An Attempt to Normalize the Hailstorm Variability for the Evaluation of Cloud Seeding

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ABSTRACT

Procedures have been developed from relationships between parameters of hailstone size distributions and storm thermodynamics to normalize the effects of storm thermodynamics of integral hail parameters observed at the ground. Hail parameters considered in this study are the number concentration, water content and kinetic energy. Results from statistical tests on these integral hail parameters obtained from various not-seeded storms showed that significant differences exist among these storms. The differences appear to be due to variations in storm intensity. However, once the normalization of the storm thermodynamic variability is applied to the data, no significant differences are found. The technique seems to reduce, if not eliminate, the variations due to the effects of storm thermodynamics on the integral hail parameters.

Statistical analyses of the normalized surface integral hail parameters, obtained from two operationally seeded storms, are carried out to test the procedure for the evaluation of silver iodide seeding effects. A storm-by-storm approach is used to illustrate anthropogenic variability such as differences in the execution of the seeding operation. Integral hail parameters obtained from a storm that is not seeded at the appropriate location as prescribed by the seeding hypothesis do not differ from those of the control storms. For the other storms seeded at the right location, a nominally significant difference is found in the hail total number concentration but not in the integral hail water content and kinetic energy. However, no conclusion as to the cause of the isolated significant difference can be drawn because the storms analyzed did not constitute a randomized experiment.

1. Introduction

In the evaluation of cloud seeding for hail suppression, one compares a control sample with a sample that has been treated. Ideally, the treated sample and the control sample should be derived through a randomized experiment. The need for randomization in seeding experiments is related to the inability to predict the natural storm behavior, which is primarily a result of the lack of understanding of cloud processes and, to some extent, of the variable environmental conditions influencing the storm. Various randomization designs have been presented and adapted (Neyman and Scott, 1961; Flueck and Mielke, 1977). However, in the past, most results from randomized and nonrandomized seeding experiments have been inconclusive because the natural variability in the response variables is simply too large to allow detection of a moderate cloud seeding effect (Iribarne and Grandoso, 1965; Schleusener, 1962; Goyer et al., 1966; Knight et al., 1979).

It might be anticipated that the same complexity that leads to the situation just described would also be mirrored in a nonuniform response of clouds to seeding, with clouds having different properties and reacting in one way or another. Recognition of this possibility leads one to examine subsets of the data in which particular effects of seeding may be more easily isolated and detected. Physical insight gained during the conduct of a seeding experiment is often extremely valuable in deciding how to stratify the dataset, so that post hoc stratification is rather common (Woodley et al., 1977; Dennis and Koscielski, 1969; Foote and Mohr, 1979). Although stratification may emphasize the effects of seeding, the sample size is reduced. An adequate size dataset is required to allow the evaluation of the stratified data leading to conclusive results. This means that a very long experiment period is needed when stratification is used.

An alternative for isolation and detection of seeding effects is to normalize the variabilities or at least the dominant variability. An advantage of normalization over stratification is that no reduction of the dataset will result. Another option is to parameterize by fitting a statistical model as suggested by Flueck and Mielke (1977). In this procedure, auxiliary or predictor variables are measured for prescreening the treatment opportunity or post-screening to damp the observed variability and better assess the treatment effect. Two major types of variability in surface hail parameters, i.e., natural and anthropogenic (man-made) variabilities, should be considered.

Meteorological factors that affect the hailstone formation processes are the climatic, synoptic and mesoscale features, and storm thermodynamics, dynamics and microphysics (Changnon, 1977; Chisholm, 1973; English, 1973; Nelson, 1983). As stated in the exhaustive review by Browning (1977), a hailstorm is a finely
balanced mechanism, controlled by the interplay of the dynamics and microphysics and poised between two extremes. At one end it may be efficient in converting cloud water to rain. At the other extreme the storm may be inefficient so that depletion in the main updraft is small. Consequently, provided some hail embryos are able to enter the updraft, they will be able to grow into large hail. Natural variability results from differences in these factors. Depending upon the time and space scales of the variables used in the evaluation, some of the differences may not occur or be important. For example, climatic differences are important for yearly crop damage data, but not so much for daily hailfall data.

Anthropogenic variability is the variations that can be controlled, such as differences in the execution of the cloud seeding operation from what was intended and in the data collection procedure. Nearly all of the hail suppression projects in the world have involved the introduction or intended introduction of ice nucleants into the regions of hail or hail embryo formation. The choice of seeding method (ice nucleant delivery system) for hail suppression is influenced by the conceptual model of the hailstorm and of the hail suppression hypotheses, as well as by the techniques and equipment available to conduct and to monitor the seeding operation. Dennis (1977) noted three general methods for hail suppression: (a) broadcast seeding, (b) seeding in the updrafts below individual storm cells, and (c) direct injection of seeding material in the hail formation region to glaciate or compete for the supercooled water. The success of each method is still to be proven.

An area that has been frequently overlooked in the evaluation of seeding programs is the assessment of how successfully the intended seeding operations were actually carried out, or whether the amount of seeding material delivered was adequate. Foose et al. (1979) studied the seeding coverage in the 1972–74 National Hail Research Experiment and found that the day-to-day variations were as great as from 0 to 100%, with a mean of about 50% and a standard deviation of about 30%. Since the seeding operations are more thorough on some storms than on others, one might reasonably expect that the seeding effects, if they exist, would vary considerably from storm to storm.

Another source of the anthropogenic variability is the representativeness of the response data used in the evaluation. Because of limitations in resources and technology, a limited set of data is commonly used. The outcome of any evaluation analysis on any set of data depends upon the representativeness of the data, which in turn is determined by the way that the measurements are obtained (i.e., objectiveness, sampling locations, sampling frequency, response time, etc.). In general, the larger the sample size, the greater the probability that a representative sample is acquired.

To minimize the anthropogenic variability in response variables used in the assessment of hail suppression, one may consider comparing storms. Measurements obtained from seeded and control storms may also be normalized for natural variability, so that particular effects of seeding may be more easily isolated and detected, providing that seeding does not greatly affect the storm macrophysics.

Normalization schemes for the surface hail parameters developed through the study of hailstone size distributions in Alberta are presented in section 2. Comparisons among control (not-seeded) storms before and after normalization will be shown in section 3. The sample representativeness will also be discussed. The evaluation of two seeded storms utilizing the normalization scheme will be given in section 4. A summary will appear in section 5.

2. Normalization scheme for storm thermodynamic variability

Cheng and English (1983) suggested a power relationship of the form \( N_0 = A \Lambda^b \), where \( A \) and \( b \) are the coefficient and exponent parameters, respectively, between the intercept parameter \( N_0 \) and the slope parameter \( \Lambda \) of the exponential approximation for hailstone size distributions, i.e.,

\[
N(D) = N_0 e^{-\Lambda D}, \quad 5 \text{ mm} \leq D \leq D_{\text{max}},
\]

where \( N(D) \) is the number of hailstones per unit volume per unit size interval and \( D \) is the hailstone diameter. Thus \( N(D) \) gives the concentration of hailstones in the size interval \( D \) to \( D + dD \) observed at a particular instant and location.

Cheng et al. (1985) confirmed the existence of the power relationship between the intercept and slope parameters of the exponential size distribution of hail and showed that such a relationship also applies to individual storms. Furthermore, they found that the exponent parameter, \( b \) (having a value of about 4), of the power relationship between \( N_0 \) and \( \Lambda \) can be considered as a constant for all storms, whereas the coefficient parameter, \( A \), relates to the storm thermodynamics. They demonstrated this by correlating \( A \) with the cloud base temperature (\( T_{\text{cb}} \)) and the maximum water mass flux per unit area (\( M_{\text{max}} \)) determined from the Loaded Moist Adiabatic cloud model (Chisholm, 1973). Note that \( M_{\text{max}} \) depends upon the cloud base temperature as well as upon the instability of the atmosphere. Cheng et al.'s findings resemble the analytical result of Passarelli (1978) for snow spectra. Passarelli's study suggested that an equilibrium relation exists between \( N_0 \) and \( \Lambda \) of the snow exponential size distribution, of the form \( N_0 = C \Lambda^b \), where \( C \) depends upon cloud thermodynamics and other crystal growth parameters. The equilibrium relation does not depend upon the initial conditions, such as ice nuclei or ice crystal concentrations. Although the dominant growth processes for hail and snow are different, the stochastic
nature of these processes is speculated to be responsible for the resemblance of the relationship between \(N_o\) and \(\Lambda\) of hail and snow size distributions. The theoretical justification of these relationships for hail spectra is within the scope of future work and is not included in the present study.

Using the relationships between \(N_o\) and \(\Lambda\) of the hailstone size distribution and between the storm thermodynamics and the coefficient parameter of the \(N_o\) and \(\Lambda\) relation, a method to normalize the storm thermodynamic variability in surface integral hail parameters can be derived. The normalized forms of these quantities of interest are discussed next.

\(\text{a. Total hail number concentration}\)

In terms of the distribution parameters, \(N_o\) and \(\Lambda\), and of the maximum, \(D_{\text{max}}\), and the minimum diameter, \(D_{\text{min}}\), of the hailstone size spectra, the definition of the total hail number concentration \(N_T\) is given by

\[
N_T = \int_{D_{\text{min}}}^{D_{\text{max}}} N(D) dD = \frac{N_o}{\Lambda} [\gamma(1, D_{\text{max}}) - \gamma(1, D_{\text{min}})].
\]  

(2)

In the above, \(\gamma(a, x)\) is the incomplete gamma function

\[
\gamma(a, x) = \int_0^x t^{a-1} e^{-t} dt,
\]  

(3)

which approaches the limiting value \(\gamma(a, \infty) = \frac{1}{a-1}\) when \(x \to \infty\). Substituting the relationship between \(N_o\) and \(\Lambda\) in (2), with \(b\) assumed to be 4, and dividing the resultant equation by the coefficient parameter, \(A\), of the \(N_o\) and \(\Lambda\) relation, one obtains

\[
\frac{N_T}{A} = \Lambda^3 [\gamma(1, \Lambda D_{\text{max}}) - \gamma(1, \Lambda D_{\text{min}})].
\]

(4)

Since \(A\) depends upon the storm thermodynamics, (4) gives a hail total number concentration, which is normalized with the storm thermodynamic characteristics. Because \(D_{\text{min}}\) is a constant by the hailstone definition, (4) shows that the normalized hail total number concentration is a function of \(\Lambda\) and \(D_{\text{max}}\).

The relationship among \(N_T/A, \Lambda\) and \(D_{\text{max}}(\geq 1 \text{ cm})\) is displayed graphically in Fig. 1. The shaded area in the figure gives the region for two standard deviations about the mean \(\Lambda D_{\text{min}}\) (7.69) obtained from 191 hail samples collected from 15 Alberta storms in 1980, 1982, and 1983. The frequency distribution of the product \(\Lambda D_{\text{max}}\), found in Alberta, is similar to that obtained for northern Colorado storms (Ulbrich and Atlas, 1982). Thus, one can consider that most of the hailfalls would have values of \(\Lambda\) and \(D_{\text{max}}\) bounded by the shaded area. Note that within this area the value of the normalized hail total number concentration depends essentially upon \(\Lambda\), and not so much upon the hailstone size. The range of the normalized hail total number concentration for the commonly observed hailfalls spans an order of magnitude.

\(\text{b. Total hail water content}\)

Another commonly used hail integral parameter is the total hail water content, \(W_H\), which is given by

\[
W_H = \frac{\pi \rho}{6} \int_{D_{\text{min}}}^{D_{\text{max}}} D^3 N(D) dD
\]  

\[
= \frac{\pi \rho N_o}{6\Lambda^3} [\gamma(4, \Lambda D_{\text{max}}) - \gamma(4, \Lambda D_{\text{min}})].
\]

(5)

The hailstones are assumed to be spherical with a mass density \(\rho\) of 0.9 g cm\(^{-3}\) (Macklin et al., 1960). Similar to the formulation of the normalized total hail number concentration in section 2a, a normalized total hail water content can be defined as

\[
\frac{6W_H}{\pi \rho A} = [\gamma(4, \Lambda D_{\text{max}}) - \gamma(4, \Lambda D_{\text{min}})].
\]

(6)

Note that this normalized total hail water content is also a function of \(\Lambda\) and \(D_{\text{max}}\) only.

The relationship among \(6W_H/(\pi \rho A), \Lambda\) and \(D_{\text{max}}(\geq 1 \text{ cm})\) is shown graphically in Fig. 2. As in Fig. 1, the shaded area represents the ranges of \(\Lambda\) and \(D_{\text{max}}\) for most of the hailfalls. Unlike the normalized total hail number concentration, the normalized total hail water content has no constraints on the value of \(\Lambda\). The normalized total hail water content also has no constraints on the value of \(\Lambda\).
where $g$ is the acceleration due to gravity (980 cm s$^{-2}$), $ho_a$ is the mass density of the air through which the hailstones are falling (1.05 $\times$ 10$^{-3}$ g cm$^{-3}$), and $C_D$ is the drag coefficient (0.55). Waldvogel et al. (1978) suggested a value of 1.396 cm$^{1/2}$ s$^{-1}$ for the coefficient $V_0$, which differs from the present calculation by only 2%.

The total hail kinetic energy content, normalized with respect to the storm thermodynamics, can be defined as

$$\frac{12E}{\pi\rho V_0^2 A} = \Lambda^{-1}[\gamma(5, \Lambda D_{\text{max}}) - \gamma(5, \Lambda D_{\text{min}})].$$

Again, this quantity is a function of the slope parameter $\Lambda$ of the hailstone size distribution and of the maximum hail diameter $D_{\text{max}}$ only. Figure 3 shows graphically the relation between these parameters, and once again the shaded area gives the ranges of $\Lambda$ and $D_{\text{max}}$ for most of the hailfalls. The dependence of the normalized total hail kinetic energy content upon the maximum hailstone diameter is even greater than that of the normalized total hail water content. This is expected because the hailstone impact energy is a function of the fourth power of the hailstone radius, whereas the hailstone water mass is a function of the third power. The range of the normalized total hail kinetic energy content commonly observed spans less than an order of magnitude also.

c. Total hail kinetic energy content

An integral hail parameter that has been frequently hypothesized as being related to the physical damage caused by hailfall is the total hail kinetic energy content $E$ given by

$$E = \frac{1}{2} \int_{D_{\text{min}}}^{D_{\text{max}}} \frac{\pi\rho}{6} D^3 v(D)^2 N(D) dD$$

$$= \frac{\pi\rho V_0^2}{12A^3} \Lambda N_2[\gamma(5, \Lambda D_{\text{max}}) - \gamma(5, \Lambda D_{\text{min}})].$$

(7)

The hailstones are assumed falling in still air, and therefore, the presence of vertical or horizontal winds is ignored. Also, it has been assumed in (7) that the fall speeds of the hailstones can be approximated by

$$v(D) = \left(\frac{4g\rho}{3\rho_a C_D}\right)^{1/2} \sqrt{D} = V_0 \sqrt{D}$$

(8)

FIG. 3. As in Fig. 1, but for the normalized hail integral kinetic energy $12E/\pi\rho V_0^2 A$ (mm).

content significantly depends upon $D_{\text{max}}$, especially at small $D_{\text{max}}$. When $D_{\text{max}}$ becomes large ($\gg 25$ mm), the dependence of $6W_h/\pi\rho A$ on $D_{\text{max}}$ is less marked. The range of the normalized total hail water contents commonly observed is much smaller than that of the normalized total hail number concentration and spans less than an order of magnitude.
3. Comparisons between control storms

Time-resolved hailstone data used in this study were collected as part of the Alberta Hail Project, during the period 1982–83. The sampling and hail size measuring methods have been described in Cheng and English (1983) and Cheng et al. (1985). Hailstone samples collected from six control (not-seeded) storms were analyzed. Table 1 gives the number of hailstone samples and the thermodynamic properties of the six storms calculated from the Loaded Moist Adiabatic cloud model (Chisholm, 1973). The rank sum of the storm properties are also shown. It is rather obvious that the 30 June 1982 storm differs significantly from the other storms.

Because of the importance of hailstone size distributions in this study, the following criteria have been applied to the data to ensure that the derived size distribution parameters are representative of the hailfall at the time of collection. Consecutive samples of less than 100 stones or of maximum size less than 1 cm (five size categories), collected at the same location, were combined to form samples of adequate size and number to satisfy the arbitrary criteria of \( D_{\text{max}} \geq 1 \) cm and of the total number \( \geq 100 \) stones.

a. Sample representativeness

The outcome of an evaluation analysis of any experiment depends heavily upon the representativeness of the dataset. The representativeness is determined by the experiment hypothesis, objectivity, sampling location, response time and other related factors. Normally, the larger the sample size the better the dataset is for statistical tests. Thus, ideally one would like to form a database that would comprise all of the hailfalls from studied storms. However, such an attempt is fairly unrealistic. To make the problem more complex, hailfalls from a storm are not random, but structured (Browning and Foote, 1976; Knight and English, 1980). The dataset for the evaluation analysis should, therefore, at least be obtained at several locations over a large portion of the storm.

A complete analysis of the sample representativeness of the data used is outside of the scope of this study. Nevertheless, the sampling locations relative to the storm have been studied to provide a qualitative picture of the sample representativeness of the data. Although the hail samples were obtained at a few fixed locations, the movement of the storm gave every hailstone sample collected at a particular point on the ground a unique location with respect to the storm. Relative distances between the sampling location and the 45 dBZ reflectivity weighted centroid of the storm at about 1° at the time of sampling were calculated. Since radar scans are discrete, the 45 dBZ centroid location at the time of resolved hailstone sample collection sometimes has to be interpolated. The choice of the 45 dBZ value for computing the reference point is based on the fact that this value is large enough to give a high probability of hail occurrence (Barge, 1974) but small enough to produce a smooth centroid track. However, a reflectivity factor of 45 dBZ does not guarantee hailfall at the ground. The results should only be taken as a general composite description of the sampling locations relative to the storm.

Figure 4 gives the sampling locations relative to the 45 dBZ centroid and the low-level PPIs (plan-position indicator) about the time that most of the samples were collected for the six not-seeded storms. It should be pointed out that some hailstone samples from a given storm were collected at a time quite different from that of the PPI shown, and the storm structure may have changed somewhat. Areas with reflectivity factor \( \geq 45 \) dBZ are shaded. The X’s represent the locations of the 45 dBZ reflectivity weighted centroids and the origins of the grid showing sampling locations. The storm motion direction for each storm is also shown. Note that, for storms with a large number of hail samples, the sampling locations extend over a large portion of the area shown by the 45 dBZ contour. On 21 July 1983 only nine hailstone samples were collected, and the sampling locations extended into the 45 and 55 dBZ reflectivity factor area. Thus, the hail samples from this storm were likely obtained from the hail core as well as from the less intense areas of the hailfall. In the 3 August 1983 storm only eight samples were collected,

<table>
<thead>
<tr>
<th>Storm no.</th>
<th>Storm date</th>
<th>Number of hailstone samples</th>
<th>( T_{\text{CB}} ) (°C)</th>
<th>( U_{\text{max}} ) (m s(^{-1}))</th>
<th>( \text{LWC}_{\text{max}} ) (g m(^{-3}))</th>
<th>( \text{PE}_{\text{max}} ) (J g(^{-1}))</th>
<th>( M_{\text{max}} ) (g s(^{-1}) m(^{-2}))</th>
<th>Rank sum</th>
</tr>
</thead>
<tbody>
<tr>
<td>N1</td>
<td>30/6/1982</td>
<td>18</td>
<td>6.0 (1)</td>
<td>21.86 (1)</td>
<td>3.69 (1)</td>
<td>0.235 (1)</td>
<td>80.48 (1)</td>
<td>5.0</td>
</tr>
<tr>
<td>N2</td>
<td>21/7/1982</td>
<td>32</td>
<td>9.4 (4)</td>
<td>33.45 (6)</td>
<td>4.55 (4.5)</td>
<td>0.555 (6)</td>
<td>128.89 (4)</td>
<td>24.5</td>
</tr>
<tr>
<td>N3</td>
<td>11/8/1982</td>
<td>21</td>
<td>9.9 (5)</td>
<td>32.03 (5)</td>
<td>4.55 (4.5)</td>
<td>0.509 (5)</td>
<td>139.44 (6)</td>
<td>25.5</td>
</tr>
<tr>
<td>N4</td>
<td>12/7/1983</td>
<td>36</td>
<td>10.3 (6)</td>
<td>30.84 (3)</td>
<td>4.63 (6)</td>
<td>0.448 (3)</td>
<td>138.79 (5)</td>
<td>23.5</td>
</tr>
<tr>
<td>N5</td>
<td>21/7/1983</td>
<td>9</td>
<td>8.1 (3)</td>
<td>31.84 (4)</td>
<td>4.23 (3)</td>
<td>0.475 (4)</td>
<td>128.34 (3)</td>
<td>17.0</td>
</tr>
<tr>
<td>N6</td>
<td>3/8/1983</td>
<td>8</td>
<td>7.9 (2)</td>
<td>29.69 (2)</td>
<td>4.10 (2)</td>
<td>0.435 (2)</td>
<td>113.55 (2)</td>
<td>10.0</td>
</tr>
</tbody>
</table>

Table 1. Storm thermodynamic properties, not-seeded storms. \( T_{\text{CB}} \) = cloud base temperature, \( U \) = vertical velocity, LWC = liquid water content, PE = parcel energy, \( M \) = water mass flux. The rankings for the individual parameters are given in parentheses.
FIG. 4. Composite sampling locations relative to the 45 dBZ reflectively weighted storm centroid and the low-level PPI at about the time when most of the hailstones were collected for the not-seeded storms. The x-axis and the y-axis in the composite sample location diagram are, respectively, the west-east and north-south distances from the centroid in kilometers. In the PPIs, areas over 45 dBZ area shaded, and the X’s indicate the 45 dBZ centroid locations and the origin of the grid for sampling locations. The radar range rings are 20 km apart. Reflectivity contours are at 10 dBZ intervals starting at 25 dBZ. The storm motion direction is indicated by an arrow.
apparently from a low reflectivity area of the storm. Thus, the samples may not consist of hailstones from the most intense part of the storm. This result suggests that the hailstone samples collected from the 3 August 1983 storm may not be as representative as those from other storms.
b. Comparison before normalization

The distribution-free Kruskal–Wallis test (Miller, 1966) shows that the total hail number concentration, hail water content and hail kinetic energy of the six control storms are significantly different, with p-values of 0.0001, 0.0001 and 0.0003, respectively. These integral hailstone parameters are calculated from the observed size and concentration data of each individual hail sample. Thus, the number of \( N_T \), \( W_H \) and \( E \) values are equal to the number of hailstone samples for a particular storm. To determine which particular storms differ from others, a multiple comparison procedure, suggested by Dunn (1964), was used. The procedure shows that two clouds are different at an error rate \( \alpha \) if

\[
| R_w - R_e | \geq Z_{\alpha/(k(k-1))} \sqrt{\frac{N(N+1)}{12}} \left[ \frac{1}{n_w} + \frac{1}{n_e} \right]^{1/2}
\]

(A)

\[
\frac{1}{n_w} + \frac{1}{n_e}
\]

(B)

where \( R_i \) is the average Kruskal–Wallis rank obtained for cloud \( i \); \( N = \sum_{i=1}^{k} n_i \) observations, with \( n_i \) observations from the \( i \)th cloud; \( i = 1, \ldots, k \); and \( Z_{\alpha/(k(k-1))} \) is the upper \( \alpha/(k(k-1)) \) percentile of the standard normal deviate. Table 2 summarizes the calculations for the procedure performed at an error rate of 0.05 for the hail number concentration, hail water content and hail kinetic energy.

Table 2 shows that the inequality in (10) is satisfied in the comparisons of total hail number concentration, hail water content and hail kinetic energy between storms N1 and N2, N1 and N3, N1 and N4. Thus at the error rate of 0.05, the 30 June 1982 storm is significantly different from the 21 July and 11 August 1982 and 12 July 1983 storms. The 30 June 1982 storm is the least intense among the studied control storms, as shown by the rank sum of the storm thermodynamic properties listed in Table 1. The maximum water mass flux of this storm is only 80 g s\(^{-1}\) m\(^{-2}\), compared to the 100 g s\(^{-1}\) m\(^{-2}\) or greater values for the other storms. Also, the parcel energy of this storm is almost half those of the other storms. Since the hail water content depends upon the amount of water supplied to the storm and hailstones, which require energy from the storm to support their growth, the results of the multiple comparisons are not very surprising.

Large variations in the hail parameters among control storms exist, and it has been demonstrated that they may be due to the natural variability. Thus, unless this variability is reduced or accounted for, the effect of seeding may be masked when seeded and not-seeded storms are compared.

c. Comparison after normalization

Least-squares regression is used to approximate the intercept \( N_0 \) and the slope \( \Lambda \) parameters of the exponential size distribution from the size and concentration data of each individual hailstone sample. The coefficient \( A \) and exponent \( b \) parameters are obtained from the \( N_0 \) and \( \Lambda \) of the hailstone samples collected from the same storms.

Using the value of the coefficient parameter \( A \), so-determined for the corresponding storm, the observed total hail number concentration, hail water content and hail kinetic energy of the six control storms are normalized, as described in section 2, and compared using the Kruskal–Wallis test. No significant difference is found in these hail parameters among the six control storms, with p-values of 0.1210, 0.1700 and 0.7912, respectively. The p-values change considerably from 0.0001 to 0.1210, from 0.0001 to 0.1700, and from 0.0003 to 0.7912 when the observed hail parameters are normalized. The results suggest that the normalization reduces somewhat, if not totally, the natural variability in the total hail number concentration, hail water content and hail kinetic energy. If this is the case, other effects such as cloud seeding effects may then be more easily detected.

4. Test of the procedure for the evaluation of seeding effects

The Alberta Hail Project's principal target area of 48 000 km\(^2\) is centered on the radar site located at the Red Deer Industrial Airport, Penhold (Fig. 5). Operational seeding was carried out only in the southern half of the project area (area A). Operational seeding could also be carried out in area C, at the discretion
of the cloud seeding monitor. Controlled seeding experiments could be conducted in any of the three areas (A, B and C). The basic hypothesis underlying the present Alberta technique for hail suppression is that large, damaging hailstones occur because of a natural deficiency of freezing nuclei in a storm containing a large quantity of water vapor and of liquid water drops. The injection of suitable numbers of artificial nuclei in the new growth region of the storm promotes the formation of a large number of hailstone embryos and hence of less damaging hailstones because of increased competition for the available water in the main storm. Thus, the regions of weak turret growth are seeded from cloud top and the regions of weak updraft at the edges of the visible cloud are seeded from cloud base.

The normalization schemes described in section 2 are applied to test the effects of cloud seeding of hailstorms. The exercise is not a rigorous statistical evaluation of seeding effects because the storms selected do not form a randomized experiment. It is intended only as a test of the normalization procedure. Two seeded storms, with significant numbers of time-resolved hailstone samples, were chosen from the data obtained from the Alberta Hail Project, namely the storms of 24 July and 1 August 1983. Anthropogenic variability produced by differences in the execution of the seeding operation as intended exists between these two storms. Judging from data available, only the 1 August 1983 storm was seeded at the correct location prior to and during the hailstone sampling period, as prescribed by the seeding hypothesis. If the normalization makes storms more homogeneous and thus may render any possible seeding effect detectable, differences should only be found in the 1 August storm, not in the 24 July storm.

The sample criteria are the same as that used for the control storms. These two storms were mainly seeded with AgI below cloud base, presumably along the upwind portion of the subcloud inflow and the weak updraft portion of the storm. End-burning (150 g AgI) flares (burn time of 4 min) and acetone generators were used. The acetone generators burned a 2% solution of AgI in acetone (output of 2 g AgI min\(^{-1}\)). In the 1 August 1983 storm, 10 (20 g AgI) droppable pencil flares were also delivered during a single cloud turret controlled hail seeding experiment in which the test cloud microphysical characteristics were measured by a research aircraft prior to and after treatment.

a. Case of 24 July 1983

Numerous cells developed along the foothills west of Rocky Mountain House in the early afternoon. They moved east over the project area through the afternoon and evening. The most intense activity, with reflectivities in the upper 60s, was in the Airdrie–Beiseker area. Twenty four time-resolved hailstone samples were collected from the storm that moved through the Airdrie–Beiseker area. Because of a lack of knowledge about the time period during which the artificial ice nuclei would affect the hailfall, a period from half an hour prior to the collection of the first hailstone sample to the time of the last sample collection was considered. Twenty-one base flares were used (3150 g AgI) and 0.5 hours of acetone burned (52.4 g AgI) during this period, giving an average seeding rate of 34.97 g min\(^{-1}\) of AgI.

The sampling locations relative to the 45 dBZ centroid of the storm and the low-level PPI obtained at about the time that most of the hailstone samples were collected are shown in Fig. 6. The storm motion direction is also shown. The area of 45 dBZ or higher reflectivity factor is shaded. The origin of the grid box for the sampling locations or the 45 dBZ reflectivity weighted centroid is also shown in the figure as an “X.” Note that although samples were mainly located at the southern portion of the storm with respect to the 45 dBZ centroid, they extended over the area of high and low reflectivity factors of the storm. Thus, the data measured in this storm were not just from the most intense or the weakest hailfall regions.

The seeding operation on 24 July 1983 is shown in Fig. 7, which shows the cloud seeding locations relative to the main storm echo at 1620, 1650 and 1720 MDT. The altitude of the PPIs is approximately 1 km, i.e., the lowest radar scan. The first hailstone sample was collected at 1646 MDT and hail sampling ended at 1748; thus the locations given represent those where seeding may have affected the growth of the hailstones.
collected. Figure 8 gives the hodographs measured at 0630 (dashed line) and 1515 MDT (solid line) northwest of Calgary (YC). The wind measurements at 1515 MDT might have been affected by the storm that was then near the weather station. However, it should be more representative of the storm environment. The height levels are shown in kilometers above mean sea level. The ground level around this area is approximately 1.1 km. Both profiles show a southerly wind component at and below cloud base. Judging from the low-level wind directions and the locations of the storm development derived from the evolution of the radar echo, cloud base seeding should have been conducted, according to the seeding hypothesis, at the south or south-southeast edge of the storm, rather than at the northeast side as indicated in Fig. 7. During this time period, the seeding aircraft may have been directed by air traffic control to the location where seeding was conducted. The seeding aircraft was not allowed to circle at low elevation in close proximity to the busy Calgary International Airport. Moreover, according to the cloud-seeding regulations issued by Environment Canada, no cloud seeding should be conducted over densely populated areas. Thus, a seeding effect is not likely to be detected if the seeding hypothesis is valid and the seeding location prescribed by the seeding hypothesis is correct.

Kruskal–Wallis tests on the normalized total hail number concentration, water content and kinetic energy measured from this storm and from those of the not-seeded storms, discussed in section 3c, gives p-values of 0.3437, 0.8028 and 0.3960, respectively. Thus, none of these hailfall parameters of this seeded storm differ significantly from those of the control storms. Either the presumed reduction is too small to be detected with the number of samples obtained, due to improper seeding (amount of AgI delivered was not sufficient and/or the seeding locations were not correct), or the hypothesis of hailfall reduction by the introduction of artificial ice nuclei is not valid.

b. Case of 1 August 1983

Storms began to develop at about 1400 MDT in the foothills southwest of Red Deer, but dissipated as they drifted east out of the foothills. Later in the afternoon, a north–south line of intense storms developed from Rocky Mountain House to Sundre. The line tracked east across the central portion of the project area, with maximum reflectivities near 70 dBZ. By 2200 MDT, the line was about 100 km east of Red Deer and had weakened to only rainshowers. Nineteen time-resolved hailstone samples were collected from a storm at the south end of this intense line from 1811 to 2021 MDT. Ten cloud-top droppable flares (200 g AgI) were injected into a cloud turret for a controlled hail experiment, during which microphysical measurements were made prior to and after seeding. In addition, 37 cloud-base wing flares (5550 g AgI) and 2.5 hours of acetone solution (259.7 g AgI) were used below the base of this storm from 1742 to 2020 MDT, giving an average seeding rate of 38.0 g min⁻¹ of AgI.

The sampling locations relative to the 45 dBZ reflectivity weighted centroid of the storm and the low-level PPI at about the time of most hailstone sample collections are shown in Fig. 9. The 45 dBZ centroid or the origin of the sampling location grid is shown as an “X.” The storm motion direction is also given. Although the hail measurements did not cover a large area, they extended across the 55-dBZ-and-greater area into the 45 dBZ regions. Thus, the integral hail parameters, derived from the 19 hailstone samples collected from this storm, should represent the hailfall from the hail core as well as from the less intense areas.
to 185°. Thus, cloud base seeding was conducted accurately in the inflow of the storm. Furthermore, the ten cloud top droppable flares, delivered during the control hail experiment, were injected inside the new growth region of the storm. Therefore, the seeding materials were injected at the intended location into the storm.

Kruskal–Wallis tests of the normalized integral hail parameters measured from the 1 August 1983 storm and the control storms discussed earlier yield $p$-values of 0.0096, 0.3504 and 0.0975 for the total hail number concentration, water content and kinetic energy, respectively. These results show that only the integral hail number concentration of this seeded storm differs significantly from that of the control storms. Using either the mean or median values of the normalized total hail number concentration, an increase of about 77% is obtained. However, the 95% confidence limits of increase in terms of the mean value of the not-seeded storms are (0.13, 1.11) from the two-sample Mann–Whitney test. Therefore, with a sample size of 19 for the seeded storm, one cannot expect, with any great degree of certainty, more than a 13% increase in the mean total hail number concentration.

An example of the exponential approximation for a hailstone number size distribution (solid line) derived from the coefficient parameter $A$ of hail spectra collected on 1 August 1983, which had a value similar to that of 21 July 1982 and an assumed $A$ of 0.08 mm$^{-1}$, is given in Fig. 10a. Two hypothetical spectra (dashed

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**Fig. 7.** Cloud-base seeding locations (asterisks) with respect to the storm radar echo at 1620, 1651, and 1712 MDT on 24 July 1983. The square represents the city boundary of Calgary (YC).

**Fig. 8.** Wind hodographs measured at 0630 (dashed line) and 1515 MDT (solid line) in Calgary on 24 July 1983. The location of cloud base (CB) is also shown.
and dotted lines) affected by cloud seeding are also shown. The assumption is made that the seeded hailstone size distributions express the same relationship between size and concentration as found in not-seeded storms (Cheng and English, 1983; Cheng et al., 1985). The hypothetically seeded number distributions represent those that resulted from different increases in hail number concentrations (different rate of cloud seeding). The integral number concentrations calculated from hailstone size distributions these control and the hypothetically seeded storms are 0.3880, 0.7588 and 2.1722 m⁻³, respectively. There are hypothetical increases of 96% and 460% in the total hail number concentrations caused by seeding. The corresponding mass content and kinetic energy distributions for the above size spectra are shown in Figs. 10b and 10c, respectively. Increases in the smaller size range and decreases in the larger size range for both hypothetically seeded water content and kinetic energy spectra are obvious. The integral hail water content for the two hypothetically seeded hail size distributions are 0.4040 and 0.2785 g m⁻³. This suggests a small increase, 4%, and a larger decrease, 28%, when compared to the 0.3870 g m⁻³ of the control spectra. The total kinetic energy values for the three size distributions are 0.0771, 0.0631 (reduction of 18%; less seeding) and 0.0208 J m⁻³ (decrease of 73%; more seeding). Thus, although there are considerable increases in the total number concentration, the integral hail water content and the kinetic energy may not be altered noteworthily. Therefore, it appears that the artificial ice nucleants, injected into the storm of 1 August 1983, were not sufficient to cause a significant reduction in the integral hail water and kinetic energy contents. However, there may still be large random effects in any particular storm such as this one that could negate this theoretical argument.

5. Summary and conclusions

Mathematical schemes have been developed to normalize integral hail measurements observed at the ground for the effects of storm thermodynamic variations. The techniques are derived from the relationship between parameters of exponential hailstone size distributions and the storm thermodynamic characteristics. The hailstone total number concentration, water content and kinetic energy are studied. Results from statistical analyses of observed integrated hail parameters show that significant differences exist among various not-seeded storms. The differences appear to be caused, to some degree, by the variations in the storm intensity. Once the normalization is applied, no significant differences are then found in the integral hail parameters of these storms.

Because hailfalls are not random but structured, a large sample size does not always ensure data representativeness. In this study the sampling locations, with respect to the 45 dBZ reflectivity weighted storm centroid, at the time of sample collection are compared with the radar echoes. Although this procedure does not produce an absolute measure of the sample representativeness, it does provide a qualitative assessment of the data assembled. A technique to objectively test the representativeness of structural data such as hailfall data should be developed.

Statistical analyses were applied to the normalized surface total hail number concentration, water content and kinetic energy obtained from two operationally seeded storms to test the normalization procedure. The storms were seeded with silver iodide particles, mainly from below cloud base. For the storm which was not seeded in the appropriate location, as prescribed by the seeding hypothesis, no significant differences were
found in the integral hail parameters studied when they were compared with those of not-seeded storms. For the other storm, which appears to have been seeded at the right location, the hail total number concentration differs significantly from those of the control storms, whereas the integral hail water content and kinetic energy do not. Using hypothetical hailstone size distributions, it has been demonstrated that a large amount of artificial ice nuclei is required to produce beneficial competition for the water available in the storm and eventually to reduce the integral hail water content and kinetic energy. If the difference in the total number concentration is due to seeding, one can speculate that though the storm studied was seeded at the right location, the amount of seeding material was not sufficient to significantly reduce the hailstone size. If the seeding rate had been increased, a reduction in total hail water content and in kinetic energy might have been observed. However, no conclusion as to the cause of the isolated significant difference can be drawn because the storms analyzed did not constitute a randomized experiment.

The normalization appears to provide a promising means for the evaluation of hail suppression efforts. Because of variations in the cloud seeding coverage, seeding effects may be masked. Assessment of storms having the same seeding coverage/rate would minimize the variability generated by the differences in the seeding operation. Furthermore, such a procedure would create an opportunity for the determination of the amount of seeding material required for effective hail suppression. This, together with the normalization procedure, should be considered in the planning of a randomized experiment for the evaluation of hail suppression.

It should be emphasized, however, that the normalization schemes were derived from empirical relationships. A physical theory for the relationships between hail size distribution parameters and storm thermodynamic characteristics should be established. Until this is done, the technique discussed in this study cannot be considered universal. Data from more storms are required to substantiate these relationships. It would be desirable to examine data from other regions to ensure that similar relationships still hold.

The results presented in this study are based on time-resolved hail measurements, which are difficult to assemble into a representative dataset without major efforts and resources. This constraint limits the usefulness of this technique for monitoring hail suppression operations. Therefore, the extension of the idea presented in this study to other types of hail measurements, such as hailpad and radar, should be vigorously pursued. Such endeavors are actively being pursued.

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