Influence of glaciation on an effective-radius parametrization

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A recent study (Blyth and Latham 1991) of ice-free, non-precipitating sub-adiabatic cumulus clouds, that involved the examination of data from multiple-penetration airborne experiments conducted in Montana and New Mexico, showed that the effective radius, \( r_{\text{eff}} \),—a parameter of central relevance to global climate models—was essentially independent, at any particular level in the clouds and for any spatial scale of measurement, of the liquid-water content, \( L \), and the droplet concentration, \( N \), (i.e. \( r_{\text{eff}} \) was independent of the extent to which the cloud was diluted by entrainment of environmental air). This finding led to the prediction that \( r_{\text{eff}} \) may be simply expressed, at any location in the clouds, by the relationship \( r_{\text{eff}} = r_{\text{ad}} \), where the ‘adiabatic radius’ \( r_{\text{ad}} \) is given by

\[
r_{\text{ad}} = \left( \frac{3}{4\pi \rho_L N_{\text{ad}}} \right)^{1/3},
\]

In this equation, \( \rho_L \) is the density of liquid water and \( N_{\text{ad}} \) and \( L_{\text{ad}} \) are the ‘adiabatic’ values of \( N \) and \( L \) respectively. Thus the predicted ‘non-dimensional effective radius’ is given by

\[
A = \frac{r_{\text{eff}}}{r_{\text{ad}}} = 1.
\]

Analysis of the airborne data for the Montana cumuli (35 cloud penetrations) revealed that \( A = 0.83 \pm 0.07 \), and for the New Mexican cumuli (25 penetrations) \( A = 0.93 \pm 0.05 \) in reasonable agreement with the prediction of constancy of \( A \), and fair agreement (considering the crudity of our model) with the prediction that \( A = 1 \).

It was mentioned in Blyth and Latham (1991) that preliminary analysis indicated that when these clouds started to glaciate (or more precisely, in those regions of the clouds where glaciation had commenced) the climatological parameter \( A \) increased.

This indication has now been subjected to further study, which forms the topic of this note.

The analysis presented herein is confined to data from airborne studies of New Mexican cumuli in the summer of 1987. These are discussed in more detail in Blyth and Latham (1993). The clouds were studied using the National Center for Atmospheric Research King Air cloud physics research airplane. Multiple penetrations were made approximately every 5 minutes, mainly near the tops of the cumulus clouds that formed over the Magdalena Mountains. (From a climatological viewpoint, the important values of \( r_{\text{eff}} \) are those confined to within a few optical depths of the cloud top.) Cloud-droplet parameters were measured with the Forward Scattering Spectrometer Probe (FSSP) at a rate of 10 Hz, corresponding to a spatial scale of about 10 m. The droplet spectra were not corrected for coincidence errors such as discussed by Cooper (1988) and Brenguier (1989). Cooper suggested that distortions of the true spectra are likely if the measured concentration, \( N \), is above 500 cm\(^{-3}\), but that there could be effects if \( N > 100 \) cm\(^{-3}\). Since the parameter discussed in this paper, \( A \), was found to be independent of \( N \), except for the lowest values, it is unlikely that coincidences significantly influenced the results. Furthermore, it was rare for the number concentration to exceed 500 cm\(^{-3}\).

We were careful to eliminate spectra that were contaminated by ice particles (Gardiner and Hallett 1985) by excluding data, in regions of cloud containing ice, when the liquid-water content measured by the FSSP was more than about one-and-a-half times that measured by the Johnson–Williams device.

The clouds were typically 2 to 3 km deep with base and summit temperatures in the ranges from 1 to 11°C and from −4 to −27°C respectively. They were generally strongly subadiabatic, the normalized liquid-water content \( L/L_{\text{ad}} \) (penetration averages) lying generally in the range 0.2 to 0.4, although a few regions of 100–500 m extent were observed where \( L/L_{\text{ad}} \approx 1 \).

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The presence of ice in these clouds—together with the concentrations, sizes and shapes of the ice particles—was established by means of the Particle Measuring Systems 2DC (cloud) and 2DP (precipitation) probes. It should be noted that regions of cloud with no detectable ice may still contain small ice particles that have been missed by the probes.

In many cases no ice particles were detected during penetrations in the early stages of the cloud development. In this situation $r_{\text{eff}}$ was found to be essentially independent of the ratio $L/L_{\text{ad}}$—which parameter is an indication of the extent to which the region of cloud traversed was diluted by entrainment—and the prediction that $A = r_{\text{eff}}/r_{\text{ad}} = 1$ was confirmed to a similar degree to that reported by Blyth and Latham (1991). This finding is illustrated in Fig. 1, which is a histogram of $A$ produced by accumulating data from many penetrations through ice-free regions of several clouds. The average value of $A$ is 0.91 and the standard deviation is 0.33. Deviations in $A$ from the adiabatic value when $L/L_{\text{ad}}$ is low is largely responsible for the large standard deviation.

While Bower and Choularton (1992) also found $r_{\text{eff}}$ to remain essentially unaffected by entrainment in a few cumulus clouds measured in Montana, they suggested that a more appropriate parametrization for the effective radius in continental cumulus clouds is given by simply setting $r_{\text{eff}}$ to a fixed value of between 9 and 10 $\mu$m. The overall average value of $A_{\text{eff}}$ in ice-free regions of New Mexican cumuli is 8.5 $\mu$m, with a standard deviation of 3 $\mu$m.

![Histogram of $A = r_{\text{eff}}/r_{\text{ad}}$ for all ice-free regions of cloud, with $L > 0.01$ g m$^{-3}$, studied during the project.](image)

Noticeable changes in the parameter $A$ began to occur when the probes revealed the presence of ice particles in the midst of the cloud. Figures 2(a) and (b) present 1 s (100 m) average plots of $L/L_{\text{ad}}$ against the effective-radius parameter $A$ for many penetrations made through 28 cumulus clouds. The crosses in Fig. 2(a) correspond to measurements made when the ice-particle concentration determined from the 2DC, $N_C$, was equal to 0, and the hollow squares in Fig. 2(b) relate to conditions where $N_C$ exceeded $10\, L^{-1}$ of air. We see from these figures that the values of $A$ calculated with the 100 m cloud sections containing ice are slightly, but significantly, greater (as indicated by the averages and standard deviations) than the values determined for the ice-free regions. Notice also that there were very few low values of $A$ in regions of cloud containing ice.

The increase in $A$ when ice is present in clouds is illustrated more clearly when data from a single cloud are examined; these data are shown in Fig. 3. The values of $A$ are larger in regions containing more than $10\,L^{-1}$ of ice than in regions containing no ice. Figure 4 illustrates the temporal variations of the droplet concentration $N$, ice-particle concentration from the 2DC, $N_C$, liquid-water content $L$ determined from the FSSP, and the vertical and horizontal wind for a typical penetration in this cloud. Notice that ice is present in regions of cloud with substantial
Figure 2. Plot of \( A \) versus \( L/L_{ad} \) for regions of cloud with \( L > 0.01 \text{ g m}^{-3} \) and with (a) \( N_c = 0 \); (b) \( N_c \geq 10 \text{ L}^{-1} \) from 28 cumulus clouds. The vertical broken line is the average, and the horizontal line represents the standard deviation \( \sigma \). The values are: (a) \( \bar{A} = 0.91, \sigma = 0.33 \); (b) \( \bar{A} = 1.12, \sigma = 0.17 \), where the overbar denotes the average value. The probable reason for the fact that some values of \( L/L_{ad} \) are larger than unity is that the altitude of the cloud base was not precisely known.
Figure 3. Plot of $A$ versus $L/L_{ad}$ for regions of a single cloud on 19 August 1987 with $L > 0.01 \text{ g m}^{-3}$ and with: $N_C = 0$ (×); $N_C \approx 10 \text{ L}^{-1}$ (□). The vertical broken and horizontal solid lines are the averages and the standard deviations, $\sigma$, respectively. The values are: for $N_C = 0$, $A = 0.82$, $\sigma = 0.31$; and for $N_C \approx 10 \text{ L}^{-1}$, $A = 1.11$, $\sigma = 0.14$, where the overbar denotes the average value.

Figure 4. Time series of, in ascending order, liquid-water content $L$, cloud-droplet concentration $N$, ice-particle concentration measured by the 2DC, $N_C$, and the vertical and horizontal wind for a penetration made on 19 August 1987. The wind vectors are constructed with the horizontal wind along the track of the aircraft and the vertical wind. Gaps in the trace of $N_C$ are due to data problems.
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amounts of liquid-water content. However, the concentration of ice \( (N_c = 20 \text{ L}^{-1}) \) is four orders of magnitude smaller than the concentration of cloud droplets \( (N = 5 \times 10^5 \text{ L}^{-1}) \). Figure 5 shows shadows of ice particles sampled by the 2DC collected during a small section (marked by ‘A’) of the penetration shown in Fig. 4. The particles are a mixture of unrimed and rimed stellaris existing in a region of cloud with negligible updraught, but substantial liquid-water content.

The physics of this effect can be understood by the following argument, which we stress is qualitative only and disregards effects (such as the fall-speeds of the ice particles) that would have to be considered in a more rigorous treatment. Ice particles in concentrations of a few per litre are about four orders of magnitude less numerous than the supercooled water droplets with which they coexist. However, in a slightly water-supersaturated environment characteristic of these cumulus clouds before glaciation is well advanced, each one of them consumes vapour from the environment much more rapidly than individual neighbouring droplets (the Bergeron–Findeisen process), in part because they are aspherical and (through growing faster) larger, but principally because the super-saturation with respect to ice is much greater than that over water. Characteristic values for these enhanced-growth-rate factors are about three orders of magnitude in the latter case (taking \(-10^3 \text{ C}\) as a typical temperature) and between one and two orders of magnitude in the former. Initially, when ice is first produced, the competition for the water vapour will be negligible, because the solid particles are too few and too small to exercise a significant effect. In this situation the value of \( A \) will be around, or somewhat less than, 1.0. However, as glaciation proceeds and ice particles grow in number and size, the competition will become significant and perhaps eventually dominant. In this circumstance the vapour pressure in the region will fall, possibly to below the saturation value with respect to water, whilst remaining above the saturation value with respect to ice. In this latter situation the ice particles will continue to grow while the droplets will evaporate.

![Figure 5](image)

Figure 5. 2D images from the 2DC and 2DP probes from the region of cloud indicated by A in Fig. 4. The text under each strip of images refers to the probe, date, and the time of the beginning and end of the images. The distance between the horizontal lines is approximately 800 and 6400 \( \mu \text{m} \) for the 2DC and 2DP probes respectively.

Strongly sub-adiabatic cumulus clouds tend to contain high concentrations of relatively small droplets most likely produced by recent activation of entrained cloud condensation nuclei (e.g. Hill and Choularton 1985). When the vapour pressure falls below the water-saturation value these small droplets will tend to evaporate completely, producing an increase in the average droplet size and associated effective radius \( r_{\text{eff}} \) (and thus \( A \)), as observed in Figs. 2 and 3. The explanation is consistent with the observation that low values of \( A \) (indicating the predominance of small cloud droplets) seldom exist in cloud regions in which there are more than 10 \( \text{ L}^{-1} \) of ice.

The foregoing scenario represents only one possible way in which developing glaciation can change the water-droplet characteristics of cumulus clouds, and thereby the effective radius. An alternative effect, a decrease in \( A \) below the idealized value rather than an increase, may be produced when the supersaturation with respect to water falls (as a consequence of depletion of water vapour by the growing ice particles) without significant total evaporation of the small cloud droplets. In this case \( r_{\text{eff}} \) and thereby \( A \), will diminish.

The overall influence of glaciation on \( r_{\text{eff}} \), and hence \( A \), will be a function of the updraught speed, ice-particle sizes and concentrations, together with the droplet size distributions, which themselves are generally influenced by entrainment. Furthermore, \( r_{\text{eff}} \) may be affected by the
concentration and size of graupel pellets growing by accretion of supercooled droplets. It is planned to explore this range of possibilities further, by means of modelling and more extensive data analysis.

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