Precipitation Identification from Radar Wind Profiler Spectral Moment Data: Vertical Velocity Histograms, Velocity Variance, and Signal Power–Vertical Velocity Correlations

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ABSTRACT

Correlations between range-corrected signal power $S_r$ and radial vertical velocity $V_r$ from the vertical beam of a UHF wind profiler can be used to distinguish between air- and precipitation-dominated echoes using an $S_r$–$V_r$ correlation diagram. While there is no clear correlation between vertical air motions and $S_r$, there is a strong correlation between the precipitation fall velocity and $S_r$ in snow, and to a lesser extent, in rain. This is illustrated through intercomparison of three types of precipitation events, and two types of clear-air events.

Using a histogram of $V_r$ from an event where there is evidence of precipitation in its $S_r$–$V_r$ correlation diagram, and from other information, it is possible to objectively determine a threshold value of $V_r$, referred to as $V_T$, that approximately identifies which measurements are dominated by Rayleigh scattering from precipitation in that event. A method is introduced that uses the histogram of observed $V_r$ from that event to provide an estimate of how many measurements are incorrectly attributed to Bragg scattering or Rayleigh scattering as a function of $V_r$. The error estimates can be used to select $V_T$ on a case-by-case basis and according to the needs of the particular application. An objective dual-optimization technique results in an estimated overall error of less than 6%, averaged over three case studies. In addition, it is shown that inclusion of velocity variance from the vertical beam in the $S_r$–$V_r$ correlation diagrams can help distinguish between rain and snow, and between convective and stratiform precipitation.

1. Introduction

While Bragg scattering from air is governed by features that may be of great interest, such as turbulence and vertical moisture gradients, precipitation can mask their signatures in spectral moment data from radar wind profilers. If such parameters are to be studied, diagnosed, or monitored in the future by radar wind profiler networks (or for archival of the full Doppler power spectrum may not be feasible), it is important to determine on a point-by-point basis whether the backscattered power observed by such radars is due to Rayleigh scattering from hydrometeors or Bragg scattering from air. It is also valuable to compare the observations of precipitation identified from the profiler data with the winds measured at the same time and place by the same instrument. This allows the kinematic features identifiable from the observed wind profiles to be related to the development of precipitation.

The value of superimposing the regions and types of precipitation on time–height profiles of winds measured by radar wind profilers has been recognized only recently, and this capability has already benefitted studies of mesoscale phenomena such as convective and stratiform rain ahead of a midlatitude cyclone (Fabry et al. 1993), the structure of complex frontal systems (Neiman et al. 1994a,b) and unique observations of a mesoscale convective system (Ralph et al. 1995). Although the dense network of precipitation radars now found throughout much of the United States provides some similar information, profilers provide a unique perspective by regularly measuring winds both within and outside of precipitation using a very different sampling strategy, and profilers are not all sited in areas covered by such weather radars. In addition, it may be desirable to extend earlier precipitation studies that used radar wind profilers, such as the interesting studies of tropical precipitation and vertical air motion by Cifelli and Rutledge (1994), Williams et al. (1995), and May and Rajopadhyaya (1996), without depending on precipitation being observed at the ground, and without having to coordinate observations from a separate weather radar with different sampling characteristics.
Although it has been well established that UHF (200–2000 MHz) and VHF (20–200 MHz) radar wind profilers can provide important information about precipitation through analysis of the full Doppler power spectrum (e.g., Fukao et al. 1985; Larsen and Röttger 1987; Wakaugui et al. 1987; Ecklund et al. 1988, 1995; Gossard 1988; Currier et al. 1992; Rogers et al. 1993; 1994; Chilson et al. 1993; Rajopadhyaya et al. 1993; Cifelli and Rutledge 1994), large profiler networks, unlike many research profilers, may be unable to retain more than the moments of the spectrum because of communication and data storage limitations (Ralph et al. 1995). For this reason, and because working with spectral moment data is less cumbersome than full spectrum data, it is useful to develop precipitation-detection and interpretation techniques that can be applied to spectral moment data alone.

Spectral moment data can be used to distinguish between measurements dominated by Bragg scattering from air from those dominated by Rayleigh scattering from hydrometeors (Steiner and Richner 1994; Gage et al. 1994; Ralph et al. 1995; Williams et al. 1995).

To date, two primary methods have been used, each of which demonstrates and uses correlations between signal power, vertical velocity, and spectral width: 1) examination of events based on precipitation observed at the surface, and 2) the climatology of clear-air reflectivity compared to reflectivity in precipitation, a method introduced in Ralph (1995). While Gage et al. (1994) and Williams et al. (1995) used the first method to study the climatology of tropical precipitation, Steiner and Richner (1994), Ralph (1995), and Ralph et al. (1995) used a combination of the two methods in studies of precipitating midlatitude weather systems. While Ralph (1995) used an extensive climatology of clear-air reflectivity to establish how vertical radial velocity could be used as a key parameter in a thresholding technique, Steiner and Richner (1994) combined signatures from all three spectral moments in a thresholding technique based on a short climatology of each, but were unable to effectively detect snow in their case study. Although results from Steiner and Richner (1994) and Ralph (1995) suggest that snow is more difficult to identify than rain in spectral moment data, observations presented in Rogers et al. (1994) and in Ralph et al. (1995) clearly illustrate that snow is readily observable by UHF radars because of its characteristic fall velocity, except in isolated regions of large vertical air motions, such as in mountain waves and deep convection. The purposes of this paper are to further explore and refine the use of correlations between the first three spectral moments, to extend the vertical velocity thresholding technique to better identify snow, and to establish objective criteria for distinguishing between measurements dominated by Bragg scattering and those dominated by Rayleigh scattering, and between different precipitation types.

2. Methodology and radar characteristics

This paper presents analyses of observations from the Wind Profiler Demonstration Network (WPDN) 404-MHz radar wind profilers from three precipitation events and two clear-air events. The precipitation events have been presented in an earlier paper (Ralph et al. 1995), where the presence of precipitation was verified in each case and the background meteorological conditions were described. These cases included a winter storm with rain and snow, snow in jet stream cirrus, and a mesoscale convective system (MCS) with stratiform and convective regions. The clear-air cases include one dominated by mountain wave activity with vertical air motion amplitude exceeding 1 m s\(^{-1}\), which was used in Ralph et al. (1992) to study mountain wave behavior, and another with no evidence of mountain wave activity. This paper will focus first on the winter storm case and then will intercompare the patterns established from that event with those of the other events.

Detailed descriptions of the spectral moment data from the WPDN profilers (signal power \(S\), radial vertical velocity \(V_r\), and velocity variance \(\sigma_r^2\)), as well as important operating parameters, are given in van de Kamp (1988), Barh et al. (1994), and Ralph et al. (1995). Note that \(V_r > 0\) represents a downward velocity. Although there are three antenna-pointing directions and two pulse modes (low and high altitude), this paper uses observations from only the low mode of the vertically pointing beam. Because \(S\) reported by the WPDN profilers is not normalized to account for the effect of range, the range-corrected signal power \(S_c\) is calculated using \(S_c = S + 20 \log_{10} r\), where \(r\) is the range in meters to the center of the range gate. Although it would be most useful to transform \(S_c\) into reflectivity factor in precipitation, it is very difficult to do so because the WPDN profilers are not calibrated; the receivers become saturated in conditions of strong reflectivity; and \(S\) is reduced by an unknown amount in the lowest 1–4 range gates by a sensitivity time control (STC) procedure so as to increase the accuracy of radial velocity measurements there. The STC is accomplished by reducing the receiver gain in the first few range gates, which greatly reduces receiver saturation in those gates. For the cases presented here, the STC procedure operated only in the two clear-air events. No editing or quality control procedures are applied to the 6-min data presented in this paper, but many of the unreliable data points will be identifiable as such in the correlation diagrams used here. These bad data points can result from several phenomena, including strong scattering in sidelobes, weak scattering near the top of the low mode, birds or airplanes in the main lobe or in sidelobes, ground clutter and its suppression, etc.

3. Distinguishing between rain, snow, and air in a winter storm

On 11 and 12 February 1993, a winter storm produced a wide area of primarily light precipitation in the
form of snow, freezing rain, and rain over the central United States, including Iowa. During this time the Slater, Iowa, wind profiler recorded evidence of precipitation in the spectral moment data. This is shown in Ralph et al. (1995), where surface observations from Des Moines (41 km south of the profiler), along with rawinsonde observations of the freezing level above Omaha, Nebraska, and Peoria, Illinois, provided verification of the interpretations based on the spectral moment data. A representative time-averaged (18 min) profile of each of the three spectral moments is shown in Fig. 1, where evidence of the melting level near 1.50–1.75 km above mean sea level (MSL) is present. These interpretations are consistent with calculations of the relative sensitivity of 404-MHz profilers to Rayleigh scattering and Bragg scattering (Ralph 1995). The calculations have shown that Rayleigh scattering should almost always dominate Bragg scattering when reflectivity-weighted terminal velocities exceed 0.5–0.9 m s$^{-1}$ in snow, and 2.5–4.5 m s$^{-1}$ in rain (Ralph 1995). These thresholds are based on the climatology of the atmospheric structure parameter $C_n^2$ and are represented by the symbol $V_T^*$. From the radar wind profiler, surface, and rawinsonde data, it is apparent that the altitude of the melting level decreased with time, eventually reaching the ground about halfway through the 18-h precipitation event, during which 6.5 mm of precipitation accumulated (4 mm as freezing rain, and 2.5 mm as snow).

Although Fig. 1 clearly illustrates the signature of the bright band in the spectral moment data, it also points out the difficulty of determining at what altitude, if any, the signal becomes dominated by Bragg scattering. While $V_T^*$ provides a useful first approximation to this transition in snow, it may be possible to determine the threshold on a case-by-case basis through correlations between $S_n$ and $V_r$. This threshold is referred to as $V_T$, so as to distinguish it from the climatology-based threshold, $V_T^*$. Under conditions where Rayleigh scattering from precipitation dominates Bragg scattering from air, there should be a correlation between $S_n$ and $V_r$ that results from the increase in reflectivity-weighted fall velocity associated with larger reflectivity factor, whereas no such correlation need be present when Bragg scattering dominates. This behavior is shown in Fig. 2, which contains 30 h of 6-min data from the low-altitude mode of the Slater, Iowa, profiler from 0600 UTC 11 February to 1200 UTC 12 February 1993. This mode covers the altitude range from 0.8 to 9.6 km MSL at 250-m gate spacing and includes the 18 h during which precipitation reached the ground at Des Moines and 12 h when no precipitation reached the ground. Figure 2 is based on an approximately 11 000-point dataset, out of which about 2300 points are noise. However, many of the noise points fall outside the range of values included in the figure.

Five major groupings of points can be identified, including four representing conditions dominated by scattering from either air, snow, rain, or melting snow, and one representing noise.

![Fig. 2. Scatterplot of unedited range-corrected signal power $S_n$ and radial vertical velocity $V_r$, measurements from the vertically pointing beam of the low-altitude mode of the Slater, Iowa, profiler from 0600 UTC 11 February to 1200 UTC 12 February 1993.](image-url)
1) One set is characterized by small $V_r$ and extends over a wide range of $S_c$. This behavior is characteristic of Bragg scattering from air, where it appears that most of the vertical air motion $w$ measurements fall between \(-0.3\) and \(+0.3\) m s\(^{-1}\) in this case. The range of $w$ measurements depends both on the measurement precision and on the temporal and spatial variability of $w$.

2) One of the largest sets of points is characterized by strong correlation between $V_r$ and $S_c$ for values of \(0.3\) m s\(^{-1}\) < $V_r$ < \(1.5\) m s\(^{-1}\). This behavior is indicative of snow, as is shown below in a quantitative way.

3) A smaller, but still significant set of points is characterized by $V_r$ values consistent with those that should be observable at 404 MHz in light rain (i.e., $V_r$ > \(3\) m s\(^{-1}\)) and $S_c$ values between 130 and 150 dB. This set of points has a slight, but noticeable correlation with $S_c$.

4) The cloud of points with similar and higher $S_c$ but with $V_r$ between 1.5 and 3 m s\(^{-1}\) likely represents melting snow.

5) Finally, noise is represented by the points with $S_c$ < \(110\) dB that extend over a wide range of $V_r$.

Although all of these data were observed during the same 30-h time interval and the same range of altitudes, the measurements dominated by Rayleigh scattering from precipitation occurred at different heights and times than those dominated by Bragg scattering from air. Thus, the characteristics of the vertical air motion corresponding to the measurements dominated by Bragg scattering might not very accurately represent the vertical air motion within the regions of precipitation.

By considering only points from altitudes above the freezing level, it should be possible to eliminate all points representative of rain. Because the radar and rainsonde measurements indicate that the 0°C level was never above 2.0 km, only points above 2.5 km are included in the $S_c$–$V_r$ scatterplot diagram in Fig. 3a. In contrast, only points from below 2.5 km are included in Fig. 3b. Comparison of Fig. 3a with Fig. 3b indicates that no points with $V_r$ indicative of rain and melting snow were present above the freezing level, except for points with $S_c$ at the noise level. Snow, melting snow, and rain are present in Fig. 3b because the melting level was within the 2-km-thick layer between the ground and 2.5 km MSL during several hours of the time interval shown. Points indicative of noise are absent in Fig. 3b because they occurred only in range gates above 2.5 km. Figures 3c and 3d show that in this case snow rarely had velocity variance greater than 1 m\(^2\) s\(^{-2}\) (Fig. 3c), while rain often did (Fig. 3d), as is discussed in section 5.

It should be expected that the points indicative of rain and snow should have slopes that are approximately consistent with theoretical relationships between reflectivity factor and reflectivity-weighted terminal velocity, as long as $w$ is small relative to the terminal velocity. Such relationships have been presented in Ralph (1995) for exponential drop size distributions in rain and snow. The relationship for snow is shown in Fig. 3a, where a value of $S_c = 144$ dB was chosen to correspond to 10 dBZ\(_e\) (effective reflectivity factor) because both corresponded to $V_r = 1.0$ m s\(^{-1}\) (for a typical snow density of 0.04 g cm\(^{-3}\)), assuming $w = 0$. Similarly, Fig. 3b shows the relationship for rain, where 16.5 dBZ\(_e\) corresponded to $S_c = 142$ dB for $V_r = 4.5$ m s\(^{-1}\) (at the 1000-hPa pressure level). Although there is only fair agreement between the theoretical curve and the measurements in rain, the large downward velocities are clearly indicative of rain, and the weak correlation with $S_c$ is in the correct sense (i.e., direct). A likely reason for the weak correlation is that the rainfall rate did not vary substantially during the rain portion of the event, which would cause the distribution of measurements in $S_c$–$V_r$ space to cover only a small region on a diagram such as Fig. 3b. It is interesting to note that the presence of a broad range of snowfall velocities in a case with nearly constant rainfall rates is consistent with conditions where precipitation formation was dominated by the growth of ice crystals and aggregates above the freezing level. Because of the wide variety of snow densities that can exist, and because the relationship between dBZ\(_e\) and $V_r$ is sensitive to such differences (Ralph 1995), curves for snow are also drawn on Fig. 3a for extreme values of snow density (i.e., 0.01 and 0.08 g cm\(^{-3}\)). The axis of the distribution of measurements that are characteristic of snow in the winter storm case is also shown in Fig. 3a, based on an approach described in the next section. The excellent agreement between the data and the theoretical curves for snow supports the conclusion that this portion of the data consists of measurements dominated by Rayleigh scattering from snow. This comparison also suggests that much of the scatter about the axis of the distribution of measurements characteristic of snow could result from variations in snow density within the sample. However, variations in $w$ and in precipitation rate could also affect the scatter and are likely responsible for some of the points falling outside the range covered by variations in snow density in this case (Fig. 3a). It is interesting to note that very few data points were correlated with these curves at values of $S_c < 120$ dB. This may reflect the greatly reduced likelihood that Rayleigh scattering can dominate Bragg scattering for very low values of dBZ\(_e\). Some other factors may influence the relationships between these variables in a given event, including varying snow particle size distributions and the relative number of clear-air, snow, and rain samples within the total number of points plotted.

4. Determining a threshold radial vertical velocity to distinguish between scattering from snow and air

In contrast to snow, scattering from rain is relatively easy to identify because the fall velocities of rain that
Fig. 3. Same data as shown in Fig. 2, but stratified by altitude and velocity variance. (a) All measurements from above 2.5 km MSL (no rain included). (b) All data below 2.5 km (includes rain and snow). (c) Data from above 2.5 km (no rain included) with velocity variance greater than 1.0 m² s⁻². (d) Data from below 2.5 km (includes rain and snow) with velocity variance greater than 1.0 m² s⁻². Theoretical relationships between $S_n$ and $V_r$ are shown in (a) for three snow densities ($\rho_s$, see text), and in (b) for rain. The radial vertical velocity $V_r$ of the peak in the $V_r$ histogram corresponding to snow, determined from Fig. 4, is marked by dots in (a).

a. $C_n^2$ climatology method

Observations made over long time periods have provided a climatology of $C_n^2$ that can be used to assess what amount of Rayleigh scattering from hydrometeors is required to dominate the Bragg scattering from air. This approach is presented in Ralph (1995), where background values of $C_n^2 = 10^{-15}$ and $10^{-13}$ m⁻²/³ were used to calculate what radar reflectivity factors $Z$ would provide equivalent amounts of backscattered power, depending on radar wavelength. By assuming exponential drop size distributions, Ralph (1995) then transformed these threshold values of $Z$ into modal terminal
velocities $V^*$, which could be used as climatology-based thresholds $V^*_r$. These values of $C_n^2$ were exceeded only 21% and 0.5% of the time, respectively, at 805 m above ground level (AGL) in Colorado during a full year (Chadwick and Moran 1980). Because $C_n^2$ decreases with height on average (Gossard 1990), the frequency of occurrence of $C_n^2 = 10^{-15}$ m$^{-2/3}$ in the middle and upper troposphere is likely similar to that of $C_n^2 = 10^{-15}$ m$^{-2/3}$ in the atmospheric boundary layer. These calculations indicated that the $V^*$ threshold for rain would vary between 2.5 and 4.5 m s$^{-1}$ for a 404-MHz radar, depending on altitude. In snow, however, the $V^*$ threshold is between 0.5 and 0.9 m s$^{-1}$, depending on the snow density. Although the simplicity of these thresholds makes them useful in a broad range of applications, it is likely that the optimal threshold for a particular event and application may differ from these average thresholds, because background $C_n^2$ can change due to seasonal and spatial variations in water vapor mixing ratio, for example. Also, this method does not readily provide an estimate of the amount of data that may be misinterpreted as due to precipitation, or to air.

b. A method based on radial vertical velocity histograms

1) A winter storm

As is evident from Figs. 2 and 3a, there is a distinct separation between the cluster of points characteristic of snow and those of air, at least for larger values of $V_r$. For smaller $V_r$, however, the two distributions merge. Because there appears to be a systematic separation between the two distributions for most values of $S_{inc}$, it is plausible to use this separation to distinguish between air and snow. To identify the value of $V_r$ that corresponds to this separation, velocity histograms of $V_r$ within different ranges of $S_{inc}$ are shown (Fig. 4). Using $V_r$ bins 0.05 m s$^{-1}$ wide, all the points within a given range of $S_{inc}$ are counted. These $V_r$ histograms, plotted for 5-dB increments of $S_{inc}$, become bimodal even for very small $S_{inc}$ (120–125 dB, Fig. 4c), with peaks corresponding to air and snow (Figs. 4c–e), and eventually for rain at large enough $S_{inc}$ and $V_r$ (Figs. 4f and 4g). The velocities corresponding to the peak representing snow are plotted in Fig. 3a. The largest values of $S_{inc}$ represent the bright band and have $V_r$ between 1.2 and 2.3 m s$^{-1}$ (Fig. 4f), values characteristic of large snow aggregates that have developed a liquid film, but have not accelerated greatly. The minimum separating the snow and air peaks in the velocity histograms, as measured from these diagrams, is at velocities between 0.25 and 0.35 m s$^{-1}$. This range in $V_r$ approximately represents the optimum velocity threshold separating air and precipitation echoes, as described later in section 4d.

This result is summarized in Figs. 5a and 5b, which combine all values of $S_{inc}$. From these figures it is apparent that the overall minimum in $V_r$ between the Rayleigh- and Bragg-scattering peaks is about $V_r = 0.35$ m s$^{-1}$. This is somewhat smaller than the smallest threshold value calculated from a climatology of $C_n^2$, that is, $V^*_r = 0.50–0.90$ m s$^{-1}$ (Ralph 1995). As can be inferred from Fig. 5b, the value of 0.35 m s$^{-1}$ would allow some measurements of vertical air motion to be interpreted as snow because the largest values of the distribution of points corresponding to air likely extend to $V_r > 0.35$ m s$^{-1}$. Thus, using $V_r = 0.50$ m s$^{-1}$ would reduce the likelihood of making this error, but at the same time would incorrectly classify some points as air, when in fact they are snow. This implies that the $V_r$ threshold for detecting snow should be selected on the basis of whether or not the application depended most on a high probability of detecting snow, or on a low false-alarm ratio.

2) Comparison with jet stream cirrus and a mesoscale convective system

To assess the variability of the significant features found in the correlation diagrams and velocity histograms, two other types of precipitation events (Figs. 5c–f) and two clear-air cases (Figs. 5g–j) are examined in addition to the winter storm case described above. The two additional precipitation cases represent both larger and smaller precipitation rates than those associated with the winter storm. Both events are described in some detail in Ralph et al. (1995). An event in which no precipitation was observed at the ground (the jet stream cirrus case), but in which there was convincing evidence of snow in the upper troposphere is shown in Figs. 5c and 5d. This case also clearly exhibits the snow and air signatures found in the winter storm case (cf. Figs. 5a and 5c) but lacks evidence of rain. In this sense it resembles virga documented by Gage et al. (1994) and possibly represents a region where some satellite-based estimations of rainfall may be in question. While certain ranges of $S_{inc}$ exhibit a minimum in the velocity histogram (inset to Fig. 5d) similar to that found in the winter storm case, the velocity histogram containing all points has no minimum, just a hump that makes the total distribution asymmetric about $V_r = 0$ (Fig. 5d). In a case dominated by the passage of a mesoscale convective system (MCS) containing approximately 42 min of convective rain and more than 2 h of stratiform rain measured at the ground, there was more intense precipitation than in the winter storm (59 mm in 3 h, compared to 6.5 mm in 18 h), there is more of a separation between the snow and air data points, and the data corresponding to rain extend to much larger fall velocities (i.e., 9 m s$^{-1}$, compared to 6 m s$^{-1}$; Fig. 5e). The velocity histogram for the MCS (Fig. 5f) has a minimum like the winter storm case, but at $V_r = 0.75$ m s$^{-1}$, rather than 0.35 m s$^{-1}$, probably because very few Rayleigh-dominated points had $V_r < 0.75$ m s$^{-1}$ in the MCS, as discussed below.
It also has a prominent maximum at \( V_r = 8.25 \text{ m s}^{-1} \) corresponding to rain.

In the MCS data, the region of snow appears as a clump of points more separated from the Bragg-scattering points than in the other two cases. This may be due to the fact that the clouds and smaller snow particles, for which \( S_e \) and \( V_r \) are small, were found mostly above the 10-km maximum height of data used in this plot. In the MCS several points with large values of \( S_e \) are found to have upward velocities, and likely represent hydrometeors within a strong convective updraft.

If minima in the partial velocity histograms (i.e., histograms constructed from a subset of \( S_e \) values, as in Figs. 4c–g, and in the insets to Figs. 5b, 5d, and 5f), or in the total velocity histograms (i.e., those containing all values of \( S_e \), as in Figs. 5b, 5d, and 5f), are interpreted as vertical velocity thresholds, then the thresholds would be 0.35, 0.35, and 0.30 or 0.75 m s\(^{-1}\) (depending on whether partial or total histograms are used), for the winter storm, jet stream cirrus, and MCS cases, respectively.

Two anomalous features resulting from receiver saturation, which is due to strong backscatter from heavy precipitation [see Ralph et al. (1995) for a discussion of this effect], are present in the correlation diagram for the MCS (Fig. 5e): 1) many points are artificially forced together near the top of the diagram, and 2) an unrealistic layering of points at particular values of \( S_e \) occurs in the rain. The layering results from correcting for range in conditions when \( S \) has been forced to a nearly uniform value over a range of heights because the backscattered power from precipitation at each
Fig. 5. Scatterplots of unedited range-corrected signal power $S_r$ and radial vertical velocity $V_r$ (a, c, e, g, and i), and $V_r$ velocity histograms (b, d, f, h, and j). (a) and (b) Winter storm case shown in Fig. 2. (c) and (d) Snow in jet stream cirrus observed by the Haskell, Oklahoma, profiler (0700–1900 UTC 6 December 1990). (e) and (f) Stratiform and convective precipitation in a mesoscale convective system observed by the Hillsboro, Kansas, profiler (0200–1300 UTC 14 May 1992). (g) and (h) Observations made by the Lanthrop, Missouri, profiler under conditions without clouds or precipitation (1200 UTC 11 October–0900 UTC 12 October 1994). (i) and (j) Observations made by the Platteville, Colorado, profiler under conditions without clouds or precipitation, but with strong trapped lee wave activity (0000–1800 UTC 8 April 1992). Insets in (b), (d), and (f) are as in Fig. 4 but for values of $S_r$ that most clearly show the two peaks identified as representing snow and air.
range has saturated the receiver. This layering effect is evident in the rain portion of Fig. 5e, where the lowest layer (near $S_C = 140$ dB) is from the lowest gate, the second layer ($S_C = 146$ dB) from the second gate, and the third layer ($S_C = 148$ dB) from the third gate, and there is some evidence of similar layering up to the sixth gate. It is important to note, however, that the total velocity histogram (Fig. 5f) is unaffected by these errors.

3) COMPARISON WITH TWO CLEAR-AIR EVENTS

The two clear-air cases were selected to represent the extreme range of clear-air conditions encountered in the WPDN. One case (Figs. 5g and 5h) consists of measurements made far from mountains, in conditions with no precipitation or clouds within 200 km. The second case (Figs. 5i, j) is from a location just downstream of the Colorado Rocky Mountains on a day when trapped lee waves with strong clear-air vertical motions (1–2 m s$^{-1}$) were present (Ralph et al. 1992), as was evidenced by trapped lee wave clouds that developed just after the end of the time interval used in Figs. 5i and 5j. In both of these cases there were no clouds overhead (based on surface observations and satellite imagery). Although these two cases cover the extreme ranges of clear-air events, from one with near-zero vertical air motion over all heights and times (Fig. 5h) to one with a wide range of vertical air motions and hence a wide distribution of points in the total velocity histogram (Fig. 5j), neither has any evidence of a group of points that has fall velocities of rain, or correlations indicative of snow. This comparison indicates that the presence of correlation between $S_C$ and $V_r$ in the scatterplots, combined with an asymmetry of the velocity histograms for $V_r$ with magnitudes greater than 0.25 m s$^{-1}$, can be used to identify when a substantial number of radar wind profiler measurements, during a given time and height range, are dominated by Rayleigh scattering, and thus indicate the presence of precipitation.
c. Error analysis

Because the threshold velocities for snow, derived either from the climatology method or from the velocity histograms, may overlap the range of frequently observed vertical air motions, it would be very useful to estimate what fraction of the data is misinterpreted for a given velocity threshold. A technique is presented here that provides just such an estimate.

For precipitation events the distribution of radial vertical velocities $n(V_r)$ within the velocity histograms (Figs. 5b, 5d, and 5f) is the sum of primarily two distributions: one corresponding to measurements dominated by Rayleigh scattering from precipitation $n_p(V_r)$, which thus represents fall velocities of hydrometeors, and another corresponding to measurements dominated by Bragg scattering from air $n_a(V_r)$, which thus represents vertical air motions. There are also likely to be a few measurements for which the backscattered power from the air and from the snow are nearly equal, and thus $V_r$ does not well represent either form of scattering. Measurements for which the backscatter power was at the noise level can be removed by correlating $V_r$ and $S$ (i.e., signal power before range correction), and finding the threshold value of $S$ below which $V_r$ is approximately randomly distributed over a wide range of $V_r$. This threshold appears to be about 28–30 dB in the five cases studied here, with some apparently good measurements discarded if 30 dB were used. By determining the average vertical air motion $V_a$ from the peak representing air in a $V_r$ histogram, it is possible to transform the measured radial vertical velocity $V_r$ into an approximately air-relative velocity $V = V_r - V_a$, which is useful when considering precipitation. This transformation is only approximate because it assumes that the vertical air motions outside the regions dominated by Rayleigh scattering are similar to those within the precipitation region. Now,

$$n(V) = n_p(V) + n_a(V).$$

(1)

This is shown in schematic form for conditions with snow and air only (Fig. 6a). Outside convective precipitation, where $V_a$ is a poor estimate of the vertical air motion relevant to a particular $V_r$ measurement, it is reasonable to assume that all measurements dominated by Rayleigh scattering have velocities such that $V > 0$ (i.e., the approximate air-relative velocity is downward). This is consistent with the fact that hydrometeors small enough to have near-zero fall velocity do not occur in large enough concentrations to produce a measurable amount of backscatter (Rogers et al. 1994). Under this assumption the distribution of measurements with $V < 0$ represents only vertical air motion measurements, that is, measurements dominated by Bragg scattering. Thus, it is possible to infer the half of the vertical air motion distribution with $V > 0$, from the half of the vertical air motion distribution with $V < 0$, a step that appears justifiable based on the ap-

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Fig. 6. (a) Schematic representation of the combination of the two primary radial vertical velocity $V_r$ distributions that form a total $V_r$ histogram (dashed) in a dataset containing a substantial number of measurements dominated by Bragg scattering from air and by Rayleigh scattering from snow. Here, $V_r$ is the threshold vertical velocity used to distinguish between measurements representative of snow and air; $V_a$ is the value of $V$, at the peak of the air portion of the distribution. Half the total number of Bragg-scattering points ($0.5N_p$, hatched), the Rayleigh-scattering points misinterpreted as due to Bragg scattering ($N^*_p$, light shading), and the Bragg-scattering points misinterpreted as due to Rayleigh scattering ($N^*_a$, dark shading) are shown. (b) The fraction of Bragg-scattering points misinterpreted as due to Rayleigh scattering ($f^*_a$ solid), and of Rayleigh-scattering points misinterpreted as due to Bragg scattering ($f_a^*$ dashed) as a function of $V_r$ for three precipitation cases shown in Fig. 5.
proximate symmetry that should be expected and that is evident in the clear-air cases (Figs. 5h and 5j). Although the vertical air motion distribution should not be expected to be perfectly symmetric, by making the sampling intervals long enough, it becomes more likely that the frequency of occurrence of upward and downward motions would nearly balance each other, except in conditions where updrafts and downdrafts systematically differ in size. The inference of symmetry can be expressed as

\[ n_a(V) = n_a(-V). \]  \hfill (2)

This result can then be used to calculate what percentage of the total number of measurements of vertical air motion have a magnitude greater than the velocity threshold \( V_T \) or \( V_T^\pm \) chosen to distinguish between Bragg-scattering- and Rayleigh-scattering-dominated measurements.

The total number of points dominated by Bragg scattering from air \( (N_a) \) is simply twice the number of points with \( V_a < V_a^\pm \), where only half the points in the velocity bin containing \( V_a \) are included before multiplying by 2 in this calculation. The total number of these points with \( V_a > V_T \), referred to as \( N^*_a(V_T) \), is equivalent to the number of points with \( V_a < V_T^\pm \), where \( V_T^\pm = V_a - (V_T - V_a^\pm) \). Thus, the fraction of Bragg-scattering points that are likely to have \( V_a > V_T \), that is, the fraction of Bragg-scattering points that are misinterpreted as Rayleigh-scattering points, is given by

\[ f_a(V_T) = \frac{N^*_a(V_T)}{N_a}. \]  \hfill (3)

Here \( N^*_a(V_T) \) is shown schematically in Fig. 6a. Table 1 gives \( f_a(V_T) \), and the supporting information for each, from the three precipitation cases shown in Fig. 5. Velocity thresholds \( V_T \) determined from the minimum in the histograms in Fig. 5, and from analyses such as shown in Fig. 4, correspond to \( f_a = 0.07, 0.04, \) and \( 0.01 \) for the freezing rain, jet stream cirrus, and MCS cases, respectively (Table 1).

Based on the knowledge of \( n_a(V) \) from Eq. (2), it is also possible to estimate the number of measurements dominated by Rayleigh scattering \( n_p(V) \) through

\[ n_p(V) = \begin{cases} n(V) - n_a(V), & V > 0, \\ 0, & V < 0. \end{cases} \]  \hfill (4a)

\[ n_p(V) = \begin{cases} 0, & V > 0, \\ n_a(V), & V < 0. \end{cases} \]  \hfill (4b)

Summing \( n_p(V) \) over all \( V > 0 \) yields an estimate of the total number of measurements dominated by Rayleigh scattering \( (N_p) \). From this information it is possible to determine the number of Rayleigh-scattering points misinterpreted as Bragg-scattering points \( N^*_p(V_T) \) for a given velocity threshold. This is shown schematically in Fig. 6a. The fraction of all Rayleigh-scattering points that are misinterpreted as Bragg-scattering points is given by

\[ f_p(V_T) = \frac{N^*_p(V_T)}{N_p}. \]  \hfill (5)

Velocity thresholds \( V_T \) determined from the minimum in the histograms in Fig. 5, and from analyses such as shown in Fig. 4, correspond to \( f_p = 0.02, 0.08, \) and \( 0.27 \) for the freezing rain, jet stream cirrus, and MCS cases, respectively.

In the MCS case it appears that \( f_a \) is overestimated slightly and \( f_p \) underestimated slightly because 17 data points dominated by Rayleigh scattering had upward velocities because they were in updrafts within the convective region and 8 points from the stratiform region had upward velocities but were contaminated by interference. These conclusions are substantiated in the next section. Because these points had \( V < 0 \), they were considered as Bragg-scattering points in the calculations performed above. Based on the total number of Bragg- and Rayleigh-scattering points, 1499 and 2201,

<table>
<thead>
<tr>
<th>Case</th>
<th>( N_a ) and ( N_p )</th>
<th>( V_T ) (m s(^{-1}))</th>
<th>( f_a )</th>
<th>( f_p )</th>
<th>( N_a^* )</th>
<th>( N_p^* )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter storm</td>
<td>4335 and 4227</td>
<td>0.35</td>
<td>0.07</td>
<td>0.02</td>
<td>295</td>
<td>95</td>
</tr>
<tr>
<td>Winter storm</td>
<td>4335 and 4227</td>
<td>( 0.39 ) (opt)</td>
<td>0.05</td>
<td>0.05</td>
<td>199</td>
<td>94</td>
</tr>
<tr>
<td>Winter storm</td>
<td>500</td>
<td>0.50</td>
<td>0.02</td>
<td>0.13</td>
<td>87</td>
<td>563</td>
</tr>
<tr>
<td>Jet stream cirrus</td>
<td>2917 and 686</td>
<td>( 0.34 ) (opt)</td>
<td>0.04</td>
<td>0.04</td>
<td>128</td>
<td>30</td>
</tr>
<tr>
<td>Jet stream cirrus</td>
<td>2917 and 686</td>
<td>0.35</td>
<td>0.04</td>
<td>0.08</td>
<td>105</td>
<td>58</td>
</tr>
<tr>
<td>Jet stream cirrus</td>
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<td>0.01</td>
<td>0.35</td>
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<td>240</td>
</tr>
<tr>
<td>MCS</td>
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<td>( 0.10 )</td>
<td>145</td>
<td>214</td>
</tr>
<tr>
<td>MCS</td>
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<td>0.09</td>
<td>0.11</td>
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<td>232</td>
</tr>
<tr>
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<td>1499 and 2201</td>
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<td>0.04</td>
<td>0.18</td>
<td>54</td>
<td>405</td>
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<tr>
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<td>0.01</td>
<td>0.27</td>
<td>14</td>
<td>592</td>
</tr>
</tbody>
</table>
respectively, in the MCS event (Table 1), it appears that 17–25 points would change \( f_a \) and \( f_p \) by 0.008–0.017.

d. Dual-optimization of \( V_T \) based on error estimates

Based on the analysis described above it is possible to adjust \( V_T \) so as to minimize \( f_a \) or \( f_p \), depending on the application. Figures 6a, 6b, and Table 1 are useful for this purpose. For example, if it is desired to capture 90% of all measurements likely to be representative of precipitation fall velocities in the winter storm case, then one would choose \( V_T = 0.46 \text{ m s}^{-1} \), because it corresponds to \( f_p = 0.10 \) in that case (Fig. 6b). It would also be possible to estimate that 2.7% of the Bragg-scattering points would be incorrectly interpreted as Rayleigh-scattering points for this choice of \( V_T \) (i.e., \( f_a = 0.027 \)). Conversely, if one needs to identify events that are due to Bragg scattering from air in the winter storm case, without including more than 2% of the measurements dominated by Rayleigh scattering, then \( V_T = 0.34 \text{ m s}^{-1} \) should be selected because \( f_p = 0.02 \) for that choice (Fig. 6b). This choice would result in capturing 92.7% of all the Bragg-scattering points (i.e., \( f_a = 0.073 \)).

An alternative optimization of the threshold could be based on finding \( V_T \) for which both \( f_a \) and \( f_p \) are minimized. This can be approximated by the velocity threshold for which \( f_a = f_p \). This is evident in Fig. 6b where the curves for \( f_a \) and \( f_p \) cross for a particular event. These results are summarized in Table 1. It appears that this dual optimization results in objectively determined thresholds between 0.28 and 0.39 m s\(^{-1}\), with \( f_a = f_p \) between 0.04 and 0.10, depending on the event. Results based on the minimum threshold value of 0.50 m s\(^{-1}\) determined from the climatological method described earlier are also shown in Table 1 for comparison. The resulting values of \( f_a \) and \( f_p \) indicate that this climatology-based threshold tended to avoid misinterpreting Bragg-scattering points as due to Rayleigh scattering (i.e., \( f_a \) was relatively small); at the expense of misidentifying more Rayleigh-scattering points as Bragg-scattering points (i.e., \( f_p \) was relatively large). Overall statistics for the three cases combined show that a total of 5.7% of all points above the noise level were incorrectly identified as due to Bragg or Rayleigh scattering when the dual optimization method was used, compared to 8.7% for the uniform 0.50 m s\(^{-1}\) climatological threshold.

It is important to recognize that these calculations should be made from datasets large enough to contain significant numbers of both Bragg- and Rayleigh-dominated measurements. At the same time, the interval chosen should not be so large so as to include too much variability in background conditions. In this paper, 12–30-h time intervals were used and included data only from the 36 range gates of the low-altitude mode of the WPDN profilers. Intervals in which contamination by migrating birds (Wilczak et al. 1995), ground clutter, etc. make up too much of the data should also be avoided. These factors will influence how well this method can be applied in real time, although once a few hours of precipitation can be identified in an event, it should be possible to use the threshold velocity estimated from that data to interpret the new data on a real-time basis until the precipitation event ends. A better approach for real-time application may be to preselect a threshold velocity based on the climatology method (Ralph 1995; Ralph et al. 1995) but adjusted based on the results of examples from this paper.

Future studies that make use of independent measurements of when precipitation is present in each pulse volume need to be carried out both to assess the accuracy of the error estimation method presented here, and to determine what the optimal time interval is for use in this method. It should be emphasized that this error estimation technique is likely to fail in conditions where large clear-air vertical motions are common, as in mountain waves. However, the vertical velocity thresholding technique itself is not intended for use in such circumstances (Ralph 1995). In addition, the optimal threshold would likely be affected by the characteristics of the particular profiler (e.g., the beamwidth and sidelobe location and strength) and characteristics of the site (ground clutter).

5. Velocity variance as an indicator of precipitation type

Although radial velocity and range-corrected signal power provide convincing evidence of the presence of precipitation and of precipitation type, additional information is available in the variance of the radial vertical velocity (\( \sigma^2 \)). The variance used here is determined from the width of the peak in the Doppler power spectrum containing the mean radial velocity (van de Kamp 1988; Barth et al. 1994; Ralph et al. 1995). Other recent work has recognized that variations of velocity variance can be closely related to characteristics of precipitation (e.g., Steiner and Richner 1994; Ralph et al. 1995; Williams et al. 1995), including the fact that it is enhanced in convective regions, and in rain. Additional factors can strongly influence \( \sigma^2 \) in the vertical beam, such as the strength and vertical shear of the horizontal wind, the width and length of the radar pulse, the duration of the averaging, inhomogeneities in the vertical velocity within the pulse volume, and the strength of turbulence within the pulse volume. Although none of these factors is accounted for in this study, the results shown below nonetheless reveal distinctive patterns in \( \sigma^2 \) that are closely related to precipitation type, suggesting that these other effects are often less significant than those due to precipitation.

To illustrate this behavior, the data shown in Figs. 5a, 5c, and 5e are stratified according to \( \sigma^2 \), using 0–1, 1–2, 2–3, 3–4, 4–5, and 5–10 m\(^2\) s\(^{-2}\) as convenient
ranges established from experience (Fig. 7). For the purpose of exploring differences between convective and stratiform precipitation, the MCS case has been split into two subsets (Figs. 8a and 8b)—one containing only the stratiform portion of the storm and another only the convective portion—as determined from observations of precipitation rate at the ground. While these $\sigma^2$ ranges are useful in the cases discussed here, it is likely that more refined ranges would be more useful. It should be noted, however, that the magnitudes of these ranges are sensitive to the radar and sampling characteristics (e.g., beamwidth, pulse length, dwell time, etc.) and hence should be considered only as rough guidelines applicable to the WPDRN 404-MHz profilers.

As discussed earlier, measurements from above 2.5 km in the winter storm case contain no rain because temperatures at those altitudes are well below freezing. Examination of $\sigma^2$ in this subset of data (Fig. 3c) reveals that very few data points in snow have $\sigma^2 > 1.0$ m$^2$ s$^{-2}$ or conversely that greater than 99% of the measurements in snow have $\sigma^2 < 1.0$ m$^2$ s$^{-2}$. A similar result is evident in the snow portion of the measurements from below 2.5 km (Fig. 3d). In contrast, approximately 90% of the measurements from below 2.5 km with $V_r$ indicative of rain (i.e., $V_r > 3$ m s$^{-1}$) have values of $\sigma^2 > 1$ m$^2$ s$^{-2}$. Comparison of this result with Figs. 7 and 8 suggests that $\sigma^2 = 1.0$ m$^2$ s$^{-2}$ can be used as an approximate threshold to distinguish between rain and snow outside convection. One of the likely reasons that the combined thresholding technique of Steiner and Richner (1994) succeeded in rain but failed to identify snow is that the technique depended too heavily on enhanced velocity variance, as well as on relatively large fall velocities.

By considering only the times during the MCS case for which precipitation was simultaneously observed at the surface, it is possible to focus on data for which there should be no measurements dominated by Bragg scattering (Fig. 8). This interpretation can be made because of the strong precipitation rates observed, and because of the great depth of the storm (i.e., clouds were observed by satellite at heights exceeding the 10-km maximum height of the profiler data used). The stratiform region (Fig. 8a) shows the characteristic signatures seen in the winter storm case for which the precipitation was also stratiform, except that a few points have $V_r$ characteristic of Bragg scattering (i.e., $V_r < 0.30$ m s$^{-1}$). Because these few points have unusually large $\sigma^2$ (i.e., $\sigma^2 > 5.0$ m$^2$ s$^{-2}$), and occurred at the highest gates of the low mode, it appears that these points result from contamination of some type, most likely due to sidelobes. Careful examination of the winter storm (Fig. 7a), jet stream cirrus (Fig. 7b), and stratiform portion of the MCS (Fig. 8a) indicate that $\sigma^2$ did not exceed 5.0 m$^2$ s$^{-2}$, except for a few outlier points that are likely due to contamination by sidelobes. Thus, it appears that $\sigma^2 > 5.0$ m$^2$ s$^{-2}$ does

Fig. 7. Scatterplots of unedited range-corrected signal power $S_r$ and radial vertical velocity $V_r$: (a) as in Fig. 5a, (b) as in Fig. 5c, and (c) as in Fig. 5e. Velocity variance is represented by six ranges, each distinguished by its own symbol.
not typically occur in stratiform precipitation. In contrast, the convective region (Fig. 8b) contained a large number of points for which $\sigma^2 > 5.0 \text{ m}^2 \text{ s}^{-2}$. These points, along with all the others are spread over a wide range of $V_r$ and show little of the patterns found in the stratiform cases. In addition, there is a tendency for the measurements with $\sigma^2 > 5.0 \text{ m}^2 \text{ s}^{-2}$ to have very large $S_{rc}$, which may be indicative of unusually large wetted hydrometeors such as hail or graupel, and even a tendency, relative to points with smaller $\sigma^2$, to have more velocities that are suggestive of upward air motion (Fig. 8b). These factors are consistent with the interpretation that points with $\sigma^2 > 5.0 \text{ m}^2 \text{ s}^{-2}$ are indicative of convective regions within precipitation.

6. Conclusions

Correlations between vertical radial velocities $V_r$ and range-corrected signal power $S_{rc}$ measured by the vertically pointing beam of 404-MHz radar wind profilers can be used to determine whether or not precipitation is present in profiler measurements within a chosen time and height range, and what type of precipitation it is. This behavior results from the fact that fall velocities and range-corrected signal power are well correlated in snow, but not in air, and that rain with large enough reflectivity factor has velocities that are outside the range of vertical air motions normally observed outside strong convective cells. Comparison of these correlation diagrams from a wide range of precipitation events—that is, light rain and snow in a winter storm, heavy rain and snow (aloft) in both convective and stratiform portions of a mesoscale convective system, and very light snow (aloft) in jet stream cirrus—with two cases of purely clear-air measurements support the conclusion that snow and rain are readily identifiable in such diagrams.

A method is introduced that objectively determines a threshold radial vertical velocity on a case-by-case basis that can be used to distinguish measurements dominated by Bragg scattering from those dominated by Rayleigh scattering. These thresholds were found to be 0.28–0.39 m s$^{-1}$ downward, which is somewhat smaller than the 0.5 m s$^{-1}$ minimum threshold calculated for 404-MHz radars from climatological measurements of $C_n^2$ (Ralph 1995). Because vertical air motions can regularly exceed this threshold, a method was developed to estimate the errors associated with using a threshold. This method is based on histograms of vertical radial velocity for each case. In precipitation events, a significant asymmetry was present about the maximum in the vertical velocity histograms, which was associated with measurements dominated by Bragg scattering. Such an asymmetry was not evident in the clear-air cases. Under the assumption that measurements with upward velocities were never dominated by Rayleigh scattering from precipitation, overall statistics for the three cases combined show that a total of 5.7% of all points above the noise level were incorrectly identified as due to Bragg or Rayleigh scattering when a dual optimization method was used to select the velocity threshold, compared to 8.7% for the uniform 0.50 m s$^{-1}$ climatological threshold. Examination of the strong convection case, where the assumptions were invalid, suggests that these error estimates underestimate the actual errors, but only by about 1%, for a 12-h-long dataset with 0.7 h of convective precipitation.

It is also shown that very few measurements of snow had velocity variance greater than 1.0 m$^2$ s$^{-2}$, while very few measurements of rain had velocity variance less than 1.0 m$^2$ s$^{-2}$. Thus, a velocity variance of 1.0 m$^2$ s$^{-2}$ is a good threshold to distinguish between rain and snow in data from the WPDN profilers, provided the two other spectral moments have unambiguous signatures of precipitation, as had been suggested by Ralph et al. (1995). Unlike the regions and cases where
stratiform precipitation dominated the signal (as defined by surface precipitation rate measurements), the convective region had many measurements with velocity variance greater than 5.0 m² s⁻². These measurements also had vertical velocities that suggested they were correlated with updrafts. Thus, this appears to be a useful threshold to identify convective precipitation in the one case observed here.

Although most of the conclusions made here are based on examination of several cases chosen to represent a range of conditions, it may be important to explore their applicability in more events. Because the error estimates are based on assumptions that may be invalid under some circumstances, it is also important to test the results using an independent method to determine the presence or absence of clouds for each range gate and time. This may be possible using data from very short wavelength radars with high vertical resolution, such as a K-band radar, which is very sensitive to clouds, and insensitive to Bragg scattering from air (e.g., Martner and Kropfli 1993).

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REFERENCES


