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CLOUD MICROPHYSICS

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Introduction

Cloud microphysics is the branch of the atmospheric sciences concerned with the many particles that make up a cloud. Relative to the cloud as a whole, the individual particles are very small and so exist on the 'microscale', that is, over distances from fractions of a micrometer to several centimeters. The microscale 'structure' of a cloud, a specification of the number concentrations, sizes, shapes, and phases of the various particles, is important to the behavior and lifetime of the cloud. The abilities of clouds to produce rain or snow, generate lightning, and alter the radiation balance of the earth, for instance, stem in large part from their individual microstructures. Cloud physicists attempt to characterize the diverse microstructures of atmospheric clouds and to understand the phenomena that cause them to change with time.

Clouds typically form in response to changes in atmospheric conditions on scales much larger than the particles, indeed, often much larger than the cloud itself. Most commonly, the upward motions of moist air, driven by synoptic-scale disturbances or convection, cause decreases in the local pressure and temperature, which lead to a lowering of the equilibrium vapor pressures of the liquid and solid phases of water. 'Excess' vapor, that amount above the equilibrium value, develops in rough proportion to the magnitude of the updraft speed. This state of disequilibrium is gradually relieved as the vapor condenses out onto suitable aerosol particles to form the liquid and solid (ice) particles of the cloud.

A cloud, especially during its early stages of formation, often exhibits the properties of a colloidal system, a suspension of tiny particles that follow the airflow and interact only weakly with one another. Whereas the individual aqueous particles may form, grow and subsequently disappear, the system as a whole remains microphysically stable for a time and well characterized in terms of the number concentrations of liquid drops and ice particles. The discipline of cloud microphysics helps us to understand the specific mechanisms needed to break such colloidal stability and to form precipitation.

Microphysical Descriptions

The microstructure of a cloud may be categorized and described statistically in a number of ways. Empirical descriptions, typically derived from *in situ* or remote measurements of clouds, facilitate communications among atmospheric scientists and provide the first glimpses of the physical processes likely to have been operative at time of measurement. The microstructure may differ substantially from one part of a cloud to another, and it evolves with time in ways that depend on the environmental setting and the physical phenomena that are active. Mathematical and numerical models, depending on their purpose, may employ the

empirical descriptions directly, or they may calculate the time evolution of the microstructure if the relevant processes are included.

Cloud particles vary in both phase and composition. The phase, whether solid or liquid, is the traditional descriptor of the aqueous particles, but the nonaqueous aerosol particles in the atmosphere are also crucial to cloud development and evolution. Liquid condensate forms preferentially on the soluble 'cloud condensation nuclei' (CCN), which often contain sulfates or nitrates. Ice particles, by contrast, typically form on 'ice nuclei' (IN), insoluble aerosol particles that contain crustal components or biogenic matter having crystalline structures related to that of ice. The dominant phase of the aqueous particles forms the basis for classifying clouds as 'warm', when only liquid drops are present, or 'cold', when ice is involved (with or without liquid drops). The 'mixed-phase' region of a cloud, throughout which both the liquid and solid (ice) phases of water may be present simultaneously, is that vertical zone between the melting level (0°C) and the –40°C isotherm, the practical lower limit for liquid water to exist in the metastable (i.e., 'supercooled') state. The relative abundance of each phase in a given cloud depends on the prevailing meteorological conditions and the microphysical processes active throughout the lifecycle of the cloud.

The sizes and shapes of the aqueous particles play important roles in cloud development. Whereas ice particles can and do appear in a wide variety of shapes, all but the largest liquid drops tend to remain spherical because of surface tension effects. **Figure 1** depicts the various categories of liquid drops based on their sizes. Note that the size of a particle is an important determinant of its terminal fallspeed and hence of its ability to fall against typical updrafts speeds ($\sim 10\,\mathrm{cm\,s^{-1}}$ in stratiform clouds; $\sim 10\,\mathrm{m\,s^{-1}}$ in convective storms). Drizzle drops represent the traditional transition between the small 'droplets' that remain suspended in the air and the larger drops that may reach the ground as rain. The shapes of bigger raindrops tend to become distorted because of the large dynamic pressure on the lower side, giving the rough appearance of a ball of dough pressed lightly onto a table top. This flattening of the underside causes the aerodynamic resistance to increase relative to that of spherical drops of equivalent volume, thus limiting the fallspeeds of raindrops to little more than about $10\,\mathrm{m\,s^{-1}}$.

The ice particles in a cloud vary enormously in both size and shape. The first ice to appear in many clouds tends to be small ($\sim$10 μm across) and monocrystalline in structure. Single crystals of ice subsequently grow into hexagonal prisms (each prism being bounded by two 'basal' faces and six 'prism' faces) with axial ratios

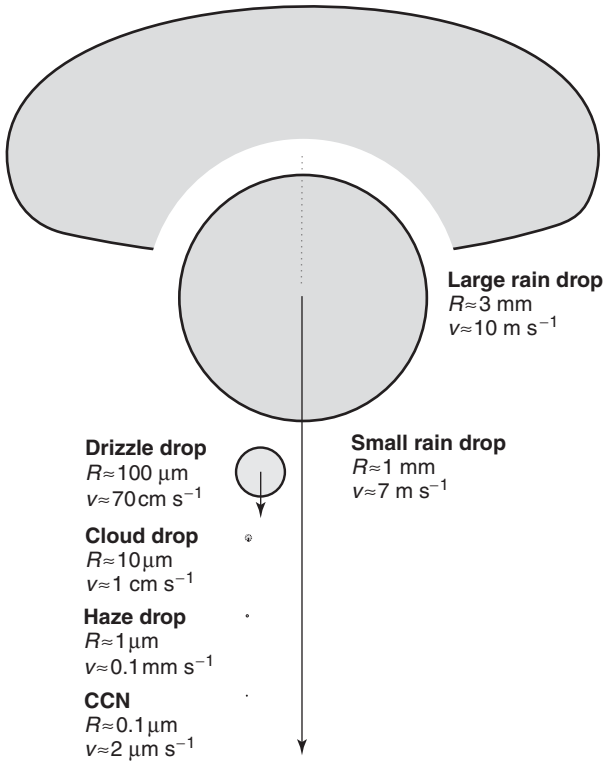

Figure 1 Various categories of liquid drops found in clouds. The indicated drop radii (R) are drawn roughly to scale, as are the arrows representing the terminal fallspeeds (v) of the various drop categories. 'CCN' represents a 'cloud condensation nucleus', a solution droplet that serves as the initial site of condensation. The large raindrop is shown distorted to represent the effect of a large dynamic pressure on its underside.

(length along the principal or 'c' axis divided by the 'a' axis, the width across the corners of the hexagon) that depend systematically on the temperature. As shown in **Figure 2**, 'plates' (c/a axial ratios less than unity) are found when the temperature is either between 0°C and about −3°C or between about −8 and −22°C. On the other hand, 'columns' ($c/a > 1$) appear in the approximate temperature ranges −3 to −8°C and less than −22°C. Deviations from simple hexagonal prisms are common and depend on the excess vapor density, as suggested by the various symbols in **Figure 2**. In addition to the many single crystals, a number of polycrystalline forms of ice are found in cold clouds. For instance, several to hundreds of single crystals may clump together to form 'aggregates' (i.e., snowflakes), and supercooled cloud droplets may freeze onto ice particles, giving rise to rimed crystals, graupel, and hail. Individual crystals seldom grow to more than a few millimeters across, but hailstones can sometimes exceed 10 cm in diameter.

The cloud microstructure is best viewed as a multidimensional specification of the number concentrations

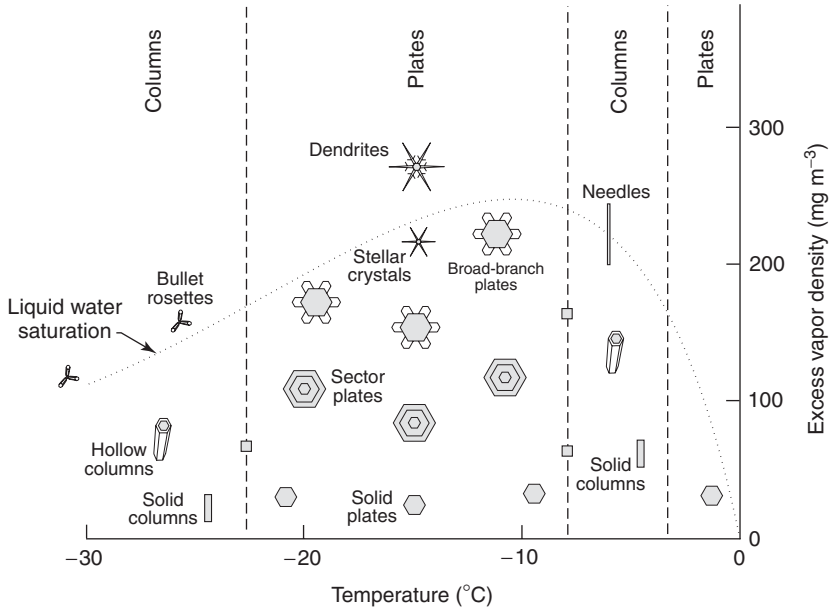

Figure 2 Schematic representation of the wide variations in the shapes (i.e., habits) of ice crystals found in clouds at the indicated temperatures and excess vapor densities (relative to the ice phase). The excess vapor density of air saturated with respect to liquid water is shown by the lightly dashed curve. The heavy dashed vertical lines identify the temperatures of the primary-habit transitions. (Reproduced with permission from Lamb D (1999) Atmospheric ice. In: Trigg GL (ed.) *Encyclopedia of Applied Physics*. Weinheim: Wiley-VCH.)

of the various particles in a cloud. In the case of liquid drops, one need consider only the size of the drops in addition to the usual dependence of number concentration on the three-dimensional location within the cloud and time. The non-spherical ice particles, by contrast, require some additional specification, such as axial ratio, to account for the shapes of the particles. 'Spectral' descriptions of the cloud particles tell us in effect how many of what kinds and sizes of particles are present at given locations within a cloud.

Often, it is useful to characterize the spectra in mathematical terms to minimize the number of variables needed to represent the microstructure. The size distributions of raindrops from convective storms, for example, can be described by analytical functions that have been fitted to observational data by specifying two or three parameters, as shown in **Figure 3**. An exponential function, one subclass of which is referred to as a 'Marshall–Palmer' distribution, has the form

$$n(D) = C \exp(-\lambda D) \quad [1]$$

where C and λ are the parameters fitted to the measured number concentration $n(D)$ of drops within a unit size interval about diameter D. Exponential distributions are used frequently because only two parameters need to be specified. However, as the dashed line in **Figure 3** shows, exponential distributions often over estimate the number of smaller drops. A more general form, the modified gamma distribution,

$$n(D) = CD^{\mu} \exp(-\lambda D) \quad [2]$$

attempts to correct this deficiency, although at the expense of requiring an additional parameter, μ. At least for the data shown in **Figure 3**, a log-normal function of the form

$$n(D) = \frac{N_T}{\sqrt{2\pi} D \ln \sigma} \exp(-\ln^2 (D/D_g)/2 \ln^2 \sigma) \quad [3]$$

works well with appropriate choices of the three parameters, N_T, σ, and D_g. It is important to recognize that all size spectra simply describe the cloud microstructure without regard to the mechanisms that produced it.

Warm-Cloud Microphysics

The liquid drops in 'warm' clouds evolve spectrally via sets of microphysical processes that interact in complicated ways with the larger-scale cloud environment. The process of condensation, for instance, starts during upward motions of the moist air and continues even as the drops interact with themselves and grow into raindrops. As the cloud updraft entrains dry environmental air, ceases or possibly reverses, evaporation may dominate for a time and change the

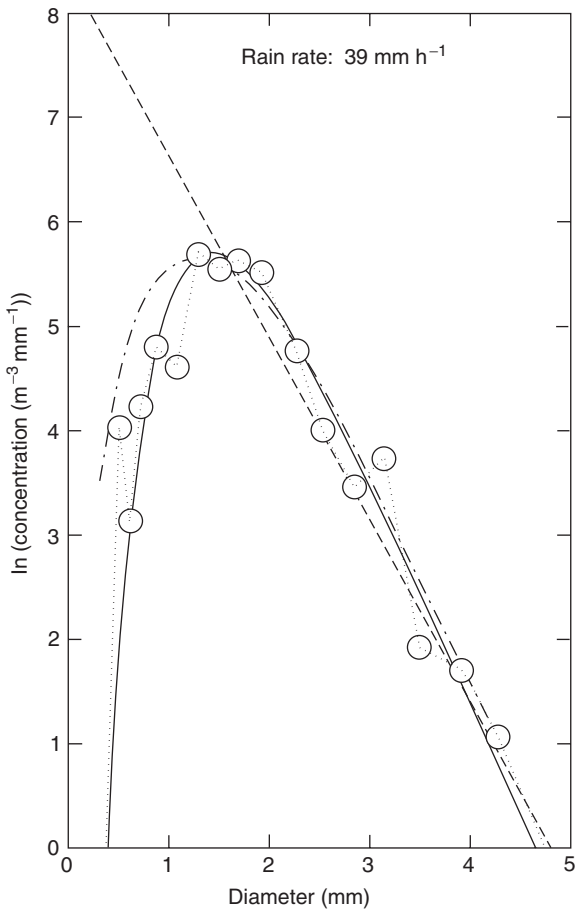

Figure 3 The size distribution of rain drops from a convective storm. Circles connected by light dotted lines: field data based on a one-minute sample from a 20-channel disdrometer. Dashed line: data fitted with an exponential function (eqn [1]). Dot-dashed curve: data fitted with a modified gamma function (eqn [2]). Solid curve: Data fitted with a log-normal function (eqn [3]). (Adapted with permission from Feingold G and Levin Z (1986) The lognormal fit to raindrop spectra from frontal convective clouds in Israel. *Journal of Climate and Applied Meteorology* 25: 1346–1363.)

microstructure in important ways. The ever-changing population of drops often influences the very atmospheric motions that spawned the drops in the first place.

Droplet Formation

The many small droplets in a cloud form initially as excess vapor condenses onto the larger, more soluble aerosol particles (i.e., onto the CCN). The solute in the CCN lowers the equilibrium vapor pressure of the liquid droplets through molecular-scale effects, whereas the droplet curvature increases it.

These two opposing effects of solute and curvature are typically combined in Köhler theory to give the equilibrium saturation ratio S_K as a function of the droplet radius r:

$$S_K = a_w \exp(A/r) \qquad [4]$$

Here, the first factor on the right-hand side of eqn [4] describes the vapor pressure-lowering effect of the solute in terms of the water activity $a_w = 1 - ix_s$, where x_s is the mole fraction of nonvolatile solute that effectively dissociates into i molecular or ionic components. The second factor in eqn [4] accounts for the vapor pressure-raising effect of droplet curvature, in which $A = 2\sigma_{LV}/(n_w RT)$ is a function of the physically relevant variables, the liquid–vapor surface free energy $\sigma_{LV}(= 72\,\mathrm{mJ\,m^{-2}})$, the liquid–water density $n_w\,(= 5.5\times10^4\,\mathrm{mol\,m^{-3}})$, the universal gas constant $R\,(= 8.31\,\mathrm{J\,mol^{-1}\,K^{-1}})$, and the temperature T. Because each droplet grows by the simple addition of water, its total solute content m_s remains constant. Equation [4] is thus most conveniently expressed in terms of the supersaturation needed to maintain equilibrium with the solution droplet:

$$s_K \equiv S_K - 1 = \frac{A}{r} - \frac{Bim_s}{r^3} \qquad [5]$$

where $B = 3/(4\pi n_w)$ and approximations suitable for initial cloud formation have been made.

The properties of the aerosol involved in cloud formation are often depicted graphically. Equation [5] can be seen to represent a family of 'Köhler' curves of constant solute content, as shown in perspective form in **Figure 4**. The competing effects of solute and curvature yield a clear maximum in the equilibrium saturation ratio, which must be overcome by the ambient supersaturation before the particle can 'activate' and grow spontaneously as a cloud droplet. The larger the salt particle, the smaller is the 'critical supersaturation' that needs to be exceeded. Note that

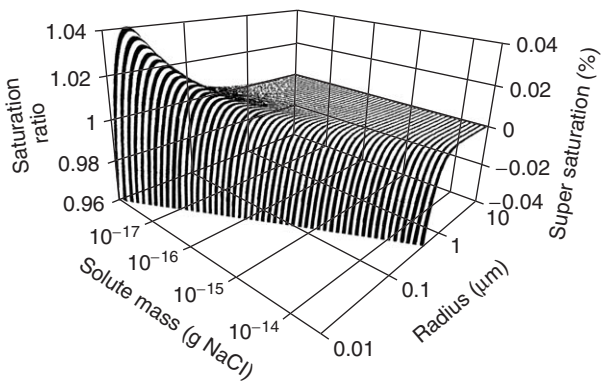

Figure 4 Perspective view of Köhler curves expanded along the solute axis. Each curve is valid for a constant mass of NaCl, the nonvolatile solute in this case.

the critical supersaturation is especially pronounced for particles with the smallest solute contents, so small particles may never experience sufficiently high ambient supersaturations to become activated. Such particles would remain as submicrometer haze droplets interstitially within the cloud.

Growth by Condensation

Individual cloud droplets that are actively growing in an updraft act in effect as tiny sinks of water vapor. During net condensation, the concentration of vapor immediately over each droplet surface (within a few mean free paths of the air molecules) is reduced relative to the average vapor concentration far from the droplet. The radial gradients of vapor concentration thus established give rise to a net flux of vapor molecules toward the drop by the process of molecular diffusion. Even though the water molecules must also be transported across the liquid–vapor interface, it is the vapor diffusion step that tends to limit the mass transport under most cloud conditions. Nevertheless, the change of phase from vapor to liquid results in a slight warming of the droplet owing to the added enthalpy of condensation, energy that must be conducted through the air and away from the droplet. This energy consequence of condensation raises the equilibrium vapor pressure of the liquid and imposes an additional limitation to the growth rate.

The theory that simultaneously accounts for the exchanges of vapor and energy between a growing droplet and the surrounding air was first developed by Maxwell in the nineteenth century. The resulting expression for the linear growth rate is

$$\frac{dr}{dt} = G(s - s_K)\frac{1}{r} \qquad [6]$$

where G is a growth parameter that varies slowly with the temperature and pressure. Note that a droplet grows only to the extent that the ambient supersaturation, s, exceeds the equilibrium value, s_K. As r becomes large, $s_K \to 0$ and the growth rate $dr/dt \propto 1/r$, indicating that the droplets grow relatively more slowly as they become bigger. Calculations based on eqn [6] show that individual droplets experiencing a supersaturation of 1% require hundreds of seconds to grow to radii much beyond 10 μm.

A population of growing cloud droplets derives its water from a common supply, namely the vapor initially carried with the rising air parcel. Competition for the available vapor among all the droplets sets up a strong interplay between the condensation kinetics and the vapor field. Results from numerical computations of droplet growth within an adiabatic parcel are shown in **Figure 5** for the case of a relatively clean maritime environment. The supersaturation (dashed curve) builds up until vapor is removed at a rate comparable to the rate that excess vapor is generated due to adiabatic ascent. Once the maximum in the ambient supersaturation is reached, no new particles can be activated, so the initial number concentration of cloud droplets is established low in the cloud. The close packing of the curves on the right-hand side of **Figure 5** indicates that the droplets tend to bunch together in radius. This narrowing of the drop spectrum is an inherent property of adiabatic condensation and poses a hindrance to the formation of precipitation.

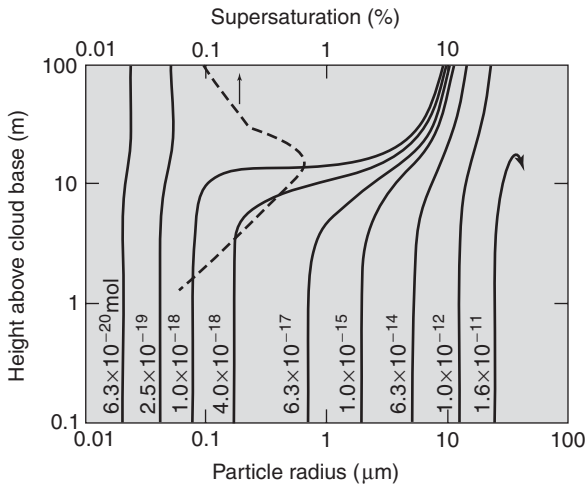

Figure 5 Results from numerical computations of droplet growth in an air parcel ascending adiabatically at the rate of 15 cm s^{-1}. The total concentration of particles in the population is 50 cm^{-3}. Solid curves: droplet radii (lower scale) at various heights above cloud base for droplets each containing the indicated number of moles of nonvolatile solute. Dashed curve: the ambient supersaturation (upper scale). (Based on calculations of Mordy (1959); figure reproduced with permission from Lamb D (2000) Rain production in convective systems. In Doswell CA. (ed.) *Severe Local Convection*. Boston: American Meteorological Society.)

Collisional Interactions

Individual pairs of cloud drops occasionally collide with one another. If two drops 'coalesce' during a particular collision, a single, larger drop replaces the two parent drops in the cloud. Repeated collision–coalescence events lead eventually to large drops that fall rapidly and become raindrops.

The growth of drops through collisional interactions may be quantified by consideration of the separate probabilities for collision and for coalescence. Most commonly, collisions result when a larger drop (the collector drop) overtakes a smaller drop

(the collected drop) during its fall through the air. However, not all drops in the geometrical path of the collector experience collisions, for the simple reason that the air deviating around the collector drop 'pushes' the smallest drops out of the way. The fraction of drops in the path of the collector that do collide with it is the 'collision efficiency', E, a complicated function of both the collector- and collected-drop sizes. The maximum collision efficiency can approach unity when the collector-drop radii are greater than about 40 μm and the collected-drop radii exceed about 5 μm. However, for collector radii less than about 20 μm, the collision efficiency becomes very small. The 'coalescence efficiency', ε, the fraction of drop–drop collisions actually resulting in the formation of a larger drop, is often less than unity for larger collected drops because of drop distortion and the trapping of air at the point of collision. The 'collection efficiency' (E_c), the product of the collision and coalescence efficiencies, $E_c = E\varepsilon$, typically reaches a weak maximum at intermediate collected-drop radii. The collection efficiency provides an overall indication of the effectiveness of drop growth by collision–coalescence.

Collision–coalescence becomes a powerful mechanism for generating raindrops under appropriate microphysical conditions. Because of the de facto thresholds that exist on both the collected and collector drop sizes, collision–coalescence tends to begin in the tail of the cloud drop size distribution, as seen in **Figure 6**. Initially, only a tiny fraction of the bigger droplets will collide and coalescence with neighboring droplets, yielding slightly larger droplets that then have enhanced probabilities of collecting additional droplets. The growth process accelerates as the collection efficiencies increase and other drops join this favored subset of the drop population. Eventually,

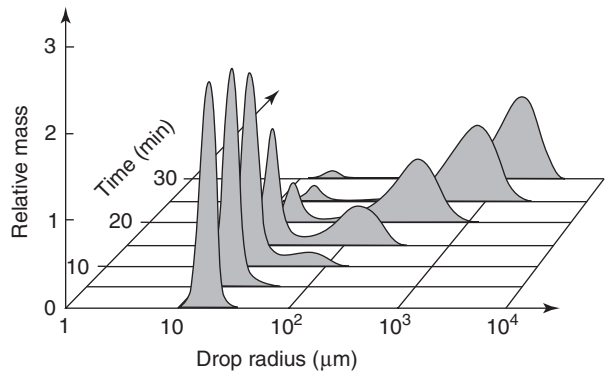

Figure 6 Evolution of the size distribution of drops due to collision–coalescence. (Based on computations by Berry and Reinhardt (1974); figure reproduced with permission from Lamb D (2000) Rain production in convective systems. In Doswell CA. (ed.) *Severe Local Convection*. Boston: American Meteorological Society.)

a new mode in the drop size distribution emerges, as shown by the maxima toward the right-hand side of **Figure 6**. Once the drops in this large-drop mode exceed a few hundred micrometers, they grow rapidly in mass at a more or less continuous rate given by

$$\frac{dm}{dt} = K(r_l, r_s)\omega_L \qquad [7]$$

where $K(r_l, r_s) = \pi(r_l + r_s)^2 E_c(v_l - v_s)$ is the collection kernel and ω_L is the liquid water content (mass per unit volume of cloudy air). The collection kernel is best viewed as the effective volume of cloudy air (containing small droplets of radius r_s, each falling at rate v_s) that is swept out in unit time by the collector drop of radius r_l having fallspeed v_l. In this continuous-growth regime (in which $r_s \ll r_l$, $v_s \ll v_l$, $v_l \propto r_l$, and $E_c = 1$), one finds to first approximation $dm/dt \propto r_l^3 \propto m$. Thus, once the collision–coalescence process gets started, the large drops increase in mass (and size) exponentially with time, until the supply of cloud droplets is exhausted or the drops rupture.

Cold-Cloud Microphysics

Clouds become 'cold' once ice particles form and become active players in the cloud microphysics. Whereas ice particles are necessary components of cold clouds, the liquid drops are nevertheless often present and important to the evolution of the cloud microstructure. The mixed-phase zone of a cloud, where the ice particles and liquid drops interact, is microphysically the most active portion of a cloud. 'Glaciation', the transformation of a cloud from supercooled liquid drops to ice particles, is complicated by the diversity of the interactions that can take place.

Ice Formation

Ice can form once the liquid drops have supercooled by at least 5°C, although the supercooled state can persist in some clouds to temperatures as low as −40°C. The first ice particles in a supercooled cloud appear most commonly when the temperature is between −10 and −15°C following 'primary nucleation', a process by which submicrometer, insoluble aerosol particles catalyze ice formation by acting as molecular templates for the crystal lattice. Such primary ice particles may form directly from the vapor phase (via 'deposition nucleation'), but more commonly they arise from the freezing of supercooled cloud droplets (via 'freezing nucleation'). The freezing of droplets at relatively high temperatures (greater than about −18°C) tends to yield single crystals that grow subsequently into

crystallographically aligned double plates. At lower temperatures, however, the probability of forming multiple crystals within a single droplet becomes large, giving rise to polycrystalline forms, such as bullet rosettes.

Some ice particles form because of the prior existence of other ice particles. Such 'secondary' ice particles arise via several mechanisms, none of which are fully understood. In some situations, crystals may 'fragment', such as when the delicate arms of dendrites break off, thereby increasing the ice particle number concentration. When conditions are just right, as when the temperature is between about -3 and -8°C, and 'graupel' particles are actively growing by riming (i.e., accreting supercooled cloud water), tiny splinters of ice may be released that subsequently grow into columnar ice crystals. The self-breeding, or 'multiplication', of ice seems to be important in clouds that glaciate rapidly.

Growth by Vapor Deposition

Individual ice particles grow initially by the deposition of vapor onto their surfaces. As with the growth of a cloud droplet by condensation, water molecules must first diffuse to the particle surface from the supersaturated vapor field surrounding the particle. However, the transport of molecules across the vapor–solid interface cannot be ignored, if for no reason other than the need to account for facets and the nonspherical shapes of the ice particles. The molecular-scale processes on the crystal surface involve migration of molecules across the surface, from the point of first contact to a step that may be a micrometer or more away. As the step gathers adsorbed molecules, it propagates across the surface, contributing an additional layer (of thickness equal to the step height) of molecules to the lattice. The rate of advancement of the crystal face is determined largely by the frequency with which the steps are generated, a factor that depends on whether the steps originate from two-dimensional layer nucleation or from the emergence of screw dislocations on the surface.

The aspect ratio of an ice crystal reflects the relative rates of growth of the basal and prism faces. For reasons that remain largely unknown, the linear growth rates vary with the temperature in complicated ways, giving rise to the observed alternation of primary habit between plates and columns with temperature (as depicted in **Figure 2**). At relatively large supersaturations with respect of ice, the vapor gradients in the vicinity of a given crystal face become important, leading to the bunching of steps, the 'hollowing' of the face near its center, and a myriad set of secondary habit features superimposed on the primary habit. Good physical reasons exist why one seldom if ever finds any two ice crystals alike in nature.

Riming and Aggregation

Riming and aggregation are both processes involving collection. As in warm clouds, collisions between particles must first occur, then the colliding particles must stick together to form a combined particle. In the case of riming, an ice particle collects supercooled droplets that freeze on contact with the ice surface. By contrast, aggregation involves the collisional interaction of two ice particles, with no change of phase. Aggregation is a complicated process, in part because snowflakes fall erratically and because two solid particles may simply bounce apart after colliding. Snow crystals are most likely to stick together at temperatures within a few degrees of the melting point (because of sintering) and when the arms of dendrites can interlock.

Growth via the riming process occurs by stages that depend on particle size and the rate at which supercooled droplets are accreted. Initially, during the 'crystal stage', the rate of accretion is slow, the collision efficiency becoming appreciable only once the vapor-grown crystal attains an *a*-axis dimension of about 150 μm for plates, 25 μm for columns. The crystal becomes lightly to moderately rimed, but the crystal morphology remains identifiable. The growth is termed 'dry' because each droplet freezes rapidly on the spot of impingement. The 'graupel stage' begins once the crystal identity becomes obscured by the shroud of 'dry' rime ice on the particle. During this and the subsequent 'hail stage', the ice particle grows in mass at rates described reasonably well by eqn [7], with suitable adjustments in the parameters. The hail stage is distinguished from the graupel stage by the formation of one or more layers of clear ice, which results when the rate of accretion exceeds the ability of the particle to dissipate the enthalpy added by the freezing of the supercooled water. Such 'wet' growth occurs when the surface temperature rises to 0°C and the accreted liquid spreads across the surface before freezing. Hailstones represent one extreme to which the microstructure of clouds can evolve.

Precipitation

Precipitation, whether as rain, snow, sleet, or hail, generally results once the aqueous particles in a cloud have grown sufficiently large to fall against the local updraft. In the case of a stratiform cloud, one

characterized by rather weak and uniform updrafts over a broad area, the precipitating particles may simply fall out through the base of the cloud, in the process depleting condensate from the cloud and depositing it on the ground. On the other hand, in convective storms, in which large local updraft speeds can aerodynamically support big particles, the precipitation itself may influence the motions of air though the cloud. The large mass associated with the precipitation commonly initiates a downdraft along the edge of the updraft, causing the microphysical and dynamical aspects of cloud evolution to become intertwined in complicated ways. As an aid to the discussion below, **Figure 7** offers a summary of the various processes operating in the 'warm' and 'cold' parts of a representative convective cloud during precipitation formation.

In the warm parts of clouds, large drops can emerge out of the stable population of cloud droplets only through collisional interactions. Condensational growth alone is too slow, but it must be recognized as a necessary process, for the numerous cloud droplets serve as the feedstock for the growth of the larger drops. The collision–coalescence process becomes an effective mechanism for breaking the colloidal stability of the cloud once the threshold size ($\sim$25 μm in diameter) for collection has been overcome. The needed 'coalescence embryos' can arise from the droplet population itself (most commonly in clean, maritime environments or during turbulent mixing), alternatively from 'giant nuclei' in dusty regions. Once the warm-rain mechanism is established, the rain drops grow rapidly by sweeping out the smaller cloud droplets until they themselves become unstable and break into fragments during their fall to earth.

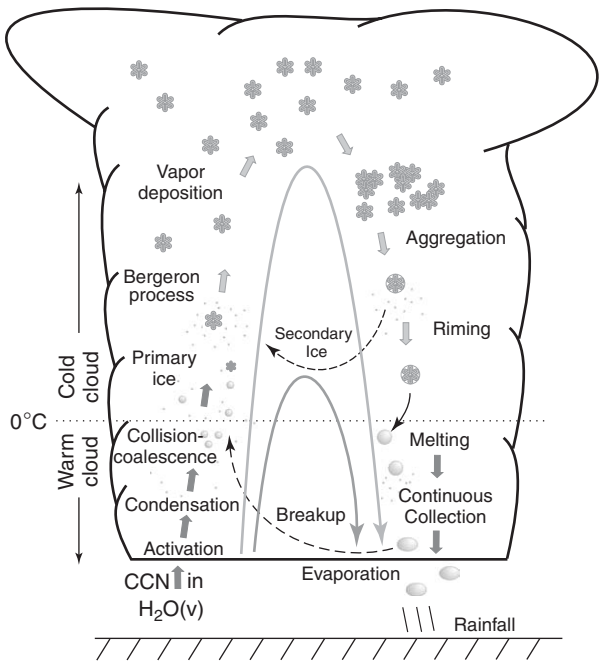

Figure 7 Summary depiction of the microphysical processes operating during the formation of precipitation in a deep convective cloud.

In the cold parts of a cloud, the colloidal stability of the cloud is broken once the ice phase has been nucleated in the presence of supercooled droplets. This 'ice crystal' mechanism, often termed the Bergeron process, arises from the inherent difference in the equilibrium vapor pressures of liquid and solid water. The relatively low vapor pressure of ice compared with that of the droplets at any given temperature gives the ice crystals a growth advantage by causing water vapor to transfer (via diffusion) from the many cloud droplets to the fewer ice crystals. The process proceeds rapidly, especially in the temperature range between -12 and -15°C, permitting the ice crystals to attain sizes sufficient to initiate the other cold-cloud growth mechanisms, such as aggregation and riming. These large ice particles eventually fall into the warm part of the cloud, where they may melt and join the population of raindrops formed by the collision–coalescence process. The ice process can be an effective initiator of precipitation in both stratiform clouds and summer thundershowers.

The efficiencies with which clouds develop precipitation depend partly on the types of microphysical processes that are active and partly on the environmental settings in which the clouds form. For instance, the relative ease with which 'maritime' clouds release precipitation compared with 'continental' clouds most likely stems from the differences in aerosol abundance found in the different airmasses. The relative absence of active sources of aerosol particles over the open oceans allows the CCN concentrations to become low, which in turn means that those droplets that do form in maritime clouds tend to be larger on average than those in continental clouds. The collision–coalescence process thus gets started early in the life cycle of maritime clouds, providing such clouds with a decisive microphysical mechanism for developing precipitation. Within a given climatic regime, storm organization on the mesoscale seems to be an important contributor to precipitation efficiency. Low magnitudes of vertical wind shear at the time of cloud formation tend to favor vertically erect storms with high precipitation efficiencies, presumably because the incipient precipitation particles can then fall directly through the condensate-rich inflow of the storm. At the same time, however, such systems tend to be short-lived and yield relatively small total amounts

of precipitation. Storms that form in environments in which the wind varies modestly with height in both speed and direction last longer and yield the most precipitation, for then synergism ensues between the dynamical time scales of the storm and the time scales for the microphysical processes to operate effectively.

See also

Cloud Chemistry. **Clouds:** Classification; Climatology; Measurement Techniques *In Situ*. **Convective Cloud Systems:** Modelling.

Further Reading

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CLOUDS

Contents

Classification

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Introduction

The classification of clouds and reports by observers of their coverage have taken on considerable value in recent years. Those reports now provide a quite long-term record in which changes in cloud coverage and type may be discerned that might not be detectable in the relatively short record of satellite data.

The classification system, was, of course, formulated and revised on the basis of visual attributes, without the benefit of in-cloud measurements. Airborne measurements would have delineated the internal composition of clouds that appear somewhat similar and have common labels but are very different in composition (e.g., stratus and altostratus). In this article, airborne measurements of cloud composition will be used where possible to augment the composition of clouds.

What is a Cloud?

As defined by the World Meteorological Organization, a cloud is an aggregate of minute, suspended particles of water or ice, or both, that are in sufficient concentrations to be visible: a collection of 'hydrometeors', a term that also includes in some cases, due to the distance of the observer, the precipitation particles that fall from them. Today the term 'cloud' also includes those clouds that are nearly invisible to the human eye but are readily detectable in satellite thermal imagery.

Clouds are tenuous and transitory; no single cloud exists for more than a few hours, and most small clouds in the lower atmosphere exist for only a few minutes. In precise numbers, the demarcation between a cloud and clear air is even hard to define: how many cloud drops per liter constitute a cloud? When are ice crystals and snow termed 'clouds' rather than precipitation? When are drops too large to be considered 'cloud' drops, but rather raindrops? These questions are difficult for scientists to answer in unanimity because the difference between cloud particles and precipitation particles, for example, is not black and white; rather they represent a continuum of fallspeeds. For some scientists, a 50 μm diameter drop represents