead>
<tr>
<th>Date</th>
<th>Number of gages</th>
<th>Storm duration (h)</th>
<th>$G$ (mm)</th>
<th>$G/R$</th>
<th>Relative dispersion about $G/R$ %</th>
<th>Average difference % (Storm bias removed)</th>
</tr>
</thead>
<tbody>
<tr>
<td>29 Apr. 1974</td>
<td>20</td>
<td>12</td>
<td>7</td>
<td>0.82</td>
<td>21</td>
<td>30</td>
</tr>
<tr>
<td>29 Apr. 1974</td>
<td>21</td>
<td>5</td>
<td>22</td>
<td>2.23</td>
<td>31</td>
<td>51</td>
</tr>
<tr>
<td>1 May 1974</td>
<td>21</td>
<td>9</td>
<td>24</td>
<td>2.41</td>
<td>20</td>
<td>57</td>
</tr>
<tr>
<td>20 May 1974</td>
<td>12</td>
<td>7</td>
<td>15</td>
<td>0.68</td>
<td>46</td>
<td>75</td>
</tr>
<tr>
<td>21 May 1974</td>
<td>10</td>
<td>3</td>
<td>5</td>
<td>0.59</td>
<td>36</td>
<td>87</td>
</tr>
<tr>
<td>23 May 1974</td>
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<td>2</td>
<td>3</td>
<td>0.64</td>
<td>10</td>
<td>60</td>
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<td>25 May 1974</td>
<td>22</td>
<td>6</td>
<td>25</td>
<td>0.91</td>
<td>27</td>
<td>30</td>
</tr>
<tr>
<td>25 May 1974</td>
<td>12</td>
<td>5</td>
<td>23</td>
<td>0.88</td>
<td>32</td>
<td>31</td>
</tr>
<tr>
<td>30 May 1974</td>
<td>22</td>
<td>7</td>
<td>25</td>
<td>1.09</td>
<td>34</td>
<td>32</td>
</tr>
<tr>
<td>3 Jun. 1974</td>
<td>15</td>
<td>5</td>
<td>7</td>
<td>0.41</td>
<td>29</td>
<td>160</td>
</tr>
<tr>
<td>6 Jun. 1974</td>
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<td>2</td>
<td>16</td>
<td>0.68</td>
<td>43</td>
<td>79</td>
</tr>
<tr>
<td>8 Jun. 1974</td>
<td>10</td>
<td>5</td>
<td>31</td>
<td>0.49</td>
<td>39</td>
<td>141</td>
</tr>
<tr>
<td>7 Apr. 1974</td>
<td>19</td>
<td>4</td>
<td>14</td>
<td>1.13</td>
<td>23</td>
<td>17</td>
</tr>
</tbody>
</table>

Average 14 cases: 1.04 30

1 Relative dispersion (coefficient of variation) about $G/R = 100% \sigma[G/R]/G/R$; where $\sigma[G/R]$ is the standard deviation of the gage-radar ratios.

2 Average difference = $100% \sum_{i=1}^{N} \frac{|G_i - R_i|}{G_i} / N$.

3 Average difference (storm bias removed) = $100% \sum_{i=1}^{N} \frac{|G_i - (G/R)R_i|}{G_i} / N$.

4 The relative dispersion among all available gage-radar comparisons (14 storms) is 63%.
FIG. 3. Comparison of hourly rainfall measurements by radar and rain gage for two storms 9 hours apart passing over the same region. Measurements are for 170 km$^2$ areas in Oklahoma. The radar data were collected by the NSSL WSR-57 on 29 May 1964.

Large wanderings in $G/R$ are not unexpected in view of the observed range in reported $Z-R$ relationships and variable nature of the drop-size distribution. Table 2 also indicates that the average difference between radar and gage point measurements for all 14 storms is 63%. By removing the mean storm bias, i.e., multiplying the radar estimates by $G/R$, the average difference is reduced to 24%. Hence, much of the radar error results from storm-to-storm differences in the relationship between radar-received power and rainfall rate. Storm differences are further illustrated in Fig. 3 for two storms, nine hours apart, passing over the same rain gage network. A significant reduction in radar error could be achieved if a procedure were found to determine the average storm bias.

Inspection of $G/R$ fields for individual storms reveals that within-storm radar errors are not random but vary with position on a scale similar to the spacing between individual rain cells. For example, Fig. 4a

It is important to point out that real-time rainfall estimates accurate to within a factor of two can be very useful for alerting weather forecasters to regions likely receiving heavy rain and which therefore should be carefully monitored for possible flash flooding.

FIG. 4. (a) Gage measured thunderstorm rainfall (mm) on 6 June 1974 from 1500-1700 CST. The density of the rain gage network was 1 gage per 23 km$^2$. Two cells, one to the south and one to the north, dominate the pattern. (b) Ratio of the gage measured ($G$) to radar estimated ($R$) rainfall for the same time period as (a). The radar data were collected by the NSSL WSR-57. Note that the $G/R$ ratios tend to be low on the edges of each storm and high toward the storm centers.
shows the rainfall pattern from two cells (one in the north half and the other in the south half) passing through a dense network of rain gages. The $G/R$ pattern in Fig. 4b shows that each storm has its own pattern. Note that low values tend to occur along the storm edges and high values toward the centers. Often contiguous storms exhibit distinctive error relationships (Fig. 5). Both Figs. 4 and 5 show a correlation between $G/R$ and gage depth; i.e., heavy rainfalls tend to be underestimated and light rainfalls overpredicted. This result, mentioned by Woodley and Herndon (1970) and noted by Desautels and Gunn (1970), is partially attributed to smoothing in the radar observations, variations in the drop-size distribution, and vertical and horizontal air motions.

It would be expected that the agreement between gage and radar measurements would improve when comparisons are made for areas rather than points. Indeed, studies by Harrold et al. (1974), Wilson (1968), and Muchnik et al. (1968) all indicate some error decrease in radar measurements as the area size increases. The Harrold et al. and Wilson studies also show error decreases as the length of the measurement period increases. However, Wilson (1968) found that a significant error reduction with increasing area size was only possible if the mean storm bias ($G/R$) was removed.

### 4. Radar rainfall adjustment techniques

The large storm-to-storm and within-storm errors in radar rainfall estimates (Table 2) dictate further adjustment on a storm-to-storm basis. Adjustments usually involve either changing the $Z-R$ relationship or keeping the $Z-R$ relationship fixed and utilizing rain gage observations to adjust the radar estimates.

**a. Changing the $Z-R$ relationship**

Some improvement in radar rainfall estimates may be realized by selecting $Z-R$ relationships based on storm type (Jones, 1966); however, the improvement is small because of large variations within storm type classifications. Atlas (1964) and Kessler (1965) suggest that the conversion relationship could be determined from statistics describing the radar echo pattern. While this technique has shown some success in categorizing storms by type (Puhakka, 1974), little correlation has been found within a storm type between the optimum relationship and echo properties such as intensity, intensity variance, storm size, and orientation of major rain areas (Wilson, 1966).

**b. “Calibration” with rain gages**

As early as 1954, Hitschfeld and Borden (1954) recommended “radar used for precipitation measurements should be calibrated against rain gages rather than by any other means.” The resulting hypothesis attributes specification of the precipitation distribution to the radar, and precipitation magnitude to the gages.

1) **Adjustment for storm-to-storm radar bias**

In the simplest “calibration” procedure, the mean radar bias is determined and a correction applied uniformly to the radar estimates. Ordinarily a number of gages ($N$) are used and a multiplicative adjustment factor ($F$), the ratio of gage observed ($G$), and radar-indicated rainfalls ($R$), are computed from either

$$F = \frac{\sum_{i=1}^{N} G_i}{\sum_{i=1}^{N} R_i} \quad (5)$$

or

$$F = \frac{1}{N} \sum_{i=1}^{N} \frac{G_i}{R_i} \quad (6)$$

With Eq. (5), observations receive a weight proportionate to depth while with Eq. (6) all gage-radar comparisons have equal weight. In each case corrections are applied for bias associated with radar hardware calibration and the $Z-R$ relationship. Indeed, the relationship selected to convert reflectivity to rainfall is of little importance and has negligible impact on the final (corrected) radar depth estimate (Brandes, 1974). On the average for Oklahoma storms a single gage-radar comparison determines the storm radar bias with an average expected error of $\pm 30\%$ (Table 2). Because this error is much less than the longer term deviation in $G/R$ of $63\%$ (14 storms), even a single gage observation can provide useful storm “calibration” information. When the mean radar bias is estimated

Grayman and Eagleson (1970) and Cain and Smith (1976) note $\log[G/R]$ tends to be more normally distributed than $G/R$; hence a more appropriate expression for Eq. (6) may be $F = \text{antilog}[\log(G/R)]$.

On the other hand, experience has shown that if the gage rainfall is light ($<2$ mm) or the rainfall variability (observed by radar) about the gage is very large, it is best not to use the gage for adjustment (Wilson, 1975a).
TABLE 3. Radar areal estimates of rainfall utilizing gages for calibration.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Location</th>
<th>Rain type</th>
<th>Radar λ(cm)/θ(deg)</th>
<th>Z-R relation</th>
<th>Number of cases</th>
<th>Radar observation frequency (min)</th>
<th>Radar range (km)</th>
<th>Area size (km²)</th>
<th>Duration</th>
<th>Adjustment type and calibrating gage density (km²/gage)</th>
<th>Error before adjustment (%)</th>
<th>Percent error using calibration gages only</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wilson (1970)</td>
<td>Oklahoma</td>
<td>Thundershowers</td>
<td>10/2</td>
<td>KR&lt;sup&gt;1.5&lt;/sup&gt;</td>
<td>b</td>
<td>23</td>
<td>5–10</td>
<td>35–100</td>
<td>3500</td>
<td>Storm</td>
<td>A(3300)</td>
<td>51 (35)</td>
</tr>
<tr>
<td>Brandes (1975)</td>
<td>Oklahoma</td>
<td>Thundershowers</td>
<td>10/2</td>
<td>200R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>9</td>
<td>5</td>
<td>45–100</td>
<td>3000</td>
<td>24 hr</td>
<td>Storm</td>
<td>V(900)</td>
<td>52 (13)</td>
</tr>
<tr>
<td>Woodley et al. (1974)</td>
<td>Florida</td>
<td>Showers</td>
<td>10/2</td>
<td>300R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>39</td>
<td>1</td>
<td>12–48</td>
<td>50–100</td>
<td>1 hr</td>
<td>A(1600)</td>
<td>—</td>
<td>—</td>
</tr>
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<td>Harrold et al. (1974)</td>
<td>England</td>
<td>Showers</td>
<td>10/2</td>
<td>200R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>27</td>
<td>5</td>
<td>85–115</td>
<td>570</td>
<td>24 hr</td>
<td>A(1600)</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Wilson (1975a)</td>
<td>New York</td>
<td>Showers</td>
<td>5/1.7</td>
<td>200R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>c</td>
<td>41</td>
<td>95–112</td>
<td>170</td>
<td>24 hr</td>
<td>V(275)</td>
<td>49 (22)</td>
<td>22</td>
</tr>
<tr>
<td>Collier et al. (1975)</td>
<td>England</td>
<td>Showers</td>
<td>10/1</td>
<td>200R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>13</td>
<td>1</td>
<td>12–48</td>
<td>700</td>
<td>3 hr</td>
<td>V(233)</td>
<td>—</td>
<td>7</td>
</tr>
<tr>
<td>Jatila and Puhakka (1973a,b)</td>
<td>Finland</td>
<td>Showers</td>
<td>3/1.8</td>
<td>200R&lt;sup&gt;3.6&lt;/sup&gt;</td>
<td>6</td>
<td>5</td>
<td>18–28</td>
<td>180</td>
<td>A(180)</td>
<td>A(180)</td>
<td>43 (23) (l,g)</td>
<td>—</td>
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<td>Huff and Towery (1978)</td>
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<td>Showers</td>
<td>10/1</td>
<td>300R&lt;sup&gt;3.5&lt;/sup&gt;</td>
<td>67 (h)</td>
<td>3</td>
<td>20–100</td>
<td>5300</td>
<td>0.5 hr</td>
<td>V(150)</td>
<td>55 (27)</td>
<td>32</td>
</tr>
</tbody>
</table>

Notes:
(a) A: average adjustment; V: variable spatial adjustment.
(b) Radar estimates adjusted to remove average bias for the total experiment.
(c) Density of rain gage clusters is approximately 8 gages per cluster. Clusters outside area of measurement.
(d) Calibrating gage not within boundaries of watersheds, but is 10–20 km distant.
(e) An additional multiplicative factor of 1.7 applied to radar estimates.
(f) Varied Z-R coefficient to match point rainfall at central gage.
(g) Error when observed drop-size distribution used is 98%.
(h) Number of 30 min periods in four storms.

from a number of gage-radar comparisons, the expected error will be smaller.

When Wilson (1970) adjusted radar-derived thunderstorm rainfalls for a 3500 km² watershed by a single centrally located gage, the average error was reduced from 51% (unadjusted measurements) to 35%. When only the single gage was compared with the watershed mean depth, the error was 60%. Single gage adjustment offered no improvement in the Jatila and Puhakka (1973a,b) study of stratiform rains; but errors in convective rainfalls were lowered from 43% to 25%. These and other studies utilizing an average storm adjustment factor are summarized in Table 3. Although procedural and meteorological differences preclude rigorous comparison, the common result is a significant reduction in radar rainfall estimate error when adjustments are made on a storm basis. Nevertheless, large spatial errors (e.g., the 30% spatial error in Table 2) remain in the radar precipitation patterns.

2) SPATIAL ADJUSTMENT

Because most within-storm errors are not random, "calibration" by one nearby gage generally produces rainfall estimate errors smaller than those from an ensemble average of all available calibration ratios. This is illustrated in Fig. 6 for two Oklahoma thunderstorms where the standard error in gage-adjusted point radar rainfall estimates is shown as a function of distance from the "calibration" gage. For comparison, the standard error for the same point rainfall estimates after adjustment by the mean radar bias (G/R) is indicated by a horizontal line. The maximum useful range of a single adjustment gage varies from storm to storm (cf. 6 June 1974 and 7 April 1975). The utility of interpolating adjustment factors from a small

![Fig. 6. Standard error of point radar rainfall measurements (total storm accumulation) versus the distance of the measurement from the calibrating gage for two thunderstorm dates.](image)
number of gages to regions beyond the effective ranges of single gages has not been sufficiently studied.

Sophisticated schemes have been devised (e.g., Brandes, 1975; Collier et al., 1975; Wilson, 1975a) to routinely combine radar observations with distributions of adjustment gages. Local adjustments are made by interrogating nearby calibration sites and then assigning weights inversely by distance. In essence, the radar observations are molded to the gage observations by a plane-fitting technique while retaining the radar-indicated precipitation variance between gages. Depending upon the adjusting gage density, potential corrections include systematic bias associated with error in the radar hardware calibration, synoptic-scale influences on the $Z-R$ relationship, small-scale (within-storm) spatial variations in the $Z-R$ relationship, and changes in the precipitation below the beam.

Although the experiments in Table 3 differ as to the length of the measurement period, area size, radar range, data collection frequency, and density of adjustment gages, the reported error range of 13–27% after spatial adjustment is considerably less than for unadjusted estimates of 43–55% and somewhat less than the 18–35% error range for uniform adjustments.

Intuitively, we expect the radar, even if only roughly "calibrated," to measure the rainfall over the entire area observed by the radar better than a single gage; while in the other extreme, if the calibrating gages were very dense, we would not expect the addition of radar to provide significant improvement. Examination of the last three columns in Table 3 confirms this tendency. Data from Oklahoma thunderstorms (Brandes and Wilson, 1979; Fig. 16) suggest that the combined radar-gage estimates are usually better than the gage only for gage densities < 1 per 300-400 km². Results from Illinois convective storms (Hildebrand et al., 1979) indicate the radar-gage estimates are no longer better than the gage-only estimates when the density increases to about 1 per 250–300 km².

c. Summary remarks on radar adjustment

It appears that the primary causes of error in radar measurements of rainfall result from variations in the $Z-R$ relationship caused by microphysical and kinematic processes that affect the drop-size distribution and drop-fall speeds. Drop-size measurements and radar rainfall error patterns indicate that the variations occur from storm to storm in a systematic and perhaps predictable manner. The radar tends to overestimate light rainfall and to underestimate heavy rainfall. The search for systematic error patterns holds promise, but until such patterns can be unambiguously established it will be necessary to use gages to adjust the radar.

Successful implementation of radar into a precipitation measurement system requires that careful electronic calibration procedures be followed and that an independent check of system biases be made by comparing radar estimates with rain gage or disdrometer measurements. Data users should be cognizant of those conditions in which the radar rainfall estimates may be erroneous or have little value, such as in the vicinity of ground targets, within shadow areas created by obstacles blocking the radar beam, at excessive distances where the beam becomes very large and elevated, and when there is extreme variability in cloud processes and, consequently, in $Z-R$ relationships.

5. Concluding remarks

With reasonable efforts, radar measurements (without gage adjustments) should be within a factor of two of the true rainfall about 75% of the time (Table 2; Wilson, 1970; Woodley et al., 1975). While this accuracy may not be sufficient for adequate stream flow forecasting, it has important potential for real-time flash flood warning. This was illustrated by the 1977 Johnstown flood disaster. The National Weather Service WSR-57 radar at Pittsburgh was equipped with test equipment to automatically digitize and accumulate rainfall. While the rainfall estimates were low (Saffle and Greene, 1978), they did indicate heavy rains were occurring and they could have been very useful in issuing flood warnings. Because heavy rainfalls may frequently be underestimated, the forecaster should take action to verify the radar estimates before they indicate rainfall amounts considered necessary for flooding. While a small number of rain gage reports can significantly reduce the error in radar estimates (Table 3), seldom are sufficient gage reports available. Thus to obtain this improvement, in real time, a means is needed for acquiring special gage observations when conditions warrant.

Storm rainfall amounts based on combined radar-gage data have average errors of about 10–30%. This figure is for gage densities similar to that of the National Weather Service Climatological Network (1 gage per 1000–2000 km²). While the error may decrease further as the gage density increases, the advantage of radar diminishes since sufficiently dense gages alone provide about the same accuracy as the combined data. It is estimated that for midwestern United States thunderstorms this becomes the case for gage densities greater than about 1 per 250–400 km².

When high accuracy rainfall measurements are desired (errors ≤ 20%), such as for weather modification experiments, the utility of the radar decreases since the number of calibrating gages required is itself sufficient to provide the desired accuracy. While the radar may not improve on the absolute magnitude of the rainfall amount in this case, it still provides valuable information on the 3-dimensional location and evolution of the activity. For monitoring rainfall over large areas not instrumented with a high density (1 per 250–400 km²) gage network, the radar-gage analysis is superior to either analysis alone.

Acknowledgments. We would like to thank Dr. Edwin Kessler for originally suggesting we write this paper and his...
helpful comments during its preparation. We would also like to thank Mr. Rit Carbone and Drs. Robert Serafin and Peter Hildebrand for their many helpful suggestions in reviewing the text.

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