NOTES AND CORRESPONDENCE

Observations of Supercooled Raindrops in New Mexico Summertime Cumuli

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ABSTRACT

Observations made in 1987 with the NCAR King Air aircraft and in 1993 with the New Mexico Institute of Mining and Technology dual-polarization radar have revealed the presence of supercooled raindrops in some New Mexico summertime cumulus clouds. In the case of the radar data, the evidence for the supercooled drops came from a column of enhanced Z_{DR} that extended well above the 0°C level. The in situ data indicated that the supercooled raindrops were observed when cloud base was warmer than about 7°C and the depth of the cloud was greater than about 2.5 km.

1. Introduction

Most of the precipitation particles in New Mexico summertime cumulus clouds develop through the ice phase (e.g., Raymond and Blyth 1989; Blyth and Latham 1993). However, supercooled raindrops, which most likely formed by collision and coalescence, have been observed in a few clouds with aircraft and radar. The purpose of this note is to report on these observations. The in situ measurements were made with the National Center for Atmospheric Research King Air aircraft in small clouds that formed over the Magdalena Mountains in central New Mexico during August 1987. The radar measurements were made in a small cloud over the Rio Grande valley near Socorro using the New Mexico Institute of Mining and Technology (hereafter New Mexico Tech) dual-polarization radar during the summer of 1993.

Times are reported in UTC (local time + 6 h).

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2. Aircraft observations

In this section, we discuss measurements made mainly with the Particle Measuring Systems's 1D, 2DC, and 2DP probes. Details of the instruments on board the aircraft and of the clouds studied in the 1987 project were reported by Blyth and Latham (1993).

The main evidence for the presence of raindrops is spherical images obtained with the 2DC probe. However, there is also evidence, albeit weaker, from the 1D probe when 2D images were not recorded. Note that the sample volumes of the instruments are small, so it is possible that raindrops were present even if they were not detected by the probes. We argue that since ice was not observed by the 2D probes until the temperature of cloud top, $T_{top} \leq -10$°C (Blyth and Latham 1993), all particles recorded by the 1D probe are most likely to be liquid when $T_{top} \geq -5$°C.

Many clouds were studied during the project and raindrops were only observed in a few of them: Table 1 summarizes the observations. The temperature of cloud base was 7°C or warmer in all clouds in which supercooled raindrops were observed. The five clouds studied on 19–27 August were penetrated late in the development of the cloud and already contained ice at the time of the first penetration. So it is not clear if raindrops ever formed in these clouds, or if they were present but were not observed.

Cloud-base heights were measured by the King Air before the start of each cloud study, or occasionally by the New Mexico Tech SPTVAR aircraft. We have no information on how cloud-base height changed with

TABLE 1. Summary of observations of the clouds studied in the 1987 aircraft project. Here, T_{base} is the cloud-base temperature; 2D and 1D drops are drops detected by the 2DC and/or 2DP and the 1D probes, respectively. The entries under T_{obs} and cloud depth are the temperature and depth of the cloud at the first time when the raindrops were observed. As discussed in the text, supercooled drops were observed on two separate occasions in the same cloud on 10 August.

Date August 1987	T_{base} (°C)	2D drops	1D drops	T_{obs} (°C)	Cloud depth (km)	Ice in first penetration
8	9.0	No	Yes	3	Unknown	No
9a	7.4	Yes	Yes	−3	Unknown	No
9b	7.4	Yes*	Yes	−7	4.0**	No
10	10.7	No	Yes	−4	2.5	No
10	10.7	Yes	Yes	−3	2.5	No
12	7.0	No	Yes	−5	2.5	No
19	1.0	No	No			Yes
20	4.7	No	No			Yes
21	6.0	No	No			Yes
22	5.0	No	No			Yes
27	3.9	No	No			Yes
28	6.5	No	No			No

* Drops were mixed with ice particles.
** Best estimate.

time for a particular cloud, but observations made of other clouds at Langmuir Laboratory suggest that the change in height is typically small. Cloud-top height was determined using video observations from a forward-looking camera on board the aircraft and is accurate to about 200 m. In the following, we present data mainly from the 10 August 1987 cloud.

Time-lapse photographs have shown that the cumulus clouds that form over the Magdalena Mountains are multithermal in nature where new turrets ascend through the debris left by their predecessors. Usually, the mass of cloud gradually increases and cloud top ascends in stages as the cloud develops into a cumulus congestus. This typically takes 1–2 h. The transition to the mature thunderstorm phase occurs when a stronger and larger thermal rises through the cloud layer to higher altitudes.

The cloud on 10 August 1987 followed this general pattern of development. Figure 1 shows the variation with time of the altitude of cloud top. Cloud-base height was slightly below 3 km above mean sea level (MSL), corresponding to a temperature $T_{base} \approx 10°\text{C}$. The height of cloud top gradually ascended to about 5.5 km MSL ($T \approx -6°\text{C}$) by 1818. It then decreased to less than 4.0 km and increased to 5.5 km again at 1910. Thereafter, cloud top continued to rise and was near 6.5 km ($T \approx -12°\text{C}$) when the aircraft had to leave at 1935. There is considerable uncertainty in the measurements of cloud top between 1840 and 1905, when penetrations were being made in the lower part of the cloud. The times when supercooled raindrops were measured by the 1D and 2DC probes are indicated in the figure. Raindrops were measured in a single penetration at 1818 by the 1D probe and later in several penetrations from 1910 to 1920. The depth of the cloud was about 2.5 km or greater when the raindrops were observed. The first 2D images of ice were observed at 1925; this time is also indicated in the figure.

Figure 2 shows an example of images gathered in this cloud at 1910 by the 2DC probe at $T \approx -3°\text{C}$ ($z \approx 5.0$ km). The majority of images are spherical indicating either liquid or solid raindrops.[1] The highest cloud top at the time was about 2.5 km above cloud base. Ice particles were not detected by the 2D probes until 15 min later, so it is most likely that the drops formed by collision and coalescence. Notice that the largest drops have a diameter of about 1 mm and thus must have undergone growth by collision and coalescence. This growth stage was missed by the aircraft. Penetrations on 10 August were always made within about 500 m of the top of the highest turret, except during the one occasion when the aircraft spiraled down through the cloud to cloud base. The descent was made just prior to the first measurement of raindrops by the 2DC at 1910, so drops could have been present in the cloud up to about 30 min earlier.

Figure 3 shows that the raindrops were observed in concentrations of up to 35 L^{-1} in substantial downdrafts where the average cloud liquid water content was less than 1 g m^{-3}. Raindrops were not detected in the central region of the cloud where the liquid water content was higher and where there was an updraft. This was the first penetration in which any particles were detected by the 2DC probe. However, particles with $D \leq 100\ \mu\text{m}$ were detected by the 1D probe during a penetration made about 50 min earlier (see Fig. 1). The particles were detected at the time when cloud top ascended to

[1] The fact that we cannot determine the phase of the particles does not influence our arguments. If solid, their spherical shape suggests that they cannot have been solid for long. The important point is that they must have been supercooled raindrops at some point.

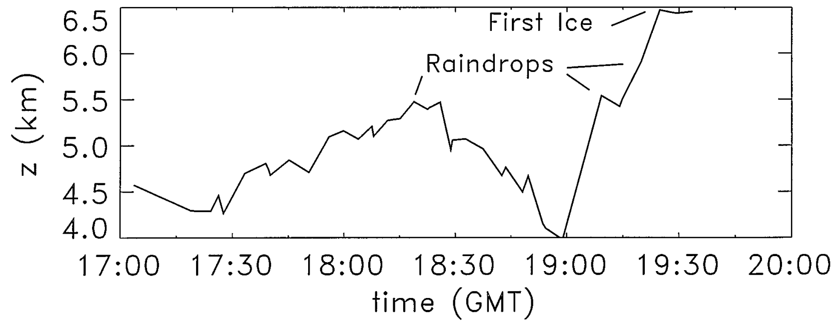

FIG. 1. Altitude of cloud top on 10 August 1987. Values of cloud-top height are uncertain between 1850 and 1905 as discussed in the text. The times when raindrops were observed are indicated. The first observation at 1818 was made with the 1D probe. Raindrops were observed with the 2DC in penetrations made between 1910 and 1920. The time of first ice is also shown.

about 2.5 km above cloud base for a brief period. Cloud top then descended to 2 km above cloud base once again, and no particles were detected by the 1D probe.

Raindrops were imaged by the 2DC probe in two other clouds during the 1987 project, both on 9 August (see Table 1). Evidence for raindrops from only the 1D probe was found in clouds sampled on 8 and 12 August.

The temperature of cloud base was 7°C or warmer in the five clouds where there were either 2D images of supercooled raindrops or particles that we believe to be supercooled raindrops detected by the 1D probe. Also, drops were not detected until the cloud depth was about 2.5 km or greater in the three clouds where there was a good measure of cloud depth. In particular, raindrops were observed on separate occasions in the 10 August cloud when the cloud depth reached about 2.5 km. It was not possible to determine the cloud top, nor cloud depth, in the 8 August cloud since the drops were observed while the aircraft was making penetrations through the lower levels of the cloud. Cloud top was obscured from view in the first cloud on 9 August.

Many other clouds were studied during the project (see Table 1), but the aircraft began penetrations of these when ice was already present. No evidence was found of supercooled (or frozen) raindrops in such clouds. This may be because cloud base was colder than 7°C in all of them or it may be that the cloud was sampled too late. Supercooled raindrops were not observed in the one cloud on 28 August that had a base colder than 7°C, but which was observed before ice formed. Supercooled raindrops were not observed in any clouds whose depth was less than about 2 km.

3. Radar observations

The radar observations are from 24 August 1993 with the New Mexico Tech dual-polarization radar (Krehbiel et al. 1996). The radar operates at a frequency of 9810 MHz (3.0-cm wavelength) with about 10-kW peak transmitted power. Alternate horizontal and vertical linear polarizations were transmitted for these observations. The presence of large liquid drops can be inferred

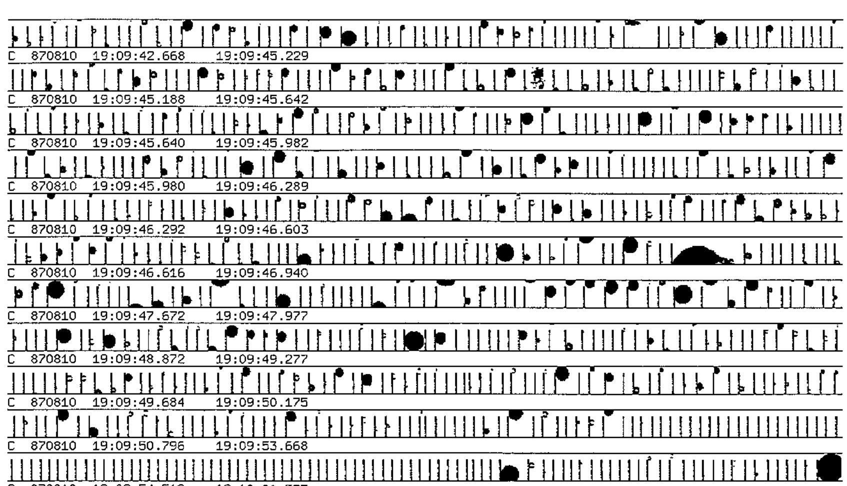

FIG. 2. 2D images made with the 2DC probe at 1909:45 on 10 August. The size represented by the space between the horizontal lines is 800 μm.

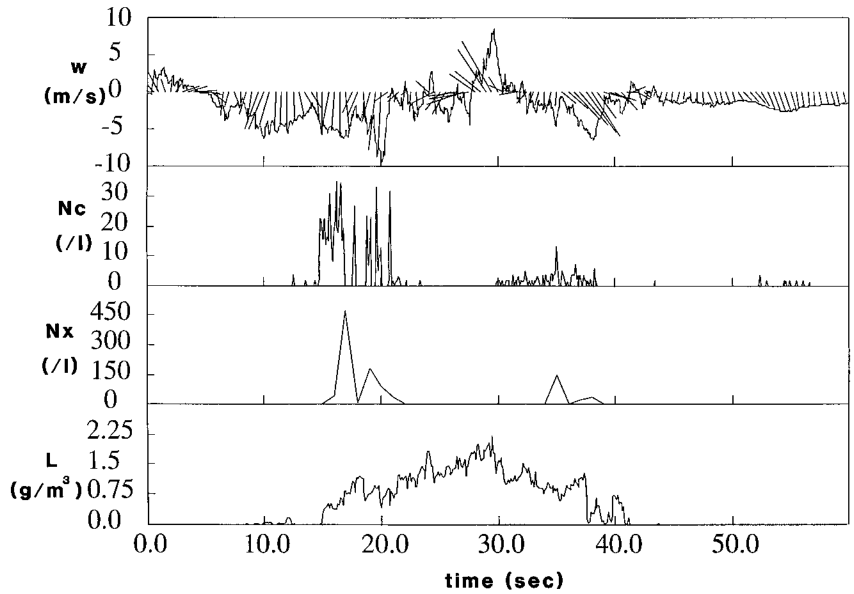

FIG. 3. Time series of data gathered on 10 August 1987. (a) The vertical wind and wind vectors constructed from the vertical wind and the horizontal wind component along the flight track; (b) the concentration of particles detected by the 2DC probe; (c) concentration of particles measured by the 260X probe; and (d) the liquid water content measured by the Forward Scattering Spectrometer Probe (FSSP).

from measurements of the differential reflectivity, defined as $Z_{DR} = 10\ \log(Z_H/Z_V)$, where Z_H and Z_V are the reflectivity factors measured with horizontally and vertically polarized radiation, respectively (e.g., Caylor and Illingworth 1987). The horizontal dimension of large, falling raindrops is greater than the vertical dimension and this results in positive values of Z_{DR}.

Figures 4 and 5 show a sequence of vertical scans through the center of a small cloud that developed about 12 km distant from the radar. These show the cloud development from the time of the first echo at 2006 to the start of cloud decay at 2022. Note that cloud base was not measured. Figure 4 shows the reflectivity values versus time and Fig. 5 shows the differential reflectivity. The reflectivity scans show that two precipitation events occurred in the cloud. The first, less intense event began in the core of the cloud at 2011 with the development of 15 dB*Z* reflectivity between 3- and 4.5-km altitude above ground level (4.5–6 km MSL; $T \approx 0 \rightarrow -10°\mathrm{C}$) (Fig. 4d). The echo from this precipitation descended in altitude and intensified to about 25 dB*Z* over the next 5 min until reaching ground by 2016 (Fig. 4i). The second precipitation event began at 2014 at 5-km altitude above ground level (AGL) (6.5 km MSL; $T \approx -13$ °C) on the upper-right side of the storm (Fig. 4g). Its echo intensified and spread more quickly than that of the initial precipitation event, increasing in intensity to 35 dB*Z* by 2020 and becoming the dominant echo in the cloud.

The differential reflectivity results of Fig. 5 show positive values for the first precipitation event but not for the second. For the first event, Z_{DR} began to exceed $+1$ dB at 2010 (Fig. 5c), one minute prior to the reflectivity reaching 15 dB*Z*. The two succeeding scans at 2011 and 2012 showed Z_{DR} values between 1 and 2 dB associated with the initial precipitation event, with the top of the enhanced Z_{DR} region extending up to 4.5 km AGL (6.0 km MSL; $T \approx -10°\mathrm{C}$) in both scans (Figs. 5d and 5e). By 2013 (Fig. 5f), the top of the enhanced Z_{DR} region had descended to about 3-km altitude AGL (4.5 km MSL), corresponding to the 0°C level, where it stayed for the remainder of the observations.

In sharp contrast, the second more intense precipitation event did not produce positive Z_{DR} values. Rather, Z_{DR} tended to be zero or even slightly negative, which is typical of graupel or hail. The difference between the two events leaves little doubt that the initial precipitation, with its column of positive Z_{DR} values above the 0°C level, contained large supercooled raindrops. The fact that the echo grew in place (Figs. 4a–c) suggests that the precipitation also grew in place, as a result of collision and coalescence with supercooled cloud drops. The possibility that the drops resulted from recirculation of melted graupel particles would require that graupel particles formed prior to 2010 (when the enhanced Z_{DR} values occurred), descended several kilometers, melted, and then reascended in the updraft back up to the $-10°\mathrm{C}$ level, all in the span of 1–2 min from initial echo detection. This is highly inconsistent both with the initial echo development and with the observations of the subsequent graupel event, which did not produce large liquid drops until at least 6 min after the start of the graupel formation.

We note that the weak positive Z_{DR} values (<1 dB) at the time of the initial echo and subsequently in weak

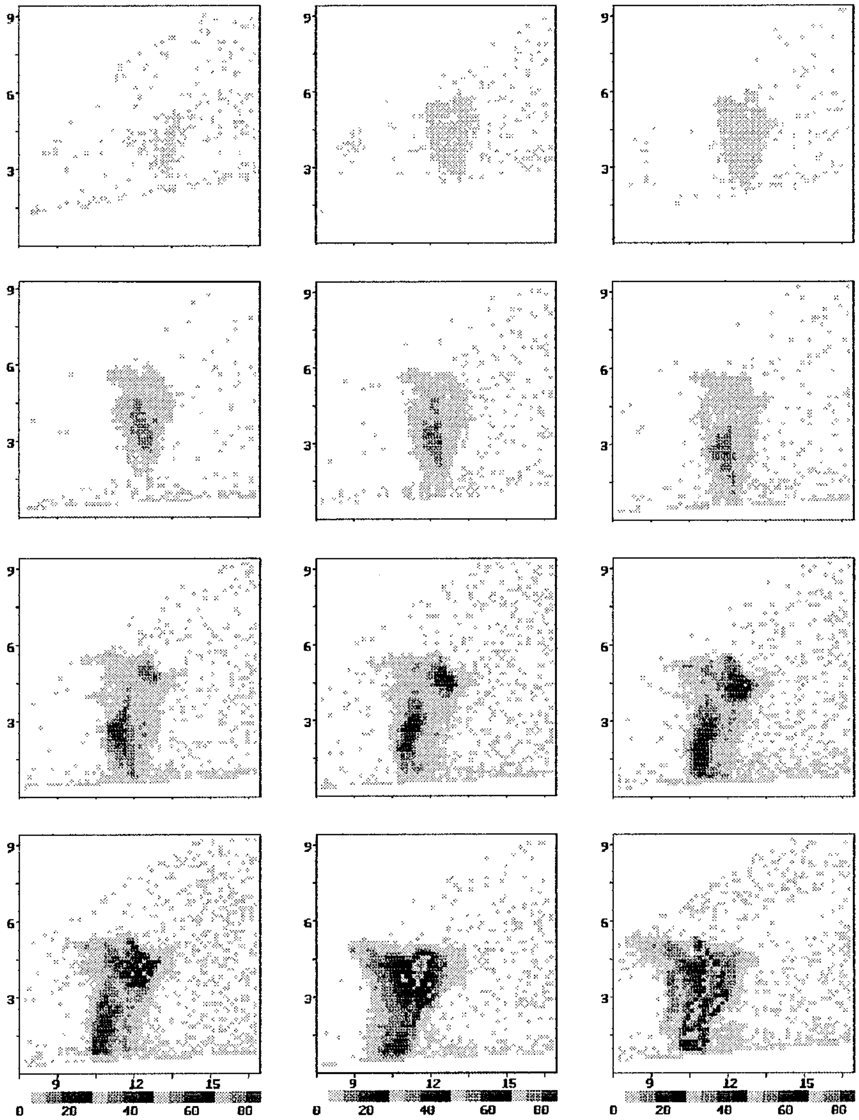

FIG. 4. Reflectivity in dBZ measured with the New Mexico Tech radar on 24 August 1993. The times for each display are (a) 2006:25, (b) 2009:11, (c) 2010:13, (d) 2011:24, (e) 2012:08, (f) 2013:04, (g) 2014:17, (h) 2015:18, (i) 2016:07, (j) 2017:08, (k) 2020:04, and (l) 2022:28.

echo regions of the cloud were due to a slight mismatch in the vertical and horizontal polarization receiver responses at low signal-to-noise ratios and are not an indication of liquid drops when the echo strength was near minimum detectable signal. As the echo intensity increased above the noise, the indicated values of differential reflectivity are correct, being positive for the liquid precipitation event and 0 for the solid precipitation (e.g., above 3 km AGL in Fig. 5l).

4. Discussion

We have presented results from aircraft and radar measurements showing the existence of supercooled raindrops in the developing stages of New Mexico summertime cumuli.

The aircraft observations indicated that supercooled drops were observed when cloud base was warmer than about 7°C and when the depth of the cloud increases to about 2.5 km. The results from the radar measurements are consistent with this latter finding. The first observation is reasonable since, according to closed parcel models of cloud droplet growth, the size of cloud droplets is greater for a particular altitude change upward from cloud base for warmer cloud bases (Rogers and Yau 1989). One possibility suggested by the latter result is that there is a sharp limit in the size of cloud droplets required for the onset of collision and coalescence below

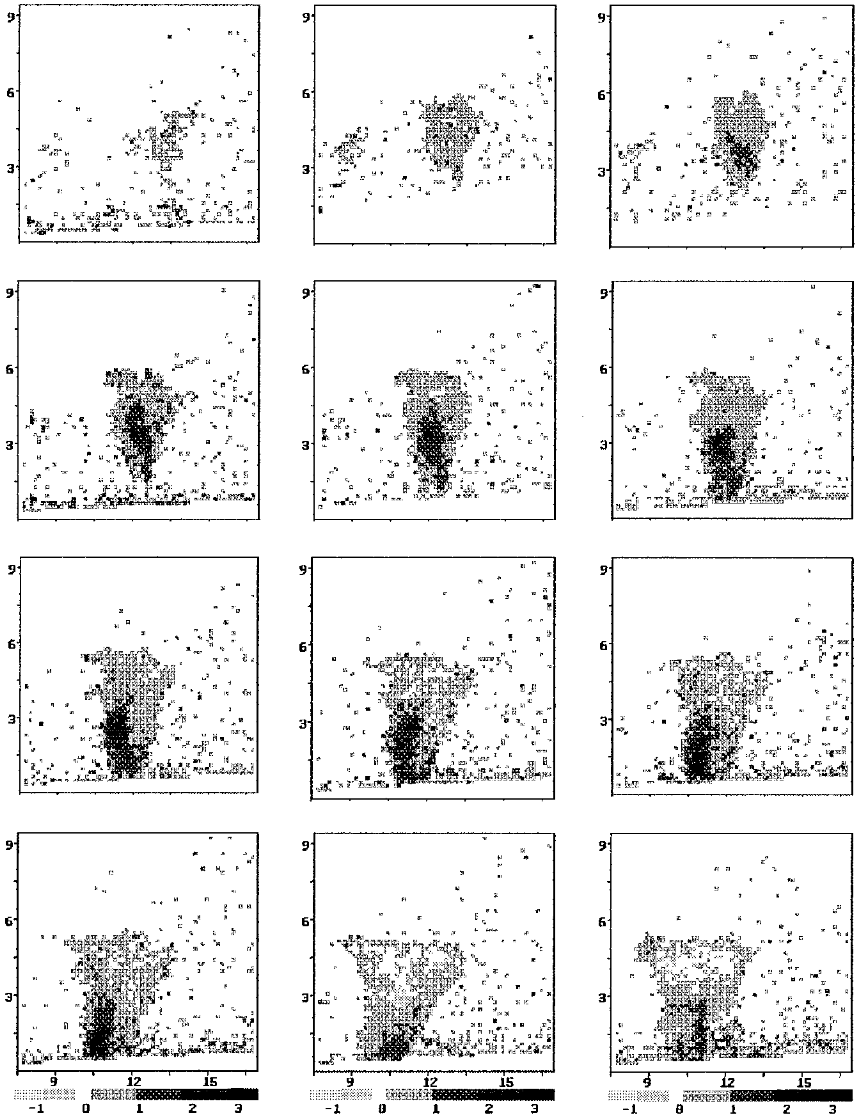

FIG. 5. As in Fig. 4 but for Z_{DR} measured in dB.

which the efficiency of collision is insignificantly small. The size of cloud droplets increases with altitude in a closed parcel, but the rate of increase decreases with altitude. Indeed, the radius typically increases by only about 1 μm in an adiabatic ascent from 2 to 2.5 km in New Mexico clouds. It is believed that the collision and coalescence process does not become significant until some cloud droplets grow to a radius of about 20 μm (Rogers and Yau 1989). However, Beard and Ochs (1993) pointed out that the collision efficiency is not well known for collector drops of about this size. So it is not really known when the collision efficiency begins to increase rapidly, or even how rapidly the increase is.

An alternative explanation for the results is that cloud drops will be exposed to significant liquid water for a longer time when the cloud is deeper, thus increasing the time for growth by collision and coalescence. This could be quite significant if giant or ultragiant nuclei are present as suggested by Johnson (1982). It is likely that the liquid water content is close to the undiluted value for much of the ascent to cloud top in the core of thermals (e.g., Blyth et al. 1988). The picture is consistent with the fact that the first raindrops measured by the 2DC on 10 August 1987 were contained in the downdraft and noticably absent from the updraft (see Fig. 3). The additional distance could also be important if the large cloud droplets that are required for the onset of collision and coalescence are produced by mixing (Baker et al. 1980; Telford and Chai 1980; Cooper 1989). The observations reported herein are also consistent with the process suggested by Twomey (1976), where a few large drops are produced by trajectories through small regions of higher than average liquid water content. Examination of these ideas will be possible with calculations and analysis of data gathered in the Small Cumulus Microphysics Experiment.

Finally, we note that raindrops continued to be observed in the two aircraft penetrations made after 1910 on 10 August 1987. In fact, raindrops were observed in a strong updraft at 1920 at $T \approx -5°C$. The presence of raindrops in an updraft at this temperature could be important for the development of ice, because of splinters that may be produced when the drops freeze (Chisnell and Latham 1976). Koenig (1963) and Braham (1964) found that supercooled raindrops played an important role in the development of precipitation via the ice phase in Missouri summertime cumuli. It is interesting to note that, in the radar data shown in Figs. 4 and 5, the graupel echo developed above and to the side of the supercooled raindrop echo subsequent to the inferred freezing of the raindrops. The graupel echo may have been produced by ice splinters ejected from the freezing raindrops.

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REFERENCES

Baker, M. B., R. G. Corbin, and J. Latham, 1980: The influence of entrainment on the evolution of cloud droplet spectra: I. A model of inhomogeneous mixing. *Quart. J. Roy. Meteor. Soc.,* **106,** 581–598.

Beard, K. V., and H. T. Ochs III, 1993: Warm-rain initiation: An overview of microphysical mechanisms. *J. Appl. Meteor.,* **32,** 608–625.

Blyth, A. M., and J. Latham, 1993: Development of ice and precipitation in New Mexican summertime cumulus clouds. *Quart. J. Roy. Meteor. Soc.,* **119,** 91–120.

——, W. A. Cooper, and J. B. Jensen, 1988: A study of the source of entrained air in Montana cumuli. *J. Atmos. Sci.,* **45,** 3944–3964.

Braham, R. R., Jr., 1964: What is the role of ice in summer rainshowers? *J. Atmos. Sci.,* **21,** 640–645.

Caylor, I. J., and A. J. Illingworth, 1987: Radar observations and modelling of warm rain initiation. *Quart. J. Roy. Meteor. Soc.,* **113,** 1171–1191.

Chisnell, R. F., and J. Latham, 1976: Ice particle multiplication in cumulus clouds. *Quart. J. Roy. Meteor. Soc.,* **102,** 133–156.

Cooper, W. A., 1989: Effects of variable growth histories on droplet size distributions. Part I: Theory. *J. Atmos. Sci.,* **46,** 1301–1311.

Johnson, D. B., 1982: The role of giant and ultragiant aerosol particles in warm rain initiation. *J. Atmos. Sci.,* **39,** 448–460.

Koenig, L. R., 1963: The glaciating behavior of small cumulonimbus clouds. *J. Atmos. Sci.,* **20,** 29–47.

Krehbiel, P. R., T. Chen, S. McCrary, W. Rison, G. Gray, and M. Brook, 1996: The use of dual-channel circular-polarization radar observations for remotely sensing storm electrification. *Meteor. Atmos. Phys,* **59,** 65–82.

Raymond, D. J., and A. M. Blyth, 1989: Precipitation development in a New Mexico thunderstorm. *Quart. J. Roy. Meteor. Soc.,* **115,** 1397–1423.

Rogers, R. R., and M. K. Yau, 1989: *A Short Course in Cloud Physics.* Pergamon Press, 293 pp.

Telford, J. W., and S. K. Chai, 1980: A new aspect of condensation theory. *Pure Appl. Geophys.,* **118,** 720–742.

Twomey, S., 1976: The effects of fluctuations in liquid water content on the evolution of large drops by coalescence. *J. Atmos. Sci.,* **33,** 720–723.