A Method for Estimating Particle Fall Velocities from Vertically Pointing Doppler Radar

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ABSTRACT

A method is presented that estimates particle fall velocities from Doppler velocity and reflectivity measurements taken with a vertically pointing Doppler radar. The method is applicable to uniform, stratified clouds and is applied here to cirrus clouds. A unique aspect of the technique consists of partitioning the Doppler velocities into discrete cloud height and cloud reflectivity bins prior to temporal averaging. The first step of the method is to temporally average the partitioned Doppler velocities over an hour or two to remove the effects of small-scale vertical air motions. This establishes relationships between particle fall velocity and radar reflectivity at various levels within the cloud. Comparisons with aircraft in situ observations from other experiments show consistency with the remote-sensing observations. These results suggest that particle fall speeds can be determined to within 5–10 cm s^{-1} by means of this technique.

1. Introduction

Short wavelength (<1 cm) cloud-sensing Doppler radars can provide detailed information on the structure and evolution of cloud systems. In the vertically pointing mode, these radars provide a direct measurement of the reflectivity-weighted velocity composed of the particle fall velocity plus vertical air motion. Because of their short wavelength, cloud radars are generally insensitive to atmospheric Bragg scattering and provide no direct measure of air motion when hydrometeors are absent. There are times, however, when it would be desirable to separate the particle fall speed in still air from air motions, especially in stratiform cloud systems in which particle and air motions are of comparable magnitude. A technique is introduced in this article that uses a vertically pointing, cloud-sensing Doppler radar to estimate the reflectivity-weighted particle terminal fall velocities in horizontally uniform cloud layers.

Data collected during the First ISCCP (International Satellite Cloud Climatology Program) Regional Experiment (FIRE II) in Coffeyville, Kansas, in 1991 are used to demonstrate the technique. The data were collected with a dual-polarization, K_a-band Doppler radar operated by the National Oceanic and Atmospheric Administration's (NOAA) Environmental Technology Laboratory (ETL). This radar and its applications have been described elsewhere (Kropfli and Kelly 1996; Martner and Kropfli 1993), and the details of its operation will not be discussed here other than to list parameters relevant to this study in Table 1.

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2. Principles of the technique

The radial velocity measured by a vertically pointing Doppler radar is the reflectivity-weighted particle vertical velocity composed of air motion and particle terminal fall velocity in still air,

$$V_D = V_a + V_t, \tag{1}$$

where V_D is the measured Doppler velocity, V_a is air motion, and V_t is the terminal fall velocity of the hydrometeors. Atmospheric hydrometeors reach their terminal fall velocity very quickly, and therefore, any reference to fall velocity in this article implies terminal fall velocity. It is assumed in (1) that V_D is from a short wavelength cloud radar (e.g., K_a band) that is not sensitive to clear-air Bragg scattering. The underlying assumption in this technique is that by averaging over a sufficiently long period of time, t_a, small-scale vertical air motions within the cloud will average to near zero and the resulting average Doppler velocity will be an estimate of the reflectivity-weighted terminal fall velocity ($V_D \approx V_t$). This necessitates a somewhat persistent cloud system such that sufficient temporal averaging at a given altitude is possible. Strongly convective cloud systems in which vertical air motions would not be expected to average down to zero are obviously excluded. This method is optimal for persistent and uniform stratiform cloud systems. Here we apply it to long-lived

TABLE 1. Operating characteristics of the ETL K_a-band Doppler radar.

Wavelength	8.7 mm
Beamwidth	0.5°
Pulse length	37.5 m
Gate spacing	37.5 m
Temporal resolution	0.25 s
Sensitivity	−30 dBZ at 10 km

cirrus clouds, but it may also be applicable to other stationary stratiform cloud systems.

The average vertical motions in stratiform clouds can be very small, approaching a few centimeters per second. Fall velocities of cloud particles can be equally small, but most often, as shown below, they are much larger. The K_a-band radar has the accuracy and precision to measure Doppler velocities of this magnitude; however, antenna-pointing angle errors become critical. If the antenna is not exactly vertical, the horizontal wind will induce an erroneous vertical velocity component. Figure 1 illustrates the magnitude of this apparent vertical velocity for a range of off-zenith pointing angles and horizontal wind speeds using the assumption that the beam is tilted in the downwind direction of horizontal wind. For a pointing error of only 0.2° and a horizontal velocity of 30 m s^{-1}, an apparent vertical velocity of 10 cm s^{-1} will be ''measured'' by the radar. For the measurements discussed here, extreme care was taken to ensure an accurate antenna-pointing angle by adjusting the level of the antenna platform daily. We feel that the pointing angle is normally within 0.1° of the vertical, and therefore, errors from this source should be less than 2–3 cm s^{-1} for typical horizontal wind speeds encountered during FIRE II.

Standing waves and slowly propagating gravity waves are also a potential source of error in the averaged vertical velocities. Standing waves are somewhat problematic without supporting measurements of the clear-air velocity of the cloud environment. This is of most concern downwind of significant topographic features. The measurements in this study were taken over a relatively flat, midcontinental region in which this source of error is minimal. In these measurements the radar was scanned in an over-the-top RHI mode every 30 min, which produces a 2D, vertical cross section of the cloud reflectivity and velocity structure. These data are useful in identifying wave motions within the cloud. Data used here did not exhibit any wave activity.

An important aspect of this method is the way in which the Doppler velocities are averaged. The Doppler velocity data are partitioned into cloud height (Δh) and radar reflectivity (ΔdBZ) bins prior to averaging. As an example, Δh might be a 500-m layer, and ΔdBZ a 2-dBZ reflectivity range. Cloud height binning (Δh) is performed between minimum and maximum absolute altitudes that bracket the cloud layer for the duration of the average period. The partitioned Doppler velocities at each layer are averaged over each reflectivity bin,

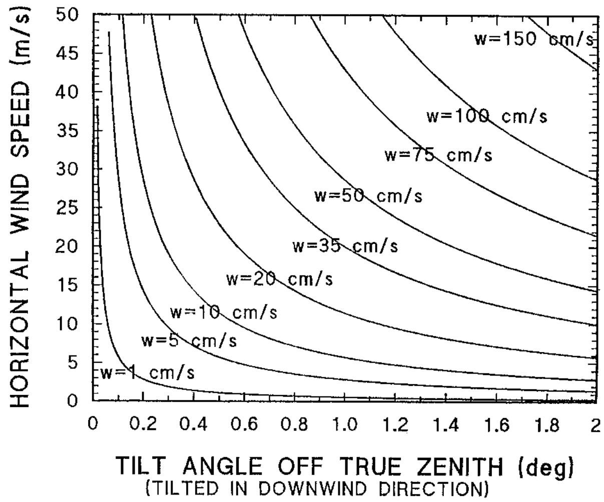

FIG. 1. Illustration of apparent vertical velocity induced by errors in zenith-pointing angle as a function of horizontal wind speed. Curves represent constant vertical velocity error (cm s^{-1}) and assume the beam is tilted in the downwind azimuth direction of the horizontal wind.

resulting in averages of velocity as a function of reflectivity at each height. This binning of the data is important for diagnosing vertical cloud structure associated with changes in cloud microphysics, especially for ice clouds, for which particle shape and density can have significant influence on fall velocity for a given reflectivity.

Although averaging over a long time period, t_a, will reduce the effects of small-scale vertical motions, there is the possibility that the small mean upward air motion required to maintain the cloud could bias the velocity averages. However, the existence of persistent mean upward air motion present within the cloud may at times be deduced from the weakest reflectivity samples representing the smallest particles. If the terminal velocities of these particles are less than the mean upward air motion, these small particles will have an upward motion. Because of the strong sensitivity of radar to large particles, weak reflectivities associated with small particles having very small fall velocities are usually observed near the cloud top in cirrus clouds where particle generation typically occurs. Our experience shows that reflectivities less than −25 dBZ are appropriate for this to work. Only a few averaging bins included ''upward''-moving particles in the highest levels of the cloud with magnitudes typically less than 8 cm s^{-1}. As cloud particles grow, both fall velocity and reflectivity increase. As a result, there were no cases of upward-biased fall velocities observed in lower levels of the cloud, as these would be masked by the larger terminal fall velocity of the hydrometeors.

To systematically correct for these upward motions, it was assumed that this upward motion existed through the entire depth of the cloud. (Data supporting this assumption are presented later.) The magnitude of the mean upward motion was determined by adjusting the upward-moving particles to zero fall velocity. Obviously, all cloud particles have a net fall velocity, and the correction to zero fall velocity is simply used as a conservative correction for large-scale upward motion. This velocity correction is then applied at all cloud levels. It is not possible to make this correction reliably if there are no reflectivities weaker than approximately -25 dBZ present. The possible existence of mean downward air motion is also problematic and cannot be corrected for using this technique, but persistent clouds are not usually associated with subsidence.

The binned averages of V_t are of little use by themselves. These averages must be condensed to a simplified and manageable form from which inferences about cloud structure and microphysics are possible. Two possible methods are suggested, each providing different yet useful results.

The first method suggested for condensing the averaged velocity data is to perform a least squares power-law fit of the averaged V_t data with respect to reflectivity and height. This process results in one equation for the average particle fall velocity at each height interval in the form

$$\langle V \rangle = \alpha(Z_e)^{\beta}, \tag{2}$$

where equivalent reflectivity (Z_e) has been used and α and β are determined from the least squares fit. These equations provide a useful means for examining variations in fall velocity with respect to Z_e at a given height. A comparison of power-law relationships at different heights can yield information about the vertical structure of the cloud system as well.

The other method of consolidating the fall speed data is to perform a multiple linear regression analysis of the entire averaging period t_a using V_t as the dependent variable and height h and reflectivity (dBZ) as the independent variables. However, since we are looking for a linear relationship, we use reflectivity in the form 10 logZ_e (dBZ) for the regression analysis. The linear regression reduces the data to one equation for a given averaging period t_a of the form

$$V_t = a(h) + b(\mathrm{dBZ}) + c, \tag{3}$$

where a, b, and c are coefficients from the regression. This equation can be used to apply an average fall speed correction to the instantaneous Doppler velocity at all heights. Using radar reflectivity as a function of range, the resulting regression equation allows one to produce time–height plots of fall speed (V_t) and/or small-scale vertical air motion (V_a) estimates for the cloud system.

3. Analysis

Analyses were performed to examine the stability and sensitivity of this averaging method. The bin sizes Δh, ΔdBZ, the averaging time, t_a, and a number of parameters specific to the ETL K_a-band radar were varied to examine their effects on the resulting averages. By making Δh and ΔdBZ reasonably small, more detail on cloud structure can be obtained. At the same time there must be a sufficiently large number of data points to obtain reliable velocity averages. The parameters ultimately chosen for these analyses were $\Delta h = 560$ m (15 range gates) and ΔdBZ $= 1$ dBZ. A smaller value of Δh can be obtained at the expense of a larger ΔdBZ; however, this study focuses on relatively weak clouds with reflectivities ranging from approximately -35 to -5 dBZ, and the small value of ΔdBZ was necessary to capture the variation in cloud structure.

Two days were selected from the FIRE II project for testing this technique. The first case was a gradually dissipating cirrus cloud system on 25 November 1991. Figure 2 is a time–height plot of vertical velocity (m s^{-1}) and reflectivity (dBZ), with time increasing from 1230 to 1500 UTC, from left to right. (All times are UTC throughout.) Downward velocities are negative. These data were obtained by pointing the radar antenna vertically while the clouds drifted over the radar. The stratified nature of the cloud reflectivity and the weak, generally downward vertical particle velocities are illustrated in the figure. The cloud base remained relatively constant with time, while the cloud top gradually lowered. Vertical Doppler velocities range from -0.6 (downward) to 0.2 m s^{-1} (upward), and reflectivities range from -35 to -5 dBZ. Sounding data indicate a cloud-base temperature of $-30°$C and cloud-top temperatures around $-50°$C. Gaps in the data record were the result of periodic scanning of the antenna to probe the horizontal structure of the cloud.

The second dataset from FIRE II was 26 November 1991. Figure 3 is a time–height plot of Doppler velocity and reflectivity from 1930 to 2130 UTC 26 November, similar to Fig. 2. This is a portion of a developing cloud system in which the cloud base and top remained relatively constant. This case is more dynamic than the 25 November case, as noted by the stronger and more variable upward and downward velocities and the more complex reflectivity structure. Rawinsonde data indicate a convectively unstable lapse rate at the cloud level. The 26 November case is analyzed using the same averaging and thresholding parameters as 25 November. The cloud-base temperature from sounding data was around $-23°$C, and the cloud top was $-50°$C.

4. Discussion

Results from the power-law fits to the V_t data at each height for the 25 November data are shown in Fig. 4a. Particle terminal velocities are plotted as positive values due to restrictions of the analysis software for both the power-law and regression analysis. The individual averages plotted in these figures have a minimum of 500 points composing each average. The requirement of a minimum of 500 points was used to increase the statistical significance of the averages and to eliminate av-

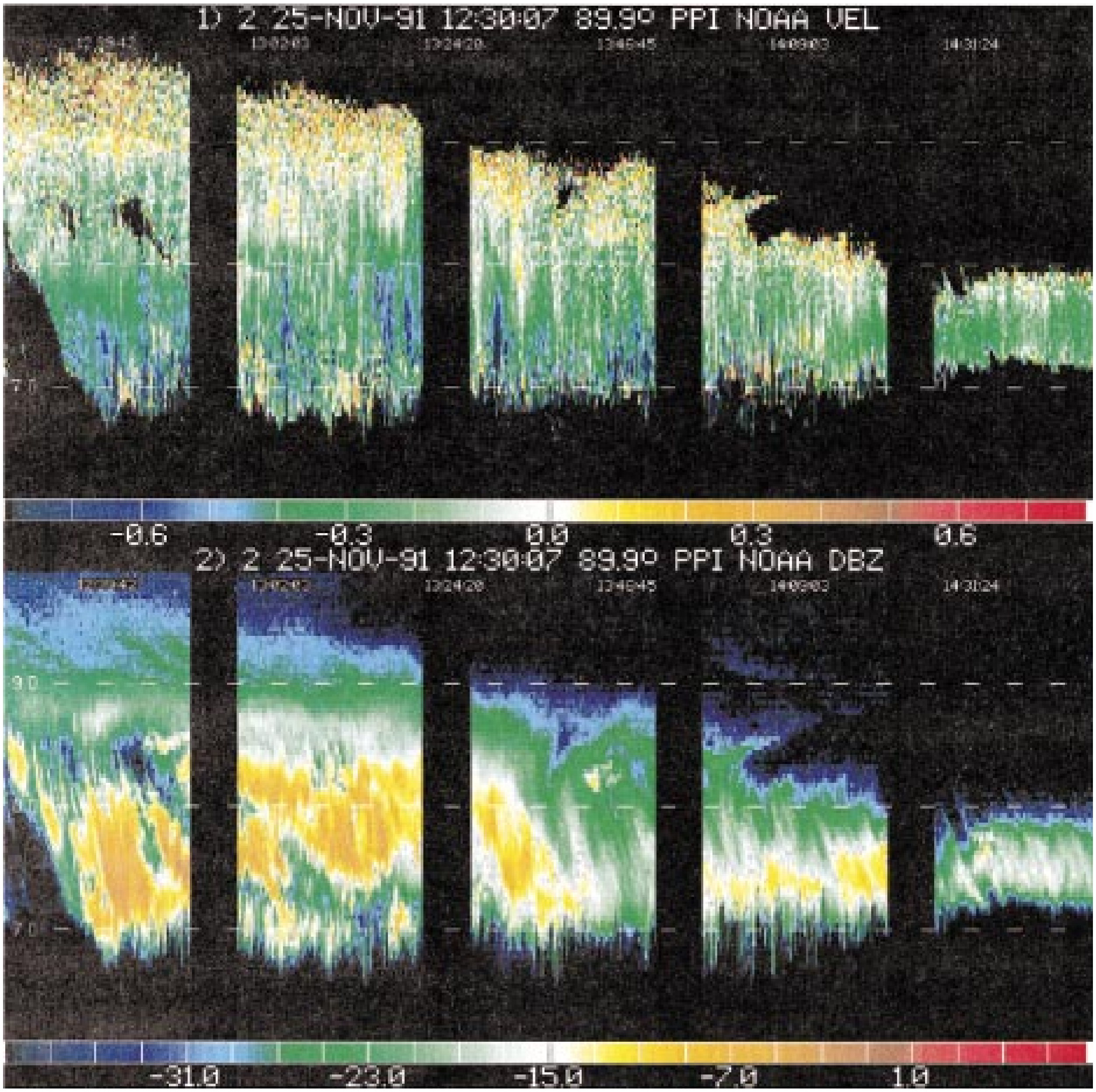

FIG. 2. Time–height plot of Doppler velocity [upper panel (m s^{-1})] and reflectivity [lower panel, (dBZ)] from 1230 to 1500 UTC 25 November 1991. Negative velocities are downward.

erages with relatively high variances, especially at cloud boundaries where the signal-to-noise ratio was low. The cloud signal near cloud boundaries also tended to be intermittent as the cloud base/cloud top would rise or fall outside the limits of the height interval. The standard deviations of the errors about the best-fit curve are less than 5 cm s^{-1}. Figure 4b shows the power-law fits for the 26 November averaged fall-speed data in which a minimum of 500 data points was also required for each averaged value. The scatter is much greater than on 25 November with standard deviations of the errors ranging from 9 to 40 cm s^{-1}.

The averaging results for 26 November had indications of upward motions in the upper layer of the cloud. A few averaging bins had upward motions around 5–8 cm s^{-1}. The radar-derived fall speeds were therefore increased by 8 cm s^{-1} at all levels prior to computing the best-fit equations under the simple assumption of constant upward motion through the depth of the cloud. To examine the validity of this correction, a nearby (but not collocated) NOAA network profiler that was observing the same cirrus cloud deck was examined. The vertical resolution of the profiler is much coarser (250 m) than the K band (37.5 m), but it can provide clear-

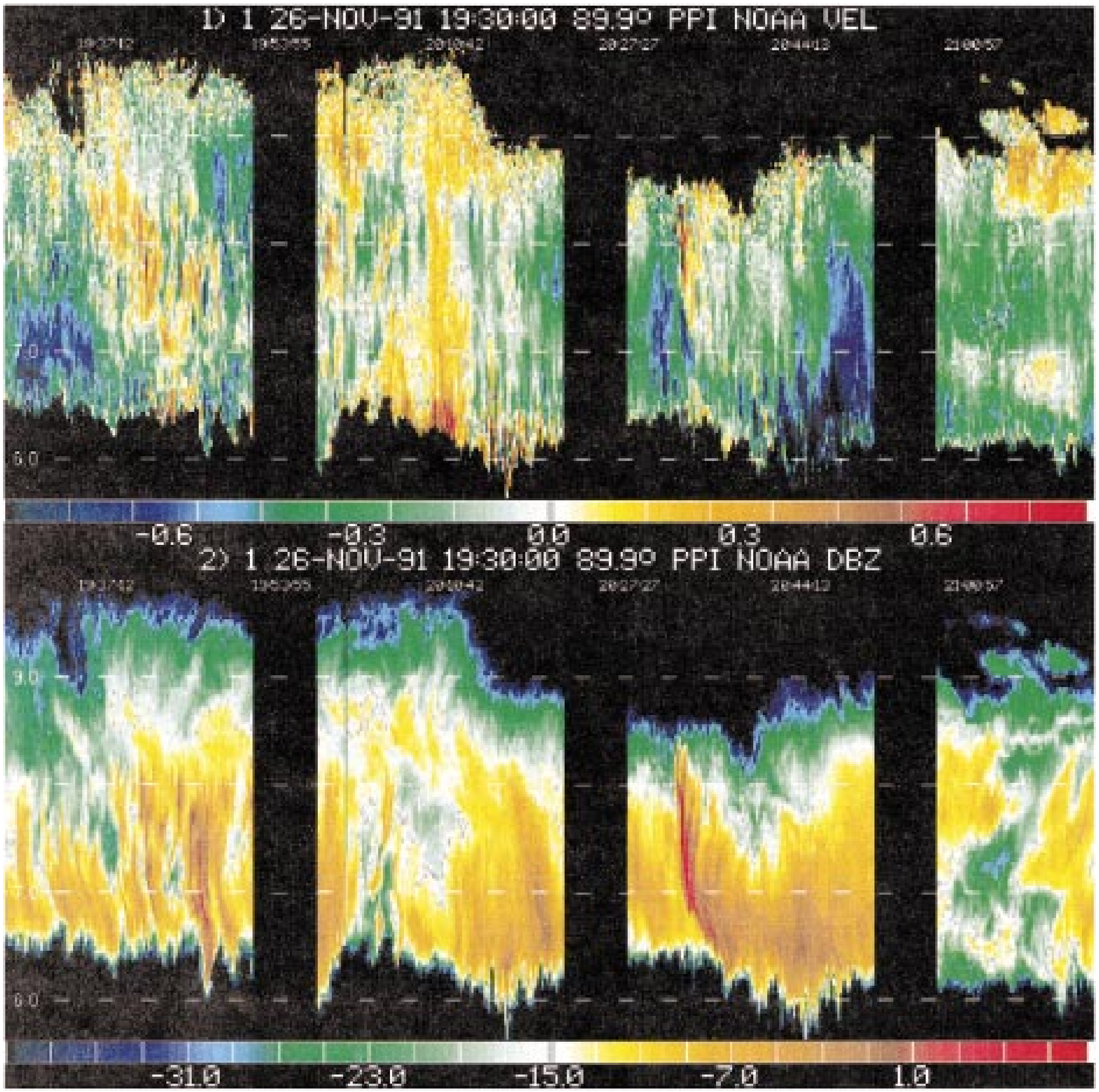

FIG. 3. Time–height plot of Doppler velocity [upper panel (m s^{-1})] and reflectivity [(lower panel (dBZ)] from 1930 to 2130 UTC 26 November 1991. Negative velocities are downward.

air velocity measurements. Other studies have shown that the NOAA network profilers are sensitive to cirrus cloud particles (e.g., Orr and Martner 1996), and as a result, the profiler cannot be used to deduce air motions within the cloud. Figure 5 is a 3-h average of profiler vertical velocity from 26 November. Due to operational parameters, the profiler produces only eight vertical beams per hour, and therefore, a 3-h average was chosen to provide a more significant average. The profiler indicates upward Doppler velocities around 10 cm s^{-1}, just above and below the height of the cirrus cloud layer. This supports the assumption of uniform upward motion for this case, and it is in close agreement with the 8 cm s^{-1} adjustment inferred from the weakest K-band reflectivities.

A number of interesting features are evident in the power-law best-fit curves for both test cases. First, the fall speeds from 25 November (Fig. 4a) are generally stratified progressively with height, exhibiting increasing fall speed with decreasing altitude. This is indicative of a relatively stable cloud system with particle nucleation occurring in the upper part of the cloud; the ice particles grow by vapor deposition as they fall through the cloud layer, producing higher fall velocities in the

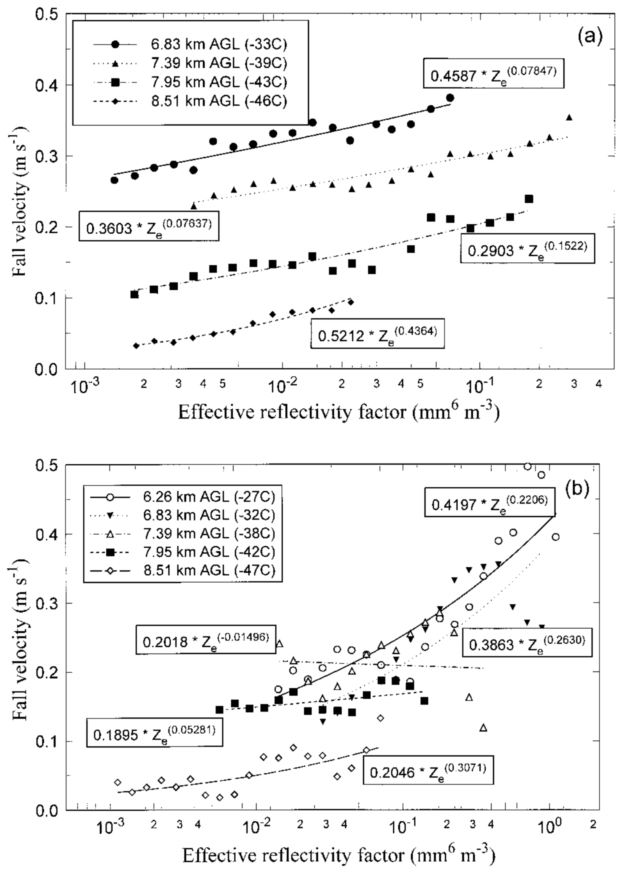

FIG. 4. Power-law fit of (a) 25 November and (b) 26 November averaged velocity data from Figs. 2 and 3. Best-fit equations are shown adjacent to the curves. Vertical scale is absolute value of terminal fall velocity.

lower portions of the cloud. Also, the slopes of the curves are similar, which suggests a similar shape to the particle size distribution as the particles grow.

In contrast to 25 November, the power-law results from 26 November (Fig. 4b) indicate two distinct regions of this cloud based on the slopes of the curves. The data from the lowest two layers of the cloud have similarly large slopes exhibiting a strong dependency of V_t on Z_e. The highest three layers of the cloud have much smaller variation of fall velocity with reflectivity and tend to resemble the curves from 25 November (Fig. 4a). This change in slope indicates microphysical changes occurring between these levels, possibly the onset of aggregation or a change to another crystal habit growth regime. This change in slope occurs between the $-32°$ and $-38°$C levels. Rawinsonde data indicate an inversion near 7.0 km above ground level (AGL), with significant drying below this level. The extremely poor data fit at 7.4 km AGL is likely a result of the observed gradients in moisture and temperature and their influence on the dynamics and microphysics near this level. Heymsfield and Miloshevich (1995) summarize their microphysical observations from aircraft and replicator

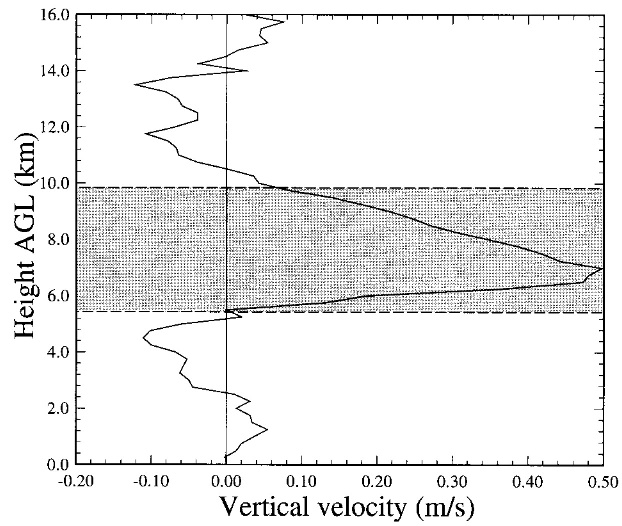

FIG. 5. Vertical profile of averaged profiler Doppler velocity from 1830 to 2130 UTC 26 November 1991, obtained with a 404-MHz wind profiler. Positive velocities are downward (opposite that of the ETL cloud radar). Shading indicates the K-band cloud echo boundaries.

measurements and find a shift in particle distribution associated with this inversion due to sublimation in the dry layer below 7.0 km AGL. Their results show the evaporation of smaller particles and enhanced aggregation in the sublimation zone, which is consistent with the change observed in radar-derived fall speeds.

The coefficients from the power-law fits can be compared with other studies of cirrus clouds. Heymsfield (1975) applied a similar relationship to a cirrus uncinus-generating cell using combined radar and aircraft data. The values found for α and β in (2) were 0.84 and 0.074, respectively. Results for stratiform ice clouds were presented in Heymsfield (1977), which had values of α ranging from 0.59 to 0.67 and values of β ranging from 0.06 to 0.095. The results from this study contain a wide range of values for α and β, with β showing the most variation. The range of α for both case studies presented in this article varies from 0.20 to 0.52, which is slightly lower than that found by Heymsfield. The values for β have a rather large spread, ranging generally from 0.06 to 0.44. Results from Heymsfield fall within the lower end of this range. These results are summarized in Table 2, excluding the extremely poor data fit on 26 November at 7.4 km AGL near the inversion layer.

It is encouraging that these results are in approximate

TABLE 2. Comparison of power-law coefficients derived from the present study and from in situ data.

Dataset	α	β
Heymsfield (1975)	0.84	0.074
Heymsfield (1977)	0.59–0.67	0.06–0.095
This study	0.20–0.52	0.05–0.44

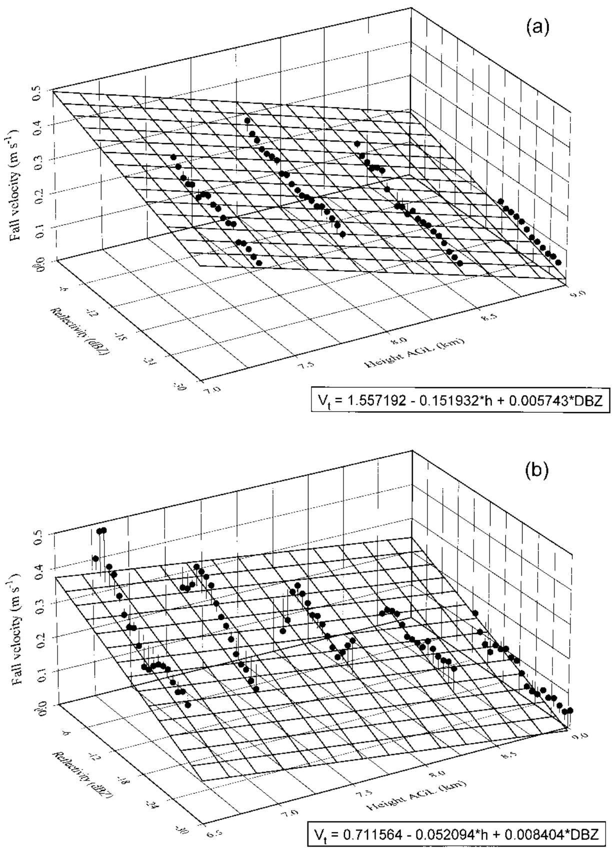

FIG. 6. Multiple linear regression analysis of averaged terminal fall velocity against height (km AGL) and reflectivity (dBZ) for (a) 25 November and (b) 26 November. Best-fit equation is shown below each graph. Vertical scale is the absolute value of terminal fall velocity.

agreement, considering that the present study and those of Heymsfield approach the problem from totally different perspectives. Most of Heymsfield's work started with explicit microphysical measurements, whereas the present study has the entire size spectrum reduced to one measurement, radar reflectivity. For a distribution of particles, the radar will respond much more strongly to the larger particles, which may provide a partial explanation for the rather large range in the values of β derived from the radar data.

Figure 6a is the result of a multiple linear regression analysis of V_t against height (kilometers AGL) and reflectivity expressed in dBZ units for 25 November, as in Fig. 2. The best-fit plane is shown drawn through the points with the equation for this plane shown at the bottom of the figure. There is surprisingly little scatter in Fig. 6a, and the variations in velocity associated with height and reflectivity are well represented; the standard deviation of the error is less than 2 cm s^{-1}. Although this error is less than the power-law fits, it must be remembered that this regression uses reflectivity on a logarithmic scale that will compress the data compared to the power-law fits that used Z_e.

Figure 6b is the multiple linear regression for 26 November. Again, there is more scatter than in the 25 November case, but the variations of V_t with height and reflectivity are nevertheless reasonably well represented. The best-fit equation is shown at the bottom of the plot. The standard deviation of the errors for the multiple linear regression is around 5 cm s^{-1}, which is slightly larger than that of 25 November.

A notable feature in the averaged data from 26 November is the large scatter about the best-fit curves. In particular, the data occasionally have a wavelike deviation about the best-fit line (Figs. 4b and 6b). This is likely a manifestation of the convective nature of this cloud. Significant deviations above the best-fit line are likely from regions of the cloud that contain a greater percentage of updrafts, the reverse holding for downdraft regions. This would imply that the underlying assumption of averaging to near-zero air motion may not be valid, at least within certain layers of the cloud. In this respect, 26 November may be a limiting case of this technique. However, considering the more complex dynamics on 26 November, the average fall-speed characteristics are still captured reasonably well, as seen in Fig. 6b.

The significant variation with height of the power-law equations precludes applying them to correct for particle fall velocity in a convenient manner. Using multiple linear regression analysis to reduce the radar data to one equation is a means of smoothing the results of the power-law equations discussed above. It provides an easily implemented equation to estimate particle fall velocity that can also be used to estimate air motions within the cloud using reflectivity and height data. The most obvious feature when examining the results of the multiple linear regression (Fig. 6) is that this method simultaneously represents the variations of fall velocity associated with reflectivity and height that were evident in the power-law results. For 25 November there is a nearly perfect match to the data with errors less than 2.0 cm s^{-1} and a coefficient of determination (R^2) of 0.98. The 26 November data, which is much more dynamic than the 25 November data, is also generally represented using this regression method with errors around 5 cm s^{-1} and an R^2 of 0.75.

The regression coefficients for 25 and 26 November are considerably different. The constant and reflectivity coefficient differ by a factor of 2, and the height coefficients differ by a factor of 3 (see Fig. 6). However, no data exist to which these coefficients can be compared, and it is difficult to directly associate a physical process to the coefficients. Currently, it is not possible to say whether this much variation should be expected from one case to another. Changes in temperature, temperature lapse rate, moisture availability, and pressure, among other things, will effect particle habit, density, and growth rates that will in turn impact one or more of the regression coefficients. Therefore, although there are numerous microphysical and dynamic processes that

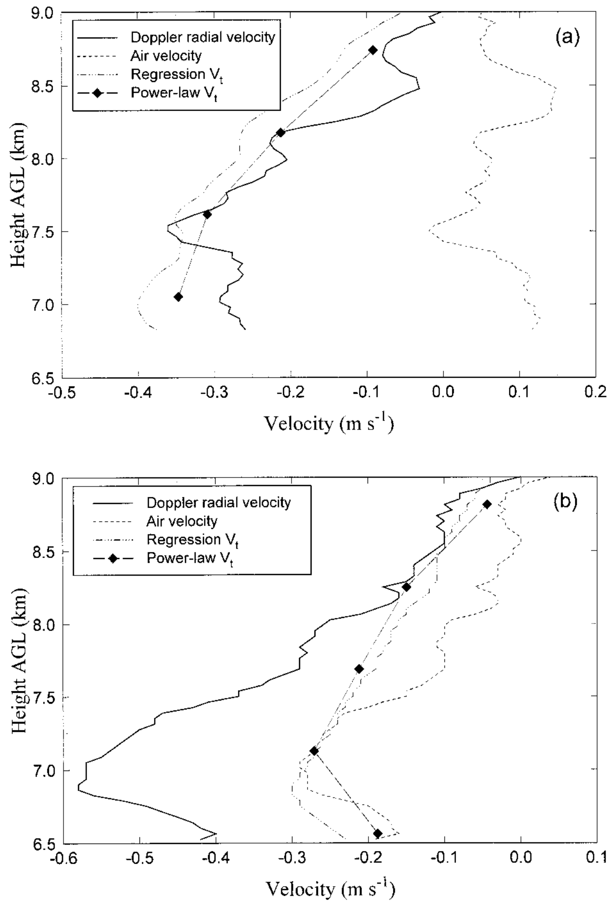

FIG. 7. Profiles of average vertical Doppler radial velocity, air motion, and terminal fall velocity (V_t) derived from both the regression and power-law analysis for (a) 1300–1310 UTC 25 November and (b) 1930–1940 UTC 26 November. Air motion is the difference between the regression-derived terminal velocity and the measured Doppler velocity. Downward velocities are negative.

can impact the multiple linear regression results, the primary use here is simply as an analysis tool without regard to its underlying physical processes.

An example of the application of the multiple linear regression results is shown in Fig. 7. Figure 7a is a 10-min average vertical profile of radial Doppler velocity, particle terminal fall velocity, and derived air motion (downward velocities are negative) from 1300 to 1310 UTC 25 November. Air motion estimates were obtained by differencing terminal fall velocity and measured Doppler velocity. Figure 7b is a similar profile from 1930 to 1940 UTC 26 November. The 25 November profile is from a region of weak rising motion, and the 26 November profile is from a region of small-scale subsidence.

Estimates of particle terminal velocity using the power-law relations are also presented in Fig. 7. Although the power-law equations are mainly intended as a means to compare the present study results with other studies, they are presented in Fig. 7 for comparison. Since each power-law equation is derived for a particular cloud layer, they cannot be used to produce profiles of particle fall speed throughout the depth of the cloud. The power-law estimates shown in Fig. 7 are produced from layer-averaged values of reflectivity. The reflectivity was averaged over the cloud layer applicable to each equation (see Fig. 4). The power-law fall-speed estimates shown in Fig. 7 are slightly less than the regression results at all levels for 25 November (Fig. 7a) and nearly the same for 26 November (Fig. 7b). Small differences between the power-law and regression methods would be expected for several reasons. First, the underlying average vertical structure in cloud reflectivity from which these equations were derived will determine how much the equations vary from one level to the next. Vertical variations in reflectivity will impact both the height and reflectivity coefficients in the regression analysis. Also, the multiple regression analysis includes data from all layers, which tends to smooth over the differences found in the layer-averaged power-law equations. As a result, the regression equations will respond differently to changes in cloud vertical reflectivity structure than will the power-law equations. It is therefore difficult to place any particular significance on the positive bias in the power-law fall-speed estimates on 25 November (Fig. 7a), which are likely specific to the chosen averaging period. The 26 November power-law fall-speed estimates exhibit small positive and negative variations around the regression results with no bias evident.

Cirrus clouds can be produced in regions of large-scale ascent with upward motions on the order of 10 cm s^{-1} up to maximums of 1–2 m s^{-1} in smaller-scale cirrus uncinus-generating cells (e.g., Heymsfield 1975). Downdrafts induced by particle drag and sublimation are typically on the order of tens of centimeters per second. The terminal velocity and derived air motion profiles in Fig. 7 are in agreement with these observations. The increase in fall velocities from cloud top to cloud base in Fig. 7 is also a common characteristic observed in cirrus clouds.

There are a number of potential sources of error that can degrade the accuracy of this technique, most of which have been discussed. The absolute accuracy and precision of the Doppler velocity measurements is obviously a controlling factor. Examination of gate-to-gate and beam-to-beam fluctuations of the radial velocity estimates from the K-band radar under strong signal-to-noise conditions indicates precision of less than 5 cm s^{-1}. This precision is improved to a fraction of a centimeter per second after the data are averaged, as described for this technique (a minimum of 500 sample points). As long as ground clutter is not present, which is the case here, there are no known sources of bias that would affect the accuracy of the observations. Other sources of error that have been discussed and can be quantified include the antenna-pointing angle (errors less than 2–3 cm s^{-1}) and the accuracy of the curve fitting. A typical application of this technique is the 25

November data that had curve-fitting errors less than 5 cm s^{-1} for power-law fits and 2 cm s^{-1} for the multiple linear regression. Bias due to average upward air motion can be identified and removed when weak signals (<-25 dBZ) are present. Application of this correction on 26 November indicates an error within a few centimeters per second. Consideration of all these errors, which will vary with each specific case, demonstrates that this technique for inferring particle fall velocities should have an absolute accuracy better than 10 cm s^{-1}. This implies that the largest relative errors occur at the cloud top where fall speeds are smallest. Relative errors decrease from the cloud top to bottom as fall speeds increase.

The most subjective part of this technique is choosing an appropriate averaging period, t_a, which depends to a large extent on the nature of the particular cloud system. The strength of the dynamics and turbulence within the cloud, the rate of advection over the radar, and the radar sampling rate are all important factors in determining t_a. Velocity averages were computed for values of t_a ranging from 90 min to 2.5 h in order to determine the sensitivity of the results to the choice of t_a. In addition, multiple averages for a given t_a, offset in time by approximately 30 min, were computed to examine the temporal stability of a particular cloud system. Systematic changes in the results would suggest that the cloud was evolving. Variations in the coefficients for both the power-law fit and regression analysis were less than 10% on 25 November. Variations in the 26 November power-law coefficients were generally less than 20%, although the highest level of the cloud had changes in the coefficients as large as 40%. Variations in the 26 November regression coefficients were less than 20%. Therefore, although the averaging period is important, it does not appear to be critical as long as a sufficiently stable portion of the cloud is selected. This also emphasizes the underlying assumption that the derived equations are generally valid only for the averaging period from which they were derived.

5. Conclusions

A method has been proposed that allows for an estimate of particle terminal fall velocity using data from a vertically pointing Doppler radar. The method is primarily intended for use in stratiform ice cloud systems that are generally uniform and long lived to facilitate averaging over time periods on the order of a few hours. This method was been implemented using two case studies of cirrus clouds from the FIRE II project. Results obtained are generally supported by other observational studies of cirrus clouds with aircraft. The method produces height- and reflectivity-stratified Doppler velocity averages that can then be analyzed using two methods. Application of a best-fit power-law analysis at each height level allows some insight into the microphysical processes and the degree of cloud stratification. Multiple linear regression analysis is used to arrive at a single equation that can be easily implemented to produce an approximate estimate of particle terminal velocity in terms of height and reflectivity. The quality of these fits, in either case, gives a good indication of how well the underlying assumptions are satisfied.

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