Rain Resulting from Melting Ice Particles[1]

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ABSTRACT

By means of a zenith-pointing radar, observations were made of the reflectivities and Doppler spectra in orthogonal planes as a dissipating shower exhibiting a bright band passed overhead. The observations have been used to test various procedures for estimating hydrometeor parameters from measurements of radar reflectivitity. They involve assumptions that the raindrop diameters were exponentially distributed, preferably in the manner prescribed by the Marshall-Palmer distribution. It is concluded that, in this case, such an assumption was not valid in regions where it was expected to be valid. As a consequence, estimates of median raindrop diameters and updraft velocities calculated from radar reflectivities were in error. The analyses indicate that raindrop size sorting under the influence of vertical wind shear can account for the observed non-exponential size distributions.

1. Introduction

On 20 August 1975 a thunderstorm in an advanced stage of dissipation moved over a zenith-pointing 3.2 cm pulsed-Doppler radar located on the grounds of Tucson International Airport. The radar, described earlier by Battan (1975), has two identical antennas and receiving systems. The antennas are arranged to measure the backscattered signals in orthogonal planes. The unique aspects of this observational system is that it can obtain Doppler spectra in either the plane of transmission or the cross-polarized plane. The radar was operated in the following manner: a sampling gate traveled upward in steps of about 150 m, dwelling at each altitude for 0.66 s during which records were made of the Doppler information in the plane of transmission, while at the same time a recording was made of the integrated backscattered signal in the cross-polarized plane. After repeating this procedure for a second set of observations (called a frame) between the ground and a preset maximum altitude at 9750 m, the operations mode was changed, and recordings were made of the Doppler data in the cross-polarized plane and of the integrated power in the plane of transmission. The operations mode was alternated every two frames, i.e., every 86 s since 43 s were required for each frame.

As will be seen from an examination of Fig. 1, which is a facsimile recording of two frames, one showing Doppler spectra in the plane of transmission followed by a second showing the data in the cross-polarized plane, there was a radar bright band. It indicated that frozen particles were falling through the freezing level and melting.

A great deal has been written about the precipitation mechanisms occurring above and below the bright band. Disagreements about the questions of particle aggregation and breakup have yet to be resolved. Lhermitte and Atlas (1963), Ekpenyong and Srivastava (1970), Zwack and Anderson (1970), Takeda (1976) and Passarelli (1976) have offered arguments in favor of the view that there is snow-particle aggregation above the bright band or raindrop breakup below it, or both. On the other hand, duToit (1967), who used Doppler radar data, and Ohtake (1969), who made observations of the hydrometeors up a mountain slope, concluded that aggregation and breakup were not important processes in explaining precipitation falling through a melting layer.

Over the years there have been a variety of theoretical studies of the evolution of raindrop spectra as a result of coalescence and breakup. A recent article by Gillespie and List (1976) lists among its conclusions the statement, "Drop size spectra initially of the Marshall-Palmer type essentially preserve that property in falling to the surface."

The purpose of the study reported here is to describe a set of unique observations, to calculate precipitation particle size spectra and to examine and explain changes in the spectra.

2. Observations and related data

The observations presented here cover a period of 9 min in the life of a decaying thunderstorm. Unlike

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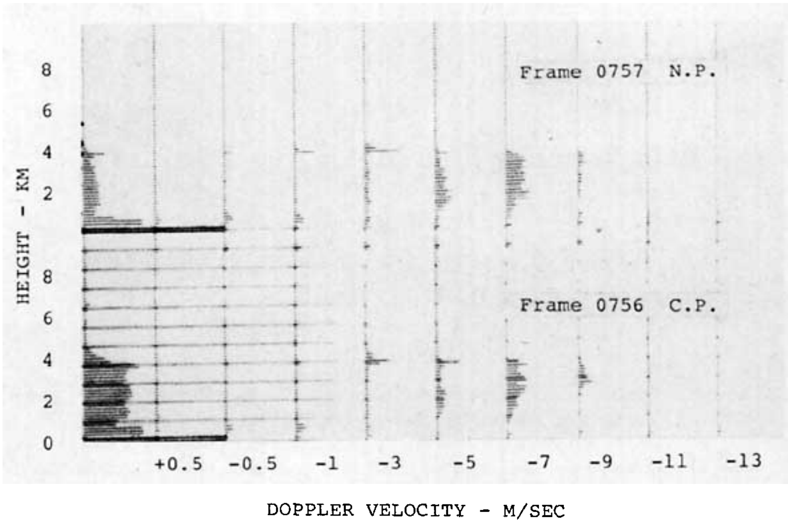

FIG. 1. Doppler spectra (signal voltages) in the plane of transmission (NP) and the cross-polarized plane (CP). The first column shows the integrated backscattered signal amplitudes.

meteorological circumstances leading to widespread rain from nimbostratus clouds where temporal and spatial rate changes often are quite small, in the case of this dissipating shower there is evidence of substantial changes over periods of several minutes.

Fig. 2 shows that the bright band, exhibiting maxima in reflectivity of just over 30 dBZ, was about 300 m below the 0°C isotherm. Note that the reflectivity in the bright band layer ranged from less than 15 dBZ to more than 30 dBZ. The pattern of echo intensity indicates that the precipitation was of a showery nature. The rainfall intensity at the ground was very light, not measurable with a weighing gage. On the basis of $Z=200\ R^{1.6}$, one obtains maximum values of 0.6 and 2.8 mm h^{-1} associated with reflectivities equal to 20 and 30 dBZ, respectively.

Fig. 3 shows the vertical distribution of temperature, dew point and wind velocity at about 1700 MST measured by a rawinsonde released about 2 km away from the radar. Of particular interest are the changes

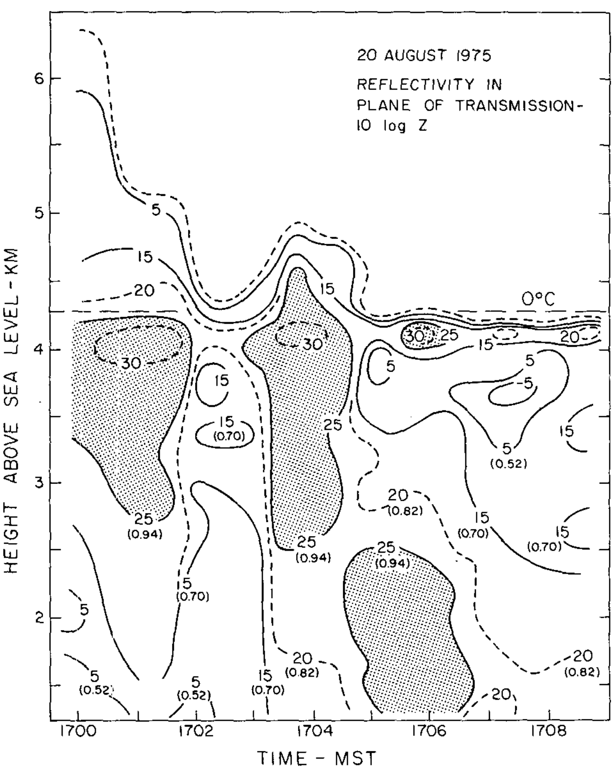

FIG. 2. Reflectivity in the plane of transmission. Numers in parentheses are median volume diameters (mm) calculated by means of Eq. (3).

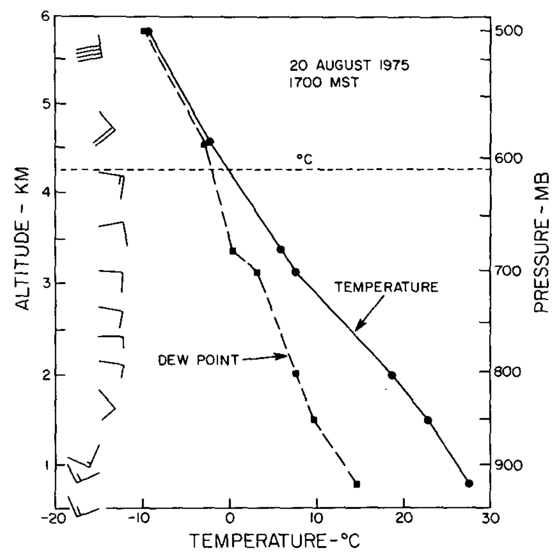

FIG. 3. Rawinsonde data at 1700 MST 20 August 1975. Each wind barb represents 5 m s^{-1}.

of the wind velocity. The wind backed with height from west-southwest at the ground to almost due east through the layer from 2 to 3.6 km. Above it, the wind veered into the south and the speed increased markedly. Through most of the region of precipitation the winds were from the east. It seems reasonable to expect that the *cloud system* was moving from the east, and therefore that changes observed in the cross sections shown in Figs. 2–8 are mostly representative of precipitation particles moving in an east–west plane. On the other hand, above about 3.6 km at the time and place of the rawinsonde, the air had a component of motion into the plane of the paper. Below about 2 km the air was moving out of the plane of the paper. Between about 2 and 4 km the particles were moving mostly in the plane of the time-height cross sections. These points should be borne in mind in interpreting the observations.

For the most part, before 1705, backscattering in the cross-polarized plane (Fig. 4) was relatively high, particularly in the bright band layer. During the last half of the period, at altitudes above 2 km, the power in the cross-polarized plane was extremely low.

As would be expected, the pattern of depolarization (Fig. 5) was similar in most respects to the pattern of power in the cross-polarized plane, except that in and below the bright band after 1704 there is evidence of considerable depolarization. It should be recognized, however, that the quantities of power in the numerators of the ratios from which depolarizations were calculated were quite small, just barely above the noise level of the radar. The indication of depolarization (Fig. 5) where no signals are indicated in the cross-polarized plane is a result of differences in rounding out values of reflectivity.

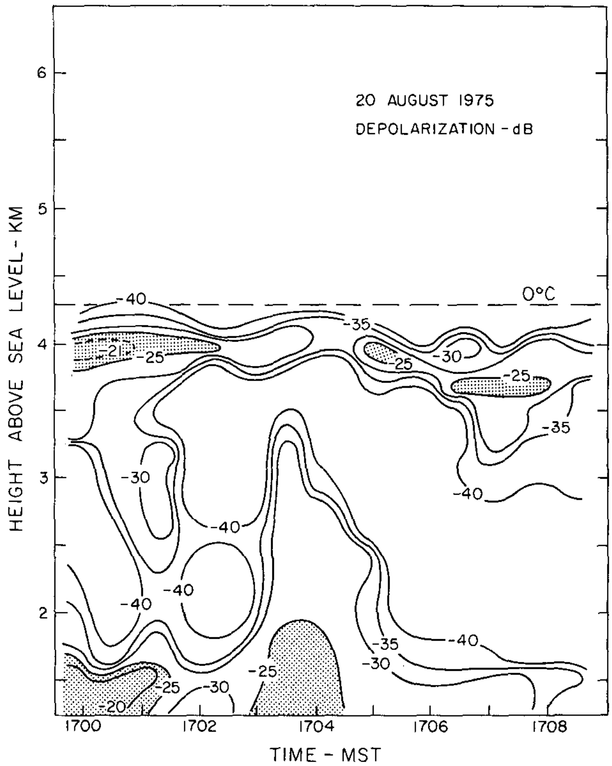

FIG. 5. Depolarization given as 10 times the logarithm of the power in the cross-polarized plane divided by the logarithm of the power in the plane of transmission.

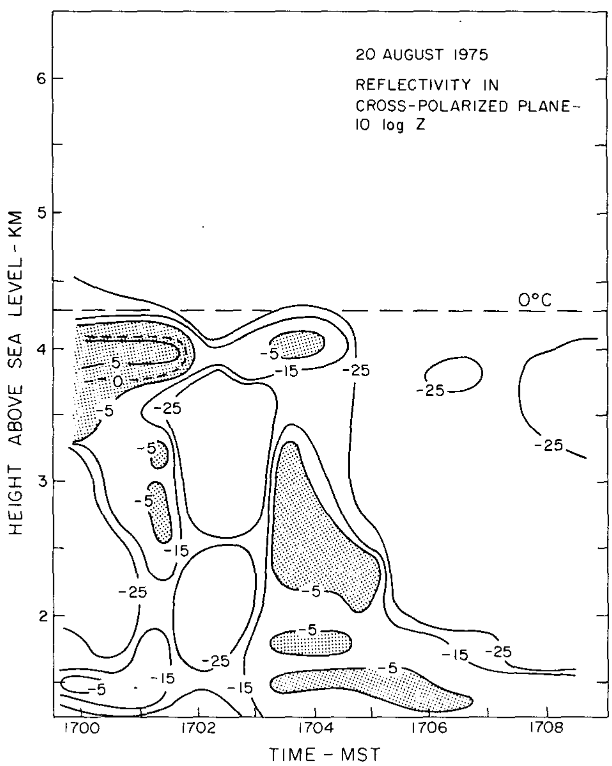

FIG. 4. Reflectivity in the cross-polarized plane.

During the first 2 min, there was a gradual increase of echo intensity downward from $\sim$5 km to 4.2 km, the center of the bright band. One can visualize, as did Lhermitte and Atlas (1963), the aggregation of snow particles above the freezing level, with a more sudden increase of reflectivity as the snow particles fell through the melting region. On the other hand, the higher reflectivities might also be explained by a plume of larger particles moving through the plane at observation at lower altitudes.

From 1703 to 1705 MST, the high-reflectivity region extended above the freezing level with a steep gradient of Z at the echo top. It can be speculated that this represents a volume of wet snow pellets or ice pellets. Note that even though the reflectivity in the plane of transmission was 30 dBZ, about the same as at 1701, the signals in the cross-polarized plane at 1704 in the bright band were $\sim$10 dBZ lower than at 1701. This would suggest that at 1704 MST, the hydrometeors were more spherical than those accounting for the earlier bright band. After 1704 the bright band was

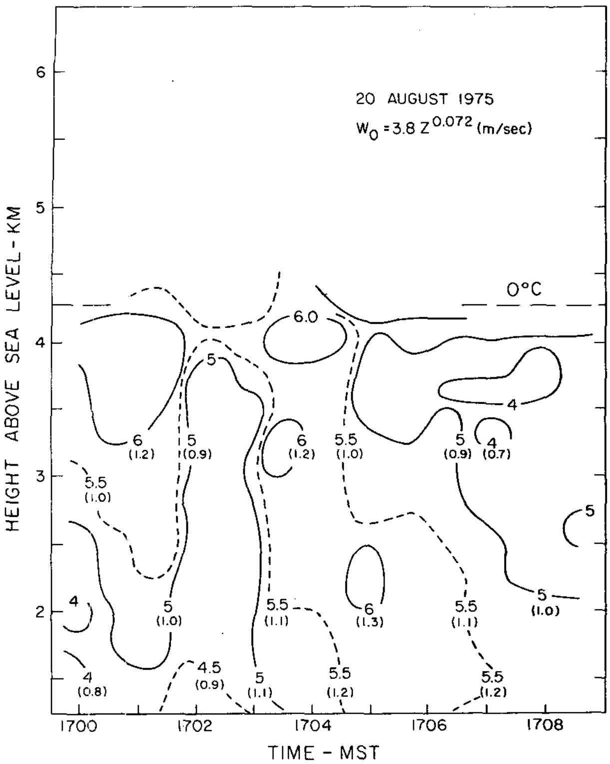

FIG. 6. Mean reflectivity-weighted terminal velocity of raindrops corresponding to the observed values of Z in Fig. 2 (Rogers, 1964). The numbers in parentheses are raindrop diameters (mm) corresponding to the indicated terminal velocities.

narrow and weak but quite distinct. It is inferred that the hydrometeors were relatively small snow pellets which melted fairly rapidly.

In Fig. 2, at heights below 3.5 km, the lines of equal echo intensity are labeled with numbers in parentheses. They are calculated median-volume diameters corresponding to the indicated values of reflectivity. The procedure for obtaining the diameters assumes that the raindrops conform to the Marshall-Palmer distribution

$$N=N_0e^{-\Lambda D}, \quad \text{where} \quad \Lambda=4.1\,R^{-0.21}, \tag{1}$$

where Λ and R are in units of mm^{-1} and $\mathrm{mm\ h}^{-1}$, respectively. According to Atlas (1964)

$$\Lambda D=3.67\frac{D}{D_0}, \tag{2}$$

where D_0 is the median-volume diameter. By suitable substitution employing these relations, and assuming $Z=200\,R^{1.6}$, one obtains

$$D_0=0.45\,Z^{0.13}, \tag{3}$$

where D_0 and Z are in units of mm and $\mathrm{mm}^6\ \mathrm{m}^{-3}$, respectively.

The calculated values of D_0 appearing in Fig. 2 suggest that during the periods 1700–1702 and 1703–1704 MST just below the bright band there were raindrops having median volume diameters of about 0.9 mm. On the other hand after 1704, median volume diameters were about 0.5 mm, indicating relatively small raindrops.

If it were assumed that in the layer from 2 to 4 km, the precipitation was falling mostly in the plane of observation, it would appear that from about 1701 until 1705, there was a net breakup of raindrops because the median volume diameters decreased with decreasing altitude. During the period 1705–1707, the calculated raindrop sizes near the 2 km level were about the same as they were aloft, indicating little net breakup or coalescence. During the last few minutes of observations there was an apparent increase of raindrop sizes with decreasing altitude, the median volume diameters increasing from 0.5 mm to about 0.8 mm.

Another estimate of a raindrop size parameter can be obtained from a knowledge of the reflectivity-weighted mean fall velocity (W_0) of the raindrops in still air. An estimate of this quantity is given by an equation developed by Rogers (1964) which assumes Marshall-Palmer distributed raindrops and relates W_0 to Z. Fig. 6 shows the pattern of W_0, and in parentheses, the raindrop diameter corresponding to a terminal velocity equal to W_0 (Beard, 1976). As might be expected, the drop sizes shown in Fig. 6 are con-

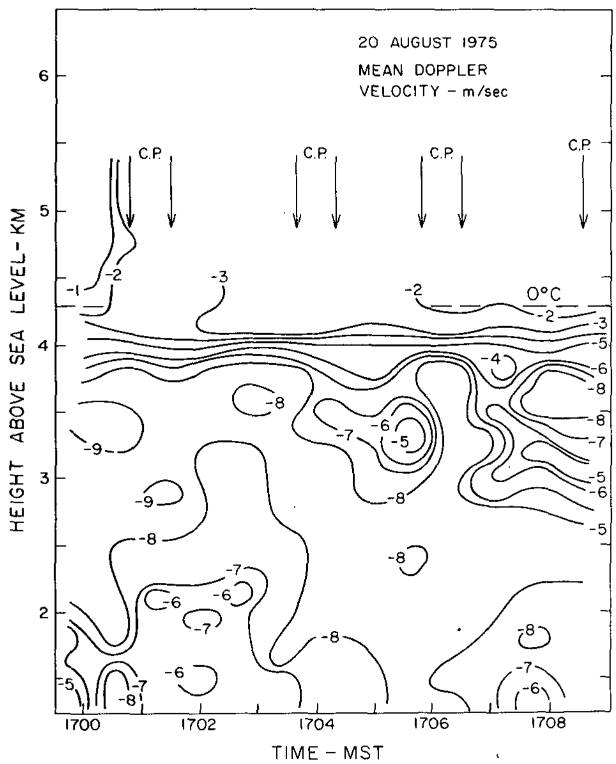

FIG. 7. Mean Doppler velocities. Arrows indicate cross-polarized data (CP). Negative values indicate downward velocities.

sistently larger than those shown in Fig. 2. Although both diameter estimates are derived from a knowledge of Z, the ones in Fig. 6 depend on the reflectivity-weighted (i.e., D^6) terminal velocities while those in Fig. 2 depend on the mass-weighted (i.e., D^3) terminal velocities. In other words, the former are more heavily biased by the larger raindrops.

The depolarization data, shown in Fig. 5, tend to confirm the just-noted inferences about changes of raindrop sizes with altitude if one is prepared to accept the notion that the greater the raindrop sizes, the greater the depolarization when viewed along the zenith. This seems like a reasonable proposition because the greater the sizes, the more extensive the drop oscillations and deformations.

A crucial point to keep in mind is that the drop parameter estimates in Figs. 2 and 6 assume that the raindrops conform to the exponential Marshall-Palmer distributions. In a later section it will be shown that this probably is not a realistic assumption at some altitudes and times. Therefore, these estimates of drop sizes are of questionable worth.

Additional information about the properties of the air and hydrometeor motions can be obtained by an examination of the Doppler spectrum parameter shown in Figs. 7 and 8. Note that the data at the times indicated by the downward pointing arrows were taken

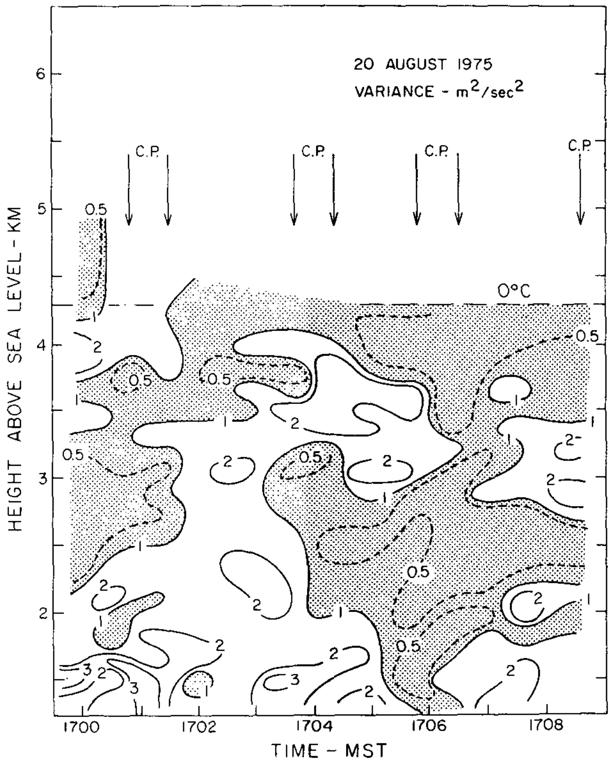

Fig. 9. Variances of the Doppler spectra.

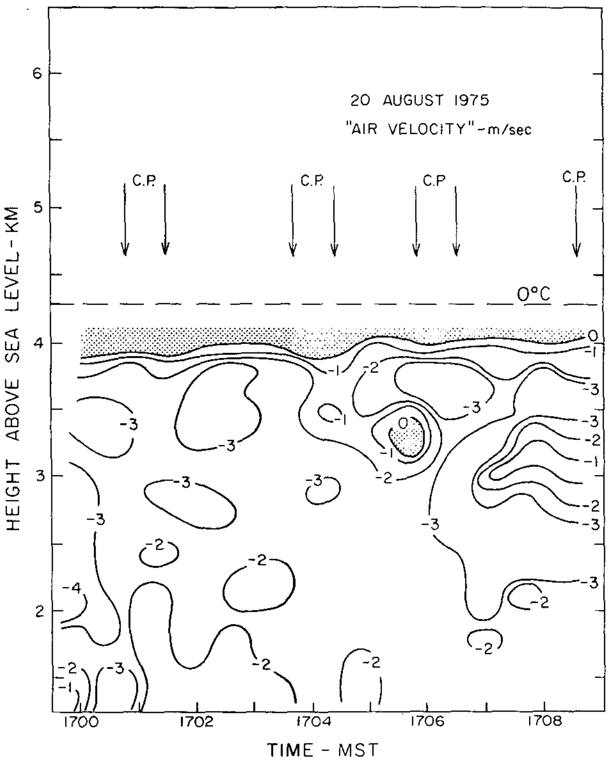

Fig. 8. Updraft velocities calculated by means of Rogers' technique which applies when rain is observed. Negative values indicate downward velocities.

from observations of Doppler spectra of the backscattered signals in the cross-polarized plane and that there is a strong correlation between the velocities measured in the two orthogonal planes.

Fig. 7 shows pronounced downward accelerations, *with respect to the ground*, of particles falling through the melting level. For the most part their downward velocities were 1–3 m s^{-1} at the level of the 0°C isotherm and reached 6 m s^{-1} about 300 m below it. If the vertical component of air velocity were zero, a Doppler velocity of 6 m s^{-1} would indicate the presence of raindrops having diameters of $\sim$1 mm (Beard, 1976). After another 150 m of fall, throughout most of the 9 min of observations, the particles, presumably raindrops, were falling mostly at $\sim$8 m s^{-1}. It is seen that over the first 3 min of observations within the rain region, the mean Doppler velocity decreased from a peak of -9.5 m s^{-1} at an altitude of 3.5 km to a minimum of -5.5 m s^{-1} near the ground. An examination of Fig. 8, which displays "updraft velocities" calculated by means of Rogers' (1964) method (the sum of the data in Figs. 6 and 7), suggests that about 1–2 m s^{-1} of this difference might be explained as a change in downdraft speed. In a later section it is concluded that just below the melting level the estimates of W_0 by means of Rogers' equation are in error because the raindrop sizes were not distributed exponentially. It is not possible to judge

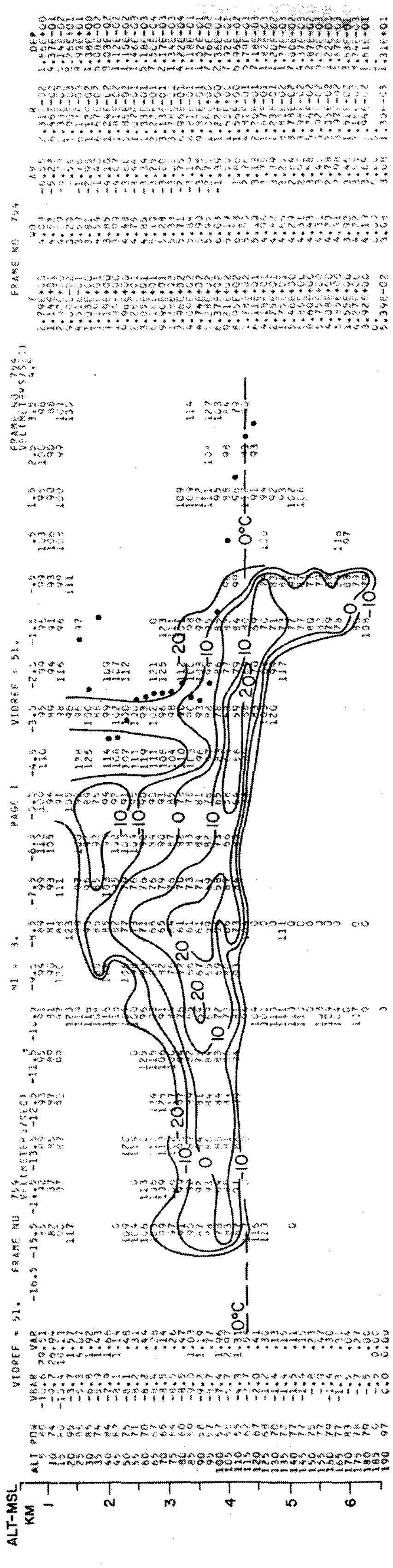

FIG. 10. One "frame," at about 1700 MST, showing a printout of available data. Note that altitude increases downward in this illustration. The following quantities are given, from left to right: ALT—altitude in kilometers and hundreds of feet; POW—backscattered power in negative dBm normalized to an altitude of 1.60 km; VBAR—mean Doppler velocity; VAR—variance; VEL—Doppler spectra with power in negative dBm normalized to 1.60 km; Z—radar reflectivity; $W_0 = 3.8\,Z^{0.072}$; AV—air velocity calculated by Rogers' (1964) equation; R—rainfall intensity from $Z = 486\,R^{1.37}$; DEP—depolarization ratio. The lines drawn across the Doppler spectra depict the field of radar reflectivity in 10 log dBZ.

how *well* the "air velocity" estimates are over most of the rainfall region.

Fig. 7 reveals two small regions between 3.0 and 3.6 km at about 1706 and 1708 MST, respectively, where the mean Doppler velocities were about -5 m s^{-1}, some 3 m s^{-1} greater than in the surrounding air. As shown in Fig. 8, Rogers' method suggests that in these regions the air motions were nearly zero or upward, surrounded by air descending at $\sim$3 m s^{-1}. A widespread weak downdraft would be a more expected state of affairs in a dissipating thunderstorm.

Fig. 9 shows the variances of the Doppler spectra. When a radar has a narrow beam pointed toward the zenith, this quantity depends mostly on the range of particle fallspeeds and small-scale turbulence. As shown by Lhermitte (1962) and others, in still air the variance attributable to rain amounts to about 1.0 m^2 s^{-2}. Greater values indicate either the effects of small-scale turbulence or the presence of frozen hydrometeors in the form of pellets falling at velocities exceeding those of raindrops. The variances >2 m^2 s^{-2} at about 1704 between 3 and 4 km, might be attributable in part to the presence of ice or snow pellets as was suggested earlier. The same argument cannot be made about the other regions where the variances were relatively large such as at 1705 and 1708 near 3 km and from 1700 to 1705 at lower altitudes. No ice particles were observed at the ground in the vicinity of the radar. If the gradients of measured velocity indicated in Figs. 7 and 8 were caused in part by differences in air velocity, they could have generated small-scale turbulence which increased the variance.

3. Hydrometeor sizes and updrafts

If one knows the nature of the hydrometeors, the Doppler spectrum, the vertical air velocities and the terminal velocities of the hydrometeors as a function of their diameters, it is possible to calculate the size distribution of the hydrometeors (e.g., duToit, 1967). Observations made by means of the Arizona radar were used to make such calculations. The form of the Doppler data and derived quantities is illustrated in Fig. 10. It shows most of the quantities obtained in a single "frame" which covered the time period 1659:54 to 1700:37 MST and the altitude range from the ground 0.76 to 8.99 km. This figure presents only the measurements between the altitudes of 0.91 and 6.55 km.

Above the 0°C isotherm, the hydrometeors had Doppler velocities mostly between about -0.5 and -1.5 m s^{-1}. The negative sign designates downward velocities. There is evidence of greater particle sizes near the melting level where the spectrum widths were greater than at higher elevations. As the particles

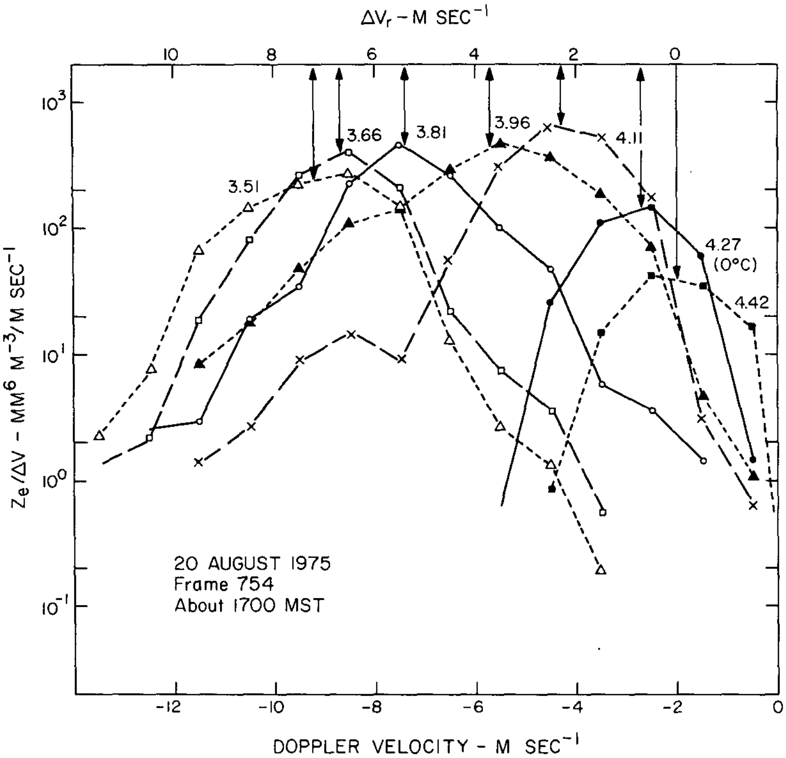

FIG. 11. Smoothed Doppler spectra at various altitudes between 3.51 and 4.42 km above sea level.

melted, becoming first wet ice particles and then raindrops, the spectrum widths increased markedly. The mean Doppler velocities and variances increased rapidly between 4.5 and 3.5 km. At lower altitudes the variances diminished before increasing again at lower elevations. The distribution of backscattered power as a function of Doppler velocity shows evidence of a bimodal drop size distribution.

In order to make meaningful calculations of hydrometeor size spectra, it is necessary to know the vertical velocity of the air. There are various techniques for estimating it. One scheme used by Battan and Theiss (1970) takes the positive tail of the Doppler spectrum and adds an appropriate amount to take account of the terminal velocity of the smallest detectable hydrometeors. On this basis, it is concluded that above ~3.8 km, the vertical velocities were essentially zero, i.e., -0.5 m s^{-1} plus the estimated terminal speed of the slowest falling detectable frozen hydrometeors. Incidentally, it is inferred, because of the nature of the record, that the signals appearing in the positive channels are noise.

Rogers' (1964) technique, which was used to obtain the data in Fig. 8, yielded the air velocities indicated by the black dots. They correspond to the values under the column AV in Fig. 10 and are equal to VBAR$+W_0$. As noted earlier, W_0 is calculated on the assumption that the hydrometeors are raindrops which conform to a Marshall-Palmer distribution. Even if not distributed precisely this way, as long as the raindrop spectrum is logarithmic Rogers' technique would be expected to be fairly reliable. Obviously, the values of AV above and in the melting level should not be accurate. On the other hand, it was anticipated that below ~3.8 km, some 500 m below the 0°C isotherm, the hydrometeors would have been mostly raindrops and that Rogers' technique would give good estimate of the vertical velocities of the air.

Fig. 8 shows that above 3.5 km, the pattern of calculated downdrafts changed little over the first half of the record. Estimates of downdraft speeds close to zero in the snow region are trustworthy. If the indicated air speeds in the rain are correct there was strong—apparently unrealistically strong—accelerations of the downdraft through the melting layer.

In an attempt to judge the value of the downdraft estimates below the 0°C isotherm and to clarify the nature of the precipitation mechanisms, the Doppler spectra were used to estimate particle size distributions.

Fig. 11 presents smoothed Doppler spectra for the data shown in Fig. 10. The vertical arrows on each curve indicate the mean Doppler velocity; the scale along the upper part of the diagram shows downward velocity with respect to the mean Doppler velocity at 4.42 km, just above the 0°C level. Fig. 11 clearly

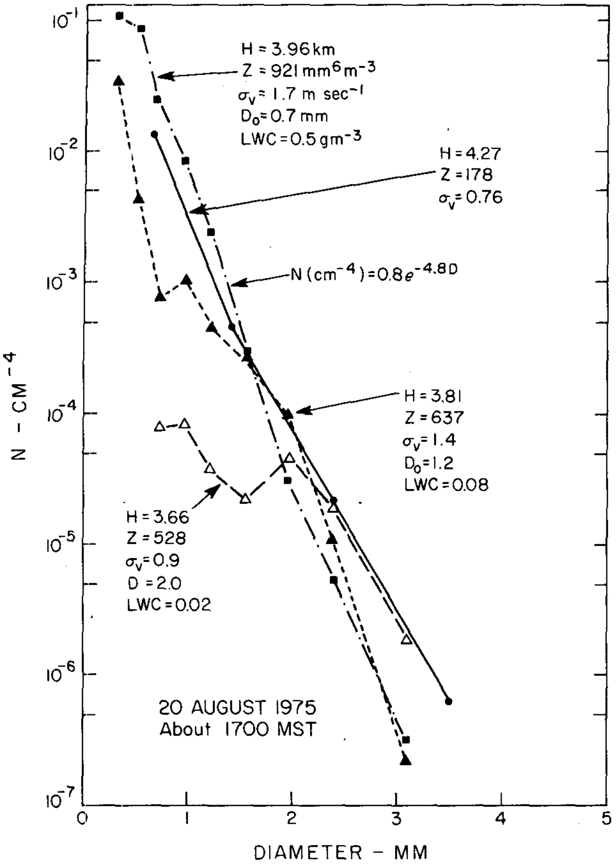

FIG. 12. Calculated size distributions at various altitudes at about 1700 MST. Note that a straight line through the data at $H=3.96$ km is described by the expression $N=0.8\ e^{-4.8D}$.

shows the shifting to greater downward velocities of the falling precipitation. At about the 0°C level which is indicated at 4.27 km but might have been slightly higher or lower in the cloud, the Doppler spectra were narrow. At altitudes of 3.51 and 3.66 km, the spectra again were narrow but were displaced about 7 m s^{-1} toward greater downward speeds. Between the 0°C level and ~3.6 km there was a fairly uniform shift of downward velocity amounting to about 1 m s^{-1} (100 m)$^{-1}$.

The tail of the spectra above about 3.8 km indicates vertical air velocities close to zero. For making calculations of particle size spectra, we assumed that vertical air speeds were zero through the layer represented by the spectra shown in Fig. 11 and that the shifts of the spectra were represented by differences in the fallspeed of the particles.

Estimates of particle sizes of frozen hydrometeors are tricky indeed. Nevertheless, they were made assuming a particle type and employing an appropriate fallspeed equation. We used the expression given by Locatelli and Hobbs (1974) for lump graupel of density 0.15 mg m^{-3}. It is $V=1.3\ D^{0.66}$, where V is in meters per second and D is in millimeters. The resulting particle spectrum corresponding to the Doppler spectrum at 4.27 km is given in Fig. 12. The diameters are those of spheres of water of masses equivalent to ice spheres.

The spectra at altitudes 3.96, 3.81 and 3.66 km were calculated assuming that the vertical air velocity was zero and that the backscatterers were raindrops having terminal velocities varying with diameter according to Beard (1976). There is considerable doubt that the hydrometeors were totally melted at 3.96 km. If some of them were partly melted ice particles, the calculations would yield excessive numbers of small drops. Just below the 0°C isotherm the calculated spectrum is essentially exponential and surprisingly close to that of the frozen hydrometeors.

At altitudes of 3.81 and 3.66 km the spectra deviated increasingly from exponential. The concentration of drops at the small end of the spectrum was less than at higher altitudes, while the number of large drops changed little with altitude. At 3.66 km there clearly was a bimodal drop size distribution.

Fig. 13 shows calculations of the size spectra of the hydrometeors about 3 min after the ones shown in Fig. 12. It also displays essentially exponential distributions just above the melting level and in the rain (presumably) just below it at 3.96 km, but a distinctly nonexponential distribution only 150 m lower down. In this case the change of Doppler spectra from the frozen hydrometeor level (4.27–4.42 km) to the rain level (3.51–3.81 km) was quite abrupt.

In order to interpret Figs. 12 and 13, it is necessary to have some idea about the trajectories of the hydrometeors. Are they falling in the plane of observation, and can it be assumed that the particles observed at elevations of 3.66 and 3.81 km evolved from distributions similar to the ones depicted at 3.96 and 4.27 km at about the same time? The answer seems to be in the negative.

Knowing the raindrop spectrum, one can calculate the liquid water content represented by the raindrops. If the spectrum is being altered as a result of breakup, coalescence or both, the water content should not change appreciably. As shown in Fig. 12, the water contents at altitudes 3.96, 3.81 and 3.66 km were 0.5, 0.08 and 0.002 g m^{-3}, respectively. This result argues that the raindrop spectrum at the lower altitude did not evolve from the one at the higher altitude. The data in Fig. 13 support the same conclusion.

A more reasonable interpretation of the curves in Figs. 12 and 13 is that, through the altitude layer between about 3.6 and 4.2 km, the hydrometeors had a component of motion through the time-height plane and were sorted by size in the manner discussed long ago by Gunn and Marshall (1955). The data in Fig. 3 show that above 3.6 km the wind increased with altitude and began turning into the south. Unfortunately, more detailed winds in the vicinity of the cloud are not available.

Although there was a bright band the precipitation was of a showery nature, and a stream of particles under the influence of wind shear would be sorted with the larger ones falling at faster speeds and being shifted toward the lower boundary of the shower. From this point of view, the calculated spectra in Figs. 11 and 12 would not be regarded as representing changes with time but rather observations at different places at the same time.

The calculated raindrop spectra below 3.81 km in Fig. 12 and below 3.96 km in Fig. 13 clearly do not conform to Marshall-Palmer distributions, and if they represent the true state of the hydrometeors, the use of Rogers' equation for calculating vertical air velocity would not be appropriate. The apparent relative scarcity of small raindrops could account for the fact that the positive tails of the Doppler spectra at altitudes below 3.81 km (in Fig. 11) do not extend below about -3 m s^{-1}. It appears to the author that the existence of nearly zero updraft in the layer extending from about 4.4 km down to about 3.5 km is more reasonable than the downward air accelerations suggested by Fig. 8.

Incidentally, calculations of raindrop sizes at altitudes $\lesssim$3.5 km are not discussed in this paper because uncertainties in vertical air velocities are so large as to render size determinations of even more questionable value than those displayed in Figs. 12 and 13.

4. Concluding remarks

This analysis illustrates the difficulties in interpreting observations of precipitation and air motions by means of a single, zenith-pointing radar. This is particularly the case when observing convective clouds.

The data suggest that immediately under the melting level the hydrometeors were in the form of raindrops whose diameters were exponentially distributed. This presumably occurred because the ice particles, which melted to form the liquid drops, had equivalent diameters which were similarly distributed.

These observations yield little information about changes of raindrop spectra *following the raindrops* as they fall. A possible explanation of the observations is that because of wind shear effects, the hydrometeors in a layer 600 m thick encompassing the melting level were moving through the height-time plane of observation and were being sorted by size. It is concluded that a few hundred meters below the melting level the raindrops were not exponentially distributed.

It is common, when using radar for the observation of rain, to assume that the raindrops conform to the Marshall-Palmer size distribution or at least to an exponential distribution. Commonly employed radar reflectivity-rainfall relations and reflectivity-mean terminal velocity relations involve such assumptions. The results cited above lead to the conclusion that sometimes they are not realistic.

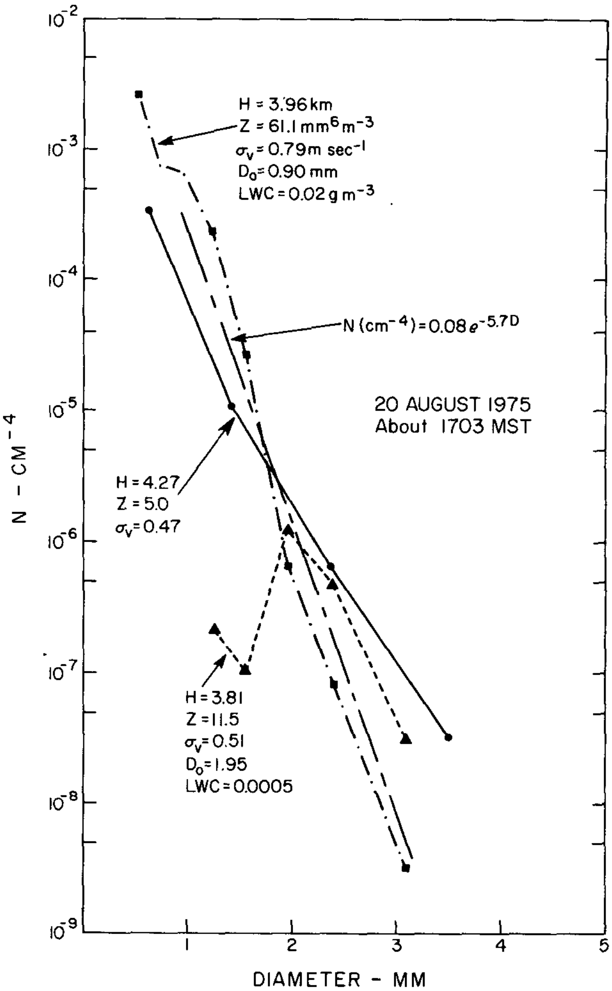

FIG. 13. Calculated size distributions at various altitudes at about 1703 MST. The line composed of short and long dashes is included as a general guide in judging the slopes of the other lines. It is described by the expression $N = 0.08\ e^{-5.7D}$.

Plans have been made to observe steady, winter rain in the same manner as was done in this investigation. It is hoped to get a situation where the pattern of precipitation is fairly steady. It would be particularly informative to have two zenith-pointing pulsed-Doppler radars located several hundred meters apart.

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