Warm-Rain Initiation: An Overview of Microphysical Mechanisms

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ABSTRACT

Rain triggering mechanisms are evaluated in three microphysical steps: droplet activation on cloud condensation nuclei, droplet growth by condensation, and droplet growth by coalescence. Although considerable progress has been made since the pioneering work of Squires, crucial questions in each of the above steps remain unresolved: Under what conditions do *giant particles* trigger rain by acting as coalescence nuclei? What is the contribution of *stochastic condensation* to the growth of large droplets in regions of entrainment? What are the *collection efficiencies* for droplet sizes critical to the onset of colaescence growth? Such questions cannot be answered without better observations. Aircraft instruments are becoming available with the potential to measure the very largest particles and cloud droplets at the concentration of raindrops. Recent advances in sampling and analysis techniques have extended observations of cloud microstructure to smaller scales, providing new insight on the growth of droplets by mixing. Continued progress in laboratory research should furnish collection efficiencies for the droplets sizes critical to warm-rain initiation. With such improved observations, careful evaluations using available microphysical and dynamical models should provide answers to key questions about warm-rain initiation.

1. Introduction

Rain in the tropics typically originates from clouds that are entirely warmer than freezing. This warm-rain process is not restricted to the tropics, however, as radar studies show that precipitation in midlatitude continental convection often develops beneath the freezing level. The propensity for rain to form in warm clouds suggests that there is an effective mechanism to transform cloud water into rainwater. Indeed, the growth of just a few cloud droplets (one per 10 L) by coalescence with smaller ones would explain the development of warm rain in 15–30 min—faster for broader cloud-droplet distributions and higher cloud-water contents [e.g., see growth-rate calculations of Braham (1968), Ochs (1978), Johnson (1982)].

The formation of raindrops can be divided by mechanisms into several microphysical steps: 1) activation of droplets on cloud condensation nuclei (CCN), 2) condensation growth (limited by vapor diffusion), and 3) coalescence growth (accelerated by increasing fall speeds and collision efficiencies). Although these steps have been studied rather intensively in the last several decades (e.g., see texts by Rogers and Yau 1989; Pruppacher and Klett 1978), there remain gaps in our knowledge that must be filled before we can accurately model the warm-rain process. A key microphysical problem is the initiation of coalescence; for example, the formation of droplets large enough to get the process going fast enough.

Patrick Squires devoted several of his earlier papers to this problem of "colloidal stability," that is, the insufficiency of droplet sizes necessary to initiate coalescence growth. Using the theory for activation on salt particles (step 1) and growth by condensation (step 2), Squires (1952a,b) determined that colloidal stability could be decreased (leading to step 3) by the activation of *only a few droplets* in maritime clouds under conditions that, "on present knowledge, seem likely to occur at times, but not normally." Maritime clouds typically have fewer CCN than continental clouds and consequently larger droplets. Thus, it follows that rain should form more readily by coalescence in maritime clouds as is generally acknowledged (e.g., see Alpert 1955). Continental clouds, however, should be entirely unsuitable for producing cloud droplets large enough to initiate coalescence growth. The rapid production of warm rain in both maritime and continental clouds lingers as one of the major problems in cloud physics. In this paper, we focus on the microphysical steps of warm-rain formation, and in particular, on the production of large cloud droplets needed to initiate coalescence growth. The state of our knowledge is summarized in the conclusions along with suggestions for future research.

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2. Droplet activation and growth by condensation

As a parcel of air rises through cloud base, the air becomes supersaturated and droplets are activated on the most effective CCN. These are generally the largest, most soluble particles. With increasing supersaturation, droplets are activated on less effective, but generally more numerous, smaller CCN. The supersaturation reaches a peak value just above cloud base, where the uptake of moisture by diffusion to the droplets exceeds that supplied by the lifting process. At this point, the droplet concentration has been determined, since the activation process ceases with subsequent lowering of the supersaturation. The distribution of particle sizes continues to evolve, however, as the activated droplets grow by diffusion, but the diameter distribution narrows because the growth rate of individual droplets decreases with diameter. Condensation within a parcel being lifted uniformly typically produces most of the droplets in a range of a few micrometers and a mode of less than 30-μm diameter (see Fig. 1 for a comparison of cloud-base observations and parcel calculations). Although continued condensation within a parcel will eventually generate droplets larger than the approximately 50-μm diameter needed to initiate coalescence (Mason and Ghosh 1957), the time for this process to produce rain is generally much too long (Mason 1959).

Various hypotheses have been suggested to accelerate production of large cloud droplets so that coalescence could result in rainfall from shallow maritime clouds within about 0.5 h. These include favorable CCN spectra, such as low concentrations of smaller nuclei and high concentrations of larger nuclei (Woodcock and Gifford 1949; Ludlam 1951; Squires 1952a,b), and mechanisms that would enhance condensation growth of a few larger cloud droplets such as mixing together of buoyant parcels (Ludlam and Saunders 1956) and fluctuations in supersaturation induced by turbulence (Belyaev 1961).

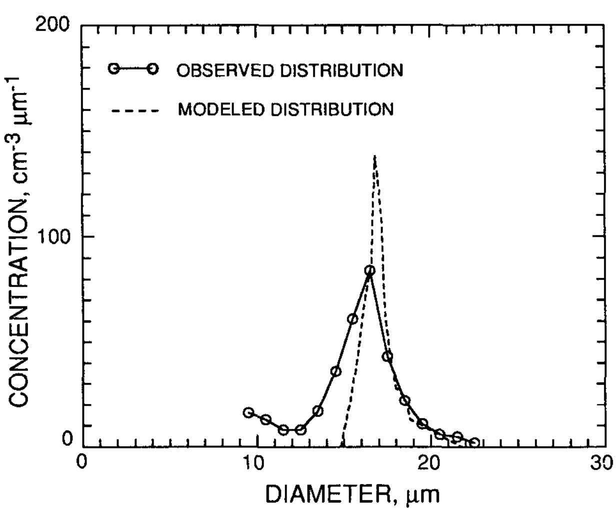

FIG. 1. Comparison of observed and computed droplet size distributions for a maritime cloud (after Fitzgerald 1972).

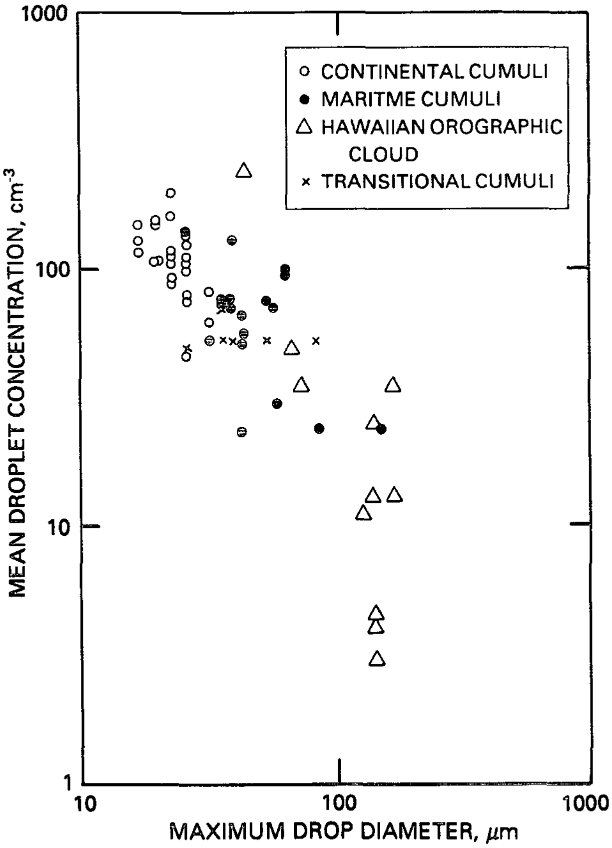

FIG. 2. Scatter diagram relating maximum droplet size to mean concentration during a cloud traverse [continental cumuli (open circles), maritime cumuli (close circles), Hawaii orographic cloud (triangles), transitional cumuli (crosses)] (after Squires 1958a).

The formation of large droplets was analyzed by Squires (1958a) based on the extensive aircraft measurements of droplet spectra in shallow clouds (Squires 1956; Squires and Warner 1957). As shown in Fig. 2 for an entire cloud traverse (about 30-cm^3 sample volume), the largest drops in maritime clouds range in diameter from 30 μm (at 30–120 cm^{-3}) for cumulus clouds to over 100 μm (at 3–70 cm^{-3}) for orographic clouds. Continental cumulus clouds had sizes less than 30-μm diameter but higher concentrations (70–200 cm^{-3}). Since a strong negative correlation was found between droplet concentration and maximum diameter, the droplet concentration served as a good indicator of cloud microstructure and colloidal stability; that is, lower concentrations were associated with larger maximum and average droplet sizes and higher potential for coalescence growth.

a. *CCN spectra favoring the production of large droplets*

Using the theory of condensation to predict the supersaturation in maritime and continental clouds, Squires (1958b) concluded that colloidal stability of continental clouds was caused by higher concentrations of large nuclei rather than insufficient concentrations of giant nuclei. [Particles are classified in diameter intervals as small (<0.2 μm), large (0.2–2 μm), and giant (2–20 μm).] Squires' theoretical predictions of colloidal stability were verified by aircraft measurements of both nuclei and droplet spectra in small nonraining cumulus clouds in Washington (Hindman et al. 1977) and corresponding calculations of droplet size distributions. They found that high concentrations of large and giant CCN led to large droplets only in the presence of low concentrations of small CCN. If high concentrations of small CCN were present, large droplets were absent, regardless of the concentration of giant CCN. In fact, the higher concentration of large droplets was associated with clouds in eastern Washington having lower concentrations of giant salt particles than in western Washington.

More recently, Johnson (1982) used a numerical parcel model with the CCN spectrum from observations to evaluate the role of giant CCN in producing warm rain. He calculated that droplets having diameters greater than 50 μm would reach concentrations of about 100 m^{-3} at 100 m above cloud base in a 2 m s^{-1} updraft (see Fig. 3). These large droplets were produced in concentrations comparable to raindrops and traceable to the giant CCN having a dry size larger than about 16-μm diameter. The existence of such "coalescence nuclei" just above cloud base is one of the simplest ideas for initiating the warm-rain process (Woodcock and Gifford 1949; Ludlum 1951; Keith and Arons 1954).

According to the extensive measurements of Woodcock (1953) in Hawaii and Florida, giant salt particles can often occur in significant concentrations (see Fig. 4). Woodcock's observations in the presence of modest surface winds of 5–10 m s^{-1} (force 3–5) show concentrations at elevations of cloud base (0.7–0.9 km) in the range of 100–1000 m^{-3} for salt particles with masses greater than 1 ng (corresponding to NaCl particles with diameters larger than 10 μm). Within the moist region below cloud base such salt particles will deliquesce into much larger solution droplets, for example, a NaCl particle of 1 ng (10 μm) has a 50-μm equilibrium diameter at 99% relative humidity. Thus, there are often significant concentrations of coalescence nuclei that can enter the base maritime clouds. [If the subcloud layer was dry, giant salt particles would not serve as coalescence nuclei because their growth by condensation is relatively slow (Ludlam 1951).]

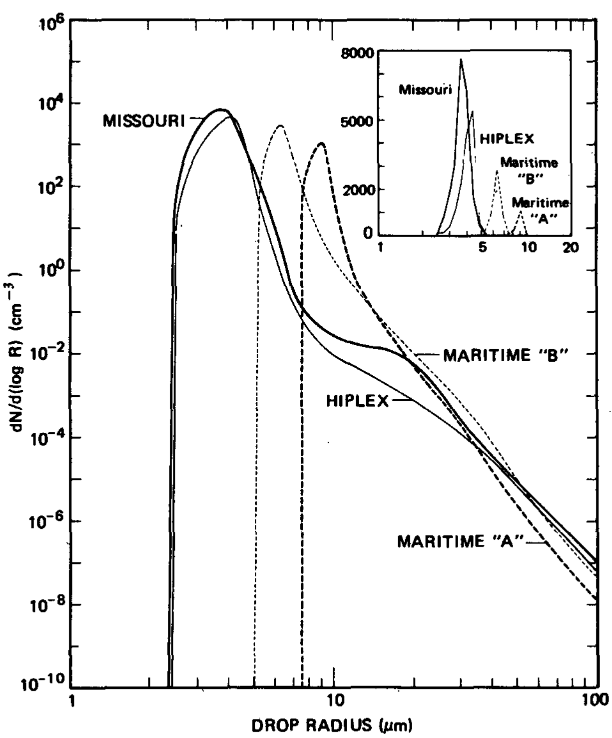

FIG. 3. Johnson's (1982) droplet size distributions for condensation in a parcel 100 m above cloud base for a 2 m s^{-1} updraft using CCN spectra based on observations.

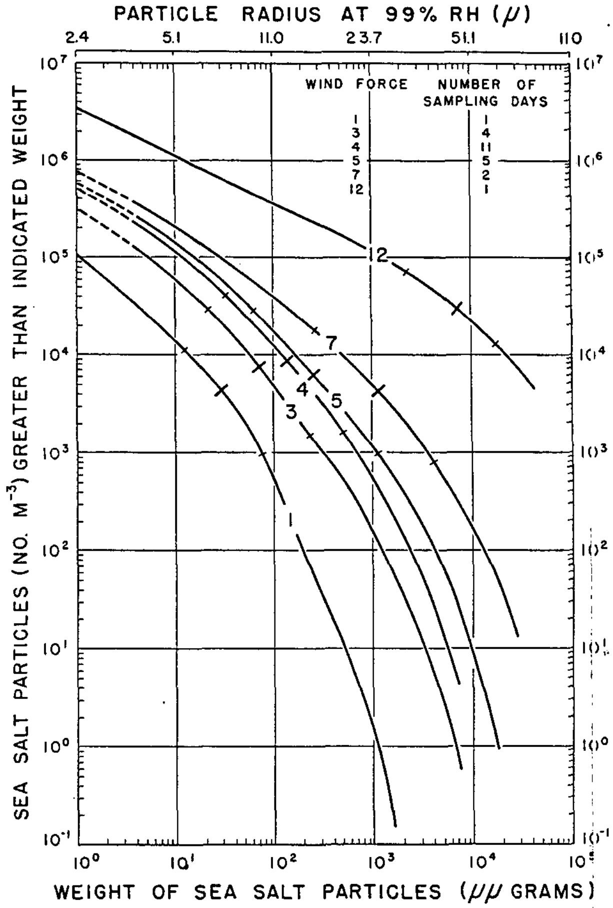

FIG. 4. Woodcock's (1953) distributions of sea salt particles over the sea in Hawaii area with curves for different wind force (force 12 "maximum" airborne salt from Florida tropical storm). The diameter classifications for dry NaCl particles are giant particles (0.2–2 μm, 10^1–10^4 $\mu\mu$g) and ultragiant particles (>0.2 μm, >10^4 $\mu\mu$g).

The concentration of giant salt particles is greatly reduced at higher elevations in the surface layer. Woodcock obtained measurements in Hawaii during surface wind up to 10 m s^{-1} at elevations above 1.5 km. His results show less than 10 m^{-3} salt particles having masses greater than 0.4 ng. This mass corresponds to a NaCl solution droplet of 30-μm diameter at 99% relative humidity. Thus, maritime clouds with bases higher than about 1.5 km, such as the orographic clouds in Hawaii, do not have significant concentrations of coalescence nuclei under typical wind conditions.

Woodcock et al. (1971) collected large and giant particles over the sea near Hawaii and found iodine to chlorine mass ratios of 10^{-3} and 10^{-4}, respectively. [The higher iodine in large particles seems to originate from organic material on the ocean surface (Blanchard 1964).] Woodcock et al. (1971) also measured the mass ratio in raindrops (10^{-3}) and rainwater (10^{-3}) collected at the base of clouds on the surface near Hilo (820 m above sea level). This finding is consistent with the growth of raindrops by accretion of cloud droplets formed on large particles. Rogers and Yau (1989) concluded from the study of Woodcock et al. that giant salt particles may not be essential for the formation of rain. In such an accretion process, however, the iodine-to-chlorine ratio is dominated by large particles even if the rain is initiated by giant salt particles because these size classes differ in mass by a factor only 10^3, whereas the number of cloud droplets collected during accretion is about 10^6. Thus, the findings of Woodcock et al. do not resolve whether or not rain is initiated on giant salt particles in such maritime clouds. A further evaluation of the role of giant nuclei in producing rain from maritime clouds will be given in the section on coalescence growth.

Measurements of particles in continental air often show high concentrations of giant and ultragiant particles (greater than 20-μm diameter; Johnson 1976) in the mixed layer that could serve as coalescence nuclei. For example, Hobbs et al. (1985) found variable concentrations of 50-μm-diameter particles (see Fig. 5), typically in the range of 100–1000 m^{-3}, from measurements taken during numerous flights around the High Plains in the mixed layer (1–3 km AGL). These ultragiant particles were mostly of soil origin, so they are relatively insoluble and will not grow appreciably as they enter cloud base. They are large enough, however, to make effective coalescence nuclei without any condensation growth (Ochs and Semonin 1979; Johnson 1982). The role of ultragiant nuclei as initiators of warm rain in continental clouds will be examined in section 3.

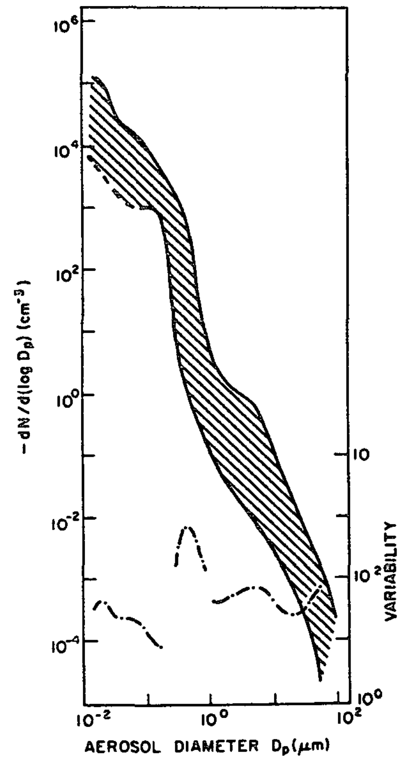

FIG. 5. Hobbs et al.'s (1985) particles concentration from flights around the High Plains in the mixed layer (1–3 km AGL) with variability denoted by dashed lines.

b. Mixing favoring the growth of large droplets

Condensation growth is often more complicated than an idealized, closed parcel in a uniform updraft—the process that produces a cloud-droplet distribution directly from the CCN spectrum. Because of turbulence, parcels are subjected to velocity fluctuations and mixing. Aircraft measurements of the vertical velocity in cumulus clouds typically show a relatively large variance in all but the updraft core. Models of the effects of velocity fluctuations on droplet spectra in a closed parcel, however, do not show appreciable broadening (Warner 1969; Bartlett and Jonas 1972) because the moisture available for droplet growth depends the change in height in a closed parcel rather than the growth time; the change in growth time between two heights is compensated by a change in growth rate (Manton 1979).

Droplet spectra produced by mixing of parcels of cloudy and clear air will lower the liquid water content by dilution, but its effect on the droplet distribution will depend on the mixing process. For example, Warner (1973) extended the model of Mason and Chien (1962) and showed that homogeneous dilution by entrainment of dry air containing CCN into a rising parcel produces a constant modal size but a decrease in mean size because of a broadening *toward smaller sizes* (e.g., see observed spectrum in Fig. 1). The small-drop tail was a consequence of nuclei being mixed in progressively. [Similar spectra near cloud base were calculated in the entrainment model of Lee and Pruppacher (1977).] In contrast to continuous-entrainment model results, the observations of maritime droplet spectra measured by Warner just above cloud base showed about five times the dispersion (standard deviation divided by mean diameter) and a tail for large drops rather than the small ones. Warner also found that the observed dispersion was uncorrelated with the mixing as determined by the measured fraction of adiabatic liquid water content. For both the observations and model results, he concluded that a simple homogeneous mixing process was unimportant in determining the droplet size distribution.

The mixing of parcels having differing supersaturation histories should lead to a broadening of the droplet spectrum because of different mode sizes. For instance, the simple addition of two spectra having differing ages but exposed to the same growth environment would be broader than either one alone. The outcome will depend on the mixing process that may include mixing with environment or with other parcels, as well as growth and decay. Mason and Jonas (1974), using a model of homogeneous entrainment and mixing of successive thermals, indeed found that bimodal spectra were produced when a second thermal developed. They showed agreement with one measurement of maritime spectra by Warner at 1.4 km above cloud base. This type of mixing, however, slows the growth of larger drops because the average supersaturation is lowered. Thus, their maritime case in absence of giant nuclei was slow, taking 30 min just to reach the coalescence stage where significant concentrations of 50-μm diameter were produced.

The notion of homogeneous mixing of parcels was brought into question in a series of papers by Latham, Baker, and coauthors stemming from the laboratory findings of Latham and Reed (1977). Mixing of undersaturated air into a laboratory cloud reduced the number of droplets in all categories without greatly affecting the shape of the size distribution. The mixing process was thought to be an inhomogeneous one whereby some portions of the cloud were completely evaporated by mixing with the unsaturated air and other portions were unaffected. Latham and Reed postulated that broadening of the droplet spectrum toward larger sizes would occur by subsequent condensation on the reduced number of activated droplets. Estimates by Baker and Latham (1979) based on idealized inhomogeneous mixing indicated an appreciable broadening toward larger droplets compared to homogeneous mixing.

There is ample evidence that mixing in clouds is incomplete from studies of High Plains cumuli (e.g., Jensen et al. 1985; Paluch and Knight 1986; Paluch and Baumgardner 1989). Although there are broad updrafts in these clouds with narrow droplet spectra, as would be expected for undiluted parcels, there are other regions with highly variable cloud properties. Such mixed regions appear to contain a finescale structure composed of clear air, undiluted cloudy air, and diluted air with partially evaporated droplet spectra. Paluch and Baumgardner found that mixed regions had droplets of highly variable concentrations down to the smallest scale (1 m) but that the large droplet peak remained at a nearly constant diameter. Paluch and Knight found that the large drop peak was nearly constant at one level but increased with altitude in a manner more consistent with a closed parcel environment than a completely mixed one. They did not find any evidence that mixing or cloud age increased the size or concentration of the largest drops. Paluch and Baumgardner found that the largest droplets occurred in regions with highest droplet concentration and water contents, not in parcels having low concentrations from entrainment. Thus, these studies of High Plains cumuli indicate that neither homogeneous nor inhomogeneous mixing of parcels enhances large droplet growth. [Note that Squires' negative correlation between maximum droplet size and mean droplet concentration (Fig. 2) was obtained from measurements in different types of clouds.]

The mixing of droplets, rather than parcels, having differing supersaturation histories should lead to a broadening of the droplet spectrum in both directions. This process would occur in regions of clouds having finescale structure (Paluch and Baumgardner 1989) with the variable droplet trajectories caused by sedimentation (Cooper et al. 1986). Analysis of variability of supersaturation in the High Plains cumulus clouds by Politovich and Cooper (1988), based on simultaneous measurements of the updraft speed and droplet size (i.e., the moisture source and sink), show a standard deviation in the supersaturation of 0.1% for unmixed regions and 0.4% for highly diluted regions. These findings indicate that there is enough variability in mixing regions to account for the observed dispersion of the droplet spectra.

Politovich and Cooper (1988) suggest that a few larger droplets in a mixed zone can grow much faster than average by falling through patches of air having excess supersaturation. Since the warm-rain process can be initiated by just a few cloud droplets, such favored growth histories can occur with very low probabilities (about 1:10 000, the ratio of the large cloud droplet to drizzle drop concentration). There was little evidence, however, for enhanced droplet growth in the shallow clouds studied in the High Plains because they decayed rapidly by entrainment.

The role of such stochastic condensation in producing large droplets was investigated by Cooper et al. (1986) using aircraft data taken in warm orographic clouds in Hawaii. Measurements of droplet spectra along parcel trajectories showed maximum droplet sizes of less than 25-μm diameter for laminar orographic clouds but a pronounced broadening of the spectra out to 80–120-μm diameter for orographic clouds with breaking waves at the top. The source of these larger droplets was not giant CCN since their concentration was much too low. Neither was it coalescence because calculations showed that the larger droplets could not have been produced even with unrealistically high collision efficiencies. The likely difference between the two sets of observations was the turbulent effect on condensation caused by the breaking waves.

Estimates made by Cooper et al. (1986) of the supersaturation for the breaking wave cases, based on simultaneous measurements of the updraft speed and droplet sizes (see Fig. 6), indicated a remarkably broad spectrum of values extending from −6% to +4%. (The supersaturation would be less than 1% for condensation in a closed parcel.) They noted that the mean supersaturation required to produce the observed 80-μm droplets was about 2%. By assuming a parcel in equilibrium, they determined that the supersaturation would be proportional to the updraft speed and inversely proportional to the mean diameter times the droplet concentration I. Since I varied more than an order of magnitude in both updrafts and downdrafts, it was possible for a larger droplet to grow rapidly at high supersaturation in updrafts containing few droplets (low I) and evaporate slowing in downdrafts containing many droplets (high I). Thus, the authors suggested that the larger droplets in the breaking wave cases could have resulted from favorable growth trajectories through regions where I is negatively correlated with the updraft. Recently, Cooper (1989) further developed the theory of stochastic condensation. He concluded that although more observations are needed to better evaluate the theory, special dynamical sequences of vertical mixing (Telford et al. 1984) and inhomogeneous mixing (Baker and Latham 1979) would not be needed to explain spectral broadening or large droplet growth during the condensation phase of the warm-rain process.

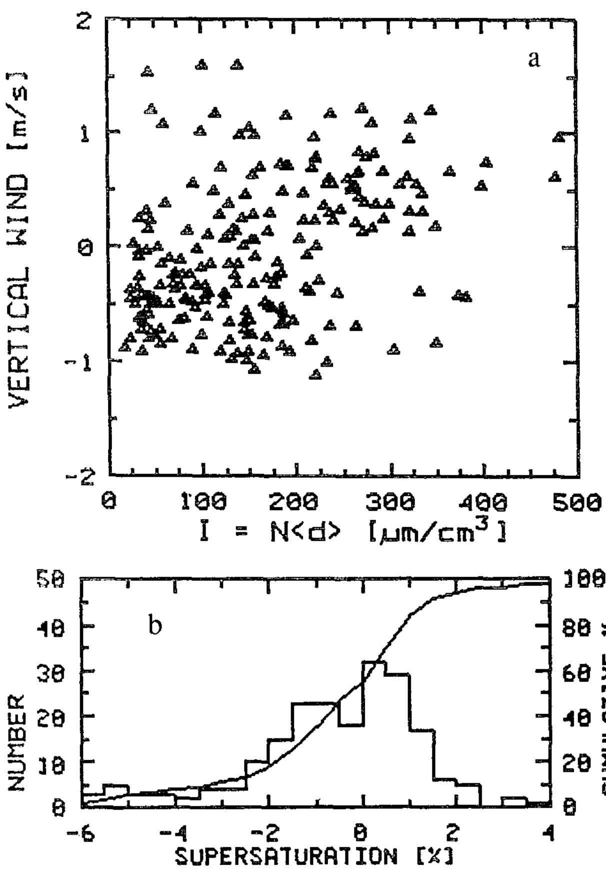

FIG. 6. Cooper et al.'s (1986) condensation parameters as determined from aircraft measurements in a region of a Hawaiian orographic cloud of breaking waves: (a) scatterplot of vertical wind versus integral diameter (mean diameter times droplet concentration) and (b) resulting supersaturation spectrum.

c. Summary on activation and condensation growth

Maritime clouds typically have fewer but larger cloud droplets than continental clouds as a consequence of fewer CCN, with a strong negative correlation between droplet concentration and maximum diameter (Squires 1956). It has been deduced from such observations that giant nuclei play no role in producing large cloud droplets, since the concentration of giant nuclei usually increases with CCN concentration, for example, as the wind increases over the ocean (Woodcock et al. 1953) or as air becomes more continental (Hindman et al. 1977). This conclusion is appropriate only for droplets measured at relatively high concentrations (1–100 cm^{-3}) compared to precipitation (100–1000 m^{-3}).

The creation of precursor raindrops, that is, the few larger droplets at concentrations similar to raindrops, should result from the ingestion at cloud base of giant and ultragiant particles. Since such particles are often found at 100–1000 m^{-3}, at least in the surface layer (Woodcock 1953; Johnson 1976; Hobbs et al. 1985), they should initiate coalescence just above cloud base (Ochs 1978; Ochs and Semonin 1978; Johnson 1982). The role of coalescence nuclei has not been directly verified by measurements because the sample volume for measuring cloud droplets is much too small—typically much less than 100 cm^3 per cloud traverse—so that the largest sizes occurring at less than 10 L^{-1} are not even detected!

Another source of large cloud droplets is the favorable condensation on a few droplets that fall through patches of high supersaturation in mixed regions. Observations implicating this mechanism were obtained in orographic clouds (Cooper et al. 1986), and subsequent theory indicated that such favored condensation may be a major cause of the broadening of the cloud-droplet size distributions (Cooper 1989). Mixing, however, can also increase colloidal stability because it broadens the spectra toward smaller sizes and lowers, by dilution, the droplet concentration and liquid water content. If precipitation develops in undiluted updrafts, then the only mechanism for reducing colloidal stability must be inherent in the nucleus spectrum, for example, giant and ultragiant particles.

3. Droplet growth by coalescence

The controlling parameters for coalescence growth are the droplet sizes. Factors such as electric charge, turbulence, temperature, and pressure, will be neglected since they would usually play a secondary role during the initial stage of coalescence growth. The particular outcome of two droplets interacting in free fall depends on their sizes and on their initial horizontal offset x. A simple result occurs for cloud droplets: they coalesce for more direct impacts but do not make contact for larger offsets. The critical offset that separates the two outcomes (x_c) defines the collection cross section (πx_c^2) and the collection efficiency $\mathscr{E} = 4x_c^2/(D+d)^2$, from the ratio of the collection cross section to the geometric cross section for large and small droplet diameters of D and d. The collection efficiency is a critical parameter for the coalescence growth since it gives the fraction of small cloud droplets that are collected in the path of the larger cloud droplet.

Theoretical calculations are based on the "simplified" hydrodynamics of interacting solid spheres, whereby a critical offset is calculated to define the collision efficiency E. (Whether or not collection will occur is discussed below.) The most complete theory for small droplets (Klett and Davis 1973) shows that the collision efficiency is negligible (<1%) for $D < 15$ μm and $d < 8$ μm and appreciable (>10%) for $D > 50$ μm and $d > 15$ μm (see Fig. 7). More recent calculations on the effects of short-range forces (e.g., Rogers and Davis 1990) have not significantly altered these efficiencies. Since most droplets in newly formed clouds are in the range of 8–15-μm diameter, $D = 50$ μm is an appropriate threshold size for coalescence growth, based on theoretical collision efficiencies. There are no reliable laboratory studies for $D < 80$ μm, however, so the critical efficiencies for the coalescence threshold remain untested (see Fig. 8).

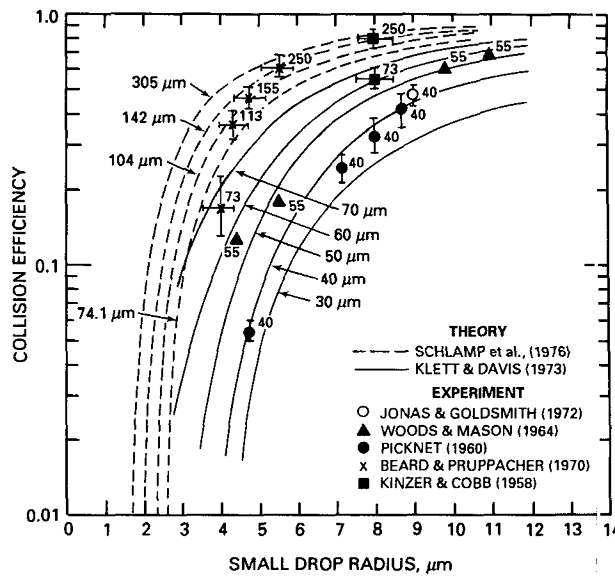

FIG. 8. Comparison of experimental collection efficiencies with theoretical collision efficiencies as a function of small droplet radius up to 12 μm for indicated large drop radius from 30 to 305 μm (after Pruppacher and Klett 1978).

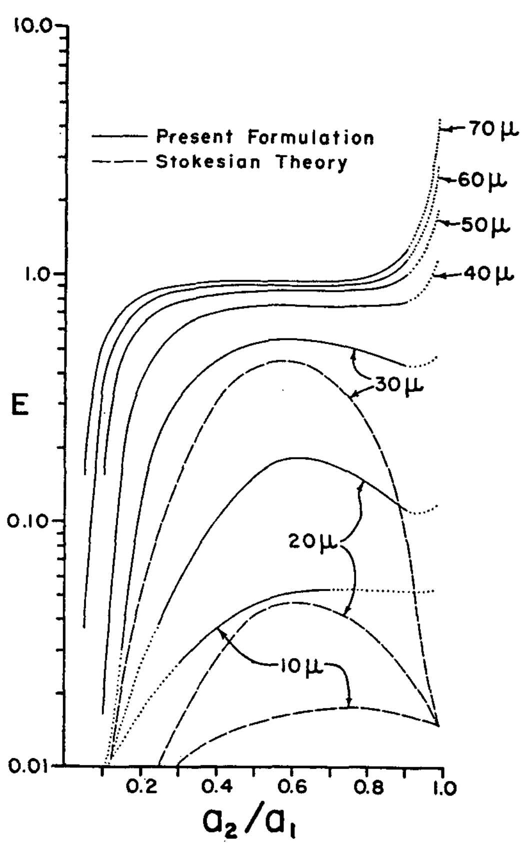

FIG. 7. Klett and Davis' (1973) theoretical collision efficiencies (solid lines) as a function of radius ratio a_2/a_1 for different large droplet radii a_1. Also shown (dash lines) are the earlier results of Davis and Sartor (1967).

Laboratory studies of droplet interactions have revealed that above this threshold the measured efficiency is less than the collision efficiency as shown in Fig. 9 (Beard and Ochs 1983; Ochs and Beard 1984). This effect is caused by droplet deformation, a complication not treated in the theory. Droplets in a certain range of horizontal offsets that should collide, according to solid sphere hydrodynamics, do not make contact because of additional air resistance between the deformed surfaces. The fraction of collisions that result in coalescence is termed the *coalescence efficiency*, $\epsilon(D, d)$, so that the measured "collection" efficiency $\mathscr{E}$ is related to the theoretical collision efficiency E by $\mathscr{E} = \epsilon E$. The semiempirical formula for coalescence efficiency, based on drop deformation (Beard and Ochs 1984), shows that ϵ decreases with increasing drop sizes from 100% for $D < 50$ μm and $d < 10$ μm to less than 50% for $D > 800$ μm and $d > 40$ μm. At still larger sizes, the collisions become more violent, resulting in various types of drop disintegration. In spite of reduced coalescence, the production of drizzle and small raindrops by the accretion process is rather effective since collection efficiencies are relatively high for most of the in-

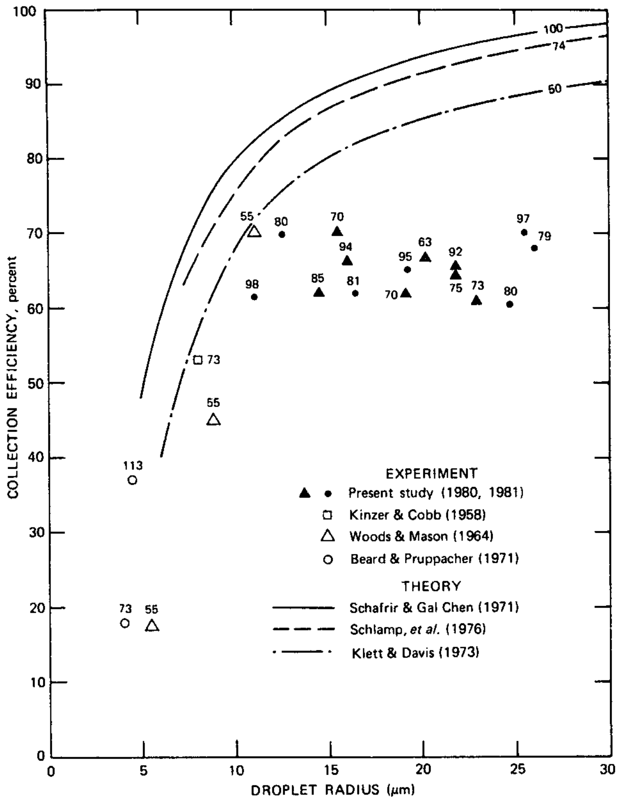

FIG. 9. Beard and Ochs' (1983) comparison of experimental collection efficiencies with theoretical collision efficiencies as a function of small droplet radius. The label for each curve gives large droplet radius in micrometers.

teractions, for example, $\mathscr{E} > 50\%$ for D = 100–1000 μm and d = 12–20 μm (see Fig. 10).

a. Calculations of coalescence growth

Two methods are generally used for computing growth rates of cloud droplets by coalescence. The simplest one is based on the collection of cloud droplets by a single, larger one. This is usually called *continuous collection* since growth occurs smoothly rather than by discrete collection events. The continuous model is only applicable for droplets large enough to retain their identity, that is, those having a negligible probability of being collected by other droplets. For continuous collection, the mass growth rate of a larger droplet, summed over smaller droplets, is given by $dM/dt = \Sigma[K(D, d)w_L(d)\Delta d]$, where K is the collection kernel and w_L is the liquid water content (per size interval Δd). The collection kernel is the volume of air (containing the small droplets) swept out per unit time by the large droplet given by $K = \mathscr{E}(\pi/4)(D + d)^2[V(D) - v(d)]$, where V and v are terminal velocities. The liquid water content is given by $w_L = n(d)(\pi/6) \times \rho_L d^3$, where n is the droplet concentration (per size interval Δd) and ρ_L is the density of water.

Continuous collection in a uniform updraft was calculated by Bowen (1950) using a cloud containing droplets of d = 20 μm at a liquid water content of W_L = 1 g m^{-3} (see Fig. 11). The growth rate in an updraft of 1 m s^{-1} for an initial droplet of D = 25 μm was found to be very slow, taking about 45 min and 2-km rise to achieve 250-μm diameter (1000 collections). Once this size was reached, the subsequent growth to raindrop sizes was more rapid (about 15 min). The Bowen model was updated by Rogers and Yau (1989) with recent collection efficiencies and an initial droplet of D = 40 μm. Growth in updrafts of 0.5–1.0 m s^{-1} reached drizzle sizes (≈150–250-μm diameter) in about 50 min at the apex of the trajectory. The subsequent coalescence growth was more rapid so that small raindrops (800–1800-μm diameter) emerge from cloud base. These calculations, using modern collision efficiencies, show that the continuous collection by droplets of D < 40 μm is relatively slow.

Braham (1968) computed continuous growth for the largest droplets at about 10 L^{-1} based on observations in cumulus clouds (see Fig. 12). For an average continental cloud in the southwest (W_L = 0.3 g m^{-3}), it took over 2 h for an initial size of D ≈ 60 μm to reach small raindrop size (D = 0.6 mm), whereas the growth in an average Caribbean cloud (W_L = 0.8 g m^{-3})

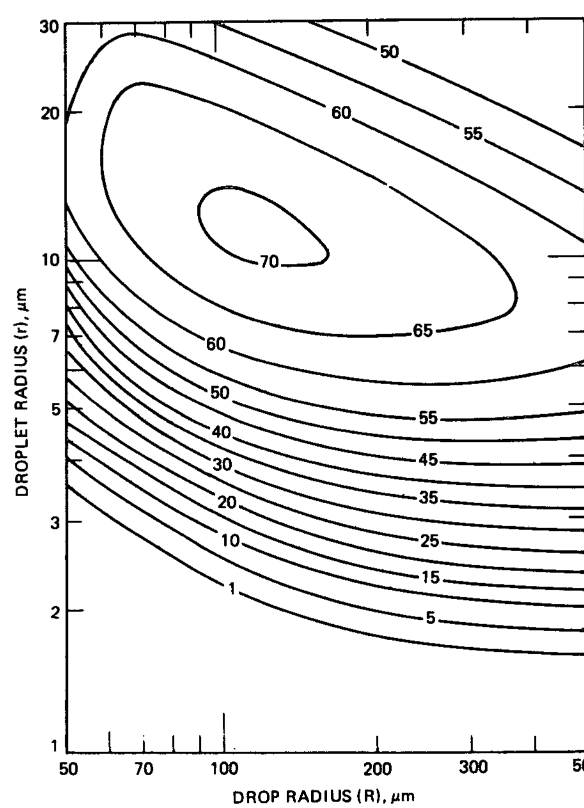

FIG. 10. Beard and Ochs' (1984) contours of semiempirical collection efficiency (ϵE) as a function of large (R) and small (r) droplet radii.

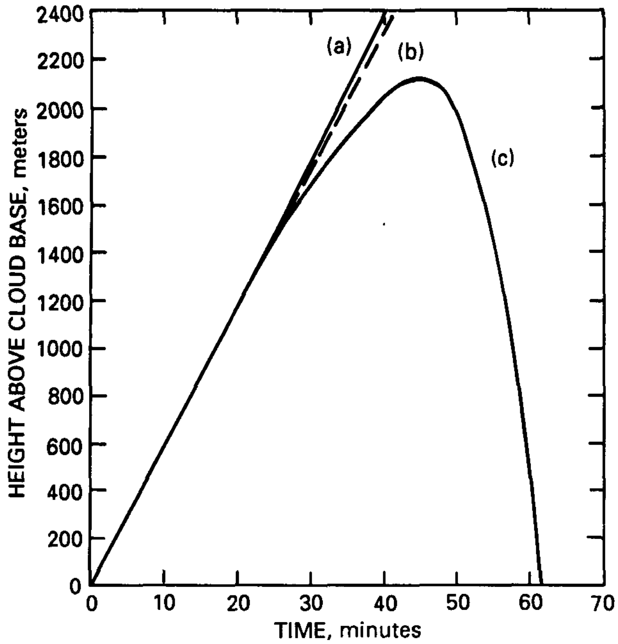

FIG. 11. Bowen's (1950) calculations showing the trajectories of (a) the air for an updraft of 1 m s^{-1}, (b) the 20-μm diameter cloud droplets, and (c) an initial 25-μm-diameter droplet grown by continuous collection.

took only 35 min for an initial size of $D \approx 80$ μm to reach 0.6 mm. For higher water contents (W_L = 1 g m^{-3}), these growth times would reduce to 1 h for the continental cloud and 20 min for the maritime cloud to produce small raindrops. Braham's calculations illustrate that initial droplets above $D = 60$ μm will speed up coalescence growth significantly and lead to the rapid onset of warm rain.

Coalescence growth was evaluated in a continuous model using collection efficiencies derived from laboratory experiments (Beard and Ochs 1984). For a droplet size of D = 100 μm with collected droplets of one size at constant W_L = 1 g m^{-3}, coalescence growth was found to be negligible for $d < 10$ μm because of low collision efficiencies but appreciable and practically independent of droplet size for $d > 15$ μm with a collection efficiency of 50%–65% (see Fig. 13). They found that coalescence growth is relatively rapid once very large cloud droplets are formed, that is, 20–30 min for the growth from an initial size of D = 100–200 μm to a final size of D = 1 mm. If rain were initiated at higher water contents (3–5 g m^{-3}), as in adiabatic cores of cumulus congestus several kilometers above cloud base, it could take less than 10 min to reach D = 1 mm. Thus, coalescence growth to raindrop size can be rapid in high liquid water contents, providing very large cloud droplets ($D \geqslant 100$ μm) are present.

The coalescence process can also be accelerated under more modest conditions by a few cloud droplets with low collection efficiencies ($D < 50$ μm) that experience a much higher than average growth rate. The variance in growth rate occurs because the time between collections in a real, discrete process is a random variable with the mean given by the continuous process. Such randomness can be modeled using the probability that a particular drop mass m_1 will collect another drop mass m in a small time interval δt, given by $P(m_1, m) = K(m_1, m)n(m)\Delta m \delta t$, where K is the collection kernel and $n(m)\Delta m$ is the drop concentration in a size interval Δm. A value of P = 0.1 for discrete collection means that out of ten interactions between m_1 and m, only one on the average will result in coalescence. In the continuous approach, these events are spread among all drops of size m_1 so that they each grow by just $m/$

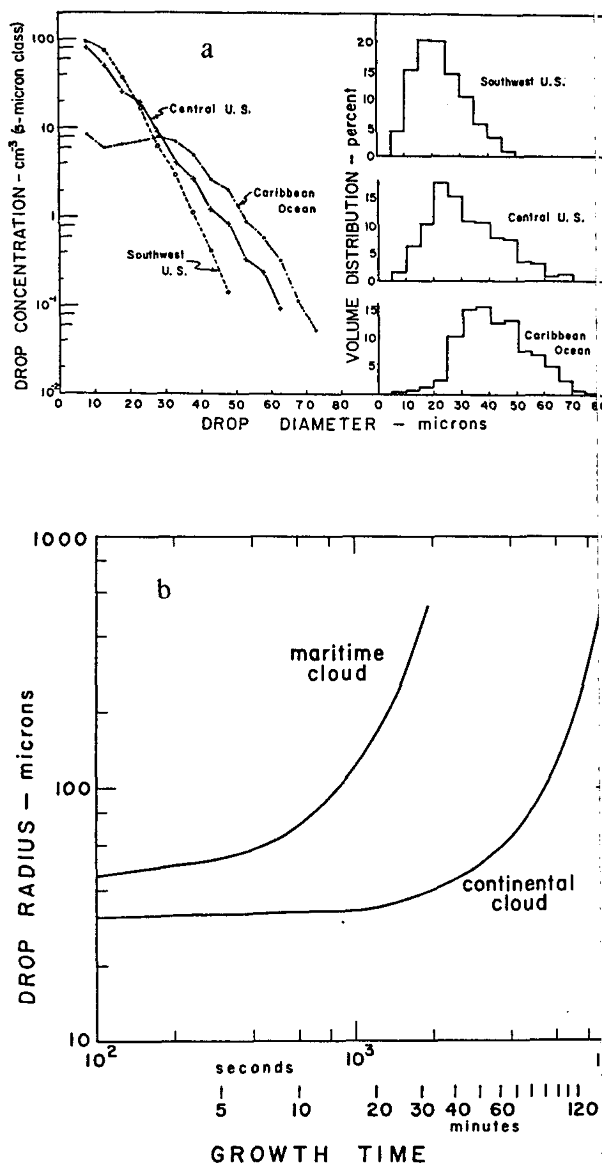

FIG. 12. Braham's (1968) droplet spectra and growth curves: (a) droplet size distributions for cumulus congestus in three regions and (b) growth of a large cloud droplet by continuous collection in two spectra.

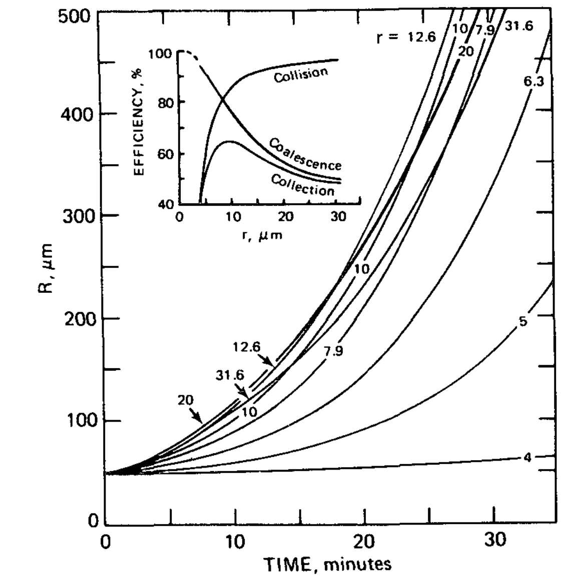

FIG. 13. Beard and Ochs' (1984) accretion growth based on semiempirical collection efficiencies ϵE for a monodisperse cloud for eight different droplet sizes. The inset shows the average efficiencies for accretion as a function of droplet size.

10. Thus, the most fortunate droplet of initial size m_1 in discrete collection can grow much faster than continuous collection.

Telford (1955) determined the variability of times for 10, 20, and 30 capture events in the discrete process starting with $m_1 = 2m$. To evaluate the formation of large droplets, he determined the growth rate of the fasting growing one in 10^5 droplets, corresponding approximately to the concentration of precipitation ($100\ m^{-3}$). He found from analytic solutions that these rare droplets grew six times faster by discrete collection than by continuous growth. Additional calculations of discrete collection were carried out by Robertson (1974) using a Monte Carlo procedure with more recent collection efficiencies for initial sizes of collector droplets. Because of a limited number of simulations, Robertson could not evaluate the fastest growth times for the fraction of droplets at the concentration of precipitation, but he did determine that growth by discrete collection varied significantly from the continuous process for droplets less than $D = 80\ \mu m$.

A more general method for evaluating coalescence growth is the use of the collection probability P to determine the number of droplets of each size contributed by collection between smaller sizes and the number that are lost by collection with all other sizes. This approach is termed *stochastic collection* since the temporal changes are controlled by collection probabilities. The evolution in the droplet spectrum is calculated from the change in the concentration of a particular drop mass from gains by coalescence of smaller pairs and losses by coalescence with other sizes. The stochastic calculation is made for each drop size and then stepped forward in time. This model gives the *mean* evolution of drop size distribution by coalescence among all drop sizes for a homogenous population of droplets, for example, evolution in a high enough column of droplets so that the fall distances are negligible during the calculation. The variability in the evolving spectra, as calculated by a probability model or a Monte Carlo scheme, is relatively small, as is the effect of random spatial distributions (Young 1975) and higher-order collisions (Pruppacher and Klett 1978). Thus, the mean stochastic process has been determined to be a reliable indicator of droplet growth across the spectrum by coalescence.

Analytic solutions to the stochastic collection equation have been obtained only for greatly simplified collection kernels. Numerical integration can be used, but considerable care must be taken to prevent numerical diffusion. If not controlled, numerical spreading will exceed the natural growth process by producing significant masses in the drizzle and raindrop categories, resulting in artificially rapid production of warm rain (Reinhardt 1972). The standard check of numerical collection is to compare results with analytical solutions based on simplified kernels (e.g., Golovin 1963). Good agreement, however, does not guarantee accuracy for realistic collection kernels. Some considerable efforts have been made to prevent numerical spreading by using tailored integration schemes for specific collection kernels (Berry and Reinhardt 1974) and more general methods (e.g., Young 1975; Ochs and Yao 1978). Thus, most earlier work with stochastic collection is confounded by numerical inaccuracies and, in particular, anomalous spreading of the droplet spectrum.

Berry and Reinhardt (1974) used special numerical integration and interpolation schemes to produce an accurate and comprehensive analysis of stochastic collection for the warm-rain process based on the collision efficiencies of Davis and Sartor (1967) and Shafrir and Neiburger (1963). The droplet sizes were specified by gamma distributions of the masses having a liquid water content of $1\ g\ m^{-3}$ with the mean sizes of 20–36-μm diameter (see Fig. 14). For spectra having mean sizes of greater than 24-μm diameter, a small but significant mass of droplets reached drizzle drop size (200 μm) within 10–22 min, with time decreasing for increasing mean size. In the case with a mean size of 20 μm, growth took well over 30 min. After reaching drizzle size, growth was relatively independent of the droplet spectrum, taking about 12 min more to grow to 1.0-mm diameter. The studies of Berry and Reinhardt indicate that warm-rain initiation can be fairly rapid (≤ 15 min) for liquid water contents greater than $1\ g\ m^{-3}$ from initial distribution having mean sizes greater than 30-μm diameter containing droplets out to 60-μm diameter.

Ryan (1974) evaluated the sensitivity of stochastic collection for the growth rate of the 100 largest droplets in a continental cloud ($200\ cm^{-3}$) with a gamma distribution and $W_L = 1\ g\ m^{-3}$. He found that the growth

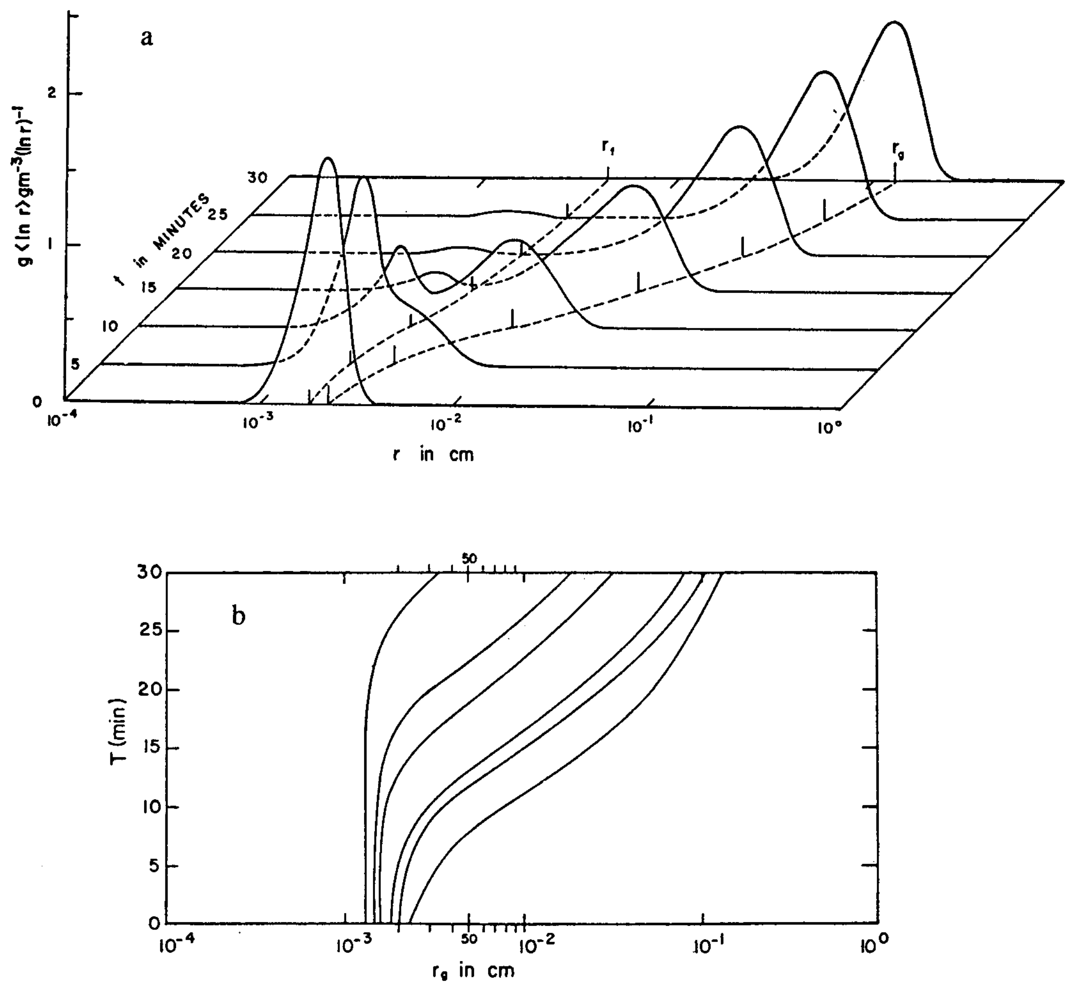

FIG. 14. Berry and Reinhardt's (1974) evolution of droplet spectra by stochastic collection: (a) time evolution of an initial size spectrum r_f^0 = 24 μm (radius of mean mass) and σ = 1.0 (relative mass variance) and (b) time change of r_g (radius of mean-square mass) in sequence from left to right for r_f^0, σ (10 μm, 1.0; 14 μm, 0.25; 12 μm, 1.0; 14 μm, 1.0; 18 μm, 0.25; 18 μm, 1.0). Note that r_f follows the cloud-droplet peak, whereas r_g follows the precipitation drop peak.

rate increased by 50% using the collection efficiencies of Klett and Davis (1973), compared with the earlier ones of Hocking and Jonas (1970), and by 500% using geometric efficiencies. When the collision efficiencies were enhanced by shear, using a scheme based on Manton (1974), the further increase in growth rate was negligible (<10%). In the subsequent stochastic calculations of Ameida (1979), the turbulence was erroneously enhanced by applying an inertial subrange scaling to interactions in the viscous subrange (Pruppacher and Klett 1978).

Improved efficiency calculations in the viscous subrange by Grover and Pruppacher (1985) have indicated a negligible effect of the vertical velocity fluctuations on collision efficiency for droplet sizes greater than 8-μm diameter at high turbulent intensities. Thus, warm-rain initiation is apparently unaffected by such one-dimensional turbulence. More recently, Reuter et al. (1988) have evaluated the effect of turbulent velocity fluctuations on the collection kernel by the enhanced probability that two droplets will approach each other. This research has been criticized by Cooper and Baumgardner (1989) because the droplet approach was greatly exaggerated by the use of a constant diffusion coefficient with an inertial subrange formulation. Corrected efficiencies were found to be enhanced by only 5% for D = 100 and d = 20 for the extreme turbulence found in thunderstorms. At present, questions about the effects of turbulence on warm-rain initiation remain largely unanswered because reliable efficiencies are yet unavailable for stochastic model calculations.

b. Calculations of warm rain initiated by giant nuclei

The possible significance of the giant nuclei in warm-rain initiation was not evaluated in the earlier stochastic collection studies since the droplet distributions were prescribed rather than determined from the CCN spec-

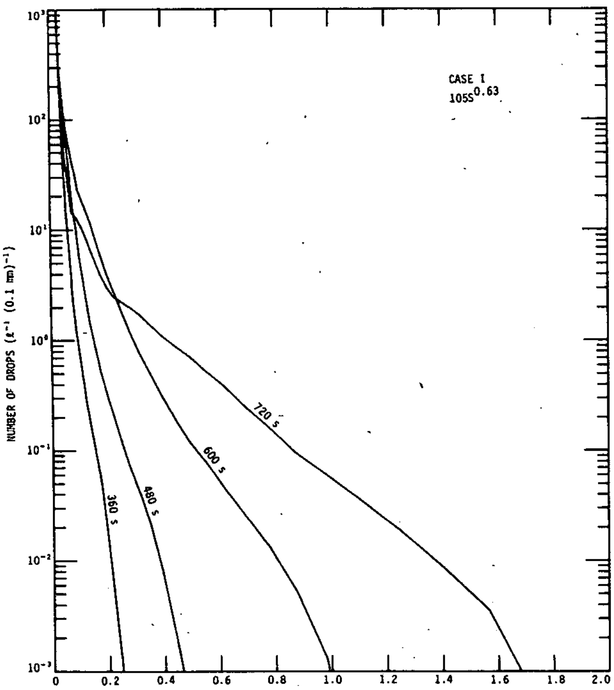

FIG. 15. Ochs' (1978) evolution of a drop spectra by stochastic collection in a parcel for maritime CCN with a 4 m s^{-1} updraft.

trum (extending to 25-μm diameter and beyond in the cases investigated by Berry and Reinhardt). Recall that theoretical studies of colloidal stability indicated that the onset of coalescence growth was retarded by high concentrations of small CCN and unaffected by the concentrations of giant particles (i.e., Squires 1952b and subsequent studies). It is obvious, however, that this is not the complete answer since larger cloud droplets will form on larger particles irrespective of nucleus concentration. (In a closed parcel it is impossible for one salt-particle size to overtake another; the largest droplets form on the largest nuclei—and at the lowest supersaturation.) Therefore, particle concentrations as low as that of raindrops (100—1000 m^{-3}) must be considered in evaluating the onset of coalescence growth. For instance, in maritime air, the concentration of NaCl particles having dry diameters greater than 10 μm would have solution diameters of greater than 50 μm in equilibrium at 99% relative humidity and therefore should serve as coalescence nuclei.

Ochs (1978) evaluated warm-rain formation using a parcel model of condensation and stochastic collection with a moment-conserving technique to prevent anomalous spreading. He used a maritime distribution containing 50-μm-diameter salt particles at a concentration comparable to raindrops (1 L^{-1}). Since such ultragiant particles are large enough to coalesce with smaller droplets without deliquescence, they should readily initiate rain above cloud base providing the cloud droplets exceed about 10-μm diameter. Ochs, indeed, found that condensation on the maritime CCN spectrum plus collection produced a significant concentration of drizzle drops (200-μm droplets at 1 L^{-1}) in about 6 min (about 1 km above cloud base, as shown in Fig. 15). Since the collection process was stochastic, it was impossible to assess the direct effect of such ultragiant particles on precipitation formation.

The development of warm rain for a continental CCN distribution, containing ultragiant particles at the same concentration as the maritime spectrum, was evaluated using two collection kernels, one based on the best available collision efficiencies (Jonas 1972; Klett and Davis 1973; Shafrir and Gal-Chen 1971; Beard and Grover 1974), and the other based on earlier, generally lower collision efficiencies (Hocking and Jonas 1970; Shafrir and Neiburger 1963). The formation of precipitation (20 dBZ) was accelerated from

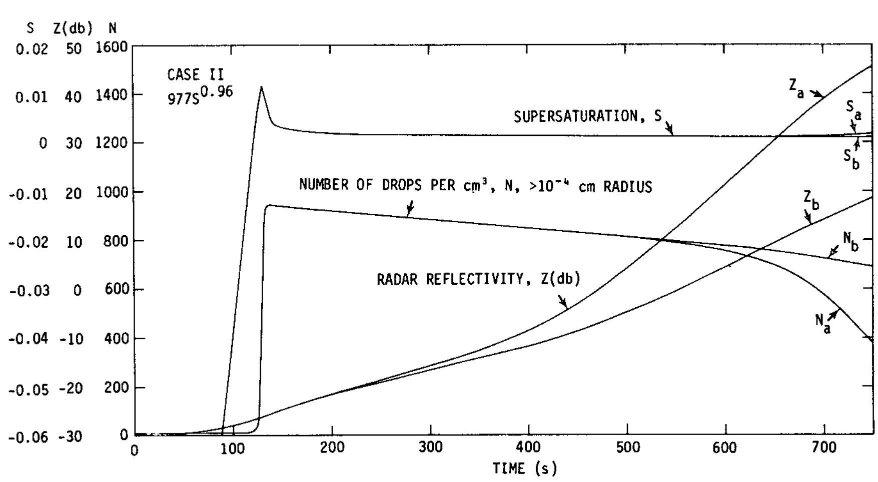

FIG. 16. Ochs' (1978) evolution of cloud properties in a parcel model for continental CCN with a 4 m s^{-1} updraft for (a) recent collision efficiencies and (b) earlier collision efficiencies (see text).

12.5 to 10 min using the more recent efficiencies (see Fig. 16). In the absence of ultragiant particles, initiation would be expected to take significantly longer.

In a subsequent study using the same parcel model, Ochs and Semonin (1979) evaluated the role of giant and ultragiant particles (composed of ammonium sulfate) in warm-rain formation for a continental CCN spectrum. They found that the time to form drizzle and raindrops, as measured by a computed radar reflectivity factor, was dramatically lengthened when the salt particle size was truncated below 50-μm diameter (as shown in Fig. 17). (There were too few particles larger than 50 μm to affect the droplet evolution significantly.) It took about 24 min (1.4 km) to form rain (20 dBZ) with the CCN truncated at 50-μm diameter and about 34 min (2.0 km) when the spectrum was truncated at 10 μm. It only took 16 min (1.0 km) to form drizzle (-20 dBZ) with the CCN truncated at 50-μm diameter and about 30 min (1.8 km) when the spectrum was truncated at 10 μm. Thus, in this particular study, giant and ultragiant salt particle were found to serve as coalescence nuclei that accelerated the initiation of warm rain.

Johnson (1982) noted that ultragiant particles are large enough to undergo coalescence growth at cloud base and that ultragiant particles seem to be a regular feature of the aerosol spectrum (Nelson and Gokhale 1968; Johnson 1976; Hindman et al. 1977). From the available aircraft observations, he developed aerosol distributions that extended out to 200-μm diameter. The typical maritime concentrations for a particle diameter of 50 μm were 1 m^{-3} (somewhat higher than generally found by Woodcock), whereas the typical continental concentration for the same size was 100 m^{-3}. More recent measurements of Hobbs et al. (1985) from numerous flights around the High Plains showed even higher concentrations of 1000 m^{-3} for 50-μm diameter particles.

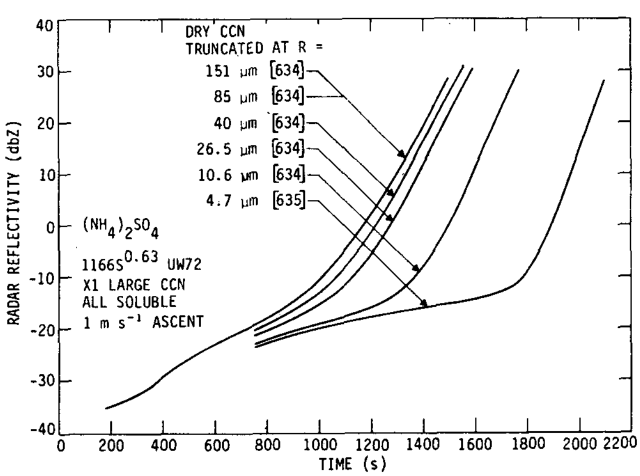

FIG. 17. Ochs and Semonin's (1979) evolution of the radar reflectivity factor in a parcel model for continental CCN with a 4 m s^{-1} updraft showing effect of truncated CCN spectra.

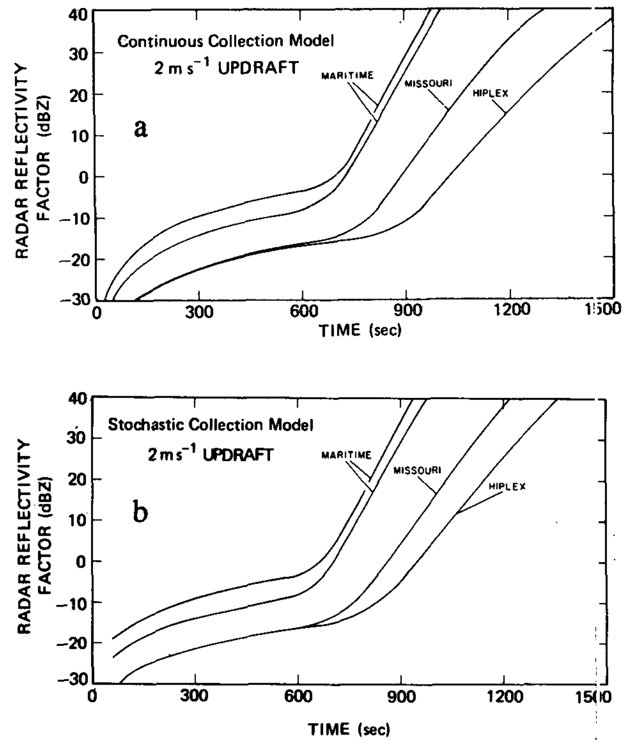

FIG. 18. Johnson's (1982) evolution of the radar reflectivity factor in a parcel model for four CCN spectra with a 2 m s^{-1} updraft: (a) continuous collection and (b) stochastic collection.

Johnson obtained the droplet size distributions near cloud base by modeling condensation using four CCN spectra and base temperatures: two maritime CCN at 15°C, Missouri continental CCN at 10°C, High Plains CCN at 2°C (see Fig. 3). The activation of droplets was calculated using a model similar to Fitzgerald (1972) with a large number of size and solubility categories (about 300 total per CCN distribution). The results for a linear plot of concentration had the narrow size distributions of activated droplets found at cloud base (Fig. 1), typically a few micrometers in width with mode sizes less than 20-μm diameter. These distributions also contained larger droplets (Fig. 3) formed on ultragiant CCN that could serve as coalescence nuclei. The highly soluble maritime CCN doubled in size by deliquescence near cloud base, whereas the relative insoluble continental CCN did not grow appreciably. The model results for cloud-base droplet distributions were used as a starting point for determining the role of ultragiant CCN on the subsequent growth by coalescence.

The growth of larger droplets was compared for the different cloud-base distributions from the evolution of the radar reflectivity in a parcel rising at 2 m s^{-1} undergoing condensation and continuous collection (see Fig. 18a). The importance of each large droplet category to warm-rain formation was evaluated (see Fig. 19a) by determining its contribution to radar reflectivity at the point where it reached a significant echo

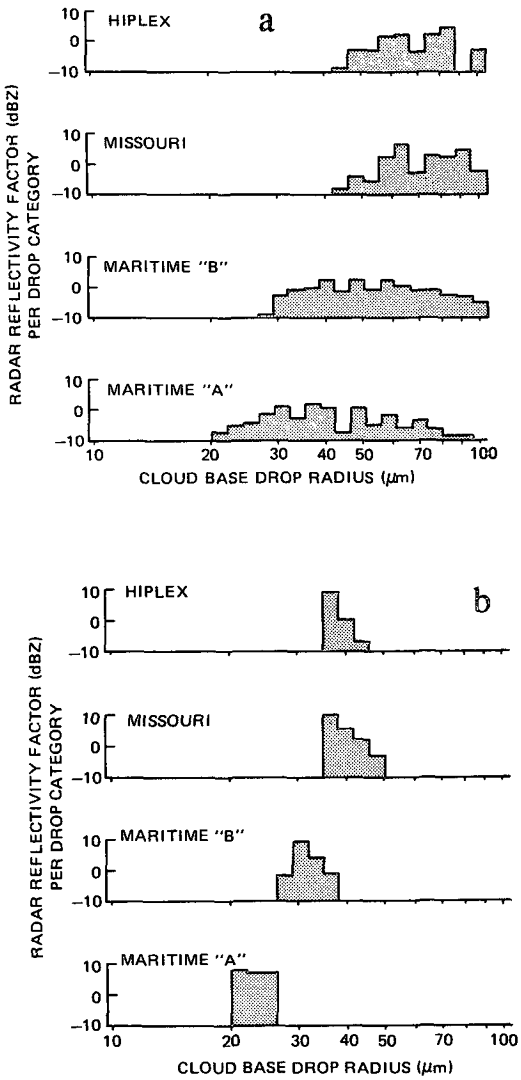

FIG. 19. Johnson's (1982) cloud-base droplet categories that contribute to the total radar reflectivity factor at precipitation onset (10 dBZ) in a continuous collection parcel model with an updraft of 2 m s^{-1}: (a) without sedimentation and (b) with sedimentation.

of 10 dBZ. For the maritime distributions, most of the contribution was from droplets with cloud-base diameters greater than about 70-μm diameter but less than 150 μm (corresponding to ultragiant CCN in the range 35–75 μm diameter). For the continental distributions, the predominant contribution to the radar reflectivity factor was from somewhat larger droplets having cloud-base diameters between about 120–160 μm (corresponding to ultragiant CCN of about the same size). Coalescence growth was also calculated using a stochastic model of Young (1975), with results for the evolution of the echo strengths (see Fig. 18b) similar to the continuous model, indicating the dominance of ultragiant nuclei in initiating and developing warm rain.

The effect of sedimentation on warm-rain formation was also assessed by Johnson using his continuous model. The spectrum of cloud droplets was assumed to be constant at any level, determined from condensation of a specified fraction of available (adiabatic) moisture. Since the concentration of each large-droplet

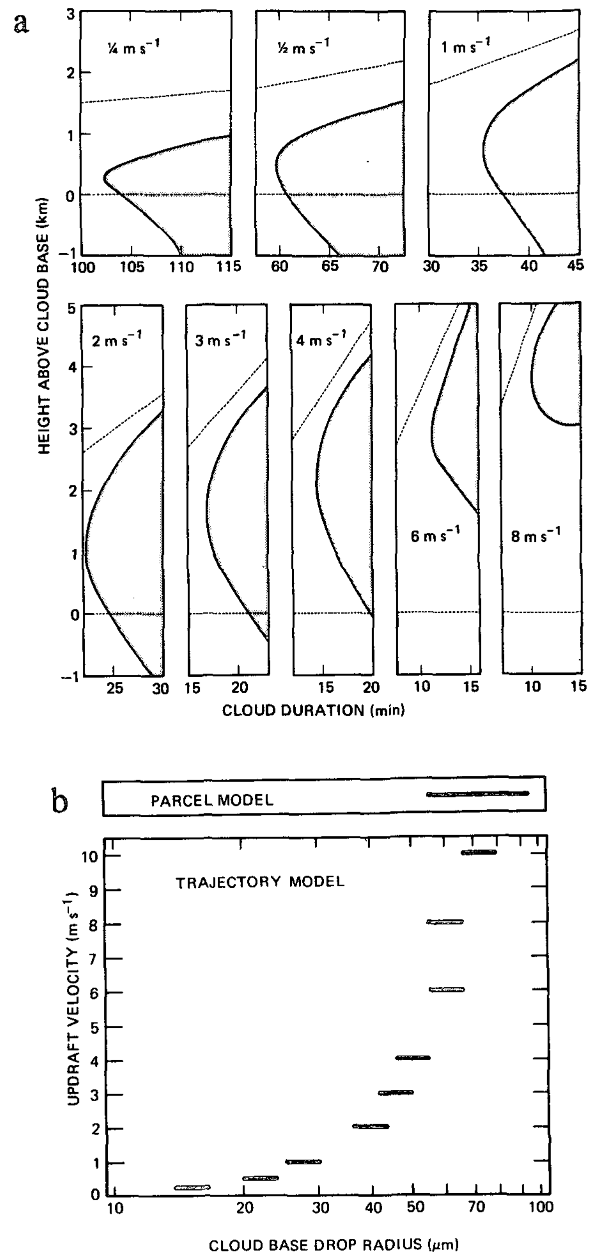

FIG. 20. Johnson's (1982) result for CCN in his continental Missouri case using a continuous collection parcel model with sedimentation: (a) evolution of radar reflectivity factor in excess of 10 dBZ (shaded area) for eight updraft speeds (cloud bases and tops indicated by dash lines) and (b) cloud-base droplet sizes contributing most to reflectivity at the location of precipitation onset (10 dBZ) for nine updraft speeds (trajectory model) in comparison to model without sedimentation (parcel model).

category could be determined as a function of time and height, it was possible to construct a time–height contour of radar reflectivity as a measure of rain initiation (see Fig. 20a). From his sedimentation results, Johnson determined the droplet sizes at cloud base that contributed the most to reflectivity (see Fig. 20b). Although the results varied with updraft speed, ultragiant particles were the source of rain initiation in the maritime and continental distributions with appreciable updrafts. For example, with a 2 m s^{-1} updraft, the sizes at cloud base contributing most to the reflectivity (see Fig. 19b) were in the range 40–70-μm diameter for the maritime distribution and 70–100 μm for the continental distributions. The sizes are larger at higher updraft speeds as shown in Fig. 20b for the continental Missouri case.

The modeling studies of Ochs and Johnson demonstrate that warm rain can be readily initiated if giant and ultragiant particles are present in the CCN distributions. These studies further show that significant rain can form by subsequent accretion of cloud water. The sensitivity of warm-rain initiation to coalescence nucleus concentrations, however, has not been adequately addressed. In the surface layer, the concentration of giant and ultragiant particles can vary by several orders of magnitude (e.g., Hobbs et al. 1985). Also, the concentration of giant and ultragiant salt particles measured by Woodcock were generally lower than those used by Ochs and Johnson so that thorough studies on their role in various maritime conditions remain to be done.

4. Conclusions

a. Large-droplet sources

The initiation of warm rain is critically dependent on the presence of large cloud droplets. Their concentration is most directly affected by the CCN distribution, but large droplets may also be produced by favorable condensation in regions of entrainment. Thus, the source of large droplets will vary with cloud type, within clouds, and over their lifetimes, for example, because mixing varies spatially and temporally during the life cycle of a cumulus cloud. Also, the concentration of very large aerosol that serve as coalescence nuclei varies widely in space and time. In the presence of rain, these particles will be removed (Squires 1952b) but may also be more vigorously generated (e.g., because of stronger surface winds). A prevalent situation for the rapid development of warm rain is within tradewind cumulus clouds in which giant salt particles produce large cloud droplets that initiate coalescence just above cloud base.

An active warm-rain process has also been identified in continental midlatitude convection from strong radar echoes that first appear below the freezing level during cloud development (Braham and Dungey 1978; Ochs and Johnson 1980). Such clouds form in humid conditions similar to tropical clouds, with high liquid water contents needed for rapid growth via coalescence. Although the CCN concentrations are much higher than maritime spectra, leading to smaller cloud droplets by condensation, the presence of ultragiant particles, even if mostly insoluble, would provide coalescence nuclei to initiate rain in updrafts. The high liquid water contents within adiabatic cores would assure that sufficient water was distributed in droplet sizes necessary for rapid accretion by coalescence nuclei.

In mixed regions of warm-based continental convection, larger cloud droplets should also result from favorable condensation trajectories. Since such large droplets grow by coalescence, some may be transported into updrafts by large eddy circulations in updraft–downdraft shear zones (see Rauber et al. 1991), where they would grow more rapidly and perhaps produce the initial radar echo. It is now clear that the favorable trajectory mechanism has the best observational and theoretical support of any mixing hypothesis for producing large droplets.

After precipitation has formed, raindrop breakup will provide embryo droplets that can grow by accretion into new raindrops (Langmuir 1948). Laboratory studies show that collisional breakup should provide numerous droplets in the range 100–200-μm diameter (Czys and Ochs 1988) and larger (Low and List 1982). In deep convection, accretion of supercooled water by ice particles well above the freezing level becomes the major precipitation process, transferring cloud water to graupel and hail. As large graupel or hail fall through the melting zone, droplets are shed that would provide numerous raindrop embryos in deep, warm-base convection. Thus, the source of large droplets in the developed stage of precipitation is likely to differ from the initial stage of warm rain.

b. Warm-rain calculations

The modeling studies discussed in this paper are based on collision efficiencies computed for rigid spheres. More recent models have used the efficiencies of Klett and Davis (1973) for smaller droplets (less than 140-μm diameter). We believe that these efficiencies are still the most reliable ones for the warm-rain initiation problem because they are based on the most appropriate treatment for the flow around interacting spheres (including both viscous and inertial terms in the Navier–Stokes equation). Experimental verification, however, is lacking for most droplet sizes critical to warm-rain initiation, for example, $D = 30$–70 μm interacting with $d = 10$–30 μm. Since the stochastic models depend critically on collision efficiencies, the published findings bearing on warm-rain initiation should be regarded as preliminary, subject to possible future adjustments to the collision efficiencies. At present, questions about the effects of turbulence on warm-rain initiation remain largely unanswered because reliable efficiencies are unavailable for stochastic model calculations.

Incorporation of warm-rain initiation into predictive models using, for example, a dynamical cloud model, remains out of reach because the problem has not been solved on the microphysical scale. Note that the rain initiation problem is similar in this respect to the cloud initiation problem because the microstructure that governs them is not used in cloud models—both are artificially initiated by adjustable model parameters. An approximation called *autoconversion* can be used for warm-rain initiation, with a rate that typically increases with mean cloud-droplet size and water content (Berry and Reinhardt 1974). Today, even this simplification of the physics is not usually included in dynamical cloud models.

c. *Where we stand*

Our understanding of warm-rain initiation will, no doubt, improve with better observations of cloud microstructure. Aerosol and cloud-droplet *instruments* having higher sample volumes are necessary to resolve the concentration of the critical larger particles and droplets. Current developments include array probes with 512 photodiodes that will expand the sample volume by 16 compared to existing 2D probes, and automated analysis of holographic camera data needed for routine particle sizing (D. Baumgardner 1992, personal communication). With such improved devices, we should be able to measure the concentration of the larger particles and droplets during cloud formation and the development of warm rain.

Better sampling and analysis *techniques* must be applied to the key parameters of warm-rain initiation. Cooper et al. (1986) took a significant step in this direction by sampling along cloud parcel trajectories. Analysis of these data provided convincing evidence for a new mechanism of large droplet production. Paluch and Baumgardner (1989) have extended cloud-probe data to smaller scales and revealed a microstructure of diluted and undiluted spectra. Such innovative research is needed to obtain the essential evidence on the initiation problem.

The role of giant and ultragiant particles in warm-rain initiation remains unappreciated, perhaps because it is not completely resolved. For example, the sensitivity of warm-rain initiation to coalescence nuclei with concentrations that vary considerably in space and time has not been fully addressed in model calculations. Although the critical importance of collection efficiencies to rain initiation has been obvious since the efficiency calculations of Hocking, we still lack solid data on collection efficiencies for critical droplet sizes. Some lab data suggest that theory is much too high (Pruppacher and Klett 1978), so that collection may be relatively weak in the absence of coalescence nuclei. This leaves model calculations of warm-rain initiation rather uncertain.

It has been many decades since Patrick Squires first began his studies of warm-rain initiation (Squires 1952a,b). He described the essential character of the problem when he stated (Squires 1958a):

> "It is well known that clouds warmer than freezing sometimes produce rain, especially in maritime air. However, the factors which determine whether a warm cloud will rain are not well understood. It is generally agreed that the raindrops form by coalescence of cloud droplets; it must be expected that the existence of large droplets in a cloud will assist this process both because the collection efficiency of large drops is greater than that of the small ones and because, even with a constant collection efficiency, the mass rate of growth of a droplet increases like r^4."

Considerable research on warm-rain initiation has been accomplished since this statement was published. Although we have established a better appreciation for the widespread significance of the warm-rain process, we are still trying to understand the source of large droplets required to initiate it. At this rate we are quite a way off from being able to predict, on firm microphysical grounds, whether or not it will rain.

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