meteorological bases for precipitation development^{1,2}

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The character of precipitation in any region, whether it be humid or arid, is the result of complex interactions between a large number of factors—physical, chemical, meteorological, geographical, topographical—having dimensions ranging from the submicron to the global, and acting on time scales from milliseconds to centuries. Study of these factors and their interactions on all space-time scales can only add perception and depth to our understanding of a region. During the past two decades, however, research has tended to emphasize those factors operating on scales of single clouds and cloud groups. These are the scales of primary concern to the cloud

physicist and these are the scales presently amenable to cloud seeding and weather modification research.

This paper will attempt to summarize the major points in our knowledge of the meteorological physics of precipitation development. For this purpose I will make extensive use of a flow diagram, Fig. 1, to keep account of the various processes involved. Then I will touch upon the interaction links between precipitation formation and cloud dynamics. Wherever possible I will emphasize those aspects of precipitation physics which seem to be especially important in arid regions. The emphasis throughout will tend somewhat toward convective clouds for the simple reason that these have received the major attention and are best known to cloud physicists. We believe, however, that in broad outline our discussion will apply equally well to stratus and cumulus clouds.

¹ Presented at AAAS Symposium on Weather Modification in Arid Lands, New York, N. Y., 30 December 1967.

² Paper prepared under sponsorship of National Science Foundation, Grant GA-740.

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27/11/67

FIG. 1. Flow diagram of precipitation processes.

Condensation mode of drop growth

The initial stage of cloud drop formation is the thermodynamic activation of suitable "cloud" nuclei in a slightly supersaturated environment. In typical clouds between 30 and 1000 such nuclei per cm^3 become activated by supersaturations of less than 1% with respect to a flat water surface. Following activation, growth of the nuclei into drops continues through the diffusion of water molecules from the supersaturated drop environment to the drop surface. This process is the well known one of condensation.

The supersaturation required to activate the cloud nuclei and to provide a continuing source of vapor for condensational growth usually is provided by adiabatic cooling of expanded air parcels. In the case of cumulus clouds, an updraft provides the expansion and cooling as air parcels are lifted to regions of lower pressure. In stratus and altostratus the expansion and cooling is due to a combination of very slow upward motion and *in-situ* pressure lowering accompanying dynamical development of the larger-scaled weather systems in which the clouds are imbedded. In mountainous regions the lifting may be orographically induced.

Nuclei of cloud formation are thought to have sizes of about 10^{-2} to 1 micron diameter and to contain a soluble salt as the active component. The exact nature of these particles, and their origins, is not completely understood but evidence suggests that they may include sea-salt residues (mainly NaCl) and soluble sulphates resulting from oxidation and hydrogen of sulphurous gases formed in the decay of biological materials. For a discussion of the processes of nuclei activation and growth the reader should consult a cloud physics text such as Byers (1965a).

Evidence indicates that there may be important variations in the cloud nuclei populations, at least on a regional basis. Twomey (1959), and Squires and Twomey (1960, 1966) have shown that a systematic difference in nuclei spectra in air of tropical marine versus arid continental origin is the root cause of systematic differences between convective cloud droplet spectra from the two regions. This is illustrated in Fig. 2 which shows cloud droplet spectra for cumulus congestus clouds of three regions: Caribbean Ocean (winter), Central United States (summer) and Southwestern United States (summer). The droplet concentrations and liquid water content for these spectra are: Caribbean Ocean, 52 cm^{-3} , 0.8 gm m^{-3} ; Central United States, 188 cm^{-3} , 0.6 gm m^{-3} ; Southwestern United States, 230 cm^{-3} , 0.3 gm m^{-3} . The first two of these spectra are from Battan and Reitan (1957), the third is previously unpublished data from the Chicago Cloud Physics Group.

Note the progressive broadening of the spectra, the shift of modal diameters to larger sizes, the decrease in drop concentration and increase in liquid water content as we go from cumulus congestus clouds of the arid Southwest, to similar clouds in the humid Central United States and over the Tropical oceans. These data

FIG. 2. Droplet spectra for cumulus congestus clouds of three regions.

are considered to be fairly representative of *average* conditions within cumulus congestus clouds of these three regions. It is known, however, that the liquid water content of these distributions is low, by a factor of 2 to 5, compared with measurements made within active updraft regions in such clouds. Measurements of drop spectra limited to active updraft regions are not yet available.

Studies carried out by Howell (1949), Mordy (1960) and Neiburger and Chien (1960) have shown that the influence of cloud nuclei in determining the number and size of cloud drops is restricted to the lowest few hundred meters above cloud base. It is only in the base region that cloud nuclei are activated, hence it is there that the concentration of cloud drops is determined. Subsequent condensation merely serves to increase the sizes of drops which are present.

The drop spectra of Fig. 2 are not to be regarded as extreme examples. They were selected for illustration since they involve clouds of common type (cumulus congestus), sampled by the same research group (University of Chicago) using a single sampling technique (multiple shot slide impactor). It should be obvious to the reader that among natural clouds one will find a complete gradation in drop spectra ranging from those more narrow than our samples from the Southwest to others still broader than our examples from the Caribbean. Moreover, it is likely that these spectra had undergone some modification by coalescence prior to observation.

Coalescence mode of drop growth

Differences in cloud drop spectra, such as those shown in Fig. 2, profoundly affect the rates of drop growth by coalescence. This process, the simplest of the drop growth processes, is one by which large drops overtake, collide, and coalesce with smaller ones. The effectiveness of this process depends upon several factors:

- a) difference in falling speeds of drops of different sizes;

- b) cloud liquid water content and distribution among drops of various sizes;
- c) inertia of the drops which limits their ability to flow around one another when caught on a collision course;
- d) surface properties of the drops and external electric field conditions which determine whether a collision will result in capture or bounce-off.

Mathematically the coalescence process can be expressed as

$$\frac{dm_2}{dt} = \pi (r_2)^2 (u_2 - u_1) \chi_1 E_{12}$$

where r denotes drop radius, u is terminal velocity, χ is fractional liquid water content and E is the appropriate collection efficiency. Subscript 2 denotes the collector drop and subscript 1 the collected drops. Summation is extended over all drops smaller than the collector drop. When all drops are about the same size this "continuous" coalescence model must be replaced by a stochastic model (Telford, 1955; Twomey, 1964; Berry, 1967).

The coalescence process has been studied extensively and we can make reasonably accurate calculations of the evolution of a few large drops growing in a field of smaller ones. Such calculations have appeared in several places (e.g., Fletcher, 1962). A calculation based upon the data of Fig. 2 is given in Fig. 3. These curves show the computed sizes, as a function of time, of those drops which were initially present in concentrations of 10^{-2} cm^{-3} (5 micron interval) in the arid region and tropical region clouds of Fig. 2. For this calculation, collision efficiencies were obtained by combining values

from Shafrir and Neiburger (1963) with those from Langmuir (1948). It was assumed that every collision resulted in a coalescence. The importance of such a calculation is not in its detail, but in its demonstration of the control exerted by the initial drop spectrum on the rate of coalescence drop growth.

Coalescence precipitation

The calculation shown in Fig. 3 suggests that our maritime cloud should develop 1-mm diameter drops in concentrations exceeding 10^4 m^{-3} in about 30 minutes whereas over two and one-half hours would be required to do this in the arid region cloud. Increasing the liquid water to 1 gm m^{-3} without changing the drop spectra would decrease these times to about 20 minutes and one hour, respectively. In this calculation the maritime cloud is favored for coalescence because it confines the liquid water into a smaller number of larger drops and because of its larger water content.

Calculations such as these suggest that the role of coalescence in initiating rain should vary considerably among clouds of different types occurring in different regions and in different seasons. They also suggest that the coalescence process will be most effective in clouds of warm humid air masses. It has indeed been established that coalescence is of great importance in the tropical climates (see for example Bergeron, 1933; Byers and Hall, 1955; Braham *et al.*, 1957; Mason, 1957; Brown and Braham, 1963). There are also many documented observations of coalescence rain from clouds in maritime polar air masses (Lewis, 1947; Singleton, 1960; Stewart, 1964; Cornford, 1967, and others).

There is no known reason why the coalescence process does not also operate in arid region clouds. However, to the extent that our calculations are valid, we would anticipate that it would operate so slowly that cloud destroying forces (turbulence and mixing plus depletion of updraft sources) would be likely to dissipate many clouds before any appreciable amounts of rain could form.

When we try to verify this prediction with observations from arid region clouds we encounter difficulties. In the first place there have been relatively few in-cloud studies of precipitation particles in arid region clouds. However, Battan (1963) made use of extensive data from a height finder radar to argue that coalescence *was* responsible for precipitation initiation in summer cumulus clouds of Arizona. MacCready and co-workers (1965) also reported finding drizzle drops at an early stage of growth of Arizona cumuli. These were attributed to giant cloud nuclei in the air below cloud bases. The calculation shown in Fig. 3 would be completely invalidated by a low concentration of large particles not included in the cloud spectra. We must await better field observations before assessing the role of coalescence in arid region clouds.

Since the principle of drop spectra control over coalescence appears valid, in our flow diagram of precipitation

FIG. 3. Size-time plots of a precipitation drop growing by coalescence in two different cloud spectra.

processes we distinguish between broad spectra clouds (from continental type nuclei) and narrow spectra clouds (from maritime type nuclei) as initial points in rain development. Continuing through the diagram we show that coalescence can readily lead to drizzle and rain in broad spectra clouds (heavy line) whereas it proceeds more slowly in narrow spectra clouds (thin line).

There is another important consequence of the early development of large coalescence drops in cumulus clouds. Such clouds can easily have updrafts strong enough to carry these drops upward through the freezing level where some of them may freeze, thus introducing the ice phase into supercooled clouds. This may have important consequences for precipitation development, a topic which we will explore after examining more general aspects of ice-crystal precipitation processes.

Rain produced by coalescence of liquid particles is commonly known as "warm rain" in deference to the fact that it was first identified in tropical clouds which produced rain without ever reaching the freezing level. It is clear, however, that the coalescence process can operate equally well in supercooled and non-supercooled clouds.

Bergeron, or ice crystal, mechanism

We turn now to precipitation processes involving ice particles. Here we start with what is classically known as the Bergeron process since it developed from the pioneering paper by Bergeron presented before the Lisbon, Portugal IUGG meetings in 1933 and published in Paris in 1935. Bergeron ruled out coalescence as an important mechanism because he assumed that cloud drops were of nearly equal size and therefore unable to have appreciable range of fall speeds and hence unable to undergo appreciable collision and coalescence. Resting his argument upon the observations that clouds supercool, and that cumuli tops commonly develop striae and fuzziness prior to rain from their bases, and upon the well known fact that the saturation vapor pressure over ice is less than that over water at the same temperature, Bergeron hypothesized that most raindrops begin as ice crystals in supercooled clouds where they grow at the expense of evaporating drops to produce snow crystals and snow flakes.

Recent research has filled in many of the details of the ice crystal mechanism. We now differentiate between primary and secondary sources of atmospheric ice particles. Non-ice particles that are thermodynamically activated and then grow by vapor deposition into small ice embryos are called primary ice nuclei. These are the classical ice nuclei, or sublimation nuclei of Bergeron. It is usually accepted that a major fraction of natural primary ice nuclei are grains of platey silicate minerals of earth origin (0.1 to 3 micron diameter clay particles); see for example Schaefer (1950); Isono (1955); Mason and Maybank (1958); Kumai (1961); Kumai and Francis (1962); and Byers (1965b). However, the possibility of an extra-terrestrial origin for some of the primary

ice nuclei has not been ruled out. (For a discussion of the meteor dust hypothesis see Fletcher, 1962, pp. 250–258.)

Measurements of the concentrations of primary ice nuclei are made in a variety of ways (Warner, 1957; Kline, 1963; Soulage, 1965) all of which give data in general agreement showing that there are important space-time variations in natural ice nuclei, Fig. 4. Moreover, and very importantly, these measurements show that natural ice nuclei are usually too sparse for effective precipitation initiation in cumulus clouds and too sparse to account for observed ice crystal concentrations in natural clouds (Mason, 1957; Koenig, 1963; Braham, 1964; Mossop *et al.*, 1967). There is a growing awareness among cloud physicists that secondary ice sources, including the possibility of crystal breakup, may often overwhelm the concentrations of primary nuclei.

Following nucleation the resulting ice crystals develop crystal habits and grow at rates which are profoundly affected by temperature. In the lower part of Fig. 5, are shown the conclusions of Kobayashi (1961), Mason *et al.* (1963) and others, regarding crystal habit as a function of growth temperature. It is now fairly clear that growth habit is a function only of temperature while the extent of crystal elaboration is controlled by the degree of vapor saturation.

Compared with that for crystal habit, our knowledge about crystal growth rates is much less satisfactory. Classically we have drawn upon a combination of thermodynamics and kinetic theory to predict that crystal growth would be given by

$$\frac{dm}{dt} = 4\pi CD\Delta\rho$$

where D is the diffusivity of water vapor in air, C is an "electrostatic capacitance" factor (function of crystal size and shape) and $\Delta\rho$ is the difference between satu-

FIG. 4. Median ice nuclei concentrations as a function of temperature for 15 ground level sites. Numbers of observational series on numbers of days shown in parenthesis. From Kline (1963).

FIG. 5. Ice crystal growth rates (top) and ice crystal habit regions (bottom) as a function of temperature.

rated vapor density over the growing crystal surface and its environment. The curve for Δp corresponding to a crystal growing in a water saturated environment is shown in the lower part of Fig. 5. Using McDonald's (1963) values for the C factor, the rate of growth of an ice crystal of major dimension of 200 microns, at a pressure of 1000 mb, is computed from the classical formula to have values shown by the dashed line in the upper part of Fig. 5.

All known empirical evidence for the rates of growth of ice crystals is shown in the upper part of Fig. 5. The solid line was derived by combining observations on the relative rate of growth of ice crystals as reported by Hallett (1965) with laboratory measurements of actual growth rates of crystals at -2.5°C and -5°C as reported by Mason (1957). Also shown are: a) a range of growth rates for crystals at -18°C as measured by Reynolds (1952), and b) growth rate values at several temperatures as derived by Todd (1964) from data reported by Nakaya (1954).

We note immediately that the laboratory observations give two peaks in the crystal growth curve—a primary maximum in the vicinity of -14°C and a secondary maximum near -5°C .

Although the amount and quality of data on crystal growth rates is less than satisfactory, it is clear that the classical growth equation does not adequately fit the

observations. Observations show a region from about -12°C to -17°C in which crystal growth rates are as much as 2 orders of magnitude larger than found at warmer and colder temperatures. This zone of rapid growth occurs in the temperature region favoring dendritic crystal habit. This coincidence of two temperature dependent effects has profound implications for precipitation development. It means that in any cloud extending from 0°C to -20°C crystal growth in the -12°C to -17°C region will overshadow that of all other levels. It helps explain the frequent observation that dendritic crystals are the most frequent crystal form of natural snows. With this combination of large growth rates and a dendritic crystal habit in the -12°C to -17°C temperature region, we should anticipate almost a step-function difference in the precipitation behavior of clouds depending upon whether or not they extend into this -12°C to -17°C level.

Before taking up the combined effects of crystal growth and riming it is instructive to consider some of the implications of a crystal growth rate which is a marked function of temperature as shown in Fig. 5. First we consider a stratus layer with a top temperature of about -8°C . The number of natural ice nuclei active at -8°C is not readily measurable with present equipment, but an extrapolation of the mean curves of Fig. 4 gives concentrations of about 100 m^{-3} . Individual measurements often suggest fewer than 10 m^{-3} ice nuclei active at -8°C . From the data in Fig. 5 we compute that the crystals resulting from such nuclei would grow into columns about 100 microns long in 10 minutes and about 1 mm long in an hour. The low concentrations of natural nuclei active at these temperatures, combined with the slow growth of the resulting crystals, gives understanding to the observations that such supercooled stratus seldom precipitates naturally.

In contrast we consider a stratus layer having a top temperature of -15°C to -20°C . At these temperatures we expect natural ice nuclei in concentrations of 10^3 to 10^4 m^{-3} . Each resulting ice crystal will grow rapidly into a dendritic crystal—reaching a diameter of 0.5 mm in less than one minute and 3 mm in less than 30 minutes. During these times these crystals will fall about 10 m or 500 m, respectively. Simple calculations such as these lend understanding to the often observed winter phenomena of snow showers from stratus and stratocumuli only a hundred feet thick.

Crystal clumping

Technically snow flakes are clumps of a large number of individual snow crystals. The process of collision and aggregation of crystals into flakes is called clumping. It is observed to be important in the production of snow. Probably everyone has observed that snow flakes frequently consist of a loosely adhering group of crystals which break into a myriad of individual crystals upon landing on a coat sleeve.

Present evidence suggests that clumping is most im-

portant in the temperature range between 0C and -10C. Hosler, Jensen, and Goldshlak (1957) found in laboratory studies that at ice saturation the aggregation between ice particles was negligible at -25C and increased to a maximum at 0C. Magono (1953) found that maximum snowflake diameters occurred with ground level temperatures of about -1C—consistent with the notion that clumping is important only at temperatures slightly below 0C.

Combined effects of crystal growth and riming

Riming is the growth process in which ice particles sweep-out supercooled cloud drops which freeze (all or in part) on contact. Mathematically it can be described with an equation similar to that for coalescence. In this case the latent heat of fusion of the accreted cloud drops is added to the riming particle, thus warming it. The riming particle dissipates this heat by convective heat transfer to the environment and, to a limited extent, by evaporation. Obviously the heat transfer process sets limits on riming growth which can be completely frozen. Spongy ice is the term applied to particles formed when the heat transfer is not adequate to permit freezing all of the accreted water. List (1959, 1960, 1961, 1963) and Macklin (1961, 1962) have investigated this ice form both in wind tunnels and theoretically. Based upon their work it appears likely that all riming particles in stratus clouds will be able to dispose of the latent heat and thus remain "dry." Snow pellets and graupels growing in the updraft regions of cumulus clouds may be either spongy or dry depending upon their size, temperature and the cloud water content. Although an interesting problem in cloud physics, consideration of details of this matter takes us beyond the purpose of this paper.

We now consider the combined effects of crystal growth and riming in producing precipitation particles in cumulus clouds. From the data of Fig. 5, combined with relationships between crystal size and mass provided by Nakaya and Terada (1934), and data on falling speeds from Nakaya (1954) and Langleben (1954), we can compute the mass of growing ice crystals as a function of time, or height, in a cloud of assumed updraft speed. For our purpose we assume an in-cloud lapse-rate which is saturated adiabatic with an equivalent potential temperature $\theta_E = 352.8$, which is typical of U. S. summer conditions (Byers, 1965a). Our first calculation, plotted in Fig. 6, assumes an updraft of 5 m sec^{-1} in the arid region cloud spectra of Fig. 2, with crystals nucleated at temperatures of -5C and -10C. The masses of the growing crystals are plotted as a function of time starting at the moment a cloud parcel passes the -5C level. We note that the crystals nucleated at -5C show a minor spurt of growth, then grow slowly as they pass through the -6C to -10C region. Growth picks up again on passing -12C and thereafter the particle grows quickly to a size large enough to begin riming.

The crystal nucleated at -10C very quickly passes

FIG. 6. Mass of a crystal growing by vapor deposition followed by riming into a graupel. Assumes an updraft of 5 m sec^{-1} and crystal nucleation at -5C or -10C. Particle mass equivalent to 0.5 mm and 1.0 mm graupel indicated by arrow.

into the high growth region and shows very rapid growth to a size large enough to rime. If we assume that riming takes place in a water content of 1 gm m^{-3} we obtain the upper, dashed curves.

The inescapable conclusion from this computation is that in a cumulus cloud, the rapid growth regime from -12C to -17C so completely dominates crystal growth that crystals nucleated at any temperature above about -12C will end up with virtually identical masses by the time they are carried up to about the -15C level. This conclusion, first pointed out by Neumann *et al.* (1967), has important consequences for cloud seeding.

If we keep the nucleation temperature at -10C but allow the updraft speed to vary we obtain the results shown in Fig. 7. Here we see that the fast updraft case is first to carry its crystals into the -12 to -17C high growth regime, but the fast updraft also carries the crystals through this high growth regime in the least time thereby limiting their total growth. The slow updraft case takes longer to get started but allows more time for the crystals to grow in the -12 to -17C region. Thus it is the slowest updraft that produces the fastest rate of growth of precipitation through vapor deposition on crystals, and their subsequent riming, in clouds occupying the -5 to -20C region.

Drop freezing

The freezing of large cloud drops and small rain drops was established as one of the important chapters in precipitation physics by the University of Chicago group on Project Whitetop. Through systematic aircraft sampling inside of clouds they found that the major pre-

FIG. 7. Mass of a crystal growing by vapor deposition followed by riming into a graupel in cloud with three different updrafts.

cipitation sequence in summer cumulus congestus and cumulonimbus clouds of Missouri was coalescence induced drizzle followed by freezing at -5°C to -10°C and subsequent riming into graupels (Koenig, 1963; Braham, 1964). It was observed that frozen drops, snow pellets and small graupels, larger than about 200 microns diameter, were present in concentrations of 10^4 m^{-3} in clouds whose tops had never been colder than -10°C . In addition, the frozen particles were accompanied by irregular ice particles which were presumed to have been formed by fracturing of the drops during freezing. These irregular particles were in marked contrast to the regular symmetry of crystals grown through vapor deposition.

The freezing onset temperature of about -5°C was much warmer than had been anticipated from prior laboratory studies of drop freezing. It had the very important consequence of introducing the ice phase into the clouds several minutes earlier and in much larger concentrations than could have occurred from crystal growth on primary nuclei. The result was to increase the precipitation efficiency of the clouds because the pellets were observed to grow by riming faster than the unfrozen drops had been growing through coalescence. Reasons for this accelerated growth were discussed by Braham (1964).

Vali and Stansbury (1965, 1966) distinguish between two alternate models of the drop freezing process. These are:

- Stochastic model—at any given instant all drops have an equal probability of freezing which is proportional to the volume and increases exponentially with the degree of supercooling. Mathematically this is stated:

$$P_{\theta} = \frac{1}{N} \frac{dN}{d\theta} = \frac{BV}{\alpha} e^{-a'\theta}.$$

- Singular model—drop freezing is due to impurities, each of which has a “characteristic” temperature at which it is certain to produce nucleation. The probability of freezing is proportional to the number of impurities in a drop (i.e., to volume) and to the distribution with temperature of the characteristic temperatures of various impurities. This model is written:

$$P_{\theta}' = \frac{1}{N} \frac{dN}{d\theta} = B'V e^{-a'\theta}.$$

In these equations P_{θ} is the probability of freezing at θ degrees of cooling, B and B' are rate constants, V is drop volume, α is cooling rate and a and a' are constants.

Laboratory studies by Bigg (1953) and Gokhale (1965) were interpreted in terms of the stochastic model. Studies by Vali and Stansbury (1966) and Vali (1967) suggest that freezing of drops of rainwater show some aspects of both models but that they most nearly fit the singular model.

Using constants derived by Vali and Stansbury and Gokhale it can be computed that typical drop freezing temperatures in natural clouds should be: 100 micron radius, -16°C to -22°C ; 500 micron radius, -8°C to -12°C ; 1000 micron radius, -5°C to -10°C . These values lend understanding to the field observations of Project Whitetop.

On the basis of present knowledge we can predict fairly well the meteorological situations in which freezing of coalescence-grown drops will be important. In regions of an active coalescence mechanism (thus far identified with the warm humid air masses) we anticipate that drop freezing will be of great importance only in cumulus clouds since, in these air masses, the stratiform clouds tend to be warm. Drop freezing near -5°C appears to require 500–1000 micron drops—thus the updraft must be 2 to 5 m sec^{-1} to insure that such drops will be carried up through the freezing level. If the updraft is too strong (say 15 to 30 m sec^{-1}) there will be insufficient time for drops to grow large enough to freeze prior to being carried high into the cloud. In this case drop freezing takes on a different role. Gliki *et al.* (1962) has shown that drops freeze as polycrystalline aggregates suitable for spatial dendritic growth. At temperatures well below -20°C , where crystal growth is very slow, the introduction of spatial dendrites even of modest size (100–300 micron diameter) can launch the riming mode of precipitation growth earlier and faster than it can follow from crystal deposition.

Winter stratus clouds frequently have top temperatures above -10°C and are topped by a strong inversion. Previously we saw that the ice crystal mechanism was ineffective in such conditions. The long life of supercooled stratus is favorable both for coalescence growth and for small drop freezing via the stochastic model. Perhaps this combination accounts for the observations of snow pellets often observed from such clouds.

Secondary ice sources

There are several ways in which the ice phase may be introduced into a supercooled cloud. Our last section dealt with drop freezing and some of its consequences. Earlier sections discussed primary ice nuclei; something of their origin, nature, and concentration. It was pointed out that concentrations of primary ice nuclei, as measured in clear air below clouds, frequently are not adequate to account for the observed concentrations of ice particles in clouds, if we assume one ice crystal per one ice nucleus. Several researchers have pointed out that this discrepancy is of the order 10^2 to 10^3 -fold in the region between -5°C and -15°C . What is the origin of these additional ice particles? Two possible sources can be suggested in addition to drop freezing previously discussed: viz., influx from higher clouds, and multiplication of existing ice particles.

The thought that cirrus clouds might serve as seeding agents for lower clouds has long been discussed. The major difficulty in accepting this idea was the uncertainty as to how long cirrus crystals could survive—how far cirrus crystals could fall—in clear air before complete evaporation. Much of this uncertainty was removed by recent studies of The University of Chicago Group (Braham, 1967; Braham and Spyers-Duran, 1957). It has been shown that certain types of cirrus clouds (cirrus uncinus and cirrus spissatus) produce crystals that can survive up to 20,000 ft of clear air descent. On six different days of visible cirrus development, during the summer of 1966 near Bemidji, Minnesota, our cloud physics airplane was operated at temperature levels of -5°C to -10°C to check for possible cirrus crystal remnants. On all six days ice crystals were found in concentrations between 10^3 and 10^6 m^{-3} at levels 10,000 ft to 20,000 ft below the cirrus with no intervening clouds visible. These concentrations are between 1 and 5 orders of magnitude higher than expected from ice nuclei measurements and more than adequate to seed any lower cloud into which it might fall. The limited evidence available suggests that stratiform cirrus clouds have a much lower potential for seeding across deep clear air layers. This probably is due to the fact that crystals in cirrostratus are much smaller than those of cirrus uncinus and fall much slower. However, these clouds might serve very effectively as sources of ice nuclei if physically joined by lower cloud layers.

The Chicago group also reported the discovery of a class of polar front rain storms which resulted from cirrus seeding of middle level clouds.³ It was observed that a disordered field of supercooled middle clouds—Ac, As, Ac cas—could be organized by cirrus seeding into a single circulation cell capable of giving light to moderate rain.

³ Braham, R. R., Jr., 1967: Cirrus cloud seeding as a trigger for storm development. Tech. Note 36, Univ. of Chicago Cloud Physics Lab., 15 August 1967. Also presented before 48th Annual Meeting, American Geophysical Union, Washington, D. C., 17 April 1967.

The only suggestion for an ice particle multiplication process which has received serious attention is that of splintering of a freezing drop. Mason and Maybank (1960), Schaefer (1952), Mason (1956), Langham and Mason (1958) and others have found that under certain conditions a freezing drop could shed many small fragments of ice. Usually the origin of these was attributed to mechanical fracturing of an outer shell of ice under the stress of inward freezing. Koenig (1966) accepted the splintering process and argued that in a cloud a few initial ice particles might collide with supercooled drops, causing them to freeze and release several splinters, thus setting up a chain reaction which could completely glaciate a cumulus cloud. More recent studies by Dye and Hobbs (1967) cast doubt on the importance of the splintering process by giving reasons for believing that the conditions of the original experiments might not apply to natural clouds. Thus we must await additional evidence as to the importance of the splintering concept.

The possibility that snow crystals and flakes might fracture and break apart in clouds has often been mentioned. But lack of evidence for a force acting on the scale of individual crystals and capable of breaking them has blocked acceptance of the idea. At the other hand collections of particles inside clouds strongly support the operation of some sort of a crystal break-up process. In our cloud studies we have collected snow pellets containing crystal fragments which can easily be distinguished from crystals that have been fractured in the process of collection. The author suggests that crystal break-up is common in mixed phase clouds as a result of collisions between crystals (probably dendritic) and large cloud drops and drizzle drops. The relative falling speed between a dendritic crystal and a drop of 500 micron diameter is about 2 m sec^{-1} . If the drop is already frozen the impact might shatter the crystal. If the drop is unfrozen both mechanical and thermal stresses resulting from the rapid freezing might be sufficient to fracture a fragile crystal.

Principle of vertical continuity

For an appreciable amount of precipitation to occur the clouds from which it falls must have vertical continuity over deep layers. In large, well organized precipitation areas each atmospheric level has a special role to play. Because of their low temperatures the upper levels provide a supply of ice crystals which serve as precipitation embryos for lower levels. The middle levels (temperature ca -15°C) provide a maximum rate of crystal growth. Clumping appears to maximize between 0°C and -10°C . The lowest, and wettest cloud levels contribute virtually all growth by riming and coalescence. It is easy to show that a particle growing by particle accretion in a uniform cloud will increase in mass in proportion to the cube of the distance fallen. This means that about one-half of the total growth will occur in the lowest 20% of the cloud while about 30% will be accreted in the bottom 10% of the cloud.

FIG. 8. Diagram suggesting interactions and feedback-loops between precipitation physics and cloud dynamics.

Interaction and feedback

This discussion began with a simple statement about the formation of a cloud and has progressed through precipitation processes of increasing complexity. It seems appropriate to conclude it with mention of a major frontier area in meteorology.

It seems obvious to me, as a cloud physicist, that we have gone about as far as we can go in studying precipitation processes by considering them only in the domains of physics and chemistry. Real progress must now involve study of the interactions and feedback loops between the physics of precipitation and the dynamics of clouds that produce precipitation. Fig. 8, which is self-explanatory, is a first feeble step toward identifying and labeling the ways in which these two areas interact.

For many years cloud physicists studying precipitation and meteorologists studying atmospheric dynamics have gone separate ways with little motivation to work toward common ends. Fortunately this situation is beginning to change. The change is being prompted by two very different developments.

The first of these was the demonstration that ice nuclei seeding can cause some cumulus clouds to grow through stable layers that otherwise would limit their vertical development. Early in the days of cloud seeding both American and Australian experimenters occasionally observed clouds which grew rapidly following seeding. But it was only after Simpson *et al.* (1965) and Davis (1966) had developed numerical models of cumulus cloud dynamics which could be applied to clouds in field experiments that we began to develop a useful understanding of the interrelations between cloud physics and cloud dynamics.

The second development which will force dialogue between cloud physicists and theoretical meteorologists was the acceptance as a national goal of the objective

of marked improvement in weather forecasting over periods of one or two weeks (Project GARP, 1967). This objective makes it mandatory that the *in-situ* atmospheric energy sources be added as an input parameter in the numerical forecasting schemes. Two of the most important of these sources are the latent heats of condensation and sublimation which precipitation leaves behind in the atmosphere. Ways must be found to extend our knowledge about clouds and cloud parcels to entire cloud fields and to frame this knowledge in ways useful to those concerned with global wind circulations.

Only through serious cooperative study by specialists in many branches of atmospheric sciences can we achieve this national goal which we have accepted.

Acknowledgments. The author is indebted to a large number of colleagues whose research has helped formulate our present understanding of precipitation physics. In a survey paper such as this it is not possible to cite all pertinent literature. References have been limited to those providing additional background to the key points. The author also wishes to express appreciation to Dwight Kline, ESSA, for permission to use Fig. 4, and to Miss Maureen Dungey who assisted in drafting figures.

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