Ice Initiation in Unmixed Updraft Cores in Northeast Colorado Cumulus Congestus Clouds

Andrew J. Heymsfield, Charles A. Knight and James E. Dye

National Center for Atmospheric Research, Boulder, CO 80307

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ABSTRACT

Ice particle concentrations have been measured from the NOAA/NCAR Explorer sailplane in unmixed and mixed updraft regions within northeast Colorado cumulus congestus clouds, and compared with the concentrations predicted from measured ice nucleus spectra. The clouds investigated were "cold, continental" cumulus with droplet populations of $-1000 \text{ cm}^{-3}$ and cloud base temperatures between $+7$ and $-7^\circ C$. The concentrations of ice particles within unmixed updraft regions, exclusive of ice particles so large that they almost certainly entered the unmixed region by sedimentation or recycling, are consistent with those expected on the basis of the ice nucleus spectra, suggesting that primary ice nucleation is the dominant mechanism active within these regions. Millimeter size ice particles found within unmixed updraft cores presumably enter through sedimentation or recycling. The data do not indicate an important role for ice multiplication in these clouds.

1. Introduction

Braham and Squires (1974), reporting on a review of cloud physics, considered clarification of the relationship between ice crystal and ice nucleus concentrations as "absolutely essential to a complete understanding of natural precipitation." The present study examines this problem, using measurements obtained in unmixed updraft regions and mixed updraft and downdraft regions in vigorous cumulus congestus clouds in northeast Colorado.

The relationship between ice nucleus and ice crystal concentrations in clouds has been the subject of numerous investigations conducted over the past 15 years. Ice crystal concentrations up to four orders of magnitude higher than ice nucleus concentrations have been noted in summertime cumulus clouds, which exhibit cloud droplet spectra typical of modified continental clouds (Koenig, 1963) or of maritime clouds (Mossop et al., 1968). A comparable enhancement in ice crystal concentration has been noted in wintertime clouds formed in maritime air (Koenig, 1968; Hobbs, 1969). These studies suggest that collisions between ice particles and droplets are responsible for initiating secondary ice crystal production, or "ice multiplication." Hobbs (1969) reported that the greatest enhancement occurs when the cloud top temperatures are warmer than $-10^\circ C$.

This is consistent with the observation by Mossop et al. (1968) of unexpectedly high ice particle concentrations in a cumulus cloud with a cloud-top temperature of $-4^\circ C$. In contrast, no significant discrepancy between the concentrations of ice nuclei and ice crystals was observed by Gagin (1975) in wintertime continental cumulus clouds in Israel.

Recent laboratory experiments by Hallett and Mossop (1974), Mossop and Hallett (1974) and Mossop (1976) show an ice multiplication effect, as yet not understood, that depends in a complex way on the drop size spectrum and the temperature. Secondary ice production via this effect may help to explain some of the field observations.

Ice particles in cumulus clouds can be initiated through a variety of mechanisms over a wide range of cloud vertical depth. Therefore, the processes of ice initiation are difficult to investigate from aircraft measurements since only limited regions of the cloud are penetrated. In the present study, measurements of ice crystal concentrations in updrafts within summertime, continental, cumulus congestus clouds are compared to measurements of ice nucleus concentrations. The primary emphasis of this comparison is on the unmixed and slightly mixed updraft regions described by Heymsfield et al. (1978, hereafter referred to as HJD) because ice initiation within undiluted updraft cores can be more easily interpreted. The ice nucleus spectrum at the cloud base or at the surface is representative of the spectrum within the core, since the air did enter the cloud through the cloud base. In addition, the time available for ice particle growth at each level
within this region can be calculated to within a factor of 2.

2. Data acquisition

The cloud physics data used in this study were acquired during the 1976 field program of the National Hail Research Experiment (NHRE), primarily with the NOAA/NCAR Explorer sailplane which entered the side or base of vigorous cumulus congestus clouds and ascended within the updraft. Supporting measurements of the cloud microstructure were obtained in some cases by the University of Wyoming Queen Air 10UW at approximately the −5°C level. Thermodynamic properties of the inflow air below cloud base were measured by the NCAR Queen Airs 304D and 306D. A more complete description of the aircraft and instrumentation employed in this study and the corresponding flight patterns is presented in HJD.

Ice particle concentrations were obtained from a Cannon particle camera (Cannon, 1975) mounted on the sailplane. The threshold size for particle detection is ≈100 μm diameter and the limit for distinguishing water from ice particles with the camera as used on the sailplane was 200 μm diameter. Because previous (Cannon et al., 1974) and continuing studies show that the clouds in the NHRE area rarely contain drizzle or larger size drops, we have assumed that all particles between 100 and 200 μm diameter were ice particles. The sample volume of the sailplane Cannon camera (Fig. 1) increases from nearly 4 ɛ for 100 μm particles to 15 ɛ for 1 mm particles over 10 s, the minimum averaging time used in this study (this corresponds to 16 frames, or 400 m of flight path). Measurements of cloud droplet size spectra were obtained with a PMS Forward Scattering Spectrometer Probe (FSSP) mounted on the sailplane. Several problems were found with the FSSP measurements. These malfunctions limited the use of the data to the first 5–10 min after the sailplane entered the cloud. These problems are discussed in more detail in Section 5.

Measurements of ice nuclei (IN) were made using membrane filter collection systems on 304D, 306D and 10UW and at the surface. Data from 304D and 306D were obtained primarily in inflow regions, and from 10UW in both inflow and outflow regions. Conventional Millipore filters were used on 304D and 306D and hydrophobic (Sartorius Type, 0.45 μm) membrane filters on 10UW, and both types were used for surface samples. Typical sample volumes were 100 ɛ. The filters exposed on 304D and 306D were processed at temperatures (T) of −16 and −20°C by G. Langer at NCAR using a rapid response thermal gradient diffusion chamber similar in design to that described by Langer and Rodgers (1975). The operation of the chamber in 1976 was improved over that described by Langer and Rodgers by modifying the airflow so that it was perpendicular to the filter and by more carefully controlling the humidity (G. Langer, private communication). The filters exposed on 10UW were processed at the University of Wyoming, at T = −16°C in a static thermal gradient diffusion chamber or in a ventilated cham-
ber over the range of ice supersaturations from 11 to 19% (see Rogers and Vali, 1978). Additional measurements of IN were obtained from samples collected on 304D and 10UW with aluminized Mylar bags. These bags were processed at −20°C in an NCAR-type ice nucleus counter; an improved version of the counter described by Langer et al. (1967). Filter samples of aerosols collected at the ground at the Tennay Ranch (near Grover, Colorado), at Sidney, Nebraska, at the Chapel Ranch (Southeast of Grover, Colorado) and at Sterling, Colorado, were processed for ice nuclei with the two chambers noted above. Intercomparisons of the two techniques showed that the two processing systems gave comparable results (Rogers and Vali, 1978).

3. Calculations of ice crystal concentrations expected within an unmixed core

To compare the measured ice crystal concentrations with those expected from ice nucleus concentrations, it was necessary to calculate the expected concentrations of ice crystals >100 μm, the lower limit of particle detection by the Cannon camera. These calculations used a simple one-dimensional parcel model described in HJD, coupled with equations describing ice crystal growth. The parcel model also provided idealized vertical profiles of temperature and vertical velocity.

In the model, ice crystals in concentrations corresponding to the ice nuclei\(^4\) (see Section 4) were initiated in 1°C increments of temperatures from −10 to −20°C. The concentration of ice crystals initiated at −10°C was the total number of nuclei active at ≥ −10°C, because no IN measurements were available for temperatures warmer than −10°C. As the IN concentrations are generally very low and the linear growth rates are relatively low at temperatures warmer than −10°C, the effect of this approximation on the derived concentrations of crystals >100 μm is negligible. The negligible increase on the maximum ice crystal dimension that would result from initiating ice crystals at −5°C rather than −10°C is exemplified by one of the warmer cloud base cases (22 July), in which the maximum dimension was found to be only 15 μm greater.

The linear and mass growth rates used in this study are based on experimental values of Ryan et al. (1976) for ice crystals with maximum dimensions between 20 and 200 μm. The measurements of Ryan et al. were made at constant temperatures over the temperature range −3 to −21°C. Since the present calculations are for growth at continually decreasing temperature, the use of laboratory measurements at constant temperatures may introduce a small error in the calculated concentrations of ice crystals >100 μm, but it is not possible to evaluate this error with present data. The ice particles are assumed to move with the air.

As the laboratory experiments reported above were conducted at a pressure of ~1000 mb, the growth rate at 400–500 mb, for example, would be underestimated if the data were applied directly. The growth rates were therefore adjusted for lower atmospheric pressures by the technique described in the Appendix. Calculations for the appropriate pressure level slightly increased the predicted concentrations of detectable ice crystals. Of all of the cases examined, the largest ice crystal diameter achieved in an updraft to a temperature of −22°C was 200 μm. This was, of course, in the case with the weakest updraft.

"Expected" ice crystal concentrations in mixed updraft regions were computed in much the same way as for unmixed regions, except that the vertical velocity and temperature profiles were computed using conventional entrainment calculations given in HJD. The same IN spectra were used in these calculations.

HJD present comparisons between the predicted and measured updraft velocities. The measured 10 s average updraft velocities, which did not reflect the peak 1 s vertical velocities, were frequently lower than the calculated values—by as much as one-half for the unmixed regions and often much more for the mixed regions. The calculated concentrations of ice crystals that might result from primary nucleation may therefore be underestimates. Using a vertical velocity profile that was in close agreement with 10 s average vertical velocities reported in HJD produced an increase in the calculated ice crystal concentrations at −20°C of at most 0.08 \(ε^{-1}\) and was generally less than 0.03 \(ε^{-1}\). This approximation is thought to be negligible for the present study.

As mentioned previously, the clouds investigated in this study are the same as those described by HJD and a more complete description of the characteristics can be found therein. The thermodynamic and dynamic measurements that are particularly relevant to this study appear in Table 1. The vertical velocities computed from the model for six of the clouds investigated appear in HJD.

a. Ice nucleus measurements

IN spectra were calculated from data obtained at Tennay Ranch at 1300 MDT, and at Sidney at 1300 and 1700 MDT. Concentrations were typically ~0.2 \(ε^{-1}\) at −16°C and 1.0 \(ε^{-1}\) at −20°C, as indicated in Fig. 2. The IN measurements obtained at Tennay Ranch and Sidney at 1300 MDT are in reasonable

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\(^4\) The IN concentration was adjusted to account for the expansion of the air in rising to the level where the nuclei were active.
agreement, suggesting that the ice nucleus spectrum was spatially homogeneous near the surface. Increased IN concentrations were occasionally noted at 1700 MDT. This increase has been attributed to thunderstorm outflow (G. Langer, private communication). The aircraft measurements, taken below cloud base and as much as possible within storm inflow, were in overall agreement with the surface measurements at $-16$ and $-20^\circ$C. Rogers and Vali (1978), by comparing measurements from aircraft with those at the ground, have concluded that the surface measurements of aerosols and nuclei in northeast Colorado are representative of the inflow at cloud base. Further study of this question is underway using data from the NCAR counter on 304D, and is yielding the same conclusion (G. Morgan, private communication).

A “representative ice nucleus spectrum” was selected for each of the days, based on proximity in space and time to the cloud sampled, and is shown in Fig. 2. These spectra were used in the model calculations.

![Fig. 2](image-url)

**Table 1. Summary of measurements.**

<table>
<thead>
<tr>
<th>Date (1976)</th>
<th>Temperature ($^\circ$C)</th>
<th>Vertical wind (m $\text{s}^{-1}$)</th>
<th>Range of sampling temperature ($^\circ$C)</th>
<th>Peak droplet concentration (cm$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>21 June, cell 1</td>
<td>1.6</td>
<td>6</td>
<td>-14 to -22</td>
<td>705</td>
</tr>
<tr>
<td>21 June, cell 2</td>
<td>1.3</td>
<td>4</td>
<td>-2 to -13</td>
<td>1010</td>
</tr>
<tr>
<td>22 June</td>
<td>2.6</td>
<td>6</td>
<td>-5 to -20</td>
<td>880</td>
</tr>
<tr>
<td>30 June</td>
<td>1.8</td>
<td>5</td>
<td>-10 to -20</td>
<td>1330</td>
</tr>
<tr>
<td>6 July</td>
<td>-1.0</td>
<td>3</td>
<td>-8 to -32</td>
<td>880</td>
</tr>
<tr>
<td>8 July, cell 1</td>
<td>-6.7</td>
<td>3</td>
<td>-5 to -14</td>
<td>830</td>
</tr>
<tr>
<td>8 July, cell 2</td>
<td>-7.0</td>
<td>3</td>
<td>0 to -14</td>
<td>Instrument malfunction</td>
</tr>
<tr>
<td>9 July</td>
<td>-7.6</td>
<td>4</td>
<td>-8 to -13</td>
<td>1025</td>
</tr>
<tr>
<td>22 July</td>
<td>6.2</td>
<td>2</td>
<td>-5 to -24</td>
<td>890</td>
</tr>
<tr>
<td>25 July, penetration 1</td>
<td>6.7</td>
<td>3</td>
<td>0 to -11</td>
<td>1310</td>
</tr>
<tr>
<td>25 July, penetration 2</td>
<td>7.1</td>
<td>3</td>
<td>-3 to -19</td>
<td>Instrument malfunction</td>
</tr>
<tr>
<td>27 July</td>
<td>0.5</td>
<td>3</td>
<td>9 to -14</td>
<td>1190</td>
</tr>
</tbody>
</table>

![Fig. 2](image-url)

**Fig. 2.** Measured ice nucleus concentrations versus temperature for collections at various sites on the days examined in the present study. The “representative” spectrum used is that used in the calculations of the expected ice crystal concentration. The IN spectra are from data at $-16^\circ$C and a range of supersaturations with respect to ice, converted to a temperature range assuming water saturation.
b. Ice crystal concentrations

The measured ice crystal concentrations were divided into two categories: 1) particles with diameters ($D_{\text{ice}}$) between 100 and 200 $\mu$m which presumably had recently nucleated within the unmixed core from ice nuclei that ascended from cloud base and 2) particles with $D_{\text{ice}} > 200$ $\mu$m which were too large to have nucleated and grown to that size in the core and thus presumably have been introduced into the unmixed core by sedimentation, with or without recycling. Concentrations from each category were averaged over temperature intervals corresponding to 3°C, within which the concentration would not be expected to vary by more than a factor of 5 as a result of primary nucleation. Averaging over smaller temperature intervals would have given poorer sampling statistics. The measured concentrations of ice particles in both categories are plotted versus temperature for unmixed regions ($\theta_v$ within $\pm 1$ K of the cloud-base value of $\theta_v$) in Fig. 3. The concentrations of particles between 100 and 200 $\mu$m diam-

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**LEGEND**

- $D \leq 200$ $\mu$m
- $D > 200$ $\mu$m
- IN ICE NUCLEUS SPECTRUM
- [] NUMBER OF ICE PARTICLES OBSERVED

**Fig. 3.** Measured ice crystal concentrations in unmixed regions versus temperature for $D_{\text{ice}} \leq 200$ $\mu$m and $D_{\text{ice}} > 200$ $\mu$m. Ice crystal concentrations are averaged over 3°C temperature intervals. The expected concentrations of ice crystals >100 $\mu$m (curve 1) based on ice crystal growth rates at the appropriate pressure level indicate a stepwise increasing concentration with decreasing temperature because of the assumed initiation of ice crystals in 1°C increments. The representative IN spectra are indicated by the bold curve.
eter are shown by solid vertical lines and the concentration of those $\geq 200 \mu m$ diameter are shown by dashed vertical lines. These lines extend over the $3^\circ C$ temperature intervals for which sailplane data were available for each of the 12 days. The number in brackets beside each bar gives the number of particles actually photographed. The figure also shows the representative IN spectrum for each day by a solid bold line. The expected concentrations of ice crystals $>100 \mu m$ ($N$, designated as 1 in Fig. 3), based on ice crystal growth rates at the appropriate pressure level indicate a steplike increasing concentration with decreasing temperature. The steps are a result of particles being nucleated in increments of $1^\circ C$ in the model, and the rapidly increasing rate of particle growth near $-15^\circ C$ as the particles ascend in the updraft.

The upper size limit of $200 \mu m$ represented the largest size that particles achieved in the model calculations, using the updrafts implied from parcel theory. As noted previously, the updraft velocities in the calculations are likely to be overestimates, and therefore the $200 \mu m$ upper limit may be an underestimate of the largest size that particles could have achieved following nucleation. To test for the sensitivity to this upper limit, the concentrations of ice particles with $200 \mu m < D \leq 400 \mu m$ in mixed regions were determined for the time periods. The ratios of these concentrations to those which use $200 \mu m$ as an upper limit are listed in Table 2. It is clear that the conclusions drawn in this paper are not sensitive to the $200 \mu m$ upper limit of ice particle size.

Concentrations of ice crystals with $D_{ice} \leq 200 \mu m$, averaged for the unmixed regions of all of the clouds investigated, are presented in Fig. 4. The limits of the measured ice nucleus and calculated ice crystal concentrations are shown for comparison. Ice crystal and ice nucleus concentrations were in agreement within a factor of 10 in the data thus averaged.

Measured ice crystal concentrations are plotted versus temperature for mixed regions in Figs. 5–7. The techniques used in averaging the data are similar to those described for unmixed regions. The data were divided into three groups to seek a correlation between different amounts of mixing and the ice crystal concentration. In one group (Fig. 5) $\theta_e$ values in the updraft were between 1 and 2 K of the cloud base value, indicating that the air was probably only slightly mixed. The second group (Fig. 6) contained the portions of the updraft regions in which $\theta_e$ was more than 2° different from the unmixed value. The third group (Fig. 7) comprised measurements obtained within downdraft regions. The concentration of ice crystals $>100 \mu m$ calculated as described above also appears in Figs. 5–7, though obviously in downdrafts (Fig. 7) these concentrations have no meaning. The predictions for the mixed regions are slightly higher than those in unmixed regions because the time available for growth is longer. The calculation may be unrealistic in that it assumes that the ice nucleus concentration is still given by the “representative” spectrum, but lacking measurements of nuclei around the side of the cloud, it is the best that can be done at present.

Ice particle concentrations comparable to those measured in unmixed updraft regions are found in regions of mixed updraft that were within 1–2°C of the unmixed $\theta_e$ value (Fig. 5). As the degree of mixing within updraft regions increases (Fig. 6), one
finds higher concentrations of ice particles with $D_{\text{ice}} > 200 \mu\text{m}$. There is satisfactory agreement between the predicted and the measured ice crystal concentrations within mixed updraft regions for ice crystals with $D_{\text{ice}} \leq 200 \mu\text{m}$, with the exception of 6 and 22 July. Concentrations of ice particles with $D_{\text{ice}} > 200 \mu\text{m}$ were about the same in heavily mixed updraft (Fig. 6) and in downdraft (Fig. 7) regions. However, concentrations in heavily mixed and in downdraft regions were significantly higher than in slightly mixed updraft regions (Fig. 5). No expected ice crystal concentrations are indicated in Figs. 5, 6 and 7 for 22 June, 30 June and 8 July, cell 1. Using the conventional entrainment calculations, the air parcels in each of these cases did not lift to altitudes that were cold enough to produce ice crystals of 100 $\mu\text{m}$ size.

The regions sampled above 7.7 km on 6 July were previously thought to be unmixed (HJD) as evidenced by the $\theta_e$ values that were within 1 K of the
moist adiabatic value. The high concentration of rimed ice particles and the absence of liquid water noted at the upper levels of this region implied that appreciable energy was released in the air parcel due to the freezing (riming) of cloud droplets. The increase in $\theta_e$ due to the freezing of cloud droplets, which could have amounted to 3 K through freezing of the entire liquid water mass within an unmixed air parcel, was estimated by assuming that the reduction in liquid water below the adiabatic value was entirely due to freezing. A comparison of the measured and “recomputed unmixed” $\theta_e$ values provided a means of distinguishing unmixed from mixed updraft regions.

4. Discussion

The idea that motivated our analysis was to take advantage of the knowledge of undiluted ascent in certain portions of the cloud (HADD) to compare ice nucleus with ice crystal concentrations. These regions are the only ones in which such comparisons can be interpreted with any simplicity, because mixing introduces new sources of ice nuclei and greatly complicates deduction of the temperature history of the air. Much of the previous work has used cloud-top temperature as a lower limit of the nucleation temperature, which is valid, but it suffers from lack of definition of the source (or sources)
of the nuclei, and of the time available for development of the observed ice crystal concentrations.

In the present study, the measured ice crystal concentrations in the unmixed updraft portions are not inconsistent with those predicted from the ice nucleus measurements. Thus, there is no reason to appeal to ice multiplication. The high concentrations of ice crystals that are sometimes found within the mixed regions of these clouds might owe their origin to mixing from above or to some type of multiplication process. On both 6 and 22 July, for instance, radar cloud tops were above the −40°C level, so nucleation itself need not be a problem, but the data are not unequivocal. The origin of the ice in the mixed regions needs further study.

Ice nucleus concentrations measured with the filter technique (two different chamber designs) and the NCAR counter were in good agreement (Rogers and Vali, 1978; NCAR Tech Note, to be published). With more extensive data it might have been possible to conclude that the ice nucleation mode(s) active in the ice nucleus measurement techniques also applied to the cloud environment. However, too much of the agreement is in the form of zero measurable concentrations, both predicted and found, to enable such conclusions to be made. A
larger sample volume for ice and/or measurements higher in the clouds are needed for such a study.

While the ice concentration data do not give any reason to invoke ice multiplication, it is of interest to examine the question of what the observations in these clouds would have shown if ice multiplication had been operating. Previous studies have shown that ice multiplication processes are most active around −5°C. Ice crystal concentrations have been reported to exceed the IN concentrations by factors of 10^2 to 10^4 near this temperature. One approach to deducing what ice multiplication might produce in the updrafts examined in northeast Colorado is to assume that the process is rapid enough to produce a 10^2–10^4 enhancement of the IN concentration at −5°C, and use the model to predict the concentration of detectable crystals in the same way it has already been used. Fig. 8 presents such an analysis, and it is clear from inspection of Figs. 3 and 5 that enhancements of 10^3 or more did not occur at −5°C in the updrafts studied.

The Hallett-Mossop ice multiplication effect has been used recently to explain some ice crystal concentration measurements in Florida cumulus (Hallett et al., 1978). Since it appears to require rather specific conditions for its operation, in terms of cloud droplet spectra and ice content, and the rate of multiplication is known approximately, it is worth estimating its potential effects in the types of clouds studied herein.

According to the experiments of Hallett and Mossop, one secondary ice crystal is produced for approximately every 250 droplets ≥24 μm diameter that are deposited on a riming particle at −5°C (Mossop, 1976). An estimated, rounded average over the temperature range −3 to −8°C, from Fig. 5 of Mossop (1976), is one crystal per 500 collisions. If there are N_R riming collectors with a diameter \( \bar{D}_a \), defined as that of the particle with mean cross-sectional area, the fraction of the volume swept out by these collectors as they fall through the depth \( h \) of the −3 to −8°C layer is \( N_R \pi \bar{D}_a^2 h/4 \). The number of collisions with droplets ≥24 μm diameter would then be \( N_{24} \pi \bar{D}_a^2 h/4 \), where \( N_{24} \) is the concentration of droplets ≥24 μm, and \( \epsilon \) is the collection efficiency of the rimed particles for such droplets. The number of secondary crystals (\( N_s \)) produced is

\[ N_s = (\epsilon/500)N_{24}N_R \pi \bar{D}_a^2 h/4. \]

Another requirement for the Hallett-Mossop effect, recently discovered (Mossop, 1978), is the presence of appreciable concentrations of droplets with diameters <10 μm. In the continental clouds of northeast Colorado, the concentration of such drops in the −3 to −8°C layer is always well in excess of 10 cm^-3. This requirement is, therefore, always met.

Use of this formula involves assuming that \( N_R \pi \bar{D}_a^2 h/4 \) is fairly constant over the −3 to −8°C layer. Depletion of \( N_{24} \) in the layer is neglected. \( N_s \) is thus the number of extra ice crystals introduced into the updraft at −8°C. Any time-dependent calculation would have to include growth of the collectors themselves and their manner of entry into the layer. The assumption about collector particles is an oversimplification, but has the virtue of being (very roughly) checkable by radar observations coordinated with the aircraft measurements.

Application of this formula requires values for \( N_{24}, N_R \) and \( \bar{D}_a \), for the −3 to −8°C layer, which is about 800 m thick. Data on \( N_{24} \) from the PMS FSSP probe from all flights have been examined, and the portions considered to be reliable\(^3\) are shown in

\(^3\) Detailed investigations of the FSSP performance have shown a number of general shortcomings of the measurements, and during 1976 the FSSP measurements from the sailplane were also
Fig. 9 plotted against temperature. It is apparent that at times (specifically 22 and 25 July) there are values of $N_{z4}$ of several per cubic centimeter in the right temperature range, but usually $N_{z4} < 1 \text{ cm}^{-3}$. Fig. 10 shows data on $N_{z4}$ in a form better suited to prediction and understanding—as a function of liquid water content. In unmixed regions $N_{z4}$ does not approach $1 \text{ cm}^{-3}$ until the liquid water content reaches $1.6 \text{ g m}^{-3}$. At the $-5^\circ\text{C}$ level, an adiabatic liquid water content of at least $1.6 \text{ g m}^{-3}$ requires a cloud-base temperature not less than $+3^\circ\text{C}$. July 22 and 25 are the two days of the sample with cloud base temperature above $+3^\circ\text{C}$, and in fact on both of these days the temperature was above $+6^\circ\text{C}$ (see HJD).

Of the days investigated the 25 July case should be by far the most favorable for secondary ice production via the Hallett-Mossop effect because $N_{z4}$ was highest on this day in the $-3$ to $-8^\circ\text{C}$ region, and, additionally, there was an exceptionally high concentration of precipitation size particles in the unmixed and slightly mixed updraft. Using the data from this day as an example of what we believe is nearly an upper limit for well-developed cumulus congestus clouds in northeast Colorado, we find $N_R = 0.3 \text{ fl}^{-1}$ and $D_a = 0.6 \text{ mm}$ in reflectivities of 40 to 45 dBZ. Using these values with $\epsilon = 1$ and $N_{z4} = 4 \text{ cm}^{-3}$, again upper limits especially when we consider spreading of the droplet spectrum.
we find that the concentration of secondary ice particles at $-8^\circ$C is $N_s = 0.5 \ell^{-1}$.

This calculated enhancement is very much an upper limit for the kinds of situations studied in northeast Colorado, and a more usual figure is likely to be one or two or more factors of 10 lower. In spite of the roughness of the approximations, we can conclude with considerable confidence that only in unusual circumstances—very warm cloud bases coupled with high graupel concentrations in the updraft—is the Hallett-Mossop effect, as detailed in the reported experimental results, likely to increase natural ice formation in northeast Colorado so that it is clearly inconsistent with measured ice nucleus concentrations.

The present results reinforce those of Gagnon (1975) and of Mossop (1978) that continental clouds are not good candidates for the Hallett-Mossop effect being important. However, in some respects they are more solidly based on observation, at least for the unmixed updraft regions. It is noteworthy that conditions in mixed regions are more favorable for the Hallett-Mossop effect in all aspects: more large drops, more graupel and slower updrafts than in unmixed regions, so that there is also more time for any secondary ice crystals to grow and become important for precipitation. The high concentration of ice crystals in some regions in the present observations is more likely caused by mixing from above (e.g., Paluch, 1979).

A comparison of the measured ice crystal concentrations with those expected from the representative ice nucleus spectra suggested that primary nucleation is the dominant mechanism of ice formation at temperatures warmer than approximately $-15^\circ$C in northeast Colorado cumulus congestus clouds. Insufficient data were available to determine the dominant mechanisms active at lower temperatures. Computations of ice crystal growth rates within the unmixed cores of these clouds indicate that ice crystals nucleated within these regions do not achieve a diameter of 200 $\mu$m until the temperature is $\approx -25^\circ$C.

Ice particles with maximum diameter in excess of 1 mm were frequently found in low concentrations within unmixed cores. These particles were growing in a region that contained nearly adiabatic liquid water contents. They presumably entered into these regions through sedimentation or recirculation after having been nucleated and grown in adjacent regions. This recirculation mechanism has been proposed by Browning (1963) and others for the generation of hail in supercellular storms.

Based on the measurements obtained in unmixed regions, the following mechanisms are envisaged for the origin of ice in undiluted updraft cores in northeast Colorado:

1) Ice crystals form on ice nuclei which are lifted from near the surface (Fig. 11a). This mechanism (primary ice crystal nucleation) probably accounts for most ice particles $<200 \mu$m.

2) Large (~3–5 mm) ice particles can be intro-
duced into strong updraft cores (Fig. 11b) from one updraft to another when a small relative difference exists between particle terminal velocity and the updraft velocity, or through turbulence. This mechanism can result in large particles ascending within the updraft.

3) Large (in excess of 1 mm) ice particles can sediment from higher levels within the same cloud into unmixed regions containing weak updrafts (Fig. 11c). These particles may have been originally introduced into the updraft by mechanism 2) above or grew entirely at the upper levels of the cloud. Ice multiplication does not appear to be a significant mechanism within these regions.

Within mixed updraft regions primary ice nucleation appears to be the dominant mechanism of ice formation at temperatures warmer than $-15^\circ$C. In these mixed regions, abundant concentrations of ice crystals and/or ice nuclei can be mixed into the updraft region, as suggested from the high concentrations of ice crystals noted near the edges of the clouds investigated on 6 and 22 July. Under certain conditions—warm cloud bases, pulsating or weakly developed updrafts or dissipating cloud regions—droplets within mixed regions may grow to be sufficiently large to enable secondary ice crystal production to occur, but in northeast Colorado is not likely to have a significant effect on ice particle production.

Ice crystals which are nucleated near the edge of the updraft within mixed regions or ice particles mixed into the updraft from adjacent areas are capable of achieving considerably larger sizes than those in unmixed regions because of the longer residency time within the updraft. This is consistent with the frequent observation of ice particles $>1$ mm within mixed updraft regions previously reported by Dye et al. (1976) and further supported herein. The higher terminal velocities of these particles favor them for subsequent reintroduction into the main updraft core.

By using the unmixed regions of the clouds investigated in northeast Colorado as a controlled cloud chamber, this study has shown that the measured ice nuclei and measured ice crystal concentrations are consistent. This is the first in-cloud evidence suggesting that the techniques employed in the processing of ice nucleus data can be related to ice crystal concentration. However, since the measured primary ice crystal concentrations in the regions investigated were near or below the limit of detection, it would be desirable to test this relationship with additional data in northeast Colorado. This will provide information on the relevance of secondary ice crystal production in clouds of differing geographical location and intensity.

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$^7$ This may occur, for example, when a small cloud exists upwind of and along the inflow direction of a larger cloud. Particles precipitating from the smaller cell could potentially fall into the updraft region of the larger cell at the lower levels where the updraft velocities are weak. In this case, particles $>1$ mm have a horizontal velocity relative to the larger cell and could be introduced into the unmixed region.

APPENDIX

Calculation of Enhanced Ice Crystal Growth at Pressures Lower than 1000 mb

Measurements of the growth rates of ice crystals smaller than 200 μm at an atmospheric pressure of ~1000 mb were obtained from laboratory experiments by Ryan et al. (1976). At atmospheric pressures below 1000 mb, the growth rates of small ice crystals increased primarily due to the higher diffusivity of water vapor in air. Assuming that there is no enhancement in the growth rate of ice crystals smaller than 200 μm due to ventilation, the ice crystal mass growth rate \( \frac{dm}{dt} \) can be written as (Fletcher, 1966)

\[
\frac{dm}{dt} \bigg|_p = 4\pi CG_p^0 S_1,
\]

where \( p \) is the atmospheric pressure, \( C \) the ice crystal shape factor, \( G_p^0 \) a variable which is dependent on the temperature and pressure and \( S_1 \) the ice supersaturation. The growth rate of an ice crystal at a pressure \( p \), expressed in terms of the growth rate at a pressure of 1000 mb, based on Eq. (1), is

\[
\frac{dm}{dt} = \frac{G_p}{G_{1000}} \frac{dm_{1000}}{dt}.
\]

The data of Ryan et al. describes the changes in linear dimensions on the \( a \) axis (\( \Delta a \)) and the \( c \) axis (\( \Delta c \)) as a function of time (\( \Delta t \)) for 1000 mb.

At atmospheric pressures below 1000 mb, it is assumed that a factor \( (X) \) exists which uniformly increases the linear growth rates on both the \( a \) and \( c \) axes. Over a time period \( \Delta t \), the dimensions of the ice particle increase by \( X\Delta a \) along the \( a \) axis and by \( X\Delta c \) along the \( c \) axis. For a hexagonal ice crystal with dimensions \( a \) and \( c \), the change in mass (\( \Delta m \)) over the time \( \Delta t \) is

\[
\Delta m = \frac{3\sqrt{3}}{8} \left[ (a + X\Delta a)^3(c + X\Delta c) - a^2c \right] \rho \Delta t,
\]

where the term in brackets represents the change in ice crystal volume and \( \rho \) is the density of the added mass. By expanding the right-hand side and by eliminating the \( X^3 \) term which is negligible, \( X \) is expressed as a quadratic equation in terms of \( \Delta m_p \) as

\[
X^2(c\Delta a^2 + 2a\Delta a\Delta c) + X(a^2\Delta c + 2ac\Delta a) - \frac{8\Delta m_p}{3\sqrt{3}\rho \Delta t} = 0.
\]

Through use of the approximation [see Eq. (2)]

\[
\frac{\Delta m_p}{\Delta t} \approx \frac{dm_p}{dt} \approx \left( \frac{G'_p}{G'_{1000}} \right) \left( \frac{\Delta m_{1000}}{\Delta t} \right),
\]

\( X \) can be solved in terms of the mass growth rate at 1000 mb. Typical values of \( X \) ranged from 1.2 to 1.5.

REFERENCES


