The Role of Giant and Ultragiant Aerosol Particles in Warm Rain Initiation

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ABSTRACT

Giant and ultragiant aerosol particles can play an important role in warm rain initiation. Recent aerosol measurements have established that particles as large as 100 μm are a regular part of the atmospheric aerosol. When ingested in growing clouds, these particles will produce a tail of large drops on the upper end of the cloud-droplet distribution. In a series of numerical computations, this tail of large drops was found to be a significant destabilizing factor which can speed precipitation development.

1. Introduction

In his classic study of the mechanisms of precipitation development, Bergeron (1935) minimized the importance of non-freezing processes. In the years following that paper, however, there has been a gradual acceptance that these "warm" processes are responsible for much of the rain that falls in tropical and maritime areas, and that these same processes operate effectively in many midlatitude continental storms.

a. Background

Warm rain development depends on the growth of cloud droplets by condensation and the subsequent growth of precipitation by coalescence of cloud droplets. The critical problem in explaining the onset of coalescence is to understand how drops of different sizes are first introduced in a cloud. It has long been recognized that the condensational growth of a cloud droplet, in terms of its radius, is inversely proportional to the radius. If, as was formerly believed, all nuclei involved in cloud formation were much smaller than typical cloud droplets, then continued condensational growth should produce an increasingly monodisperse distribution. Since these drops would fall at about the same speed, collisions between them would be exceedingly rare. This was the difficulty that led Bergeron to underestimate the significance of drop collision and coalescence. Houghton (1938) was one of the first to emphasize the potential importance of the coalescence mechanism and suggested several possible explanations for the initial production of different size drops. A little over a decade later, Bowen (1950) and Ludlam (1951) performed detailed calculations illustrating the rapid development of precipitation-sized particles once drops of different sizes were introduced in a cloud. Several years later, East (1957) explained the role of condensation as the driving force behind rapid coalescence growth. Taken jointly, these studies still constitute the classical core of our knowledge about coalescence processes. The major result of these studies is an understanding of precipitation development in "warm" clouds once sufficient numbers of drops with radii >20–25 μm are formed. The mechanism for the initial production of these first large drops, however, remains unclear.

There have been many attempts to explain the formation of drops capable of triggering coalescence. Condensation on giant sea-salt particles suspended in the subcloud air was the first source of these large drops to be studied extensively. In 1938, Houghton had identified these particles as likely candidates to initiate coalescence. Subsequent measurements of the concentrations of these salt-nuclei in maritime air have generally confirmed their presence in adequate numbers to account for rain development. Although great effort has been spent trying to identify comparable sources of soluble nuclei in continental air, such sources have not been identified and it is generally agreed that these particles do not play a major role in continental clouds (Mason, 1971).

Telford (1955) argued that the random nature of collisional processes causes a small number of fortunate drops to grow much faster than average. Collisions between the numerous small cloud droplets, although inefficient on average, may thus produce a few large drops which initiate rain development. Although this basic idea was not new (Schumann, 1940; Melzak, 1953), it offered an attractive alternative to the giant nuclei hypothesis and generated...
a great deal of work by a number of investigators. Computer simulations of this "stochastic" collection process have been reasonably successful in explaining precipitation development in relatively unstable maritime clouds, but have been less successful in stable continental clouds. At present, in spite of the many advances in this area, the origin of the first large drops that trigger coalescence growth is still largely unresolved (Telford, 1975).

Difficulties in explaining the rapid development of large drops in continental clouds have led investigators to consider more elaborate mechanisms. Mason and Jonas (1974), for example, examined drop growth in successive small thermals, where the residue of an initial thermal was allowed to interact with later ones. Recently, other scientists have turned to studies of turbulence to explain the initial production of large drops. Woods et al. (1972) and de Almeida (1979) studied the possible role of small-scale turbulence in increasing the collision efficiencies between small drops. Manton (1979), Baker and Latham (1979), and Telford and Chai (1980) have investigated another aspect of the problem, examining the role of turbulence in broadening the cloud droplet distribution by mixing cloud elements from different portions of the cloud or between the cloud and its surrounding environment. While these studies, and others, have had some notable successes, it is not clear to what extent they provide a satisfactory general explanation for the origin of the first large drops.

b. An old aspect of coalescence theory

This paper returns to the simple hypothesis that large drops grow from large nuclei. The essential requirement to test this idea is a complete description of the initial aerosol distribution over the full range of particle concentrations that are potentially of importance in precipitation development. Fig. 1 shows one way of estimating the range of concentrations which should be considered. This figure illustrates the size and number concentration of raindrops required to meet several different criteria for precipitation onset. In particular, if the first few raindrops reach several millimeters radius, concentrations as low as $10^{-3} \text{m}^{-3}$ are capable of producing significant radar echoes. To properly test the potential of the largest aerosol particles to initiate precipitation, the initial aerosol distribution must be extended until particle concentrations fall significantly below this level. Previous studies have failed to perform an adequate test of this hypothesis by not extending aerosol size distributions to large enough sizes. Recently, a number of investigators have measured the concentrations of atmospheric aerosol particles out to large enough sizes to make such a test possible. This paper performs that test.

The upcoming section will examine the initial evolution of the cloud droplet distribution by condensation. Subsequent sections will then use the condensation-produced drop spectra as the starting point for rain initiation studies using models that include both condensation and coalescence.

2. Big aerosol particles make big drops

Meteorologists have traditionally considered 10 \(\mu\text{m}\) to be an upper size-limit for atmospheric aerosol particles (Ludlam, 1980). Observational studies, however, have shown that particles as large as 100 \(\mu\text{m}\) or more are a regular part of the atmospheric aerosol. For example, Okita (1955), Jaenicke and Junge (1967), Noll and Pilat (1971), Schütz and Jaenicke (1974), and Jaenicke and Schütz (1978) have measured the surface concentrations of aerosol particles $>10 \mu\text{m}$ in a number of different locations. Airborne measurements (Nelson and Gokhale, 1968; Hindman, 1975; Johnson, 1976; Hobbs et al., 1976, 1977, 1978; Hindman et al., 1977a,b) have also confirmed the widespread presence of these unexpected large particles in the air below cloud base. While giant insoluble particles have been known to be carried aloft into thunderstorms (Rosinski, 1967; Rosinski and Kerrigan, 1969; Rosinski et al., 1976, 1979), current aerosol studies suggest that this is a common occurrence, which is not limited to severe storms.

Johnson (1976) introduced the term ultragiant for aerosol particles larger than 10 \(\mu\text{m}\) in order to call attention to their extreme size and to distinguish

\footnotesize{
\textsuperscript{2}Bergeron (1935) discussed the possibility that turbulence might promote coalescence, but discounted its importance.

\textsuperscript{3}This general aspect of turbulent mixing was first suggested by Houghton (1938).
}
them from giant aerosol particles, which have often been implicitly assumed to have an upper size limit of 10 µm. Over continental areas, these ultragiant particles seem to be mostly soil fragments carried aloft by the wind. While such particles will be rather complex mixtures of soluble and insoluble material, they are predominantly insoluble.

The widespread existence of particles >10 µm radius radically changes our understanding of the initial phases of cloud microstructure. In order for a 10 µm radius particle to grow larger than 20–25 µm and initiate coalescence, it must be completely soluble. Over the years, great effort has been spent looking for possible sources of such soluble nuclei. If, on the other hand, particles larger than 20–25 µm radius are a regular component of the atmosphere, then the solubility, and even the surface properties of these particles, may be irrelevant. Such particles can initiate coalescence growth without prior growth by condensation and may be termed coalescence nuclei (Ludlam, 1951; Johnson, 1979).

Although these particles are apparently a natural part of the atmospheric aerosol (Pruppacher and Klett, 1978; Jaenicke, 1980; Podzimek, 1980), it is not immediately clear if there are ever enough of them present to be able to significantly affect the clouds that ingest them.

a. Condensation model

The most logical starting point for any study of warm rain initiation is the formation of the cloud-base drop-size distribution by condensation of water on a spectrum of aerosol particles. There have been a number of detailed studies of this initial phase of cloud development (e.g., Howell, 1949; Mordy, 1959; Neiburger and Chien, 1960; Fitzgerald, 1972; Lee and Pruppacher, 1977). None of these studies, however, considered aerosol particles larger than 10 µm. The condensation model used in this study is based on Fitzgerald's (1972, 1974) formulation of the basic equations. In this type of model, the growth of an individual aerosol particle (or group of aerosol particles assigned to a single category) is followed in Lagrangian fashion by direct calculation of the expected growth or evaporation in a specified time step. This usually means that extremely short time steps have to be used. By limiting the growth or evaporation of unactivated droplets to that actually required for them to reach their equilibrium size, as suggested by Mordy (1959) and Hjelmfelt et al. (1978), substantial increases in the time step can be achieved. This, in turn, allows explicit consideration of a wider size-range of aerosol particles than would be practicable with earlier models. For a more detailed discussion of the condensation model, see Johnson (1979, 1980).

Four different initial aerosol distributions were used in this study (see Fig. 2). The “Missouri” distribution was based on observations of cloud condensation nuclei (CCN) and aerosol particles upwind of the St. Louis urban area that were collected as part of Project METROMEX. The HIPEX distribution was based on airborne observations taken over the Great Plains as part of the High Plains Cooperative Project (Hobbs et al., 1977, 1978). The two maritime distributions were adapted from Hindman's (1975) compilation of maritime aerosol data. For uniformity, and to ensure that all particles that might potentially influence precipitation development were considered, the large particle end of each distribution was extended to 100 µm radius.

Prior to actual use in the condensation model, the individual aerosol distributions were broken into discrete categories based on their size and fractional solubility. Both continental distributions (Missouri and HIPEX) were divided into 70 different logarithmically-spaced size-categories based on the radii of the dry aerosol particles. Since the maritime aerosols contain significantly fewer particles, a coarser division was used having only 50 different categories. To permit particles of a given size to have different fractional solubilities, each size-category was subdivided into five different solubility categories. This resulted in a total of 250 separate size or solubility categories for the maritime distributions, and 350 categories for the continental distributions. For any one size-category, each of the five associated solubility categories received 20% of the total number of particles of that size. Although a number of investigators have studied the solubility of atmospheric aerosol particles, no data set covers the entire range of particle sizes. In general, however, the smallest particles seem to be predominantly soluble. In continental air the largest particles are relatively insoluble, while in maritime air, sea-salt often dominates the large particle end of the spectrum and helps maintain relatively high solubilities over the entire size-range of particles. Table 1 shows the size-de-
Table 1. Aerosol solubilities.

<table>
<thead>
<tr>
<th>Initial radius (μm)</th>
<th>Maritime aerosols</th>
<th>Continental aerosols</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$e_1$</td>
<td>$e_2$</td>
</tr>
<tr>
<td>$r_0 &lt; 0.04$</td>
<td>0.40</td>
<td>0.60</td>
</tr>
<tr>
<td>$0.04 \leq r_0 &lt; 0.10$</td>
<td>0.40</td>
<td>0.60</td>
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<td>$0.10 \leq r_0 &lt; 1.00$</td>
<td>0.40</td>
<td>0.60</td>
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<tr>
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<td>0.40</td>
<td>0.60</td>
</tr>
<tr>
<td>$10.0 \leq r_0$</td>
<td>0.40</td>
<td>0.60</td>
</tr>
</tbody>
</table>

Dependent fractional solubilities ($e_i$) assigned to each of the initial continental (Missouri and HIPLEX) and maritime (A and B) aerosol distributions used in these investigations.

b. Results

Computations with the condensation model identify which of the initial aerosol particles grow into cloud droplets, and what the resultant cloud-droplet spectrum looks like. Fig. 3, for example, shows a smoothed presentation of the drop spectra expected one hundred meters above cloud base for each of the four aerosol distributions shown in Fig. 2. In each case, computations started ~400 m below cloud base (80% relative humidity) with a starting temperature and pressure chosen to produce the desired cloud base temperature. Both maritime cases started with environmental conditions that resulted in a cloud-base temperature of 15°C. The Missouri case had a 10°C cloud-base temperature, while the HIPLEX cloud base was 2°C. In each case, a uniform 2 m s⁻¹ updraft was assumed. While the logarithmic scale used in this figure is necessary to see the “tail” of large drops, it distorts the relative numbers of drops in each case. The graphical insert in the upper right-hand corner shows exactly the same spectra, but drawn to a linear scale. With the linear scale, the area under each curve is proportional to the total number of drops. In this example, the maritime calculations resulted in the activation of 63 (maritime A) and 197 (maritime B) drops per cubic centimeter. The Missouri case activated 907 drops cm⁻³, while the HIPLEX cloud resulted in 577 drops cm⁻³.

Precipitation development, of course, requires more than just condensation. In the next section, drop spectra from near cloud-base, such as those shown in Fig. 3, will be used as the starting point for calculations of the initial development of precipitation in models that include drop collision and coalescence.

3. Big cloud drops grow into raindrops

In this section, three different models of precipitation development are used to investigate the role of the largest cloud drops in rain initiation. The first model examines the growth of these drops by condensation and continuous collection in a rising parcel of cloudy air. The second model employs the same parcel framework, but allows a fuller treatment of the overall evolution of the cloud droplet distribution by using a sophisticated “stochastic” formulation of drop coalescence. The third model returns to the simpler continuous collection approach, but this time considers the differential sedimentation of the growing drops through calculations of the complete particle trajectories. In each case, the initial growth and activation of cloud droplets from nuclei is bypassed by using the results of the detailed condensation model discussed in the previous section as the starting point for all subsequent calculations.
a. Continuous collection parcel model

Perhaps the simplest and most direct way to evaluate the significance of the drops formed on the largest nuclei is to follow their growth by continuous coalescence with smaller cloud droplets in an undiluted parcel rising at constant velocity. In this type of model, the cloud droplet distribution is partitioned into two general classes: large drops and small drops. The concentration of small drops and a number of different categories of large drops are specified at the start of the computation, as are the cloud-base temperature, pressure and updraft velocity. Growth of a large drop of radius \( R_l \) and mass \( M_l \), by collection of smaller cloud droplets of radius \( r \) (and mass \( m \)) is given by

\[
dM_l = \pi R_l^2 E(R_l, r) \chi \rho_a [v(R_l) - v(r)], \tag{1}
\]

where \( v(R_l) \) and \( v(r) \) are the terminal velocities of water drops of radii \( R_l \) and \( r \), respectively, \( E(R_l, r) \) is an appropriately defined collection efficiency, \( \chi \) is the liquid water mixing ratio of the small drops being collected, and \( \rho_a \) is the density of the air. Collection efficiencies, \( E(R_l, r) \), are obtained by four-point interpolation from tables based on Young’s (1973) compilation of collection efficiency data.

Assuming all vapor in excess of saturation is condensed on the growing water drops, then condensation growth can be expressed as

\[
dm = \frac{d\omega_z}{dt} \frac{r}{nr + \sum N_i R_i}, \tag{2a}
\]

\[
dM_l = \frac{d\omega_z}{dt} \frac{R_l}{nr + \sum N_i R_i}, \tag{2b}
\]

where \( N_i \) and \( n \) are the number of large drops (of class \( i \)) and small drops, respectively, and \( \omega_z \) is the saturation mixing ratio of the rising parcel of moist air. In all cases, the drop concentrations are maintained as mixing ratios (number of drops per gram of dry air). The time rate of change of the saturation mixing ratio is given by

\[
\frac{d\omega_z}{dt} = \frac{gU}{RT} \left[ \frac{\varepsilon_z + \gamma}{P + \rho_a g} \frac{d\epsilon_z}{dT} \right], \tag{3}
\]

where \( U \) is the vertical velocity of the rising parcel, \( T \) is the temperature, \( P \) is the pressure, \( \rho_a \) is the air density, \( g \) is the acceleration of gravity, \( \kappa \) is the specific gas constant for water vapor, \( \varepsilon_z \) is the saturation vapor pressure, and \( \gamma \) is the appropriate (pseudo-adiabatic) lapse rate. The saturation vapor pressure, and its derivative with respect to temperature may be obtained from the Clausius-Clapeyron equation, or by suitable empirical equations such as those suggested by Lowe (1977).

The terminal velocity for small droplets (\( r < 35.9 \mu m \)) can be approximated by

\[
v = C_1 r^2.
\]

Larger drops (35.9 \( \mu m < r < 300 \mu m \)) have terminal velocities describable by a linear relation,

\[
v = C_2 r - C_3.
\]

For still larger drops (\( r > 300 \mu m \)) an expression suggested by Atlas et al. (1973) can be used,

\[
v = C_4 - C_5 \exp(-12r).
\]

For \( r \) in cm and \( v \) in cm s\(^{-1}\), the appropriate constants for the preceding equations are: \( C_1 = 1.20 \times 10^6 \), \( C_2 = 8.62 \times 10^3 \), \( C_3 = 1.55 \times 10^4 \), \( C_4 = 9.62 \times 10^2 \), and \( C_5 = 1.03 \times 10^3 \). Although this simple set of relations neglects the temperature and pressure dependence of the terminal velocities, it does adequately describe the general features of the variation of fall speed with size. This particular set of equations is especially useful since both the velocities and their derivatives with respect to radius are matched at the transition points between equations.

If a drop grows larger than a specified maximum size (typically 0.25 cm radius), it is assumed to break into a number (typically four) of uniform-sized fragments. Since large drops are not allowed to collide with each other, breakup is the only way the number of large drops can change. The number of small cloud droplets, on the other hand, is continually being depleted by collisions with the larger drops. In this version of the model, drop sedimentation relative to the rising parcel is ignored and all drops are assumed to stay with their initial volume of air. A complete model description is given by Johnson (1979).

Fig. 4 illustrates the model-predicted evolution of the radar reflectivity factor for the four drop distributions shown in Fig. 3. In each case, an undiluted parcel rising at 2 m s\(^{-1}\) achieves radar detectable reflectivities (~10 dBZ) within 1.5 to 2.5 km above cloud base. The differences in the predicted evolution of reflectivity in the four cases illustrated reflect both differences in the respective drop spectra and differences in cloud-base temperature.
Since the continuous collection model preserves the growth history of each collecting drop, it is possible to see which of the drops in the initial droplet distribution eventually grow to contribute the most to the total radar reflectivity. At the beginning of each run, the number of drops in each of 26 logarithmically spaced categories between 10 and 100 \( \mu m \) radius were obtained from the condensationaly produced drop spectra 100 m above cloud base (i.e., Fig. 3). Fig. 5 shows the contributions from these initial drop categories toward the total reflectivity factor at precipitation onset (10 dBZ) for each drop distribution investigated. Since the number of drops in each category was obtained by a rigorous partitioning of drops from a finite number of nuclei categories into discrete size-intervals, the histograms shown in this figure are not particularly smooth. Comparing these results with the initial drop distributions shown in Fig. 3, it is apparent that drops far out on the tail of the distribution can contribute significantly to the initial development of radar detectable reflectivities.

b. Stochastic collection parcel model

In addition to the tail of large drops formed directly on giant and ultragiant aerosol particles, collisions between the numerous small cloud droplets can also produce drops that are large enough to continue to grow effectively by coalescence. In order to evaluate the rate that additional large drops are produced by collisions among smaller droplets, it is necessary to employ the more complete stochastic model of drop coalescence. This type of model generally utilizes a fixed set of Eulerian drop-size categories, redistributing the cloud drops among these categories as they grow. Unfortunately, there is an inherent difficulty with this approach since, in general, coalescence of any two drops, or condensation growth of a single drop, will not produce a drop that is exactly the right size to fit in a new category. The subsequent splitting of the drop concentrations into the adjacent categories or interpolation between categories introduces numerical inaccuracies. When modeling precipitation development, however, the numerical accuracy required is truly awesome. While there may be a total of a thousand cloud droplets or more per cubic centimeter, drops in concentrations of one or less per cubic meter can dominate the onset of precipitation. Thus, the numerical techniques employed in these models must be capable of handling a range of drop concentrations of more than nine or ten orders of magnitude without appreciable numerical diffusion. Recently, several investigators (Young, 1975; Ochs and Yao, 1978) have produced models of condensation and coalescence that incorporate improved methods of controlling numerical diffusion that seem to be adequate for studies of precipitation initiation.

Fig. 6 shows the evolution of the radar reflectivity factor predicted by Young’s (1975) parcel model with stochastic collection for the same four sets of initial conditions used in Fig. 4. Although this model is more sophisticated than the simple continuous collection model employed previously, the results are essentially the same.

In the full stochastic model, all drops are redistributed among the fixed categories during each time step. Since the growth history of any individual drop is lost, it is impossible to tell directly which of the initial drops contribute the most toward the onset of precipitation. In this case, the only way to infer the importance of any particular size of nuclei or drop is to repeat the calculation removing the relevant aerosol particles or drops. Fig. 7 shows the effect of progressive truncations of the largest drops for each of the four cases shown in Fig. 6. The leftmost (unlabeled) curves use the entire drop spectrum, including drops as large as 100 \( \mu m \) radius. These curves are exactly the same as those shown in Fig. 6. Each subsequent curve shows the evolution of the radar

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure5.png}
\caption{Contributions from each initial “cloud-base” drop category toward the total reflectivity factor at precipitation onset (continuous collection parcel model).}
\end{figure}

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure6.png}
\caption{Time evolution of radar reflectivity factor (dBZ) in a parcel model with stochastic collection.}
\end{figure}
reflectivity factor for the same general conditions, but with the initial distribution truncated at the indicated drop radius. The two continental cases (Figs. 7a and 7b) are very similar. Truncation of drops >70 \( \mu \)m radius has relatively little effect on the time required to reach a given reflectivity threshold such as 10 dBZ. As progressively smaller drops are truncated, however, the time to reach that threshold increases sharply. The maritime cases (Figs. 7c and 7d) show much less sensitivity to truncation of the large drops. This is particularly true of the maritime “A” distribution (Fig. 7d), with its drop concentration of only 63 cm\(^{-3}\).

These results suggest that, with the exception of drops forming directly on giant or ultragiant aerosol particles, continental clouds are inefficient in producing drops that are large enough to continue to grow effectively by coalescence. In such cases, the original concentration of large drops in the tail of the droplet distribution is critical to the development of precipitation. On the other hand, maritime clouds, with relatively low droplet concentrations and high water contents, seem to be much more effective in producing additional large drops, and show less sensitivity to the large drop tail. In both maritime and continental cases, however, measured concentrations of giant and ultragiant aerosol particles are adequate to produce enough large drops to account for the initial development of precipitation.

c. Continuous collection trajectory model

The preceding sections suggest a major role for giant and ultragiant aerosol particles in initiating warm rain. These results, however, were obtained with models that neglect drop sedimentation. Sedimentation can limit the effectiveness of the largest drops in initiating rainfall. If the updraft is weak, the largest drops may simply fall out of the updraft. Even if the updraft is strong enough to carry all sizes of drops aloft, the net rate of rise of the largest drops will be less than that of smaller drops. This means that the drops that are initially the largest will not be the first to reach the high liquid water content regions of the cloud that are most favorable to rapid coalescence growth. Therefore, the largest drops at cloud base are not necessarily the first to grow into raindrops.

Continuous collection models, such as discussed in Section 3a, can be easily modified to consider the differential sedimentation of the growing drops by adding an explicit calculation of the drop trajectories. If the updraft velocity is \( U \), then the trajectory of a large drop falling at terminal velocity \( v(R) \) can be calculated from

\[
\frac{dH_i}{dt} = U - v(R),
\]

where \( H_i \) is the height of the large drop relative to
cloud base. The smaller cloud droplets, which grow through condensation alone, are assumed to move with the rising air. If the model is restricted to examining precipitation initiation, depletion of these small drops is minimal and may be neglected (Johnson, 1979). This means that, using the nomenclature already introduced, \( n \gamma \gg \frac{N_i}{n \gamma} \) and the condensation equations become

\[
\frac{dm}{dt} = (\frac{d\omega}{dt}) n^{-1}, \tag{5a}
\]
\[
\frac{dM_i}{dt} = (\frac{d\omega}{dt}) R_i \frac{M_i}{n \gamma}. \tag{5b}
\]

Thus, to a first approximation, the small cloud droplets' condensational growth is not affected by the large drop tail, and the droplet radius \( r \), at any height above cloud base can be calculated directly from the adiabatic water content for that height. In effect, these modifications just turn the continuous collection model into an updated version of Bowen's (1950) model for calculating drop trajectories. By calculating a large number of different trajectories, however, it is possible to construct time–height cross sections of the rainfall rate, radar reflectivity factor, or other properties of the cloud. If the initial concentrations of large and small drops are continually replenished at cloud base, then the concentration of particles will vary with height above cloud base according to

\[
N_i = N_0 [U - v(R_o)] / [U - v(R_i)], \tag{6}
\]

where \( N_0 \) and \( R_o \) are the cloud-base number concentration and radius of drops of category "\( n \)". To prevent difficulties at the balance point, where \( U = v(R_i) \), the predicted concentration \( N_i \) is arbitrarily restricted so that \( N_i \ll 4N_0 \). Since this restriction only applies in the immediate vicinity of the balance point, it does not significantly modify the overall results.

Fig. 8 illustrates sample trajectories for three different collector drops. In this example, the drop that was initially the largest was the first to grow big enough to fall out of the updraft. The other two drops stayed in the cloud longer, were carried higher, and eventually grew bigger than the drop that was initially the largest. The shaded area indicates the region of the cloud for which drop growth calculations suggest a radar reflectivity in excess of 10 dBZ. Although only three of the drop trajectories are shown, 26 categories of large drops were used to define the limits of this high reflectivity region. Once precipitation forms, the depletion of small droplets becomes increasingly important and the accuracy of the model deteriorates. Accordingly, all calculations were stopped abruptly whenever the radar reflectivity factor, at any height, reached 40 dBZ.

Fig. 9 shows the time–height evolution of the high-reflectivity region (>10 dBZ) predicted using the continuous collection trajectory model for each of the four drop distributions shown in Fig. 3. As in previous calculations, both maritime cases have cloud bases at 15°C, while the Missouri and HIPLEx cases have bases at 10 and 2°C, respectively. Compared with the parcel model results shown in Figs. 4 and 6, the inclusion of sedimentation means that slightly more time is required to generate precipitation and it forms lower in the cloud. In the Missouri calculations, for example, the parcel models predicted it would take slightly over 16 min to develop.
results from the parcel model (Fig. 5), there is a clear shift in importance toward smaller drops and away from drops forming on the very largest nuclei. In each case, however, the nuclei that are important are considerably larger than nuclei that have been included in previous investigations.

Both entrainment of dry air and depletion of small droplets by growing raindrops will reduce the amount of liquid available for collection. In order to estimate the sensitivity of the model results to such reductions, the calculations used to generate Fig. 9 were repeated with the liquid water contents at every height arbitrarily reduced to half of the undiluted (adiabatic) values without making any corresponding adjustments in drop concentration. As would be expected, this slows precipitation development (see Fig. 11). Even in these conditions, however, the maritime and Missouri cases still produce radar detectable reflectivities in relatively thin clouds. The HIPLEX case, with its colder cloud base of 2°C, appears more sensitive to the reduction in water contents and might be expected to form precipitation by ice-phase processes before warm rain could develop.

The location of radar echoes within the growing cloud is critically dependent on the updraft velocity. Table 2 shows the time and height at which the radar reflectivity factor first surpasses 10 dBZ for a number of different updrafts. In each case the Missouri aerosol distribution, with a cloud-base temperature of 10°C, was used as the starting point for the calculations. The total number of nuclei that activate increases sharply with increasing updraft velocity. Fig. 12 shows the corresponding time-height evolution of the 10 dBZ reflectivity region for eight of these cases. For each updraft, the contributions toward the 10 dBZ threshold from each initial drop category were analyzed. Fig. 13 indicates which of these categories were principally responsible for the initial production of the radar echo. This figure also shows the corresponding results from the parcel model. The faster the updraft, the more difficult it is for the large drops

<p>| Table 2. Updraft influences on echo onset (continuous collection trajectory model). |
|---------------------------------|-----------------|-----------------|-----------------|</p>
<table>
<thead>
<tr>
<th>Updraft (m s⁻¹)</th>
<th>Total droplet concentration (cm⁻³)</th>
<th>Time (min)</th>
<th>Height (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.25</td>
<td>397.2</td>
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<td>4.20</td>
</tr>
</tbody>
</table>
to fall out of the rising current of air. Eventually, as the updraft approaches 8–10 m s\(^{-1}\), the behavior of the trajectory model begins to approximate that of a parcel model which neglects the differential sedimentation of the growing drops.

4. Discussion

The principal conclusion of this study is that naturally occurring giant and ultragiant aerosol particles can play an important role in precipitation initiation. When ingested in developing clouds, these particles will produce a tail of large drops on the upper end of the droplet distribution. In a favorable environment, this tail of large drops is capable of rapid growth to precipitable sizes. If the aerosol particles are large enough, even insoluble particles can initiate the required coalescence growth.

Measurements of the concentrations of giant and ultragiant aerosol particles in the air below cloud base have shown that they are a regular component of the atmospheric aerosol. The theoretical studies that make up the bulk of this paper indicate that typically observed concentrations of these particles seem to be fully adequate to explain the rapid development of radar-detectable reflectivities in warm clouds. Furthermore, so long as the natural concentration of giant particles is even close to the measured concentrations that were used to construct the aerosol distributions used in this study, the time or location of precipitation onset is not particularly sensi-tive to the actual number of giant particles present. For example, increasing or decreasing the concentration of giant and ultragiant aerosol particles by an order of magnitude or more was found to have only minimal effects on the predicted 10 dBZ first echo. Although at first this seems a bit paradoxical, it just reflects the greater sensitivity of most measures of precipitation onset (such as a specified radar reflectivity factor) to drop sizes and rates of growth than to number concentration (see Fig. 1). This sensitivity is also demonstrated in the parcel model results (e.g., Fig. 6), which suggest that an order of magnitude reduction in the number of growing raindrops, which would reduce reflectivities by 10 dBZ, would only delay the build up of radar-detectable reflectivities by \(~1\) min.

As drops grow large, they become increasingly susceptible to breakup. If, following breakup, the fragments produced grow large enough to break again, they can create a breakup “chain reaction” (Langmuir, 1948). If continued, this can result in a rapid increase in the number of large drops and, in principle, could magnify the relative importance of the drops initiating the chain reaction. Young’s (1975) stochastic collection model, which was used extensively in the previous section, includes a detailed calculation of collisional breakup based on the work of Brazier-Smith et al. (1972, 1973). On the other hand, the continuous collection models used in the preceding section handled breakup by the relatively simple expedient of arbitrarily splitting any drop exceeding a given threshold into a specified number of equal-sized fragments. In no case, however, did any of the models indicate breakup played a significant role in initiating precipitation.

In relatively unstable maritime clouds, small cloud droplets seem capable of readily undergoing the multiple stochastic collisions necessary to produce
Table 3. Influence of subadiabatic water contents on precipitation development in a continuous collection parcel model (Missouri cloud, 2 m s\(^{-1}\) updraft).

<table>
<thead>
<tr>
<th>Fraction of adiabatic water content</th>
<th>Height (km) of 10 dBZ first echo</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0</td>
<td>1.96</td>
</tr>
<tr>
<td>0.9</td>
<td>2.09</td>
</tr>
<tr>
<td>0.8</td>
<td>2.24</td>
</tr>
<tr>
<td>0.7</td>
<td>2.44</td>
</tr>
<tr>
<td>0.6</td>
<td>2.68</td>
</tr>
<tr>
<td>0.5</td>
<td>3.02</td>
</tr>
<tr>
<td>0.4</td>
<td>3.50</td>
</tr>
<tr>
<td>0.3</td>
<td>4.28</td>
</tr>
<tr>
<td>0.2</td>
<td>5.86</td>
</tr>
</tbody>
</table>

drops large enough to continue effective coalescence growth. In this case, the presence of large drops formed directly on giant and ultragiant aerosol particles may not be critical for precipitation development. The more stable a cloud gets, however, the more important the original tail of large drops becomes. In high drop concentration continental clouds, the presence of these initial large drops may well be the single most important factor in the initial development of precipitation.

Of course, not all clouds that ingest ultragiant particles will develop warm rain. Clouds that extend above the freezing level are also susceptible to precipitation initiation by processes involving the ice-phase. In particular, clouds with base temperatures in the vicinity of the 0°C isotherm (such as the HIPLEX case discussed in the previous sections) are likely to be dominated by ice-phase processes before drop coalescence mechanisms have an opportunity to take hold. In other cases, excessive entrainment of dry air may prevent precipitation formation by any mechanism, warm or cold. In the absence of entrainment, however, warm-based clouds ingesting ultragiant particles should be expected to develop precipitation within a few kilometers of cloud base.

In addition to explaining the rapid development of coalescence rainfall in seemingly stable continental clouds, the existence of a natural tail of large drops has a number of important corollaries. First, adiabatic water contents do not need to be invoked to explain rapid precipitation development. To be sure, high water contents speed rain development. The preexistence of the large drop tail, however, removes the need to explain the initial development of drops large enough to continue effective coalescence growth. Traditionally, this has been the rate-limiting step in precipitation calculations and has been found to be particularly sensitive to subadiabatic water contents. Table 3 illustrates the reduced sensitivity to subadiabatic water contents resulting from the presence of the large drop tail. In these runs, the continuous collection parcel model discussed in Section 3a has been used to calculate the height at which the reflectivity factor first exceeds 10 dBZ. Each run is based on the same cloud-base drop distribution (Missouri aerosol, 2 m s\(^{-1}\) updraft). The subsequent condensational growth of the cloud drops, both large and small, is then arbitrarily restricted to a specified fraction of the expected adiabatic values. Every reduction in water content delays the formation of the 10 dBZ first echo. Until the water contents drop below 0.5–0.6 of their adiabatic values, these delays are gradual and relatively modest. Further reductions in water content, however, significantly retard precipitation development.

Another result of the existence of a natural large drop tail produced by giant and ultragiant aerosol particles is a sharp reduction in the significance of the total droplet concentration as an indication of colloidal stability. Table 4, for example, shows the estimated height of the 10 dBZ first echo predicted by the continuous collection parcel model for five different cloud droplet concentrations. For each droplet concentration, there are two estimates of the first echo height. The first estimate is based on the large drop tail appropriate for the Missouri aerosol distribution with a 2 m s\(^{-1}\) updraft, while the total concentration of cloud droplets is arbitrarily set at the specified values. In reality, the total drop concentration and number of giant and ultragiant aerosol particles should be expected to be highly correlated. This is reflected in the second estimate of the first echo height in which the number of drops in the large drop tail has been adjusted by the same amount that the calculated droplet concentration had to be adjusted to reach the specified value. Estimate I should be expected to exaggerate the significance of changes in the cloud droplet concentration, while estimate II should underestimate the effect. Neither estimate I nor II, however, indicate any critical dependence on cloud droplet concentration. *High droplet concentrations do not preclude precipitation development.* While this result is clearly at odds with most other studies, it is not unprecedented. Ochs and Semonin (1979), in the only other study to include a full range of giant and ultragiant aerosol particles, found similar results.

While the presence of significant concentrations

Table 4. Influence of cloud droplet concentration on precipitation development in a continuous collection parcel model (Missouri cloud, 2 m s\(^{-1}\) updraft).

<table>
<thead>
<tr>
<th>Cloud droplet concentration (cm(^{-3}))</th>
<th>Height (km) of 10 dBZ first echo</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Estimate I</td>
</tr>
<tr>
<td>31.6</td>
<td>1.58</td>
</tr>
<tr>
<td>100.0</td>
<td>1.63</td>
</tr>
<tr>
<td>316.2</td>
<td>1.75</td>
</tr>
<tr>
<td>1000.0</td>
<td>1.98</td>
</tr>
<tr>
<td>3162.3</td>
<td>2.23</td>
</tr>
</tbody>
</table>
of giant and ultragiant aerosol particles in the air entering cloud base seems well established, more measurements are needed to refine our knowledge of natural variations in the concentration of these particles. This is particularly true at the extreme upper end of the aerosol distribution where few measurements exist. None of the aerosol studies used to construct the initial distributions shown in Fig. 2, for example, reported concentrations of particles $>$100 $\mu$m diameter. Although the final results were not critically dependent on the presence of particles of this size or larger, the estimated concentrations of these particles that were used to fill in the upper end of the aerosol distribution were found to be capable of contributing to the initial build up of reflectivity, especially in those cases having updrafts larger than 5–6 $m/s$.

This study should not be interpreted as saying that all raindrops form on giant and ultragiant aerosol particles. It does show that these particles can be of great importance in the initial development of precipitation-sized drops. In a mature storm there will be many other processes at work. Even in the initial periods of precipitation development it is difficult to predict which of the many suggested mechanisms of large drop production will dominate in any given situation. The important point is that giant and ultragiant aerosol particles act as an inherent destabilizing factor which will operate in all clouds that ingest a broad spectrum of aerosol particles.

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