PHYSICS OF PRECIPITATION

PROCEEDINGS OF THE CLOUD PHYSICS CONFERENCE
WOODS HOLE, MASSACHUSETTS
June 3–5, 1959

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Edited by
Helmut Weickmann

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Preface

The Second Woods Hole Conference on Cloud Physics, June 3-5, 1959, was devoted to the subject Physics of Precipitation. This volume contains the papers which were presented during the sessions as well as the edited discussion remarks.

Two antagonistic requirements exist which make a planning of a conference very difficult: (1) one has to be able to discuss the details of the subject matter and (2) one would like to understand the matter in a larger frame and in its general significance. In modern science and perhaps especially in cloud physics the details of the subject branch out widely into other fields in which the scale and movements of atoms and molecules form the principal size parameters, whereas the larger frame of cloud physics, which is mainly connected to meteorology, of necessity ends in patterns determined by the scale and movements of the world-wide atmospheric circulation. In the past the large-scale and macrophysical aspects of cloud physics had often been neglected on account of the emphasis which was put on microphysical processes of clouds, and in clouds and cloud particles. In the planning of this Conference, an effort was therefore made to do justice to macrophysics and to reconcile this range of \(10^{15}\) orders of magnitude by illuminating the subject matter from three different vantage points: (1) from the scale of synoptic meteorology, (2) from the scale of clouds and cloud systems, and (3) from the scale of microphysics. In agreement with this planning three main sessions evolved: Morphology of Precipitation, Clouds and Cloud Systems, Morphology of Precipitation and Precipitation Particles, and Fundamental Precipitation Processes.

It was then considered that, with the groundwork prepared, a fruitful discussion of the problems connected with Artificial Precipitation Control will evolve which would give a fourth session.

The letter of invitation sent to the participants outlined this plan. After the proposed papers had been received, it turned out that a special session on Hail Formation could easily be formed.

The location chosen for the meeting was again the Woods Hole Oceanographic Institution. This was done in order to emphasize the significance of cloud processes at the water-air interface for the supply of water vapor in the planetary circulation. These processes are one of the objectives of the research work of that Institution which has brought about a series of significant studies on microphysical as well as macrophysical processes connected with the formation of clouds or precipitation. The fine hospitality received from this Institution is gratefully acknowledged.

With the stage thus set for an interesting conference, it was exceedingly fortunate to have not only many of the cloud-physics experts of the United States but also a number of the foremost scientists from abroad participating. There is, of course, one scientist who is most intimately connected to the subject matter since he is not only one of its foundings fathers but also one who is able to speak with authority on all three scales involved: Professor Dr. Tor Bergeron. It was therefore a privilege that he agreed to serve as Honorary Chairman of the Conference. We are glad to express our sincere gratitude for his most inspiring and stimulating participation.

We also want especially to mention and express thanks to Dr. G. Wolff and Prof. Dr. A. Goetz for their participation. These scientists are not immediately connected with the subject matter, but serve as consulting experts, the former in crystal physics and the latter in physical chemistry. Prof. Goetz was also representative of the American Physical Society.
Our gratitude is also due to Dr. H. Landsberg, then President of the Section of Meteorology, American Geophysical Union, Dr. Alan Waterman, Director, National Science Foundation, and Dr. Earl Droessler, Director, Atmospheric Sciences Program, National Science Foundation, for their support in making this Conference possible. Finally, we like to express our thanks to Mr. Waldo E. Smith, Executive Secretary, American Geophysical Union, for his untiring support and efficient help in all phases of the Conference from the initial planning to the final editing phase.

A word should be said with regard to the form of this publication. The practice of publishing the complete proceedings of a conference in book form is becoming increasingly popular in recent years, but is nevertheless also subject to criticism. It is pointed out that the papers could be published in current journals and that the discussions quite often and of necessity contain thoughts or ideas that are not always well ripened. In spite of this we considered this publication a worth-while undertaking mainly for two reasons: (1) The publication of the papers in one volume will convey to the reader not only much of the present state of the art of precipitation physics, but also of the special scope of this Conference, namely, the significance of this field within the larger scales of atmospheric flow patterns. (2) Many of the discussions greatly contributed to this aim and form therefore important links between the sessions, therefore, much of the stimulus for further research would be lost through their omission. This is particularly true of the very interesting discussions related to the last session on Artificial Precipitation Control.

It is therefore hoped that this volume conveys to the reader something of the stimulating atmosphere of this Conference which in the consensus of the participants was a full success. Professor Bergeron, the Honorary Chairman, called it in his closing remarks one of the best meteorological conferences which he had ever attended throughout his 40 years of professional work. The merits for this success are his own in view of his stimulating papers and his inspiring discussion remarks, and those of the authors for their interesting papers, of the chairmen for their able direction of the sessions (always being under pressure of time), of the participants for their stimulating discussions and endurance (one discussion period lasted until midnight), and last, but not least, of all who took part in arranging and setting up facilities and schedules.

On the background of this cooperative effort, the goal of the Conference was achieved in a way which surpassed our expectations. Justice was not only done, in the lectures, to the various scales of precipitation physics, as outlined above, but justice was also done to the next higher level of information, namely, to show how tightly interwoven these scales can be with one another. What we mean is illuminated in the contributions which showed the close relationship between the size distribution and concentration of nuclei and the size distribution and concentration of the cloud droplets (W. A. Mordy, M. Neiburger and C. W. Chien, P. Squires and S. Twomey); and between the latter and the size distribution and concentration of raindrops precipitated from these clouds (P. Squires and S. Twomey). Thus for certain types of rain, no link is missing in the chain which connects the microscale with the mesoscale.

Another example from the papers which lead from the study of snow crystals to the fine structure of precipitation processes (U. Nakaya and K. Higuchi, J. Grunow, Ch. Magono), or from a microscopic study of the hailstone structure to an understanding of its life history within the hailstorm, and thus, ultimately, to an understanding of the hail mechanism (R. List, R. Sänger). These research projects open
new avenues in precipitation research where snow crystals and hail particles are being used as inexpensive aerological sondes of great resolving power. This may lead to a degree of understanding of the accompanying storms which otherwise cannot be achieved.

In other papers different scales were connected: the areal microstructure of the rate of precipitation (mesoscale) reveals the structure of hurricanes or other rain storms (T. Bergeron), and the arrangement of radar echoes points to the structure of orographic convection (B. Ackerman) and to the morphology of squall lines or fronts (P. Austin).

Finally, there are the papers and comments which brought the important connection to the synoptic scale, as in the study of cloud distributions over the tropical oceans (C. Ronne and J. Malkus), in the synoptic analysis of the fine structure of convective storms (T. Fujita), or in the analysis of large scale flow patterns which lead to rain or drought (J. Namias), or in the very interesting problems which precipitation processes present to numerical forecasting (J. Smagorinsky).

The reader will have noticed that here the significance of cloud physics is illuminated not just in the light of two scales—microphysics and macrophysics—but in three according to Bergeron's definition (see page 61): (1) microphysics dealing with processes related to cloud elements, (2) mesophysics dealing with processes ranging from individual clouds to a cloud system, and (3) macrophysics dealing with processes of synoptic scale. Such a definition would be, as Bergeron pointed out during the Conference, in better agreement with current use in meteorology of the terms micro, meso and macro.

Since it is virtually impossible to discuss the amount of valuable information and inspiration which has come from this Conference, the post-Conference status of some key problems which we had mentioned in our letter of invitation shall shortly be appraised. These were:

(1) Can we prove that true sublimation nuclei do not exist and that AgI acts only as freezing nuclei?

B. J. Mason's paper gave an almost complete answer: he showed that AgI not only acts as a freezing nucleus initiating nucleation at water saturation, but also as a sublimation nucleus at temperatures below \(-12^\circ\text{C}\). It is not an ideal sublimation nucleus but still requires 10–15% supersaturation with respect to ice. (This result was qualitatively anticipated in Anderson's paper, Re-evaporation Ice Nuclei, which he gave at the First Woods Hole Conference.) Mason's investigations also explain why hygroscopic substances cannot act as freezing nuclei but may act as sublimation nuclei. It appears now that one of the last unsolved questions is how Nature performs in bulk water, the act of crystallization at the freezing point proper.

In addition to this work, investigations on concentration and efficiency of freezing nuclei have been reported by H. W. Georgii, D. B. Kline, and S. J. Birstein. The diligent work of these scientists who are hampered by bulky, unconventional, and heavy equipment which demands great observing skill leads us into an ever-deepening understanding of the microcosmos around us which occupies a key position in the life history of raindrops or snowflakes. Their work furthers not only our understanding of the fundamental significance of nuclei for the precipitation process but also for an appraisal of the influence of man-made air pollution on precipitation processes (see also the paper of J. P. Lodge).

(2) How much does the aerosol spectrum influence cloud structure and therefore also precipitation processes?
This question was answered through the contributions of W. A. Mordy, M. Neiburger and C. W. Chien, Cl. Rooth, and P. Squires and S. Twomey. While the authors of the first three papers showed theoretically and through the application of electronic-computer techniques the close relationship between the aerosol spectrum and the cloud droplet spectrum, the last-named authors confirmed this through investigations of the cloud-droplet spectrum in Cumulus clouds over continents and over the ocean. They were furthermore able to show that not so much the presence of giant nuclei as this difference in the microstructure of the clouds is the determining factor in the formation of warm rain. These papers must be regarded as mile stones for our understanding of cloud formation and of the warm-rain mechanism.

As these papers will no doubt lead to an increased research effort into the nature and spectrum of condensation nuclei it was very timely that in the same session Dr. Goetz discussed his aerosol spectograph. It is an ingenuously designed centrifuge that yields just that size range of aerosols which is crucial for the condensation process.

(3) Is the evolution of a soft-hail, hail, or warm-rain particle a process which is traceable back to a distinct initial particle or to just one process, or is it a process whose trail we lose in a great number of eligible particles which have accidentally met after having experienced various life histories?

This question has been answered for the warm-rain particle in the aforementioned paper by P. Squires and S. Twomey. According to these authors the warm-rain drops do not go back to a distinct particle (giant nucleus), but to a population of cloud drops which is favorable for the evolution of a warm-rain particle through coalescence, or, in other words, for the release of colloidal instability in form of rain. In the case of the soft-hail and hail particles R. Sänger and R. List showed convincingly in beautiful cross sections through hailstones and graupel that both, large droplets as well as snow crystals may constitute the initial particle around which coalescence proceeds.

(4) What do we know of Nature's efficiency of the precipitation processes?

This problem was illuminated in R. Wexler's stimulating lecture on this subject and in the ensuing discussion. The problem of the efficiency of the natural rain process is of prime importance not only for our understanding of the rain mechanism, but more so for our appraisal of artificial precipitation control. It is one of the central problems of cloud physics and of meteorology in general, and it was in order to underline its significance that Wexler's lecture appeared in a central position in the program. Bergeron's Operation Pluvius, Grunow's, Magono's and Nakaya's raindrop and snow-crystal analyses, Hallgren and Hosler's, and List's laboratory investigations will ultimately contribute to its solution as well as the excellent work on the physical evaluation of seeding activities in Santa Barbara, California, on which Todd reported. Elliott's discussion of the Wexler paper indicated again the great importance of a simultaneous consideration of various scales as the efficiency of the natural rain mechanism may come out quite different if the budget of the water vapor is considered for a whole cloud system or for a micro-element within the system.

(5) What do we know of the collection efficiency of snow crystals and flakes which seems to play a very important role for the efficiency of both, continuous and convective, precipitation?

The collection efficiency of ice spheres for crystals has been discussed in the paper presented by Hallgren and Hosler. In their observations the importance of the temperature interval near the freezing level was emphasized owing to the formation
of a water film on the crystal surface which facilitates sticking of two crystals to one another after they have collided. We may conclude, then, that aggregation of snow crystals is also enhanced inside a water cloud at low temperatures, which is in agreement with the observation that the structure of the largest snow flakes is always a mixture of crystals and frozen or unforzen cloud or drizzle droplets. In a water cloud, of course, the attraction between ice and water particles caused by the vapor-pressure difference according to Vierhout’s theory improves aggregation between the snow crystal and the cloud droplets. After the capture of a cloud droplet, Nakaya explained (p. 270) that this may either freeze to the ice surface or evaporate on to it in a strange process that resembles the floating of little water drops over a clean water surface which was first described by O. Reynolds (Papers on Mechanical and Physical Subjects, 1869–1882).

The growth by accretion in the ice phase was also the subject of a theoretical paper given by Douglas. He investigated the growth of spherical particles by sublimation and accretion. It would be most interesting if similar investigations could be carried out for simple geometric forms as they actually occur, for instance, stars, prisms, several ‘stars’ interlocked with each other, etc. As in falling, air will pass through the stars, they may collect many more cloud droplets than a spherical particle and therefore attain a considerable greater growth rate than was found by these authors.

(6) What do we know of hail formation; is it due to a very intensive or very persistent updraft?

Very little definite evidence is available on this subject. When one compares the characteristics of hail clouds with simple thunderstorms using radar data as was done by Donaldson, Chmela, and Shackford, evidence points to the significance of very powerful updrafts. This is also expressed in Beckwith’s report on severe hail damage to aircraft at altitudes at or near 40,000 ft. The extent and pattern of hail damage, derived from crop insurance records, has been successfully used by Stout, Blackmer, and Wilk for a study of path and duration of hail storms, but the data are not detailed enough to yield the life history of a single cell.

(7) What is the significance of the zones of high winds (jet streams) which appear to be a typical feature of hailstorms?

This subject was discussed in a paper by Dessens which was followed by a stimulating discussion. No doubt the problem is still with us to search for a physical explanation if the relationship between hailstorms and jet streams can be verified. Data given by Beckwith indicate that only 18% of hailstorms occurred simultaneously with the jet stream. No criterion exists so far how to judge the significance of this figure. In this search, investigations as reported by Hitschfeld will be of immense value as they will help us to analyze the true trajectories of particles which are potential hailstones. Of particular interest was here the discovery that the storm column remains upright in even a strong wind shear in contrary to the typical bending over of trade wind Cumulus clouds. The influence of vertical wind shear on thunderstorm systems has been discussed in an interesting theoretical paper by Newton who found that small clouds may be destroyed whereas large convective systems may be maintained by vertical shear. Also, Anderson discusses theoretically the formation of self-sustaining storms, while Cunningham showed beautiful pictures of a hailstorm which had been taken from high-flying aircraft.

This completes the discussion of the specific problems which we had raised in the letter of invitation, but we call the readers’ attention also to the stimulating papers given by Essenwanger, Volz, Vonnegut, and Moore. These papers give valuable
background information to the scope of the Conference: Essenwanger discusses the possibilities of a physical interpretation of frequency distributions of precipitation, Volz shows what the rainbow can tell about oscillations in freely falling raindrops, and the papers by Vonnegut and Moore discuss the always stimulating problems of precipitation and atmospheric electricity.

The reader is also referred to the extremely interesting session on Artificial Precipitation Control (papers by Orville, Sänger, Dessens, Todd, Battan and Kassander, Howell, and Semonin) which was one of the highlights of the Conference not only on account of the stimulating scheduled papers but also on account of the inspiring discussions. While it is impossible to do justice to the various new ideas and concepts that evolved in this session, this appears to be the place to recall the suggestion, made repeatedly by the Honorary Chairman throughout the Conference, to include in the evaluation of cloud physics or weather-radar projects a good synoptic analysis of the corresponding weather situations. We feel that this suggestion is of special significance in the evaluation of rainmaking or weather-control experiments into which statistical methods have entered deeply. It was therefore refreshing to find that such work is already well underway in the physical evaluation of seeding activities in Santa Barbara, California. Indeed, these investigations will lay the ground work for a new phase in the design and evaluation of seeding experiments which will be greatly superior to and finally end the era where design and evaluation of such experiments were governed by statistical methods only.

It is of necessity that all problems discussed in this Conference are still more or less in the basic-research state; and that they cannot as yet be applied to the chief objective of meteorology, namely towards improving weather forecasts in general and precipitation forecast in particular. The fact, however, that the discussions were not only inspired by the participating cloud physicists, but also, and sometimes even more so, by the scientists working in synoptic meteorology and numerical forecasting methods indicates that a most valuable cross fertilization between cloud physics and synoptic meteorology is underway and was achieved to a high degree during this Conference.

The Editor, as Chairman of the Cloud Physics Committee, acknowledges the splendid support and assistance given to him by the Committee on Cloud Physics, namely, Charles E. Anderson, Roseee R. Braham, Jr., Dwight B. Kline, J. E. McDonald, Joanne S. Malkus, and Vincent J. Schaefker.

Dr. Helmut Weickmann, Editor
Chairman, Cloud Physics Committee
American Geophysical Union

Asbury Park, N. J.
December 1959
Welcoming Address on Behalf of the American Geophysical Union

HELMUT WEICKMANN

Chairman, Cloud Physics Committee

Ladies and Gentlemen: I am very happy to welcome all of you on behalf of the American Geophysical Union to our meeting on cloud physics. As I look around in our big family here, I am especially happy to see Dr. Tor Bergeron, our Honorary Chairman. He has inspired modern meteorology for over 40 years, and we appreciate his coming over from Sweden on this long trip. We also have representatives of Australia, Canada, England, France, Germany, Japan, and Switzerland, and others from Sweden. I would like to ask all of you to get acquainted quickly. Let us be together as one big family of scientists, of researchers, and let us not forget the ones who are unable to participate because of a very involved procedure of invitation: our friends and fellow scientists from behind the so-called Iron Curtain.

I am very happy to see some experts in fields related to the subject of the Conference, such as crystal physicist Dr. G. Wolff of the U.S. Army Signal Laboratories, and chemical physicist Dr. A. Goetz of the Physics Dept., California Institute of Technology, who is also representing the American Physical Society. I might also call Dr. J. Namias and Dr. J. Smagorinsky related experts. Both are well known to us. By inviting them we hope to extend a little bit our view of the precipitation processes of clouds, and to consider the problem of precipitation within the synoptic scale.

I am especially happy that the Woods Hole Oceanographic Institution has played the host again. I wish to thank its Director, Dr. Paul Fye, for his fine hospitality and for his and his associates' assistance in the organization of this large meeting in a relatively small community.

There certainly is an idea behind having the meeting here in Woods Hole. The Woods Hole Oceanographic Institution is one of the few institutions which is actively engaged in research of both micro- and macro-physical processes of cloud and precipitation mechanisms. Moreover, it is the water-air interface around which the studies center. This interface is not only very much larger than the ground-air interface, but it is also an extremely important one as the rain which we study ultimately stems from it. The important role which clouds on this interface in the trade-wind region have for the water vapor supply in the large-scale circulation is one of the outstanding results of this Institution's studies of trade-wind cumulus clouds. The Woods Hole Oceanographic Institution is therefore a most fitting meeting place for our Conference as it will provide a scientific atmosphere which shall greatly stimulate the discussions and aid in achieving the purpose of this Conference. This purpose is, I should say, to look for a true science adventure and not just for a science fiction story.

In order to achieve this goal we have tried to arrange the program in a
special order: we proceed in three circles to the fundamental problems. In the outer ring we get acquainted with processes in the large synoptic scale which are conducive to precipitation; in the next ring we will discuss the morphology of the precipitation on a medium scale and, thus prepared, we proceed to the inner ring with discussions of the fundamental precipitation processes.

Imbedded here is a separate discussion of the processes of hail formation. As the knowledge of fundamental principles always has to precede the practical application of these principles it is only logical that the discussion of artificial precipitation control is the concluding session of our Conference.

This Conference was made possible through a grant from the National Science Foundation to the American Geophysical Union. We are therefore very much indebted to the Program Director for Atmospheric Sciences, Dr. Earl G. Droessler, for his share in the realization of this endeavor.
Welcoming Address on Behalf of the National Science Foundation

E. G. Droessler
Program Director for Atmospheric Sciences, National Science Foundation

Ladies and Gentlemen: It is my special privilege to join in these words of welcome to you and to bring you the greetings of Dr. Alan T. Waterman, the Director of the National Science Foundation. All of us who in any way helped with the arrangements for this Conference are highly gratified by the enthusiastic response you have shown by your attendance. I believe we can certainly predict a most successful meeting.

Early suggestions for this meeting grew naturally out of the very fruitful First Conference on the Physics of Cloud Precipitation Particles held about three years ago at Woods Hole. Considerable has happened in the interim.

In Washington we have seen the work of the President's Advisory Committee on Weather Control brought to an orderly close with a final report which stressed the importance of basic research and recommended that the Government give full encouragement and support to the widest possible competent research as the surest, most direct way to success in any attempt at modifying the weather.

In July of 1958, central responsibility for Federal support of research and evaluation of weather modification was given by the Congress to the National Science Foundation. A new program was established with the objective of studying more intensively and systematically the scientific basis of weather modification, through support of competent scientists working in cloud physics and allied fields.

Accordingly, we were pleased when one of the first grants approved by NSF under the new program was awarded to the American Geophysical Union for the support of this Woods Hole Conference; for we believe the bringing together of the world's foremost researchers in the field at a meeting such as this represents an important step forward toward the program objective.
Welcoming Address on Behalf of Woods Hole Oceanographic Institution

PAUL FYE

Director, Woods Hole Oceanographic Institution

Dr. Bergeron, Ladies and Gentlemen: May I welcome you most heartily to the Woods Hole Oceanographic Institution. We are delighted to have you in Woods Hole and hope you will all come back again often.

Permit me to tell you a story about some of our proceedings yesterday when we entertained the Congressional Committee which is studying oceanography. We had, in addition, a number of reporters who were most impressed by the presentations of our staff and, in particular, the one made by Dr. Joanne Malkus. In addition to attending the meeting of the Congressional Committee they learned that there was a distinguished group of world scientists meeting here today. They were anxious to talk to me, and find out something about your meeting. When they realized this was a group of experts in cloud physics they wanted to know if they could say in the papers that you chose Woods Hole because of its well-known beautiful weather and because of the beauties of Cape Cod. About that time the rain clouds began to gather and they appeared to get a little worried. It was then pointed out to them that this group had met in Woods Hole several years ago and that the weather you had for your meeting was the exact opposite of what had been experienced during the rest of that summer. So I understand that during the night the Chamber of Commerce has been breathing easier, that they hope the cloudburst we have had for you so far is only an indication of better things to come for the rest of the summer. What will be said in the papers, I have no idea. Over this, we exercise no control, artificial or otherwise.

If I were to try to tell you why the Woods Hole Oceanographic Institution is interested in meteorology or why we have groups like those working with Dr. Malkus and Mr. Woodcock, it would be like an amateur shoemaker trying to tell professional cloggers how to make shoes. I have no intention of doing that. We hope you will return and that your conference is a very profitable one.
Mr. Chairman, Ladies, and Gentlemen:

It is a great honor and pleasure for me to have the opportunity of taking part in this Conference on Precipitation, a field that has always been of primary interest to me. I want to thank the American Geophysical Union most fervently for the confidence they have shown by making me Honorary Chairman, thereby enabling me to come to this interesting Conference.

I state the Conference is to be very important, since its theme, as we know, is of utmost importance to mankind, and because the papers and abstracts already available show that a mighty attack on the named problem is imminent. The names of the authors and the fact that the program allows time for proper discussion of each paper seems to ensure, if this plan is followed, a very good result of our joint attack.

In 1875 the great physicist and physiologist H. von Helmholtz gave a lecture on "Cyclones and Thunderstorms." He felt that the impossibility of predicting the beginning and end of rain touched a sore spot in the minds of physicists. He concludes thus:

"Under the same sky in which the stars pursue their orbits, as symbols of the unchangeable lawfulness of Nature, we see the clouds towering, the rain pouring, and the winds changing, as symbols—as it were—of just the opposite extreme, the most capricious of all processes in Nature, impossible to bring inside the fence of its Laws."

However, at Helmholtz's time, nearly a hundred years ago, one only vaguely realized the vital role of these phenomena to mankind. Now, we know how the clouds and their precipitation determine unambiguously the life conditions of the better part of the so-called biosphere, the sphere of living beings. We know, for instance, that the photosynthesis, which is controlled by actinic radiation, and on land by the fresh-water supply from precipitating clouds, is by far the biggest industry on our planet. It outrivals by several orders of magnitude any kind of artificial production of energy on our Earth. We dread the catastrophic of soil erosion, caused both by the excess and lack of rain, as you know very well in the United States. We badly need the hydroelectric power, and we now fear the sinking ground-water level; not to speak of other nearby effects of rain or drought.

But today, our colleagues the physicists have, on the whole, very successfully repressed that sore point from their conscience, leaving us meteorologists almost alone with these vast, intricate and vital problems, which we evidently still have not mastered. (I need only to refer to the last 24-hour rain that we have all experienced here, or perhaps in Boston. It may have been forecast to some extent, probably as some showers; but it was certainly not forecast in this amount and of this kind, that is, as a 24-hour abundant rain.) Intense or steady precipitation will generally be the real vital and fatal type, of course. Therefore, it seems reasonable that we now devote our attention to mechanisms producing such precipitation, even if the weak or intermittent type also may be of considerable interest. We may hope to find the most clean-cut and typical mechanisms, the most conspicuous factors, behind the intense weather phenomena. This will help the scientific treatment and solution of these meteorological problems, which otherwise often are too complex. These typical cases will also, if treated artificially, point to the best methods of getting artificial precipitation, or to prevent certain kinds of precipitation, as for instance hail.

Abundant precipitation will only be produced by what one might call sustained mechanisms; that is, mechanisms producing more or less orderly lifting of air, bringing new moisture and new releasing particles (appropriate nuclei, etc.) continually through a certain limited space of the lower troposphere. These limited spaces will either be fixed geographically, or they will remain stationary with respect to the leading airflow, as is the case with frontal rain and rain of convective character. In any case, the main
thing is going on in the lower troposphere as far as the precipitation amount is concerned. Also other forms of precipitation are important to study: especially those that give unexpectedly small amounts, for example, smaller amounts than one would expect under conditions that generally would guarantee a substantial rain.

One has hitherto distinguished between microphysics, macrophysics, and synoptics of clouds. However, I think it would be better to use the terms micro-, meso-, and macrophysics (corresponding roughly to cloud particles, individual clouds, and cloud systems) since the name 'synoptics of clouds' introduces an asymmetry, and since one is apt to think that 'micro' and 'macro' are for us, but 'synoptics of clouds,' that is for the weathermen, so we will not bother with that. In fact, these things hang so intimately together that they cannot be solved separately. We must all collaborate with these problems.

Then, another point that I believe is important. Every cloud or cloud system giving abundant rain or snow can be divided into a releasing part and a spending part. I call it the releaser and spender part of the cloud or cloud system, or the 'seeder' and the 'feeder.' They may not even form parts of one and the same cloud; they may be two different clouds, one above the other, cooperating in some way or other; and they must really cooperate in order to insure efficient precipitation. A cloud without a spender will give very little or no precipitation. A cloud without a releaser will generally give no precipitation at all, or only drizzle, even if the condensation is

Fig. 1—Main precipitation mechanisms; schematic vertical cross sections
Fig. 2—Weather map with convective system over eastern U. S., April 10, 1947, 18h 30m CST
Hourly rainfall
10. IV. 1947
18-19h L.T.

Fig. 3—Precipitation intensity, April 10, 1947, 18h 30m CST
rather intense. And that is again a reason why our problems cannot be solved by micro-, meso-, and macrophysics of clouds working separately.

I am now going to discuss some important types of precipitation and their mechanisms. Figure 1 tries to show schematic vertical cross sections of the main mechanisms of cloud and precipitation (excluding the drizzle from fog and Stratus); and by chance the size of the different small pictures roughly corresponds to their relative importance on the globe.—The second row shows the convective system, which, together with the ordinary convective clouds of the first row, gives the better part of the precipitation in tropical regions, and even in middle latitudes on land in summer. Correspondingly, over the middle-latitude oceans, the convective clouds give the bulk of the winter precipitation. (In the Middle West, where the convective systems or 'squall-lines' bring most of the spring and autumn rains, the warm-fronts then seem to bring relatively small amounts of precipitation.)—The third row contains the opposite kind of mechanism, the tropical hurricane, which brings great amounts of precipitation on the oceans and adjacent coasts in late summer and fall. In this figure there is an outward similarity between the convective system and the tropical hurricane, but at the same time they are opposite. Here I only want to point out that they contain two 'circulation wheels' each, a big one $C_1$, and a smaller one $C_2$ that is caused by the rainfall cooling. If they cooperate, as within the tropical hurricane, a very efficient mechanism is set up producing a cyclonic vortex, often with extremely low pressure at the center. In the convective system, on the other hand, these two circulations counteract each other; $C_1$ causes a

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**Figure 4**—Isochrones of the pressure jump of June 1-2, 1951, according to M. Tepper
small high in the interior of the convective system, thereby preventing the formation of any cyclone. Instead, a low-pressure trough is formed just outside this small circulation wheel, as shown lately in detailed studies by Dr. Fujita.—The lowest row of Figure 1 displays two kinds of orographic cloud and precipitation systems, the high-reaching and the low one, and also the frontal precipitation mechanism. However, integrated over the whole globe these three mechanisms give much less precipitation than the mechanisms of the three upper rows.—The next figures give examples of the mechanisms in Figure 1.

The synoptic map of the eastern United States (Fig. 2) shows a convective system that is bordered at the eastern edge by what you call a 'squall line'. This is, in fact, a pseudo-cold front since the cold air is produced by the intense rain-cooling caused by the convective system that has formed within the tropical-air warm-sector of a cyclone. The main precipitation of this cyclone came from that exceedingly important mechanism. When, in 1947, I first had made the analysis shown in the cross section of Figure 1, second row, I 'forecast' that the maximum precipitation intensity would occur along the leading edge of this convective system, followed by a minimum in the interior because of cloud decay caused by the rain-cooling there; then
the precipitation intensity would pick up again at the rear edge of it.

Some years later I was able to confirm this assumption or 'forecast' as seen in Figure 3, which shows the distribution of rainfall during one hour from this very convective system. At the leading edge there is a rainfall intensity of up to two inches per hour. In the interior of the convective system the rain intensity goes down to zero over considerable areas, and further west it reaches values of at least one tenth inch per hour, proving the expected increase at the rear edge of the system. This whole mechanism moved across the eastern part of the United States, giving from one to eight inches of rain within a few hours. Seemingly, this kind of mechanism is the main rain-spender during the summer-half of the year in the Middle West. Thus, the convective system is a most important mechanism, worthy of all those very minute and ingenious investigations that Dr. Fujita has made; and also, of course, of being tackled by many more investigators.

From Dr. M. Tepper I got data on the 'pres-
sure jump' of another case with similar characteristics: June 1-2, 1951. Figure 4 shows his isochrones of this pressure jump over Nebraska, Iowa, Kansas, Missouri, and Oklahoma; thus it covered quite a vast region. After having studied the numerous barograms from this case, I do not doubt the reality of this pressure jump; although, naturally one needs not accept the various explanations given. In the United States there exists a wonderful network of pluviographs, and their registrations have been worked up and published in the shape of hourly precipitation data, this being the only country in the world that has done anything similar. With the aid of these data, I was able to analyse, in 1953, a series of maps, of which Figure 5 shows a sample. As you see, the results are similar to those reached by Dr. Fujita in 1955, so our works corroborate each other quite nicely. In this case there was a long quasistationary Polar front, with a cold-front section extending towards SW over Kansas. The warm-front section, over Nebraska and Iowa, gave only weak rain. At the top of the warm sector within the Tropical air, in SE Nebraska, there was, however, a circular convective system with intense rain along its outer edge, but with very little or no precipitation in the interior, and a small High at the very center. All this could be traced by means of the above-mentioned hourly precipitation data. The maps also show the position of the pressure jump taken from Dr. Tepper's map (Fig. 4).

We do not know for certain the origin of convective systems; but they generally form in the warm sector, and often at its top. Sometimes the 'squall line' is parallel to the cold front; but in such a case as this, with a circular pseudo-front, I am not able to accept the old explanation that a 'squall line,' or convective system, should be caused by the overflowing of cold air at a higher level. Then, why does it not extend down through Texas in this case? In fact, the explanation of the convective system may be quite another one in this case, and we are on the safe side if we start by observing it, then digest and carefully analyse it, and save our explanation for a later stage. What I wanted to underline here is that the pressure jump at the stage of the 00h 30m z map (Fig. 5a) lies far behind the 'squall line' and also behind the cold front. Whereas, nine hours later (Fig. 5b), the pressure jump has mostly overtaken both the ordinary cold front and the pseudo front. The convective system now forms an almost circular rain area, with hardly any rain at the center. On later maps, the pressure jump even went further into the Tropical air and left both the ordinary cold front and the pseudo front behind it. Thus, one may conclude that in this case the pressure jump was produced, neither by the 'squall line,' nor by the cold front: they were in this case three more or less independent agents. Certainly, these phenomena are worthy of many more investigations in the future.

Palmén has shown that the tropical hurricanes form and are maintained according to the convective theory. In Figure 6 the curve of convective lifting, in February, when there are no hurricanes, lies mainly to the left of the curve of stratification. The corresponding curves for September, in the middle of the hurricane season, show that there is then a great positive area between the curve of lifting and the curve showing the stratification of the air. What I especially wanted to emphasize is that the curves of lifting diverge with height. Thus, a certain difference of temperature at the Earth's surface corresponds to a two or three times greater difference of temperature at the tropopause, or say at the 300-mb level. This implies that a surplus of 2°C in the rising air at the cloud base will give
In left part of vertical cross-section

Air 1°C cooler than the average for this level
Precipitation/Cloud
Air-current at heat source
Air-current at heat sink

In central and right part of vertical cross-section and in ground plan

Main updraught region
Heavy rain
Light to moderate rain
Isotherm °C
Isentropic line °C

Fig. 7—Mean vertical cross section of the major tropical hurricane passing SE U.S., Sept. 17-20, 1947, according to E. Palmén and T. Bergeron
a surplus of 4 to 6°C at the top of the hurricane cloud, thereby increasing greatly the lability energy. If, on the other hand, the temperature is depressed a little at 0-level, there will be a great loss of energy.

Now, we come to the fact that there is generally a very low pressure at the center of a mature tropical hurricane. This is an important point, to which enough attention has not been paid hitherto. Thus, if the central pressure is lowered, the curve of lifting will move to the right in the diagram, and the lability energy will be increased, provided the condensation level does not rise or the temperature fall. In fact, the air flowing into the Low within the frictional layer over the warm sea will conserve its high temperature in spite of the expansion. Its temperature will keep near that of the sea-surface, thanks to the very rapid transfer of heat from this surface up into the air, and the sea-surface cannot by any means be cooled appreciably in a short time. Thereby the wet-bulb potential temperature of the air is raised. By making, so to say, the pressure very low at the center, the hurricane will dispose over a greater store of lability energy.

Figure 7 shows a vertical cross-section of the major hurricane that passed Miami and New Orleans in September 1947, which Palmax and I have both been treating. A decisive thing with the hurricane mechanism, as I see it, is that the updraft of the inflowing air takes place inside the region of the most intense precipitation, as indicated in the small map-sketch at the bottom of Figure 7 and in the third row of Figure 1. The central hurricane cloud-mass is often funnel shaped; thereby it will not attain enough height to produce an efficient precipitation release (through ice-nuclei or otherwise) until at some distance from the 'eye,' presumably outside the ring of maximum updraft, a good example of cooperation between micro-, meso- and macro-physical cloud factors. In the opposite case the rain cooling, driving $C_1$, would counteract the mechanism $C_s$, and the hurricane would transform into one or more convective systems instead; this was probably what happened to the hurricane of Figure 8.

Unfortunately, I was not able to treat just the major hurricane of September 1947 from this viewpoint. Figure 8 shows another hurricane from the same year. When lying just outside the coast

Fig. 8.—Decay of a tropical hurricane over Georgia and Alabama, Oct. 14-16, 1947; $S$ indicates areas of rising pressure
Fig. 9—Maps of hourly rainfall from the major hurricane of Sept. 1947 when crossing Florida
of Georgia, it was still quite vigorous. At its center the sea-temperature was about +26°C, and the temperature of the air that was rising from the sea-surface was probably very near this value, corresponding to a great amount of lability energy. Twenty-four hours later the hurricane center had entered land and began to fill rapidly. This effect is generally ascribed to friction. According to my view, however, the main reason for the filling is the precipitative cooling of the air over land within the friction layer, that is, of the only air that can flow into the hurricane and form an updraft. That air was in this case cooled by the rain itself down to +19°C near the center, which implies that the lability energy had become negative. That is like putting a very powerful brake on the whole mechanism, similar to braking a motor car by the engine.

This is a precipitation conference, and the hurricanes were mentioned mainly because they also represent a very important precipitation problem. In fact, a lot of the hurricane damage is done by precipitation. Figure 9 displays the distribution of the precipitation intensity within the major hurricane of September 1947 when passing Florida. It is, of course, analyzed keeping the idea in mind that it should have an 'eye' without precipitation, and that the region of intense precipitation should form a ring; evidently it does. The several small maps show how the precipitation area moves across Florida, and also how it widens during this passage. Gradually it gets two eyes, or even multiple eyes, and parts where the precipitation in the ring is much more intense. Instead of a complete ring of maximum rainfall all sorts or irregularities occur, presumably because the precipitation area lies over land

Fig. 10—Maps of hourly rainfall from the same hurricane as in Fig. 9, developing spiral arms of precipitation
and is then unevenly affected by the rain-cooling. This might also partly explain why many writers have found irregular hurricane paths. In fact, a path need not be so irregular, it may only seem so because of the multiple eyes.

Radar has revealed to us the spiral structure of the hurricane precipitation. The hurricane-cloud spirals might at first be composed by separate convective cells, arranged in rows that are gradually dragged spirally into the Low, at last joining up into real bands. Figure 10 contains a continuation of the series of precipitation maps of Figure 9, showing that the precipitation pattern based on rain-gage data also may have a spiral structure. Certainly, it is founded on a rather loose network, so that in some cases one might be able to combine the places that have got precipitation in a different way and thereby fail to attain a clear spiral rain pattern. However, in the map of September 19, 02h EST, which also has isobars and winds, the direction of the spiral arms lies somewhere in between the direction of the wind at the Earth's surface and the gradient wind. It seems quite reasonable,

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**Fig. 11—Distribution of precipitation, wind, pressure, and temperature in SW Scandinavia during the period March 24–27, 1927 (72 hr), showing marked coastal maxima of precipitation mainly due to orographically conditioned convergence**
then, that they originated from showers that became arranged parallel to the direction of the general flow, merging into solid rain bands of spiral shape and eventually into one ring-shaped rain-area as they came under the influence of the general updraft. I think that detailed precipitation studies of tropical hurricanes might form a useful complement to the radar scanning of them, since, after all, the radar does not render the rain intensity as truly as the gages do. Moreover, the radar patterns do not show the total extent of clouds, but only the patches with efficient release of precipitation. Thus, they permit no estimate of the amount of latent heat released. In other words, the release of latent heat need not be so discontinuous as it may seem from the radar pictures.

The problem of orographic precipitation is old and much treated. Therefore, I shall here only enter on a most remarkable case of exceedingly abundant orographic precipitation that occurred over southern Scandinavia in the last week of March 1927, when a front lay quasistationary from the Black Sea over Poland and Denmark to Iceland, and there was frontal precipitation all over southern Scandinavia together with southeasterly winds. Figure 11 gives a close-up of the weather situation over southern Norway and the immense amount of precipitation, 200 mm (8 in.) that fell in exactly three days on the coast. That could not be ordinary orographic precipitation. In this case the static stability of the general southeasterly current made part of it bend outside the mountains and meet the direct southeasterly branch as a northeasterly current. Thereby an orographically conditioned

![Diagram](image-url)

**Fig. 12**—Vertical distribution of maximum condensation intensity $I_w$ in an upslide cloud system
convergence was set up at the SE coast of Norway, with a mighty updraft just at the coast, and consequently a very intense condensation inside the orographic cloud system.

Figure 12 shows the amount of precipitation to expect from an upside cloud system according to Fulk's formula, assuming +10°C at a cloud base lying very near the zero level. Such diagrams give a measure of the maximum condensation intensity $I$ under the assumption made as to the vertical motion shown by the curves $w$ in the figures. The maximum condensation in the cloud mechanisms in question in each unit layer (100 m thick) is obtained by the $Iw$ curve. Integrating the area enclosed by this curve and the zero line gives the maximum amount of precipitation available under the assumption that there was no entrainment, and that the precipitation release was 100%, that is, total and immediate. Naturally, in reality one gets much less, generally only about 50 or 60% of the theoretical maximum, which in this case is 0.9 mm/h. Since the temperature at the cloud base in this case was about 0°C (instead of +10°C) and the distance to the front line at the Earth's surface 100-200 km, one can expect at most 30% of the maximum value or not more than about 20 mm precipitation on the Norwegian coast from the upside surface during those three days. So we have to explain the remaining 180 mm by the above-mentioned orographically conditioned mechanism. A cross section SE-NW over Skagerak and the southern Norwegian mountains (Fig. 13) shows the frontal upside surface with the upside cloud system and snow falling from it. Within the cold air lies the orographically conditioned, local mechanism that causes a much stronger updraft and a much more intense condensation: the feeder cloud. The water condensed is brought down promptly by snow falling through this cloud from the overlying releaser cloud.

Fig. 13—Schematic vertical cross section of frontal and orographic precipitation mechanisms in Southern Norway, March 24-27, 1927
The curve \( I \) represents the specific intensity of condensation and \( w \) the assumption made as to the vertical motion, being zero at the ground and at a front surface, 1.4 km above sea level, and 1 m/sec halfway in between. We then get an \( Iw \) curve that corresponds to a precipitation of 5 mm/hr or double the amount needed, since there are 180 mm in three days to be explained if the other assumptions are correct. Thus, the updraft would have been of the order 0.5 m/sec instead of 1 m/sec (as assumed in Fig. 14), a value that goes well with the surface convergence observed.

Figure 15 illustrates about 50% of the rain in January, 1937, that produced the famous Ohio-Mississippi flood, which began in the Ohio River drainage basin. This isohyetal map only renders the rainfall during the period January 20-25, reaching 10-14 inches along the river; the maximum total sum for the month was about 24 inches. This is a terrific amount, and the flood was, as you know, one of the biggest on record, rising more than 25 m, or 85 ft, above normal.

Fig. 14—\( Iw \) diagram of orographically conditioned precipitation

Figure 14 is an \( Iw \) diagram for this case with two alternative temperatures at the cloud base, +10°C and 0°C, the latter applying to our case. The curve \( I \) represents the specific intensity of condensation and \( w \) the assumption made as to the vertical motion, being zero at the ground and at a front surface, 1.4 km above sea level, and 1 m/sec halfway in between. We then get an \( Iw \) curve that corresponds to a precipitation of 5 mm/hr or double the amount needed, since there are 180 mm in three days to be explained if the other assumptions are correct. Thus, the updraft would have been of the order 0.5 m/sec instead of 1 m/sec (as assumed in Fig. 14), a value that goes well with the surface convergence observed.

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Fig. 15—Rainfall in the Ohio-Mississippi Basin, Jan. 20-25, 1937
Fig. 16—Synoptic map (a) Jan. 20, 1937, (b) Jan. 21, 1937 (from Mon. Wea. Rev. Suppl. No. 37) (cold air-mass and front symbols)

LEGEND
- COLD FRONT
- WARM FRONT
- STATIONARY FRONT
- UPPER AIR WARM FRONT
- UPPER AIR COLD FRONT
- OCCLUDED FRONT
SHADED AREA—PRECIPITATION OCCURRING AT TIME OF OBSERVATION.
SMALL FIGURES REPRESENT TEMPERATURE AND 24 HOUR PRECIPITATION, RESPECTIVELY.

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water-level in some sections. Supplement 37 of the *Monthly Weather Review* is a report which contains many rainfall maps and tables, pictures and accounts of damage, descriptions of the weather and its changes, and also weather maps, relating to this catastrophe. But, as far as I can see, there is no real explanation, in all that big report, of the very striking fact that this precipitation maximum lies along the Ohio River Valley, whereas there is only little precipitation (down to 1 inch) up in the Appalachian mountains on the SE side of the valley. The Report
PROBLEMS AND METHODS OF RAINFALL INVESTIGATION

Fig. 18—Rainfall in Sweden, July 26-27, 1950, 07z-07z
seems to assume that this rain maximum was caused by a stationary front along the Ohio Valley, but as far as I can see from the weathermaps the front was not stationary. Instead, I venture to say that what we have here may be a special orographic effect that is very complicated. Time will not allow me to enter in detail upon this problem, but I want to direct your attention to the possible existence of even such queer things as this one in our atmosphere. I think that when the bulk of the rain fell, the gradient wind was southwesterly and very moist; and then there was an overflow of air from the south or southeast across the Appalachians because of friction, causing a convergence on the lee side of the mountain range. In other words: a convergence might occur between a direct SW flow (corresponding to a SE-NW pressure gradient) over the Ohio Valley and a SSE flow through the Appalachian region (following the gradient). Naturally it would be important that also the moisture and lability conditions prevailing at the occasion were favorable for some precipitation mechanism. There may, of course, also be other explanations.

Figure 16 shows two weather maps from the period in question, taken from the Report cited. Those two days the general flow was an easterly to southwesterly, and there was no stationary front over the Ohio Valley, since in this case the Polar front had come down to the Appalachians and even further south. I sincerely hope that some American colleague will collaborate with me on this very interesting case, which I have had in my mind ever since I found it.

Here is another case, from Sweden, with a stationary cold front extending N-S over central Sweden and a postfrontal precipitation area (Fig. 17). That is not unexpected in itself, but the rain also extends far west, to the lee of the Scandinavian mountain range, and it stayed there for three days, giving flooding rains on the lee side of that mountain range. How would you explain that? In fact, the stream lines of Figure

![Map](image)

**Legend:**
- Scandes
- Coast
- Radiosonde station
- Front
- Limit of Ast-system
- Limit of Cloud
- Slight Wst-precip.
- Mod.-heavy "
- Precipitation
- Cold air flow
- Warm air flow

**Map:**

**Vertical Section along line AA:**

**Fig. 19—Tentative schematic map and vertical cross section of rain mechanisms over Scandinavia, July 26-28, 1930**
DISCUSSION

17 show a convergence along the southern edge of the extended rain area, between the N flow descending from the divide near 63°N and the WSW flow to the south of the convergence line. The air ascending at this line is carried towards NW by the predominant SE flow aloft (see Fig. 19). This may explain why the main part of the rain area lies to the north of the convergence of the flow in the lowest layers. The dynamic trough normally forming to the lee of a mountain range will favor a frictional inflow at the Earth's surface in this region. Usually, however, this inflow will not be powerful enough to cause and sustain any large-scale lifting of this very dry 'föhn' air, since this air will never reach the condensation level, the lifting cooling thus being dry-adiabatic. In the case studied here, however, the air in the frictional layer had been moistened by the weak but steady rain from the overhanging Altostratus-Nimbostratus sheet behind the stationary front. When the condensation level within this postfrontal mP became low enough, the tendency to a general convergence at low levels would favor an orderly overturning within this conditionally unstable cold mass, leading to the formation of the marked convergence line and rain area in the trough. A scrutiny of the individual vorticity changes in this stationary trough confirms the assumed existence of a marked convergence within it.

Figure 18 displays the ensuing precipitation distribution with up to 50 millimeters in the region of the downward current at the Earth's surface. The area of great precipitation extends even over eastern Norway. This case is only partly analogous to the Ohio case, but it shows that there are precipitation systems that have not yet been explained, and occurring where one would not expect them according to conventional theories. They may have been of orographic or frontal origin, but they are now simply conditioned by an orderly convergence and lifting, which is of extreme importance because it may give so much precipitation. A cross section from northern to southern Sweden (Fig. 19) shows the down-flow of the northerly wind, the convergence, and the resulting lifting, producing this very astonishing precipitation where there is no front and where one would expect a warm and dry föhn instead.

Last, I want to show that I think that there are two main precipitative effects superimposed on all different kinds of precipitation: those of the wavy kind and those of the convective kind, or stationary lee-wave precipitation and convective streaks of precipitation. Figure 20 shows the precipitation distribution in Holland measured in the morning of October 26, 1945, after 24 hr with a west southwest gradient wind. It shows a succession of rainfall maxima and minima from W to E over Holland. Excepting the dunes, northwestern Holland is absolutely flat, lying between one and five meters below sea level. Thus, it must be the difference of friction between sea and land that causes the first maximum, just above the dunes, and I presume that then a series of stationary lee waves is set up at the level of low clouds, as shown by Figure 21. One day later, on October 27 (Fig. 22), when the gradient wind had hardly changed at all, but the static stability had changed into instability, the stationary lee-wave system broke down. Instead, there were convective streaks of precipitation more or less parallel with the wind, again reaching great amounts. In the first case a warm front passed Holland; the next day a back-bent occlusion passed over the country.

Figure 23 shows the change that took place in the upper air during these two days. Figure 23a is the 500-mb map at the middle of the first precipitation day, with WSW wind aloft and -18 to -20°C at this level over Holland. The next morning (Fig. 23b), still with the same wind, the temperature at 500-mb had sunk to -28°C, this being the reason why the stability pattern had broken down. I presume that these two patterns will be more or less superimposed on all other precipitation mechanisms.

Discussion

Dr. Helmut Weickmann—Thank you, Doctor. It has been a very interesting lecture, which felt like a cool breeze after the heat of many years in which it was tried to increase rain by increasing the number of nuclei. It showed very conclusively that the updraft is of prime importance in the generation of rain.

Dr. Horace R. Byers—In your discussion of the ascent of air in tropical hurricanes, you said something about the augmentation of the tem-
Fig. 20—Rainfall over Holland, Oct. 25-26, 1945, 07z-07z

Fig. 21—Tentative vertical cross section of precipitative lee-wave mechanisms over Holland during the rainfall of Fig. 20
Temperature difference that occurs at the surface in contrast with that aloft. I think I misunderstood you. Could you elaborate on that principle a bit?

Dr. Tor Bergeron—Well, of course, as you noticed, I had to pass superficially over each item. The fact is that not only the temperature is increased, but also the contents of humidity, assuming with Palmén that the relative humidity remains, for instance, 85% at the ocean surface in both cases. Then, the warmer parcel may reach the cloud base with a temperature that is, say, 1°C higher and follow another moist adiabatic line that is diverging from the former one towards higher temperatures.

Mr. Jerome Namias—In other words, you assume the same relative humidity but an increased specific humidity?

Dr. Bergeron—Yes. The equivalent temperature increases more than 1°C, of course.

Mr. Namias—I am puzzled about your statement regarding the inability of radar to depict the release or the areas of release of latent energy as well as do precipitation maps. Did you mean the amount of it, or the area of it?

Dr. Bergeron—I know that I was a little unclear on that point; I beg you to excuse me. I do not claim that precipitation maps give a better picture of the areas of release of latent heat than...
the radar. But precipitation maps will probably give a better idea of the rain amount than the radar maps; and then direct cloud observations from aloft and from the ground together with radar patterns will perhaps give the very best pictures of the regions where there is release of latent heat. That is what I meant.

Dr. Weickmann—I wonder if the Ohio rains can be completely explained as simple orographic rains without having to consider the release of instability showers?

Mr. Namias—I will say something about that in my paper. Of course, it is rather unlikely that the mechanism for long continuous or repetitive rains can be ascribed to any simple orographic factors without considering the whole general circulation. The 1937 case is associated with still larger scale phenomena which Dr. Bergeron did not mention specifically, although he indicated them and lumped them together as 'synoptic phenomena.' In other words, first the stage must be set for the cyclone, which in turn provides for the release of small-scale phenomena which, as he pointed out, are amplified by orographic factors, instability, and the like.

Dr. Weickmann—I think we should emphasize still more that they may frequently be associated with thunderstorms and with quite intense instability lines.

Dr. Bergeron—May I answer this last question? I had the intention of also showing you the normal January rainfall for North America. Referring to that map, the maximum January rainfall in the eastern states lies roughly where that Ohio flood occurred and peculiarly enough to the west of the Appalachians and not on the coast, or in the mountains, or down in Florida. Now, in this special case, the maximum, not only

Fig. 23—500 mb map over NW Europe (a) Oct. 25, 1945, 18z; (b) Oct. 26, 1945, 07z
for that five-day period, but also for the rest of the month, was so closely patterned within the Ohio Valley that it would be very difficult to assume that there could be a large-scale phenomenon that should, just by chance, direct the whole thing, so that you got the maximum just in that valley. I cannot get away from feeling that there must be some rather local orographic effect. Moreover, the Appalachians have an average height of, say, 1000 meters. Clouds in that region will miss the lowest thousand meters of air, which contain the bulk of humidity, and thereby the best opportunity of getting abundant convective rain is lost. It might start better above the mountains, but once it has got started it grows better on the plains, as far as I can see. There may be other orographic effects also; but I want especially to direct your attention to the fact that the average maximum precipitation in January also falls near the Ohio Valley.

Dr. C. W. Newton—I studied an example (J. Met. Soc. Japan, 75th Anniversary Vol., pp. 243–245, 1957) that looked very much like the one in January 1937, and I just want to describe what the precipitation looked like on an hourly basis. In that case there was a cyclone that looked very much like this one, moving regularly eastward as in this case. The long streak of heavy total precipitation in that case was due to repeated movements of four or five heavy convective rainstorms over the same region, which I imagine is true also in your example. These storms formed every four hours, and the peculiar thing was that they moved successively over nearly the same track. It was not so surprising that their paths were almost in the same direction since the upper wind direction did not change much, but the odd thing is the rainstorms all started in nearly the same place. This was down in the southwestern corner of Louisiana, and at the time I saw this, I thought of your suggestion on convergence near the coast caused by differences in frictional effects, because I cannot see anything else that can possibly account for it. There is no great mountain chain in southwestern Louisiana so I think it was an orographic effect of the kind that you discussed, not due to mountains, but to differences in friction.

Dr. Donald M. Swingle—I had occasion to observe the release of rainfall in the vertical for a year and a half, and the contrast between the fairly orderly release of rainfall in either continuous rain or light showers, and the explosive periodic mechanism in the thunder showers is very clear in the data we have. This also stands out in the size distribution of raindrops, which is almost the same for shower rain and continuous rain. Therefore, relationships between rainfall and radar echo intensities are also almost the same. I think there is quite a contrast between the somewhat constrained semicontinuous orderly released precipitation as you had in these orographic cases, and the rather sporadic release one has in the strongly convective precipitation process.

Dr. Bergeron—So you mean if one knows the character of the precipitation, one would be able to evaluate radar pictures into rainfall amounts?

Dr. Swingle—Yes.

Dr. R. Wexler—I had occasion to do it by radar. It is very difficult to ascertain in advance where new storms are going to appear. Quite frequently, with a cyclonic system moving into New England, conditions appear to be the same with regard to instability all over the area. Nevertheless, the thunderstorms are bled in a few favored areas which do not seem to be too dependent on orography, but vary from day to day depending on the other conditions. One frequently finds that the outbreak of these thunderstorms is associated with mesoscale pressure features which are not detected on large-scale maps; but I find it difficult to ascribe these smallscale features entirely to orography. It appears as if the convergence in the area was channeled into a few favored Cumulus which then developed into thunderstorms.

Dr. Bergeron—I have now ordered my thoughts a little more concerning that Ohio flood problem, taking into consideration all that has been said here, by Dr. Wexler and others. I agree with Mr. Namas that the stage was set on the whole by a large-scale arrangement. In fact, there is a map showing the average pressure distribution for January 1937, and also the wind and temperature distribution. All those maps show that the trough and convergence between the Bermuda high and another high in the western United States did not lie over the Ohio region, but on an average further to the west. The stage for the activity was set in another way in the east. First of all, during those five days frontal cloud areas moved constantly across the Ohio region; thus, the upper air was loaded with any amount of release cloud, so that any spindly cloud would have been released. Secondly, there was so much moisture coming in
from the southwest that spender clouds could easily form. Then, many of them, perhaps, were simply convective formations at favored spots. We know little about them; but there are certainly such favored spots, and the convective systems probably moved as Dr. Newton said, up the Ohio Valley, maybe because their formation was favored by the lee-side convergence I mentioned before. That would be my way of summing up what has been said. Perhaps you do not agree, though.

Mr. Namias—I am inclined to think that the orography must play a role. There undoubtedly is some mountain effect, but how much remains to be shown.

Dr. Bergeron—I agree with Mr. Namias. I am quite aware of these facts. What I wanted to point out is that the bulk of the precipitation is formed within the Gulf air; I am opposed to the view that it was formed at a stationary front. No, it originated within the Gulf air, convective systems forming there; and a preferred locality is the Ohio Valley because of the reasons we both mentioned. As to the convective systems, I am sorry to say, Dr. Fujita, that you do not need a mesoscale network to find them. They may be very well located with the network that we have had in Europe for 50 years and even in the network of the United States, although that is much looser; but in order to find the finest details of the convective systems, one needs the time sections that Dr. Fujita has utilized so ingeniously, and one also needs the densest network possible.
Morphology of Precipitation Clouds and Cloud Systems

Chairman: JOANNE MALKUS

Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

INTRODUCTORY REMARKS

The first session of this Conference on Physics of Precipitation is intended to set the stage or develop the context for the subject matter and papers which are to come. In the atmosphere, there are many different scales of phenomena associated with precipitation, from the macro-scale over the mesoscale down to single bubbles or eddies within an individual cloud. This is certainly one reason why in all meteorological studies—and I think most of us will agree that cloud physics is a meteorological study—we must consider the context of the other scales of motion in which this one is operating, and in particular we generally find that we must look up the scale to larger scale phenomena. Therefore, it is extremely appropriate that the context of the first session include a discussion on the synoptic and larger scale flow patterns and their relation to the precipitation process. Our first speaker on this session is particularly well suited to begin this discussion starting with a very large scale motion before we gradually work down. I think it is a landmark in cloud physics certainly, and perhaps also even in synoptic meteorology that one of the world's leading synoptic meteorologists present a paper at a session such as this. I take pleasure in introducing Mr. Namias.
Synoptic and Planetary Scale Phenomena Leading to the Formation and Recurrence of Precipitation

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Abstract—This paper describes certain macroscale features (in both space and time) that play a dominant role in setting the stage for vertical motions, without which it is impossible to consider precipitation problems. First, an empirical relationship of broad-scale total precipitation fields to the geometry of time-averaged flow patterns in midtroposphere is discussed and to some extent interpreted in terms of the interaction of synoptic and planetary scale systems. Contrasting 'regimes' of the order of a month, in which weather processes persistently recur, are illustrated. The amazingly stable regime of October 1952, the driest month in the meteorological history of the United States, is one of these. Another is the period December 1958 through January 1959, when a prevailing dry regime changed abruptly into a wet one. The large-scale physical and synoptic characteristics of these phenomena are discussed in the light of the fine balance of the planetary circulation and against the background of seasonal change. Finally, the various precipitation forms of one of the major storms associated with the January rain-producing pattern which led to flood rains over the Ohio Valley are interpreted with the help of electronically computed charts of vertical motion.

Introduction

Of all the problems of prediction with which the synoptic meteorologist or forecaster has to deal, those associated with the formation of precipitation are probably the most formidable and frustrating. For many years, precipitation was (and frequently still is) predicted mainly by relating it to the geometry of isobaric patterns at one or more levels as well as to the anticipated evolution of these patterns. Alternatively, polar front and air mass concepts provided a more explicitly physical basis for rainfall prognosis. These concepts forced the forecaster to think in terms of vertical motion, the main causitive factor, although this thinking was still indirect and qualitative. During the past decade, as a result of a great world-wide scientific endeavor to objective weather prediction, vertical motion has begun to receive attention as a direct forecasting parameter. The development of high-speed computing methods is greatly expediting progress in this effort.

At the conclusion of this report some mention will be made of the current use of computed vertical motions in explaining areas of precipitation; a detailed account of this basic problem will be found in the following paper by Smagorinsky. The present paper will be concerned mainly with the large-scale or macro-scale features of the general circulation, the centers of action, which are perhaps best brought into focus by averaging over periods of time long compared to the life of an individual cyclone. When such a class of slowly-evolving atmospheric motions are examined against the background of climatology they throw a great deal of light upon fundamental problems underlying the recurrence of dry or wet regimes and the often sharp breaks that occur as one regime changes to another of entirely different character. Furthermore, these large-scale time-averaged states appear to set the stage for the birth and growth of cyclones in certain areas, but not in others. For complete understanding of the physics of precipitation, certainly of its duration and its longer period characteristics, it is difficult to see how the large-scale background and its historical development can be ignored.

Precipitation Fields Related to Mid-Tropospheric Time-Averaged Flow Patterns

A general model—When time averages of midtropospheric contour charts for periods ranging from a few days to a month or a season are prepared and related to abnormalities of precipitation, it soon is apparent that over most areas in mid-latitudes heavy amounts are usually found on the forward side of troughs and light on the back side. A schematic model resulting from statistical studies by Klein [1948] employing five-
day mean maps is reproduced in Figure 1. There is some carry-over of heavy precipitation into the confluent or frontogenetic zone associated with the out-of-phase wave trains. A thorough explanation of the features of this model is difficult because it is empirically derived and is the net result of many factors and scales of motion. For example, the band of heavy precipitation east of the trough is associated with: (1) the gradual upslide motions found to the east of progressive planetary trough systems, (2) the more concentrated regions of upward motion associated with the cyclones which in turn are generally related to, and interactive with, the planetary wave systems, and (3) the very concentrated vertical ascent associated with local convective cells within the unstable air masses of the cyclones. The magnitudes of the vertical motions produced by these three differing scale phenomena are respectively of the order of <1 cm/sec, 2-20 cm/sec, and 5 to 50 m/sec.

We shall now discuss special cases illustrating the role of mean planetary circulations in creating abnormal precipitation regimes.

The driest month in the meteorological history of the United States—In October 1952 the total precipitation which fell over the entire United States was the lowest of some 60 years of record [Winston, 1952]. This fact was determined by examining a routinely computed statistic obtained by combining state-averaged rainfall into a country-wide average, using appropriate weights for areas. The relation of October 1952 to the annual course of monthly normals (averages for 60 years) is graphically represented in Figure 2. Thus the rainfall of October 1952 was only about one-fourth the amount normally received over the United States in October, the deficit amounting to about \(3.3 \times 10^3\) tons of water lost to the nation. It is not surprising to see that this record dryness occurred at the time of year when the normal minimum is reached and when the westerlies and cyclone belts are usually far north and vertical stability is at a maximum.

The drought regime of this month is nicely related to the macro-scale average monthly features shown in Figure 3 by the mean 700-mb contour pattern. This wave pattern has the same phase as the normal October pattern but its amplitude is much greater. The net effect of this amplification is to: (1) deploy Canadian Polar air masses rapidly southeastward into eastern United States, resulting in anomalous cold air in the east with the polar front frequently found off the eastern seaboard; (2) effectively shut off most of the nation (except Florida) from access to moist air from the Gulf of Mexico; and (3) shunt cyclones away from borders of the United States.

The last effect is strikingly illustrated by the tracks of cyclone centers (Fig. 4). There was an
almost complete absence of cyclones over the country: the few which did form or enter were very weak and soon died (ends of arrows).

Another of the interesting characteristics of the October 1952 regime was the small variability of five-day mean patterns during the month. In other words, there was a pattern not only recurrent but also persistent. Yet there was considerable day-to-day activity in the form of rapidly moving fronts and anticyclones into and through the United States.

Even though it was well predicted, it is not possible in this short report to describe how this peculiar pattern became established, for this was contingent upon a long sequence of events interwoven with the fabric of climatological (seasonal) response. However, it is quite clear that during this particular month anomalous factors of many kinds must have been operating in concert with normal climatological factors. Thus we are dealing with a type of resonance phenomenon. Such cases obviously deserve much greater study from those who aspire to modify weather on a large scale.

A sharp change from a dry to a wet regime—Occasionally extended dry periods terminate abruptly and a regime appreciably wetter than normal ensues. The layman used to look upon these events as nature’s way of providing compensation; but in recent years such events are as likely to be attributed to man’s interference with the atmosphere, perhaps by atomic bombs or through the machinations of rainmakers. It is therefore important that the meteorologist, at least, be aware of the fact that nature frequently provides mechanisms for establishing in sequence two quite contrasting weather regimes.

In the following case an abnormally dry situation over the Ohio Valley in December 1958 (Fig. 5) was replaced by a flood-producing regime in January. Actually most of the rainfall occurred in the latter part of January. We shall first describe, in terms of 15-day means, the development of the mid-tropospheric general circulation which led to both regimes, next show how a general instability in pattern was created, and finally indicate how the pattern was transformed into a fairly stable but radically different one.

The sequence of 15-day mean 700-mb maps from the last half of November 1958 to the last half of January 1959 is shown in Figure 6. Wind speed profiles corresponding to these charts are shown in Figure 7. Note that the wave length between United States and east Pacific troughs gradually increased from longitude 73°W (at 40°N) to 90° as the peak strength of the westerlies increased up through the last half of December. Following the sharp decline in the westerlies during the first half of January, a radically different wave pattern emerged over the western hemisphere during the last half of January, when a new trough formed in the area extending from the Great Lakes southwestward to Texas. Apparently, in accord with Rossby’s [1939] expression of stationary wave lengths related to zonal
Fig. 3—Mean 700-mb contours, temperature anomaly and precipitation patterns for October 1952

index, the climatologically expected increase in zonal westerlies from November to December (which occurred strikingly in 1958) required expansion of the wave troughs spanning North America. After an abrupt drop in zonal westerlies (onset of an index cycle) in January, however, a rearrangement of waves with increased wave number occurred, representing nature's way of
achieving a more balanced state. The continuity (determined with the help of charts not reproduced) was that the trough found near the Azores during the last half of January is the one formerly present at 60° W during the first half of the month. It is interesting to note that the breakdown (formation of the new trough) was not simultaneous with the drop in zonal index but required some time (about two weeks). This delayed response appears to find analogy in dishpan experiments by Fultz [1953] in which a quasi-steady pattern of long waves set up with one rate of rotation and of differential heating (between rim and center) requires time to break into another pattern when these values are altered.

To sum up, the instability was essentially created through the interplay between climatological (insolation) factors, which cause the fall to winter increase in strength of the zonal westerlies, and the initial wave patterns, which are the outgrowth of earlier evolutions and have a lifetime of their own. The patterns first adjusted to this increase in westerlies by increasing wave length. Later on, the new extended wavelength pattern was unable to accommodate to a rapid diminution in speed of the westerlies, and an increase in planetary wave number took place. The time scale of this series of events was about 2½ months.

The December flow patterns (roughly similar
over North America to those of October 1952) were responsible for the abnormally dry weather observed in the Midwest (Fig. 5). This dryness resulted from the prevailing northwest flow and descending motion behind the trough, and the consequent lack of access to moisture from the Gulf of Mexico. During the last half of January, however, the new trough over the Midwest was
Fig. 6—Series of 15-day mean 700-mb charts illustrating lengthening wave length response to strengthening zonal index (upper three charts) and then increase in wave number (lower two charts) associated with sharply falling index.
associated with frequent overrunning (ascending) moist air masses from the Gulf. Thus, precipitation amounts about 25% of normal in December changed to over 200% of normal in January.

Of course, this heavy precipitation must also be associated with synoptic-scale systems (cyclones) interactive with the mean planetary trough, as well as the meso-scale convective activity released in those portions of the tropical air rendered unstable by ascent imposed by the larger systems. There was a dramatic change in the tracks of the cyclones during January incident to the development of the new trough, as shown in Figure 8. Note that the precipitation-producing type of storm (the Colorado or Southwest disturbance) became prevalent during the latter half of January.

The strongest development occurred from January 20 to 22, when a cyclone first appearing in New Mexico on the 20th deepened from 994 mb to 968 mb by the 22nd. This storm is of great interest not only as a manifestation of the great change in the general circulation, but also in its role as a factory for many types of precipitation ranging from snow and sleet to drizzle and thunderstorm. The period of this storm also coincides with the January thaw documented by Wahl [1952] and the associated precipitation singularly noted by Bowen [1956].

Precipitation associated with the January 21–22, 1950 cyclone over the Midwest—The great cyclone of January 21–22 over the Midwest produced floods in the Ohio Valley. The surface maps for 12h 00m GMT (1200Z) on both days are shown in Figures 9 and 10. While the electronically computed vertical motions are compiled for the 600 mb level (from a baroclinic model employing 850 and 500 mb data), these may be considered to represent a mean vertical motion in mid-troposphere (Figs. 11 and 12). Details of this model and the vertical motion computations may be found in Thompson [1957].

A glance at the charts of vertical motion and areas where precipitation is and is not occurring shows a reasonably good general correspondence between ascent and precipitation, and descent and lack of precipitation. The large area dominated by more vigorous ascent over the Great Plains and the Northeast on the 21st is related to the precipitation occurring ahead of the warm front and behind the cold front. An area of no rain in the warm sector appears in an area of lesser, but still upward, motion. On the 22nd, the instability snow showers over the southern Great Lakes and adjacent areas are indicated to be part of a larger-scale system of upward motion (induced by a lagging upper air trough) and not random flurries. Then again, the frontal showers in the south are associated with induced ascent.

If these charts are scrutinized more carefully, however, it becomes clear that the correspondence is only general and that there are large discrepancies. For example, on the 21st directly in the area of greatest upward motion there is a pre-
cipitation-free area. Also in central and east Texas no precipitation is occurring under ascending motions. On the 22nd there is a large precipitation-free area to the southeast of the cyclone where the upward motions are almost as vigorous as in the rain areas. From this and other similar comparisons it must be realized that our present understanding of the vertical motion field, including how to measure it (let alone predict it) is very incomplete. Nevertheless, these computations represent a great step forward from the often vague qualitative reasoning of only a decade ago. Then again, it should be noted that upward vertical motion alone does not necessarily imply precipitation. The air masses partaking in the lifting may be too dry or too stable. It is also possible that the nature of nuclei of condensation must be considered. These considerations highlight the great importance of further quantitative studies relating vertical motion, stability, and moisture to precipitation, for only when the impact of these obviously significant factors are isolated will it be possible to evaluate the roles of other possible influences.

Before closing this report, we should comment on the diverse nature of the precipitations spawned by the January storm. On the 21st, for example, we observe: (1) continuous snow in a broad belt some distance west of the cyclone (overrunning of Arctic by maritime tropical air);
Fig. 9—Surface weather map for 1200 GMT, January 21, 1959.
Fig. 10—Surface weather map for 1200 GMT, January 22, 1959
Fig. 11—Vertical velocity (cm/sec) for 600-mb level for 12h 00m, GMT, January 21, 1959

Fig. 12—Vertical velocity (cm/sec) for 600-mb level for 12h 00m, GMT, January 22, 1959

(2) sleet just behind the cyclone center where rain falling out of the tropical air falls through a sub-freezing layer; (3) continuous rain ahead of the warm front; (4) showers and thunder-showers in a band extending from the cyclone center southward (developing into a squall line); and (5) fog and drizzle in the northward-moving tropical air east of the Appalachians.

References
DISCUSSION


Thompson, P. D., A two-level model with effects of vertical velocity advection, irregular terrain, and variable static stability, JNWP Office Note No. 6, U. S. Weather Bureau (unpublished), 1957.


Discussion

Dr. Tor Bergeron—Did you extend your study of October 1952 to Europe?

Mr. Jerome Namias—Only casually, of course. I did not mean to infer that this was a world-wide drought. On the contrary, generally one gets these compensations in space so that the heavy rains in this case were in the oceans—in the East Pacific, particularly, and the Atlantic and Canada had more than their share. I do not re-member precisely what went on in the European areas during that period.

Dr. Bergeron—This was only a question, and since your answer is negative, I may continue. Otherwise, I would have stopped. In fact, it is queer that October 1952 is just a month which has interested us in my Institution, because there was almost constant blocking over Europe. The result was that northern Sweden had in that month the drought conditions of the United States, whereas southern Sweden was like Florida—200 millimeters of rain in certain localities with northeast winds. In northern Sweden there was practically no rain. I think it would be very interesting, really, to take up this thing circum-polarly.

Mr. Namias—I want to underline Dr. Ber-geron’s comment. Of course, it is well known that there are teleconnections between one large center of action and another, so that if we have establishment of one pattern, as we did over the United States, there will be a corresponding generation of resonance waves at other places. So, the abnormalities in precipitation over Sweden may be associated with those over the United States.

The second item: Although I had discussed the January case (the break in the Ohio Valley floods) completely and selected as a case of dramatic break of circulation and weather regime, it turns out that the particular storm did most of the damage on January 21, which happens to be one of the dates indicated by Bowen as being especially vulnerable for world-wide precipita-tion. It is also the date of the great January thaw which Eberhard Wahl (Singularities and the general circulation, J. Met. 10, 42-45, 1953) discussed in his papers. The interesting thing to note is that originally this was picked out and explained as a special case, and without consider-ation of Bowen’s data. It seems to be a good example of how the January-thaw singularity may come about.
Cloud Distributions over the Tropical Oceans in Relation to Large-Scale Flow Patterns

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Introductory Remarks by Dr. Joanne Malkus—We will come down one step in the hierarchy of scale motions and discuss the relations between individual clouds and precipitation and traveling cloud systems. We enter now into the synoptic scale of motions over the tropical oceans. As Dr. Namias pointed out, these scales are associated with and certainly influenced by the large planetary scale. This next paper is a condensation of the results on the relationship of the clouds over the tropical Pacific Ocean to large-scale flow patterns.

Abstract—The relations between tropical cloud patterns and the three-dimensional synoptic structure have been studied by quantitative time-lapse photography from the regularly scheduled MATS aircraft in the North Pacific. From the films of three circuits of this oceanic area a "Tropical Whole Sky Code" was devised. Its application proved relatively objective and reproducible; useful deductions concerning the largescale flow from the code numbers and vice versa are possible. The code numbers were used as underlays in the synoptic analyses; in addition detailed cloud maps were constructed by simplified photogrammetry in particularly interesting areas. Preliminary results show first that in the tropics, large Cumulus clouds and precipitation are highly concentrated in convergent, or disturbed zones, and second that organization of clouds into rows parallel to the low-level flow is common, with increasing intensity and wider spacing between rows as the degree of disturbance increases. The importance of studying cloud physics and dynamics in the context of the overall synoptic situation is brought out.

Tropical Pacific clouds in relation to the large-scale structure of the atmosphere formed the subject of this investigation. The method employed was quantitative time-lapse photography from the scheduled military aircraft flying between Hawaii and the Philippines, from which measurements and cloud maps were made by triangulation, together with a detailed synoptic analysis at all levels between the surface and 100 mb which was carried out at the University of Chicago by H. Richl and William Gray.

Our results, as far as cloud physics is concerned, may be summarized in one sentence: In the tropics the precipitation process must be studied in the context of the synoptic patterns, which regulate the size, structure, development, and organization of the clouds and cloud groups observed. We plan to illustrate this point and our method of utilizing our photographic data quantitatively, by showing one small portion of the film, the corresponding synoptic charts and deductions drawn from these jointly, but first we must lead up to the questions we formulated by presenting some background material.

The classical picture of convection in the tropics might suggest a statistically both random and, on the average, invariant distribution of clouds over low-latitude oceans. In the vertical, the tropical atmosphere is usually conditionally unstable from cloud base upwards, frequently to the tropopause; buoyant energy would thus nearly always be available for a hypothetical non-interacting convective parcel. Horizontally, sharp air-mass contrasts are virtually non-existent; the low-level atmosphere is well-loaded with moisture and closely coupled by vertical
stirring with an almost infinite extent of homogeneous sea surface. One might expect, a priori, uniform or random bunching of tropical Cumulus clouds, with Cumulonimbus build-ups and precipitation distributed here, there, and everywhere, and cloud patterns which statistically show little diurnal, weekly or even seasonal variations. The organization, if any, might be expected to be the hexagonal or roll convection typical of laboratory experiments over a heated surface.

This classical picture of tropical convection has gradually faded. The first blow came with the Wyman-Woodcock trade-wind expedition, discussed by Bunker and others [1949] and the entrainment theory of Stommel [1947] evolved from its data. Clouds were found to exchange air with their surroundings and consequently the degree of instability of the temperature lapse rate lost its unique significance in determining cloud growth. The moisture distribution of the ambient air was found to affect in-cloud parameters, and most tropical Cumuli lost buoyancy by drying out well below the level where a stable layer was reached. Cumulonimbus towers reaching the high troposphere were, to be sure, occasionally observed but conditions permitting their growth proved rare. The evolution of the Woods Hole Studies (during the period 1947–1957) further emphasized the controlling effects of the environment and the inhibitory factors against penetrative development. Runaway cloud growth to the Cumulonimbus stage has been found to require special circumstances. Among these are a deep moist layer, a large horizontal dimension, and probably synoptic-scale convergence in the low levels. The correlation between vertical development and degree of conditional instability proves not just to be low but generally inverse; a more unstable lapse rate is found where our films show a sky of widely separated stunted trade Cumuli, than in easterly wave and hurricane, when Cumulonimbi are rampant.

The classical convection picture collapsed entirely with the work of Riehl and collaborators which carried tropical meteorology both upward and equatorward from the lower trades, where the synoptic disturbance is least evident. In contrast to the steadiness of the lower trades, the upper tropical atmosphere is characterized by extreme restlessness. Even in the upstream inversion-dominated portions, the constancy vanishes above the moist layer, and large space and time variations in wind, humidity, and stability aloft — governed by the dynamics and thermodynamics of phenomena on the synoptic scale. Equatorward from the trades, the inversion weakens and disappears and inconstancy at all levels becomes pronounced; the steady trade flow becomes perturbed into the deepening waves and vortices characteristic of the equatorial trough zone, which itself operates in a fluctuating manner. Throughout the tropics, over land and sea the reputed clockwork regularity of showers proved an erroneous myth. Nothing could be more unreliable than tropical rain. In fact, tropical rainfall studies revealed an enormously skewed distribution, with the major fraction of monthly precipitation falling on two or three days and being almost entirely concentrated in major disturbances, again pointing the finger toward a high degree of cloud, especially precipitating cloud, organization on the synoptic scale.

Our study was designed to pursue this point in terms of specific questions, for which it was hoped to gain quantitative answers. Three circuits of the North Pacific area were accomplished in July and August 1957 and 8000 ft of film successfully exposed, only about one-third of which has been analyzed to date because of the exorbitant labor involved. The three overall conclusions of most pertinence to cloud physics are:

1. The degree of regulation and control upon cloud forms exerted by the synoptic scale of motion, which must be considered in its entire three-dimensional extent throughout the troposphere (including up to, at least, the 200 mb level).

2. The extreme rarity and concentration of precipitating Cumulus and Cumulonimbus clouds, which on the first flight (which has been completely analyzed) never occurred except in a disturbance or a region where low-level synoptic-scale convergence could be readily inferred from the charts.

3. The organization of Cumulus clouds into rows parallel to the low-level flow. On the 6300 km of Flight I, such organization was detected (and the angle determined by quantitative measurement on the films) 91% of the time; it was coded as moderate or strong 62% of the time. Organization was only absent in trade-inversion dominated circumstances and became more pronounced as the degree of disturbance increased. We have measured the orientation of
cloud lines wherever they occurred and their spacing on numerous occasions.

As a corollary to the above conclusions it was found possible to devise a code for the state of the tropical sky, with relatively few numbers, so that the code number chosen provides significant information about the synoptic pattern and physical processes at work. It consists of 17 basic sky types, ten for trade-wind skies and seven for disturbed skies. The code itself and an illustration of each number (a single frame chosen from our films) is contained in the Appendix. The code was used by us to characterize long strips of prints which were made from the movies at 15-min intervals; the appropriate code numbers are then plotted along maps of the flight path and used as underlays for the synoptic charts, as will be seen in the figures. A person can obtain a good visual image of the cloud forms during any portion of any flight by looking at the code number and referring to the sample print thereof, a result which is, in itself, of some interest. Conversely it may be hazarded that a meteorologist well acquainted with the three-dimensional synoptic picture could, with practice, enter in anticipation the code numbers along a projected flight path with little error.

We shall now illustrate our techniques, approaches, and some of our results by showing first the analysis and lastly the movies themselves of a small portion of Flight 1. Figure 1 shows the path and times of this flight, which passed through an undisturbed trade regime from Hawaii to Johnston Island, into and through a moderate easterly wave just east of Kwajalein (in the Marshalls) where the first overnight stop was made. The wave passed Kwajalein during the night and thus was encountered again by the aircraft the following morning. After this the flight proceeded on to Guam and Manila, where on this last leg it passed through the typhoon Wendy. We shall be concerned here, however, only with the two passages through the easterly wave. The remainder of Flight 1 is analyzed in detail in the Report [Riehl and others, 1959]. Figure 2 shows the flight path of the first leg with the cloud code numbers entered along it used as an underlay for the low-level streamline chart. All code numbers entered refer to "trade-wind skies" (code numbers 1-4) or "trade-wind disturbance" (code numbers 6-9), the latter in this case associated with the moderate-strength Riehl-type easterly wave, whose trough line was centered between Majuro and Kwajalein, and which was moving westward at approximately ten knots, or more slowly than the trade flow.

We would thus expect the convection and precipitation zone to be located east of the troughline; the pictures confirmed this in a striking manner. Note the code numbers 8 and 9 (Cumulus congestus, high Cumulus tops and showers: the only difference between the two is that 9 contains upper clouds, broken to overcast) appearing up to the wave trough, and then cutting off sharply ahead, going over to ordinary or even suppressed trade-wind skies. The quantitative measurements from the films bring out even more clearly the effects of the disturbance upon the cloud structure. Figure 3 shows the cloud cross section of this leg constructed from the film analysis, superimposed upon a relative humidity cross section composited quite independently from the radiosondes and synoptic data.

The horizontal dimensions of the sketched Cumuli (as well as the hatched under curve below) gives the cloud amount, while the dashed cloud outline indicates the maximum height of tops; the average top height is shown by the darker outlines. The effects of the inversion domination near flight outset, its relief at about Johnston Island and gradual approach to the wave convergence zone are pronounced. Note the upper regions of high humidity in connection with the subtropical jet over Hawaii and with the level of maximum wave amplitude (15,000-20,000 ft) in the Marshalls area and the association of these with middle and upper cloudiness. It is worthwhile to remark here that Flights II and III crossed this area without a wave disturbance and only trade-wind or suppressed skies were photographed.

Figure 4 shows arrows denoting the orientation of the cloud lines, and vertical marks the number of which is proportional to the intensity of this organization into rows. We see that the cloud lines are parallel to the low-level streamlines, within the accuracy of our ability to draw these (except perhaps just west of the wave trough where the cloud lines were backed almost north-south and no wind information existed to test whether or not the flow was so oriented). We also see that the organization into rows is weak or absent in the inversion-dominated portions of the flight; namely, just out of Hawaii and just west (ahead) of the
Fig. 1—Routes, dates, and times for Flight I, July 10-13, 1957
Fig. 2—Mean winds (knots) of lower layer with approximately uniform wind, July 11, 1957, 00Z; also top of layer (above station denoted by T and reported in 1000's of ft) and height of strongest wind in layer (below station denoted by M, height in 1000's of ft and maximum wind in knots); underlay is sky code (see Appendix); heavy solid line marks position of wave trough in easterlies; solid lines with arrows are low-level streamlines.
Fig. 3—Cross section of relative humidity along Leg 1, Flight I, with cloud cross section as underlay, isopleths are relative humidity in per cent; regions marked M are moister than the surroundings while regions marked D are drier; the wave disturbance in the easterlies lay between the longitudes of Majuro and Kwajalein; the cloud cross section was constructed from measurements on the motion picture film; lower cloudiness (Cumulus) amount indicated, to nearest one-quarter coverage, by bottom hatched under curve, middle cloudiness by black columns at mid-levels, and upper cloudiness by black columns at top; inversion Stratus is not included; width of schematic Cumuli indicate coverage (narrowest, one-quarter) not cloud size; arrows are direction of measured cloud shear (north upward); mean heights of Cumuli are solid clouds, maximum heights in the 15-min interval are dashed; precipitation is noted coming from cloud base in every 15-min interval that it was observed; the tall tower near the wave trough (close to Majuro) was measured to reach 50,000 ft; several others exceeded 40,000 ft far to the south of the aircraft path at the trough line.
Fig. 4—Organization of Cumulus, Leg 1, Flight I, superposed on lower-layer winds; straight arrows give measured direction of cloud lines; code symbols are organization intensity: x absent, I weak, II moderate, III strong, M no measurement; solid wavy lines with arrowheads are streamlines.

Fig. 5—Schematic distribution of cumuliform clouds near wave trough in easterlies, Leg 1, Flight I; orientation of cloud rows and spacing between them was measured exactly from the film, as was the distance between the amplified cloud groups, denoted schematically in the figure as a single large cloud; individual clouds have not been entered nor drawn to scale.
Fig. 6—Mean winds and altimeter corrections with respect to mean tropical atmosphere for layer 250-150 mb, July 11, 1957, 00Z; isopleths are of the latter in tens of ft; the thickness of the layer in tens of ft (first digit omitted) is written to the right of the stations.
Fig. 7—Mean winds (knots) of lower layer with approximately uniform wind, July 12, 1957, 00Z; also top of layer (circled, in 1000's of ft) and height of strongest wind in layer (maximum height in 1000's of ft; direction and velocity of maximum in code); code notation at tail of arrow denotes mean wind
Fig. 8—Selected frames from Flight 1; camera aimed south from 8000 ft altitude; (a) frame just to east of wave trough, in active convection zone, Leg 1; (b) frame just to west of wave trough, Leg 1; cloud line same as that just to west of trough in Fig. 5; after the third of these, conditions became completely suppressed; (c) wave trough convection zone on following day after wave has passed under upper cold low; the second encounter of the aircraft with the wave took place 200 miles west of Kwajalein, on Leg 2; note contrast in clouds as compared with (a) and (b)
wave trough, while it is pronounced in the convection zone of the disturbance. A more detailed study of organization in the region of the easterly wave trough is shown in Figure 5 where we have mapped schematically the cloud lines ahead and to the rear of the disturbance. The spacing of the cloud lines is to scale, but not individual clouds. The large blobs located about 100 km apart in the convection zone are actually large groups of huge buildups which were regularly spaced at these intervals along the parallel cloud lines, giving the impression of a normal mode of organization. Beginning at about halfway between Honolulu and Kwajalein, we measured the spacing of these cloud lines; it began at about 4 km between rows, comparable to the depth of the moist layer. As the wave trough was approached, the row spacing became wider until it reached about 25 km just east of the wave trough and about 30 km just west of it. It appeared that the spacing was being widened by superposition of a larger-scale organization upon the smaller (4 km) scale and that some of the cloud rows were being enhanced while those between were being suppressed.

Finally, Figure 6 shows the high-level chart, the D-values and average winds for the 250-150-mb layer which prevailed at the time of Leg 1. The important point to note is that the easterly wave trough lies just to the west of the upper anticyclone and to the east of an upper cyclone, which was stationary. While the observer and aircraft spent the night at Kwajalein the wave trough passed by and moved underneath this upper cold low. On the next day's flight it should have been encountered about 200 miles out of Kwajalein, but only faint remnants of the previously vigorous wave were detected. There was hardly any traces of disturbance left along the path of the aircraft except five or six very distant Cumulonimbi on the southern horizon, which may have been associated with the equatorial trough. The synoptic study showed the air rounding the southern end of the upper trough was likely to gain cyclonic relative and absolute vorticity. Upper level convergence and consequent sinking presumably destroyed the easterly wave and Figure 7 shows its total demise in the streamline field in the Marshall area. The cloud motion picture films showed its death far more dramatically, however, than any of our words or diagrams. Figure 8 shows three selected frames from the motion picture film; a and b were made to the west and east, respectively, of the wave trough on its first crossing on Leg 1. The cloud line seen in Figure 8b appears as the first one to the west of the trough line in Figure 5, and the large cloud build-up, in the distance reached 50,000 ft. Figure 8e shows the 'convection zone' of the trough on its second crossing by the aircraft on Leg 2 (after the overnight stop at Kwajalein). Note the complete disappearance of the disturbance, except for the very distant build-ups far to the south.

In conclusion, this work suggests that microphysical studies of tropical clouds and precipitation would be severely restricted if studied out of context with the synoptic scale flow patterns. If we wish eventually to predict, understand, or in any sense modify tropical rain, the interaction between large-scale flows, convection dynamics, and droplet formation and growth must be studied.

Note—At the Conference this paper was concluded by showing the portion of the cloud films from which the figures shown here were constructed. These films were made from military aircraft flying at 8000-ft altitude, and were taken in kodachrome at one frame per second. These films are on file at the Woods Hole Oceanographic Institution and at the Department of Meteorology, University of Chicago. The methods of photography and reduction are described in detail in the Report [Riehl and others, 1959] of which this paper is a partial condensation.

Acknowledgments—The writers wish to acknowledge the essential contribution made to this study by their colleagues at the University of Chicago, Herbert Riehl and William Gray, who carried out the synoptic analysis described herein. The work when completed is to be published under joint authorship with them.

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References
Appendix

Tropical Whole Sky Code

JOANNE S. MALKUS AND HERBERT RIEHL

This code is for use primarily in coding aerial photographs (or visual observations) over the open seas in the tropics far from land masses. Categories 0-4 denote undisturbed trade-wind skies. Categories 5-9 cover weak or trade-wind type disturbances, while categories 10-16 relate to strong disturbances (deep easterly waves, equatorial trough disturbances or convergence, through typhoons). The letters (to be placed after the appropriate numbers) permit some further refinements or sub-specifications.

TRADE-WIND SKIES

0 No clouds at all, or a few suppressed, or small numbers of very suppressed Cumulus. Upper cloud 1/10 or less.

1 Inversion-dominated: trade Cumulus spreading out into Stratocumulus, definite signs of stratiform but total sky cover less than about one-half.

2 Extensive Stratocumulus decks, more than about one-half total sky cover.

3 Average trade Cumulus sky, no particular organization into lines. No significant upper clouds. No marked shear. Tops 8000-10,000 or less, precipitation unlikely.

4 Average trade Cumulus sky, but with marked organization into rows or lines. No significant upper clouds. No marked shear. Tops 8000-10,000 or less. Precipitation unlikely.

5 Average trade Cumulus sky with upper sheet clouds up to broken. No marked organization or shear. No precipitation.

6 Marked concentration of convection, rather large convection clouds (not Cumulonimbus) in view, with extensive clear or almost clear areas. Upper sheets to broken may be present.

7 Suppressed trade Cumulus sky with upper sheets broken or overcast but not precipitating. No marked shear or organization in Cumulus.

8 Above average Cumulus activity, high tops, occasional Cumulus congestus. Showers suspected or seen. Upper cloudiness less than broken. No marked shear or organization of Cumulus.

9 Above average Cumulus activity, high tops, occasional Cumulus congestus. Showers suspected or seen. Upper cloudiness broken to overcast. No marked shear or organization.

10 Occasional Cumulonimbus, no obvious organization, no significant or extensive independent upper sheet clouds. Indicate component of shear in plane of film.

11 Rows or suspected organization of Cumulonimbus, no extensive or significant independent upper clouds. Indicate component of shear in plane of film.

12 Cumulonimbus and partial coverage of independent upper sheet clouds. Cirrus, Altostratus and/or Altocumulus.

13 Substantial upper cloud coverage, suppressed convection over wide clear areas, except some Cumulonimbus visible. Precipitation from upper clouds absent or unlikely.

14 Rain from upper sheets (overcast or almost overcast) Cumulonimbus indistinct, picture dark.

15 Rain from upper overcast, Cumulonimbus still suspected but not clearly visible, picture dark and disturbed.

16 Upper overcast raining, no Cumulonimbus seen or suspected, suppressed low convective activity; picture dark and disturbed.

POSTSCRIPTS (PRIMARILY FOR TRADE-WIND SKIES)

0 Marked organization into rows or lines.

s Shear obvious; indicate component in plane of film.

c Apply to 1 or 2 if Cumulus tops are seen poking through the Stratocumulus.

d Distant disturbance—evident from high clouds low on horizon.

— Weaker, more suppressed or less development than normal for category.
Fig. 9—Illustrations of sky code, undisturbed trade-wind skies

Fig. 10—Illustrations of sky code, organized or trade-wind disturbance or both
Typical illustrations of each sky type (and most of the postscript modifications thereof) are shown in Figs. 9-11. These were chosen from all the prints of the flight films for a combination of best representativeness and reproducible photographic quality.

Discussion

Dr. Malkus—I think this is an excellent illustration of how a picture can say much more than many words. It was with considerable surprise that we did not find the wave again after it had passed over us during the night and was presumably a hundred miles beyond Kwajalein. It simply was not there anymore. This illustrates, as we saw in Figure 8, that in order to make an analysis of precipitating systems in the tropics, we not only have to know about the synoptic patterns at all levels, but also about the intense coupling between the high tropic patterns and those in the low troposphere.

I will open the discussion by making one comment. It is a provoking question: namely, how well could we predict, understand, or even regulate the microphysics of tropical rain, if we were in an aircraft in the Marshall Islands area and, out of context, made a microstudy of the clouds? Other aircraft studies might be made in the area of Kwajalein and Majuro, which was on the other side of the wave. The result would probably be that one group of people says tropical clouds are less than 10,000 ft high and totally disorganized, while the other group says the observations are wrong, the clouds extend to 50,000 ft and are well organized. I have not heard any arguments quite that preposterous, but, of course, in the different sections and under the different conditions under which these observations were made, we would arrive at different conclusions.

Dr. Horace R. Byers—I think that the greatest sin that cloud physicists commit is to draw conclusions from small samples.

Dr. Morris Neiburger—Well, I will not say that what Drs. Malkus and Byers have just said may not be true, but it seems to me that the case is a rather different one. Rather than assuming that all clouds have the same micro-
structure the cloud physicists have been attempting to determine what the differences are which decided whether a cloud will precipitate or not.

Synoptic meteorologists had been puzzled for many years as to why apparently similar situations under some circumstances gave rain and under other circumstances did not. Then Professor Bergeron came forward with his very remarkable explanation in terms of the ice-crystal theory. That handled a good many of the problems. People began to look at the temperatures and degree of supercooling of clouds, and to consider the possibility that the supply of ice-crystal nuclei might, under some conditions, be inadequate.

Then we became aware that there are many cases of rain when clouds are not supercooled; in these cases the ice-crystal theory could not explain the precipitation. There apparently are significant differences even in the micro-structure of warm clouds, and we are now trying to see what those differences are. Far from concluding that every cloud has the same structure, liquid content, or nuclei count, we are seeking to find out what are the critical differences in them.

In all this we are aware that it is necessary to seek the key to the differences in micro-structure, and that this key may be found in the larger scale air mass properties and flow patterns. Our hope is that by knowing what differences in micro-structure are involved, the synoptic and dynamic meteorologist will know what to look for in the large scale structure.

Dr. Henri Dessens—This is very interesting. Especially the organization of the cloud groups is striking. They have been observed and described in the equatorial region by V. Schaefler (Transactions, New York Academy of Science, Oceanography and Meteorology, pp. 535-540, 1958). In the Congo Basin the organization of clouds is one of the most remarkable facts. Observations regarding the vertical organization of the atmospheric layers in which these clouds form have recently been made at Lukoléla (1°S, 17°E) in the Belgian Congo. The basic very persistent fact in rainy or dry seasons is the existence of a moist layer at the ground. This layer is called the ‘monsoon’ and is in motion from WNW to ESE. It is topped by the trade-wind layer which moves from ESE to WNW. The trade-wind layer is very dry in the dry season and sometimes humid in the rainy season. During fine weather the humid layer is invariable in thickness at about 1200 meters; but during the rainy weather great variations are observed. Within a few hours one finds successively: 600 meters, 3200 meters, 600 meters, 2200 meters, etc. Deepening of the humid layer permits the formation of cumulus-type clouds. If Cumulonimbus clouds form, the layer may be as thick as 7000 meters. The word ‘deepening’ has been selected instead of ‘wave’ in order not to anticipate a wave mechanism for this phenomenon which appears so important in the equatorial region.

Mr. Richard G. Semomin—Incidentally, it also exists elsewhere; Dr. Fujita and I for instance, observed from the aircraft several of these streaks over central Michigan, so they are not limited to tropical regions.

Dr. Tor Bergeron—With respect to what Dr. Dessens said I wonder if he knows that British Colonel Keeling as early as 1908 made a pilot balloon expedition to Khartoum and Mongalla in Sudan. During a dry week in August at Mongalla (5°N) the monsoon formed a very shallow layer, only one to two kilometers thick, but otherwise it was six to seven kilometers deep. I have worked a little with these data and I have formed some ideas as to the structure of the southwest monsoon over Africa, and maybe I shall have an opportunity to come back to this thing during this conference. (See Dr. Bergeron’s comments, pages 399–401—Ed.)

Dr. Malkus—The skewed distribution of tropical rainfall is well known. The question remained in the minds of some investigators, however, as to whether this concentration of rainfall into disturbances would be equally pronounced over the much vaster tropical oceans as compared with that over the relatively restricted island and continental masses. The opportunity to make a fairly conclusive test was offered by the Pacific Marshall area data collected and distributed under ‘Operation Redwing.’ Our study was begun using the radiosonde and six-hourly rainfall data from the atoll of Kwajalein (8°43’N, 167°44’E) which extended in time from April 15 through July 24, 1956. It is interesting to note that the ‘equatorial trough’ had a mean position in July running right
Fig. 12—Average soundings on the rainy and dry days

Table 1—Rainy versus dry soundings for Kwajalein Atoll, April 15—July 24, 1956

<table>
<thead>
<tr>
<th>No. of days</th>
<th>24-hour rain</th>
<th>Aver. lapse</th>
<th>Aver. precipitable water</th>
<th>Standard deviation precipitable water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rainy days, 7</td>
<td>1.24</td>
<td>0.56</td>
<td>5.35</td>
<td>0.40</td>
</tr>
<tr>
<td>Dry days, 13</td>
<td>0</td>
<td>0.68</td>
<td>4.99</td>
<td>0.44</td>
</tr>
</tbody>
</table>

through Kwajalein, and although we have not yet determined it for April—June, it probably did not lie very far to the south of the Marshalls.

The procedure was first to calculate the 24-hr rainfall occurring between each midnight and the following, and arrange the days in a descending order.

In a total of 100 days, S7 experienced rain (from a trace upward) and a total of 35.21 inches of rain fell. The maximum rain falling in any 24-hr period was 3.26 inches on May 28, and a total of ten days (fairly evenly distributed throughout the period; only two consecutive) experienced more than one inch. On the highest seven of these (R = 1.24 inches) the rain fell fairly evenly around the clock, and these were designated for the remainder of the study as 'rainy' days. The thirteen days on which no rain at all fell were designated as 'dry' days. A mean sounding was then constructed for the rainy and dry days separately (shown in Fig. 12) with rather striking results. First, the soundings are fantastically similar, and day-to-day variations within the 'rainy' and 'dry' classes far exceeded the difference between the two averages. This clearly demonstrates the utter futility of attempting to forecast or explain rainfall in this area in terms of the stability or moisture structure of vertical soundings alone; we must look also either to large-scale dynamics or to microphysics, or to both in interaction. Table 1 summarizes some of the essential features of the soundings.

It will be noted that the tropospheric lapse rate (950 mb—tropopause) is five per cent steeper on the dry than on the rainy days, in good agreement with other findings in the tropics of an inverse relation between instability and penetrative clouds. The difference in average precipitable water (computed both from individual soundings, then averaging, and from the average soundings of Fig. 12) is very slight, and the standard deviation of this property within each category of days (last column) is greater than the difference in the averages, which suggests that the latter is not too meaningful.

Thus we must look for reasons for the enormous difference in rain in the two categories either in terms of the dynamics of the large-scale flows and their effects on penetrative cloud growth, or just possibly in terms of the differences in microfeatures of clouds (such as drop spectrum, nuclei count and spectrum, etc.).

All evidence to date convinces us that the dynamics plays the controlling role, and if microphysical differences are at all significant they too are controlled by the large-scale dynamics and the convection itself by some as yet unknown 'feedback' mechanism.

In retrospect, it is not surprising that the concentration of rainfall may prove more pronounced over oceans, since islands and land masses, particularly mountainous ones, create both some large-scale convergence themselves (due to the diurnal heating-cooling cycle) and provide critical 'concentration points' for convergence which would locally intensify that of a very weak traveling disturbance.
Structure of Convective Storms

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Abstract—The features of the Fargo tornadoes and the meteorological situations producing them are studied in three scales: macroscale, dealing with the precipitation and tornado distribution over three-fourths of the United States; mesoscale analysis of North and South Dakota and of the rotating cloud which produced four tornadoes as it moved over the Fargo area; and, finally, microscale features of the tornado itself. These analyses are presented in an animated 16 mm motion picture. Included in this report are the results of the computed cyclostrophic wind speed and rotational speed of the funnel of the Fargo tornado. A proposed mechanism of an irreversible process taking place inside the inflowing air toward the tornado funnel is also presented.

Life cycle of the Fargo tornado—By using the still photographs and movie films collected through WDAY-TV, Fargo, and the U. S. Weather Bureau at Hector Airport, Fargo, changes in the tornado funnel throughout its life span were obtained. Figure 1 reveals that the funnel dropped from the cloud base to the ground within a matter of minutes. As soon as the tip of the funnel reached the surface, the lower portion of the funnel was sheared off, and then rounded; meanwhile, the funnel diameter above the rounded bottom kept increasing. Soon after, the tornado funnel as a whole started shrinking at about 18h 30m CST.

The diameter of the funnel is also given in Figure 1 as a function of time and the height above the surface.

Funnel diameter and centrifugal acceleration—An expanded analysis of the funnel diameter is made for the period of only 2½ minutes during which the tornado funnel reached the ground and was rounded at its bottom. The rounded funnel bottom shows an extremely high rate of increase in diameter along the vertical (Fig. 2).

Assuming that the condensation is taking place at the funnel edge and that the condensation pressure is the same everywhere at the smooth edge of the funnel, the hydrostatic assumption and the cyclostrophic wind equation enable us to describe

\[ \frac{1}{\rho} \frac{\Delta P}{\Delta R} = \frac{V^2}{R} \]

\[ \Delta P = -\rho g \Delta Z \]

Therefore, we have

\[ \frac{\Delta Z}{\Delta R} = \frac{V^2}{g R} \]

where \( V \) is the cyclostrophic wind speed; \( \rho \), the density; \( R \), the radius; \( P \), the pressure. This equation shows that the tangent of the funnel slope is proportional to the centrifugal acceleration in the unit of gravitational acceleration. It is of interest to see the maximum centrifugal acceleration 10.8 \( g \) occurring at about 18h 28.5m at 120 m above the ground (Fig. 3).

Cyclostrophic wind speed—The cyclostrophic wind speed computed from the centrifugal acceleration is shown in Figure 4 as a function of time and the height above the ground. The diameter of the funnel is also indicated by broken lines. The computed wind speed is the cyclostrophic wind speed at the funnel edge. No speed was computed at the bottom of the rounded funnel because its gradient is so small that it would give less than 10 m/sec, while heavy damage was taking place directly beneath the funnel.

In any case, the tangential speed at the edge of the funnel above the rounded bottom gives a fairly high value, which would be expected from the tornado damage. The maximum speed of 103 m/sec, or 230 mph, occurred slightly before 18h 29m when the funnel diameter, 130 m above the ground, was 200 m.

A film taken by WDAY-TV, Fargo, at the time of the Fargo tornado permits us to make an independent computation of the rotation speed of the funnel. Figure 5 indicates the tangential speeds computed by cyclostrophic assumption (double lines) and tornado film (heavy lines) for three different times: 18h 28.0m, 18h 28.9m, and 18h 29.6m CST. The computation is made along the dotted line, along the funnels shown at the top of the figure. At the present time, so far as the Fargo
Tornado is concerned, no way of computing wind speed inside or outside the funnel has been found.

Tangential wind speed at the funnel edge is summarized in Figure 6 as a function of time and the radius of the funnel. The dashed line in the figure denotes the radius of the rounded portion of the funnel which appeared at 18h 2Sm CST, then increased rapidly in diameter.

The slope of the rounded bottom funnel, which would have maintained the cyclostrophic wind speed obtained by using the movie taken by WDAY-TV, should be at least five times steeper than it actually was. At one time the rounded bottom of the funnel was almost flat, yet it was rotating at over 100 mph. This suggested that the constant pressure surface should cut through the bottom of the funnel.

Irreversible process taking place inside tornado inflow—The previous discussion indicates that the constant pressure surfaces maintaining the high-speed cyclostrophic wind at the base of the rounded bottom funnel must have been much steeper than the funnel itself.

A schematic figure of constant pressure surfaces in relation to a rounded bottom funnel is shown in Figure 7. We should reconsider the thermodynamical process taking place inside the air flowing into the tornado funnel.

The dry adiabatic process in the parcel method is an adiabatic reversible process in thermodynamics. No heat should be added or produced along the course of the expansion of a parcel. This would result in the same condensation pressure regardless of the path of the parcel so long as it starts expanding under the same initial conditions.

Through an irreversible adiabatic process, however, heat is continuously produced internally while the parcel expands. Parcel A, for instance, will show a small rate of cooling when it flows into a tornado, because the parcel moves near the ground under the influence of frictional force which produces heat; meanwhile the parcel loses its mechanical energy. Thermodynamically, the line along which the parcel expands lies between isentrope and isenthalpe on adiabatic charts.
Fig. 2—Detailed description of the cone-shaped funnel in its early stage; the bottom of the funnel was rounded, and lifted upon reaching the ground; isopleths are funnel diameters in meters.

Fig. 3—Vertical and time change in centrifugal force acting upon the parcel circling around the funnel; wind is assumed to be cyclostrophic.

Now, the illustration shows that the parcels expanding along stream lines $A_o-A_i-C_i$ and $A_o-A_f-C_f$ are friction free; therefore, the moisture inside the parcel would condense at $C_i$ and $C_f$ located at the surface of the same condensation pressure. On the other hand, the parcels following the paths $A_i-A_f-C_f$ and $A_o-A_f-C_f$ would increase their entropy through their irreversible expan-
sion, reaching their condensation pressures at $C_3$ and $C_4$, respectively. Assuming that the expansion along the surface is given by a straight line $A_0-A_y-A_x$ on the adiabatic diagram, and that the expansion along $A_x-C_3$ and $A_y-C_4$ is dry adiabatic and reversible, the temperature and pressure change of parcels flowing into the bottom of the rounded bottom funnel were qualitatively described.

Conclusions—It was found that the cyclostrophic wind speed computed from the shape of the funnel with the combined use of hydrostatic
Fig. 6—Tangential wind speed shown as a function of time and radius of the funnel; the dashed line indicates the radius of the rounded bottom funnel.

Fig. 7—The nonadiabatic process taking place beneath a tornado funnel after it is rounded; note that the edge of the rounded funnel no longer represents a surface of condensation pressure and condensation pressure in the parcel method shows reasonable values. However, as soon as the funnel reaches a certain diameter, large enough to provide a long spiral inflow path of the parcels near the ground, the lower portion of the funnel is rounded. At this stage the bottom of the funnel no longer maintains the same condensation pressure. A dry adiabatic irreversible process associated with the near ground inflow was found to be one of the explanations for this.

Acknowledgment—The research reported in this paper has been sponsored by the U. S. Weather Bureau under Contract Cwb 9530,
Editor’s Note—After having presented his lecture Dr. Fujita showed an animated cinema of the Fargo Tornado on June 20, 1957. The film is divided into three parts: (1) the synoptic pattern with isobars and moving precipitation systems, (2) the meso-scale analysis of the Tornado proper, and (3) micro-scale features of the funnel and the movement of clouds near and around it.

Discussion

Dr. Malkus—I am sure we would all like to congratulate Dr. Fujita for a fascinating physical study of tornadoes in his marvelous animated film which illustrates it. Only those of us who tried to make films ourselves realize the enormous amount of effort that goes into it.
Energetics and the Creation of a Self-Sustaining Local Storm

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Abstract—The dynamics of the evolution of Cumulus congestus and Cumulonimbus clouds may be described in terms of V. Bjerknes' second circulation theorem wherein a frequency factor, $v^2$, pertaining to the buoyancy restoring force per unit length, is defined as a function of the temperature lapse rate and the scale of the disturbance. It is shown that either exponential or sinusoidal development of large local storms in the Southwest is described, together with the associated vertical velocity field and precipitation release.

Although no direct evidence is available that supports an exponential or runaway growth for local thunderstorms, it is suggested that such may be the case for intense local storms of the cloudburst variety. Several examples of this are shown where an exponential solution is possible. The paper concludes with speculations on the creation of self-sustaining local storms by altering the scale of the disturbance through the use of solar energy converting chemicals.

On observing the development of Cumulus clouds, one often wonders if it is possible for a small Cumulus to grow to thunderstorm size by continuously enlarging itself. Experience shows how ordinary fair weather Cunuli disappear after fifteen minutes or so, and the maximum size attained during their lifetimes is quite small compared with a Cumulonimbus cloud. Except in regions of strong orography, one has little opportunity to watch the birth and death of a thunderstorm. The observer usually notes that the thunderstorm drifts into his field of view rather than develops before his eyes. However, in the Southwest of the United States, light prevailing winds and pronounced orography combine to afford the study of the life cycles of all ranges of Cumulus types, from the very small to the gigantic Cumulonimbus. It was in this region that data on the growth of Cumulus clouds were obtained for this study.

The data consist of stereo pairs of cloud photographs, 9 x 9 inches, which were analyzed stereoscopically for the rate of vertical growth of various clouds. These photographs were made at Tucson, Arizona, by the Institute of Atmospheric Physics, University of Arizona, in 1956; and at Flagstaff, Arizona, by personnel of the Cloud Physics Branch, Geophysics Research Directorate, in 1958.

From the limited number of clouds thus far studied, it appears that thunderstorm-size clouds do not develop by the continuous growth of small Cumuli. The earliest precursor of a thunderstorm is itself a vigorously growing Cumulus congestus which may have resulted when the flow fields of several isolated Cumuli merged to form a large and active convective area.

Viewing Cumulus growth as a cellular circulation, one can introduce the idea of the horizontal scale of the convection cell as having a bearing on its lifetime and vigor. This is accomplished by using a modification of V. Bjerknes' Second Circulation Theorem developed by Holand [1939] and Eliassen and Kleinschmidt [1957]. For a cellular circulating system, the circulation of the acceleration is equal to the circulation of the vertical restoring force

$$\oint \rho (\frac{\partial u}{\partial t} dx + \frac{\partial w}{\partial t} dz) = -\oint v^2 dz$$

where

$$v^2 = \frac{\beta}{T} (\gamma - \gamma_d)$$

$\gamma = $ environment lapse rate

$\gamma_d = $ dry adiabatic lapse rate

The other symbols have the usual meteorological meanings. As used in (1), $v^2$ is the static stability or vertical restoring force per unit length. For a closed streamline, the motion will be either a stable standing oscillation or an unstable cellular circulation. At a point where the closed streamline of the cell is tangent to the vertical axis of the cloud, we may determine the nature of the vertical velocity variation with time because the frequency of the circulation accelera-
tion applies to all portions of the fluid along the streamline. From (1) we may write for this point
\[
\frac{\partial v}{\partial t} = -v^2 z
\]  
(3)

or

\[
\frac{\partial^2 v}{\partial z^2} = -v^2 w
\]  
(4)

When \( -v^2 < 0 \), \( w = \lambda \sinh \nu t \), and when \( -v^2 > 0 \), \( w = \lambda \sinh \nu t \) if \( w = 0 \) at \( t = 0 \).

When \( -v^2 > 0 \), we have an exponential increase in circulation with time and when \( -v^2 < 0 \), harmonic oscillations about an equilibrium value of period \( 2\pi/\nu \) are expected. In this latter case, \( \nu \) may be regarded as a frequency factor. An expanded form of (2) may be employed to demonstrate the influence of horizontal scale on the stability of the circulation. Pettersen [1936] and Beers [1945] have shown that the frequency factor can be expressed

\[
-\nu^2 = \left( g/T_0 \right) [ (\gamma_s - \gamma) + (\gamma_d - \gamma) A_+ / A_- ]
\]  
(5)

Here \( \gamma_s \) = saturated adiabatic lapse rate
\( T_0 \) = temperature of the environment
\( A_+ / A_- \) = ratio of updraft area to downdraft area

The horizontal scale of the convection cell may make its influence felt through the parameter \( A_+ / A_- \). Ordinarily, \( \gamma_d > \gamma > \gamma_s \), so that

\[
| \gamma_d - \gamma | > | \gamma_s - \gamma |
\]

and stable solutions are to be expected in (4) if \( A_+ / A_- = 1 \). Unstable solutions are possible if

\[
| (\gamma_d - \gamma_s) A_+ / A_- | < | (\gamma_s - \gamma) |
\]

and this may be realized whenever the ratio \( A_+ / A_- \) is small enough to make this possible.

It is apparent that this parameter has a critical bearing on the stability of convection and varies with different meteorological conditions. One object of the cloud studies made in Arizona was to get an estimate of this ratio. For the clouds studied, a low frequency oscillation was noted in the vertical growth rate. This oscillation was manifested by periods of rapid upward growth followed by periods of slow upward growth. By using the observed frequency together with Raob data, (5) was solved for \( A_+ / A_- \). The results are given in Table 1. The ratio \( A_+ / A_- \), thus computed, suggests for these somewhat isolated Cumulus clouds a convective cell whose updraft area is about equal to its downdraft area. The clouds are individual ones whose fields of motion do not extend much beyond their visual borders. This concept should be reasonable for ordinary fair-weather Cumulus development and air-mass showers.

It is conceivable that in a meteorological situation of synoptic-map scale low-level convergence and high-level divergence, \( A_+ / A_- \) may be very small if the updrafts are confined to a few very intense storms. This provides for rather narrow regions of strongly ascending air surrounded by broad regions of gently descending air. Under these conditions one might expect to find unstable solutions to (4). If the circulation increased exponentially with time, one would expect to find rather spectacular results in the local weather. Although there are no data which will allow a direct verification of (5), as in the case of stable oscillations, one might turn to instances of record-breaking rainfalls to determine if the meteorological conditions were right to expect \( A_+ / A_- \) to be small and thus lead to runaway convection. In Table 2 is a list of outstanding cloud-burst rainfalls.

The two greatest recorded rainfalls in the United States occurred at Thrall, Texas, in 1921 and at Hallett, Oklahoma, in 1940. Lott [1953a] describes these storms and the attending meteorological situations. The Thrall storm was most remarkable in that the bulk of the rain fell in two bursts only lasting about four hours each. Lott attributes the bursts to the passage of an isobaric low-pressure center which was the remnant of a small hurricane which entered the Mexican coast. The conditionally unstable lapse rate, high moisture content, and convergence-divergence pattern agree very well with the conditions one should expect for runaway convection.

<table>
<thead>
<tr>
<th>Cloud location</th>
<th>Date</th>
<th>Average observed frequency</th>
<th>( A_+ / A_- ) (computed)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tucson, Arizona</td>
<td>July 23, 1956</td>
<td>11</td>
<td>1.0</td>
</tr>
<tr>
<td>Tucson, Arizona</td>
<td>July 24, 1956</td>
<td>11</td>
<td>0.7</td>
</tr>
<tr>
<td>Flagstaff, Arizona</td>
<td>Aug. 22, 1958</td>
<td>10</td>
<td>1.0</td>
</tr>
<tr>
<td>Mt. Withington, N. M.*</td>
<td>Aug. 16, 1957</td>
<td>10.3</td>
<td>1.34</td>
</tr>
<tr>
<td>Mt. Withington, N. M.*</td>
<td>Aug. 20, 1957</td>
<td>8.5</td>
<td>1.33</td>
</tr>
</tbody>
</table>

* Vonnegut and others [1959].
DISCUSSION

Table 2—Cloudburst rainfall in the United States

<table>
<thead>
<tr>
<th>Location</th>
<th>Date</th>
<th>Lapse rate</th>
<th>Vertical moisture</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Franconia, Va.</td>
<td>Sep. 1, 1952</td>
<td>Unstable</td>
<td>Surface to 550 mb</td>
<td>Tornadoes associated with hurricane Able</td>
</tr>
<tr>
<td>Thrall, Texas</td>
<td>Sep. 9-10, 1921</td>
<td>Unstable</td>
<td>Surface to 700 mb</td>
<td>Record 19.65 inches in 12 hr in two bursts (remnants of Gulf hurricane)</td>
</tr>
<tr>
<td>Hallett, Okla.</td>
<td>Sep. 4, 1940</td>
<td>Unstable</td>
<td>Surface to 300 mb</td>
<td>15.5 inches in 9 hr., non-frontal air mass</td>
</tr>
<tr>
<td>Chicago, Ill.</td>
<td>Oct. 9-19, 1954</td>
<td>Unstable</td>
<td>Surface to 500 mb</td>
<td>6.72 inches in 9 hr., record, all thunderstorm rain, south of warm front</td>
</tr>
</tbody>
</table>

The Hallett storm occurred as an air-mass type in very moist, conditionally unstable air. It is noteworthy that the rain fell between 0245h and 11h00m CST which signifies that this was a nocturnal storm which had been growing, perhaps, from the previous afternoon and thus must have built into an enormous-sized single thunderstorm. It, too, qualifies as an example satisfying the conditions for runaway convection. The Franconia storm was similar to the Thrall storm in that it was associated with the remnants of a dying hurricane low center. Although no record rains accompanied this storm, it was notable for the tornadoes it spawned.

The Chicago storms are additional examples of the conditions resulting in runaway convection. Here one had low-level convergence, high moisture, and conditionally unstable air south of a warm front so that it was possible to have small but intense updraft areas and widely distributed downdraft areas [Means, 1956].

Following this line of argument, one may inquire into the possibilities for the creation of intense, self-sustaining, local storms through artificial means. One feature which must be present is the large areal inflow to and outflow from the storm. In addition, high moisture, at least in the inflowing air, should be present. Nature provides ready-made situations of this type in the vicinity of low-pressure areas and one would need only to intensify certain updraft regions to set off runaway convection since the other elements for a self-sustaining storm are present already. One would expect to find the most favorable opportunity in a warm sector where moist tropical air is present with appreciable sunshine. Under these conditions, it seems favorable to encourage the growth of large Cumulus congestus clouds by dusting smaller Cumuli with solar energy converting materials such as carbon black. The dusting should be done so as to create a larger flow field by merging several adjacent Cumuli. This seems to be the way nature accomplishes the formation of thunderstorms and we might well begin by trying to imitate her.

REFERENCES


Discussion

Mr. Douglas K. Lilly—I'd like to propose another possible interpretation of the evidently periodic characteristics of those observations. It seems that there must be some instability in order
for the cloud to rise in the first place. I think possibly that the individual elements, both the small rapidly changing ones and the larger ones are all unstable but have a life cycle because of precipitation processes. It is this life cycle which gives them a semiperiodic appearance, somewhat similar to the occlusion processes that give periodic appearance to cyclones.

*Mr. C. E. Anderson*—In regard to the problem of bubble versus periodic motions, I thought the data I showed from Vonnegut’s work would indicate that if these were discrete elements, they could not act over a large segment of the atmosphere simultaneously. Since we find the motions near the cloud base, within the interior of the cloud, and at the top to be in phase, it is very hard to account for this by assuming some type of discrete element moving upwards.
On the Dynamical Prediction of Large-Scale Condensation by Numerical Methods

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Abstract—The paper discusses properties of the hydrothermodynamic frameworks thus far employed, the approximations regarding the microphysics of precipitation, the quality of results of numerical integrations for different models, further work in the construction of more sophisticated dynamical models, and investigations of the relation of large-scale liquid water content and of water vapor, and the implications for the dynamical prediction of cloud formation and dissipation.

A brief survey of the state of the art—Attempts to make dynamical precipitation forecasts by numerical means began over 5 years ago. The first efforts were merely to employ the vertical motions calculated during the course of baroclinic numerical forecasts [Smagorinsky and Collins, 1955; Miyakoda, 1956; and Smeebye, 1958]. The hydrodynamics were quasi-geostrophic, the baroclinic structure was described by information at three levels, and the potential vorticity was linearized when it appeared undifferentiated. Furthermore, the released latent heat was not permitted to add energy to the system.

It was assumed that precipitation occurred upon attaining saturation, but it was already realized that the space-averaged relative humidity need not be 100% for condensation or precipitation to occur. The possibility of supersaturation, supercooling, evaporation from falling drops, or inadequate nucleation was ignored. The results were reasonably encouraging, but further work suggested that departures from observation were to a large extent a result of errors in the large-scale hydrothermodynamics. The most obvious defect was the neglect of released latent heat, which is a destabilizing effect [Smagorinsky, 1956; Aubert, 1957]. This alone can amplify the large-scale upward vertical motions by as much as an order of magnitude giving maxima as large as 50 cm/sec. The degree of destabilization increases with decreasing scale and decreasing static stability.

It was also possible to remove the mathematical limitations of quasi-linearization and to add the barotropic effects of large-scale mountains. Since these models now possessed energy sources and moisture sinks it was desirable to provide a pseudo-boundary layer which would allow for surface friction and evaporation depending on land or sea. The quasi-geostrophic model equations have been recast to be governed instead by the balance (or quasi-non-divergent) condition [Smagorinsky and collaborators, 1959]. The results are often better, especially in the movement of the systems, but also suffer because of new limitations introduced. The relatively smaller characteristic scale of moisture distributions present special difficulties and somewhat special numerical techniques have been devised to reduce truncation error.

Although very distinct progress has been made, the remaining hydrodynamic degeneracies leave 24-hour precipitation forecasts with much to be desired. It is quite obvious that the geostrophic approximation as well as the balance condition are really valid for the very large-scale quasi-barotropic components of the motion. Much of the validity is lost when trying to describe the dynamics of the smaller-scale baroclinic developments which occur sporadically as extratropical cyclogenesis. This is probably the major reason why geostrophic and balanced baroclinic models on the average give no better wind forecasts at 500 mb than do barotropic models. The effects of released latent heat are on still a smaller scale, and the inertial-gravitational modes of atmospheric motion become even more important, if not essential. The divergent components appear to be of consequence not only in dynamical interactions but also for a proper accounting of the moisture budget.

There therefore seems to be no question that further progress will depend on our ability to construct an adequate dynamical framework. Until quite recently, attempts to integrate numerically the primitive equations had not suc-
ceeded. However some progress has now been achieved in devising a stable system for numerically integrating the primitive equations for baroclinic flow [Smagorinsky, 1958; Hinkelmann, 1959] as well as for barotropic flow [Phillips, 1959]. This experience is now being applied to the construction of a near-hemispheric, four-level model allowing moist adiabatic processes. This model will include the baroclinic as well as barotropic orographic effects and also a simple accounting of boundary layer processes.

It is apparent that the much smaller scale convective motions pose a special problem. It would, of course, be impractical to consider describing them by explicit dynamics. Ideally desirable is an adequate statistical-dynamical theory of moist convection which can define the classes of unstable ambient states and account for the systematic nonlinear interaction between the convective motions and the larger scale motions resolvable by explicit dynamics. Some work in this direction has been done by Malkus and Witt [1958] for dry convection. Moist convection, on the other hand, seems to be inherently different mechanistically and considerably more difficult to cope with. However, there is promise that numerical model experiments will yield further insight into the moist convective process, and work is now being undertaken in this direction.

Some gross properties of the macrophysics of condensation and precipitation—Until now the limitations of the hydrodynamic contexts did not warrant refinements in the assumptions regarding the physics of condensation. However, contiguous studies have indicated the way toward a somewhat more adequate linkage of the large-scale hydrodynamics and the condensation process. In particular, it would be desirable to allow for the non-precipitating cloud stage, since until now only a distinction between clear sky and precipitation has been attempted.

It is generally known that cloudiness and even precipitation are found to occur at space-averaged relative humidities considerably less than 100%. This cannot be dismissed as a purely instrumental aberration. Humidity as measured by the instrument and averaged from the sounding, represents the mean of a frequency distribution of smaller-scale humidity variations with considerable standard deviation. This must mean that for values of the average humidity considerably less than 100%, some condensation may be occurring due to saturation at the high end of the distribution. One would then expect that the amount or density of condensation (that is cloudiness) will increase with increasing mean humidity. Furthermore, one may view precipitation as resulting from sustained and very dense condensation, sufficient to create large enough particles, say for example by coalescence or the ice crystal process, to overcome the upward vertical currents.

Indeed one does find empirically that non-convective cloud amount, classed as low, middle, and high, is highly correlated with the average relative humidity in the respective layers. Precipitation, if interpreted as corresponding to a cloud amount somewhat greater than 1.0, also fits such a correlation. In fact, the simple linear relation

$$c = \beta h - \alpha \geq 0$$

(1)

for each layer yields an excellent fit. Here $c$ is the cloud amount, $h$ is the relative humidity in percent, and $\alpha$ and $\beta$ are empirical coefficients. The fact that the instantaneous value of $c$ does not appear to depend on the instantaneous vertical velocity is not surprising. One would expect non-precipitating condensation to depend only on the accumulated history of the vertical motion, which after all is reflected in the humidity.

For the purpose of establishing the coefficients $\alpha$ and $\beta$, it was assumed that the mean relative humidity in the 1000–500 mb layer corresponded to the span of low cloudiness, 500–550 mb to middle cloudiness, and 550–300 mb to high cloudiness. A graph of the linear relations is shown in Figure 1. (The writer is grateful to S. Hellerman for his assistance in determining this relation from careful analysis of a substantial volume of synoptic data.) It is of interest that all three levels tend to converge to $c = 1.3$ for $h = 1.0$.

![Figure 1](image-url)
It is well known that in the lowest 100 mb next to the Earth’s surface humidities close to 100% are necessary before condensation is observed. This of course is due to the relatively intense mixing in the boundary layer. When surface winds are light, condensation in the form of fog may occur with lower humidities, but still considerably higher than in the ‘free’ atmosphere. It is of interest that non-industrial haze, which may be regarded as low-density fog, also occurs at high humidity. Under normal surface wind conditions, the intensely mixed layer is capped by an inversion through which the turbulence subsides almost discontinuously, and it is above the inversion that some condensation may occur at humidities as low as 60%.

The empirical relations found in no way are intended to reflect this discontinuous turbulence structure in the 1000–800 mb layer, but rather to give a measure of the integrated effect of turbulence in the entire layer. To account adequately for the finer grain structure of turbulence would require far greater vertical resolution and more refined techniques than have been employed here. One could guess as to what Figure 1 would look like if ‘low clouds’ were stratified according to the mean relative humidity: (a) between 800 mb and cloud base, and (b) between cloud base and 1000 mb. For (a) the standard deviation of relative humidity would be larger than the mean in the 1000–800 mb layer due to weaker mixing so that condensation could occur at lower mean relative humidities. Curve (a) would then intercept the c = 0 axis at $h = 0.45$. Since the most intense mixing would be confined to the layer next to the ground, the frequency distribution of relative humidity would be very peaked. The curve (b) would intercept c = 0 at $h = 0.85$ or 0.90. Of course in this boundary layer c no longer corresponds to cloud amount but rather to the visibility which in the absence of industrial pollutants is fairly good measure of liquid water content in clouds as well as in fog [Houghton and Radford, 1938]. The fact that the visibility decreases with increasing relative humidity for humidities over 70% [see, for example, Neiburger and Wurtele, 1949] tends to support the above supposition.

An effective means for demonstrating the ‘goodness of fit’ of Figure 1 is to deduce cloud amount from synoptic radiosonde data only and to compare with ‘actual’ cloud observations, recognizing that the upper-level clouds when obscured from below must be estimated. All possible data including airways reports were employed. The comparisons are shown in Figures 2 and 3, for two cases, late spring and late fall. Included also are the geopotential fields at the 1000-, 700-, and 500-mb levels. The comparisons in the vicinity of the mountains must be ignored since the low-level humidities are fictitious and the observed cloud layers correspond to lower pressures than do sea-level observations.

The two cases shown in Figures 2 and 3 are from wholly independent data. However, the empirical linear relations of Figure 1 have been found extremely useful as an analysis aid in regions of sparse radiosonde data such as over the oceans. Moreover, even over continental U. S. the radiosonde network is often inadequate to fix the phase of smaller-scale distributions of humidity such as are associated with frontal zones.

The surprisingly good relation between liquid water content and water vapor suggests a means for incorporating the cloud stage in the water budget, and in fact will lead to a measure of the efficiency of moist adiabatic processes in large-scale condensation. Since cloud cover is a relative two-dimensional measure of liquid water, if we assume the vertical extent of large-scale (non-convective) clouds to be proportional to a linear measure of the horizontal dimension, then the volumetric measure of liquid water $W$ is proportional to $c^2$. Defining $W_1$ as the minimum liquid water content necessary for precipitation, which according to Fig. 1 corresponds to $c = c_1 = 1$, then

$$W = W_1 c^{1/2}$$

(2)

We denote by the subscript 2 the condition when $h = 1$, so that $c_2 = 1.3$. We may now write the continuity equations for mixing ratio r, mass of liquid (cloud) water per unit mass of air W, and mass of precipitating water per unit mass of air $W_P$, assuming that water vapor may change by expansional condensation or compressional evaporation, but that precipitating water does not evaporate

$$\frac{d\gamma}{dt} = \frac{\gamma}{\rho} \frac{d\omega}{dt}$$

(3)

$$\frac{dW}{dt} = (\delta - \delta) \frac{\gamma}{\rho} \frac{d\omega}{dt}$$

(4)

$$\frac{dW_P}{dt} = -\delta \frac{\gamma}{\rho} \frac{d\omega}{dt}$$

(5)
Fig. 2—From right to left: cloudiness deduced from humidity soundings and the empirical relations in Figure 1, observed cloudiness, observed geopotential, late spring.
DYNAMICAL PREDICTION OF LARGE-SCALE CONDENSATION
and the precipitation rate is

$$P = \int \left(1 / \rho_r \right) \frac{dW_p}{dt} \frac{dp}{d}\theta$$

(6)

The potential temperature $\theta$ change due to such condensation or evaporation is given by

$$\frac{d \ln \theta}{dt} = \frac{\Gamma}{\rho}$$

(7)

The notation is as follows: $p$ is the pressure, $g$ is the acceleration of gravity, $\rho_r$ is the density of water, $\omega = dp/dt$, $r_s$ is the saturation mixing ratio, $\delta$ is the fraction of mass undergoing moist adiabatic processes, and $\delta^*$ is the fraction of condensing water being precipitated. Also

$$\gamma_m(T, p) = p \left[ \frac{d \ln r_s}{dp} \right]_{th}$$

(8)

$$\Gamma(T, p) = p \left[ \frac{d \ln \theta}{dp} \right]_{th}$$

(9)

where

$$\alpha(T, p) = \frac{L_{f} c_s}{\epsilon \rho T}$$

(10)

$$\gamma(T) = \frac{L}{R_{v} T}$$

(11)

and $T$ is the absolute temperature, $\theta_{th}$ the equivalent potential temperature, $L$ is the latent heat of condensation or sublimation, $(1 - \kappa) = c_s/c_p$ is the ratio of the specific heat of air at constant volume to that at constant pressure, $R^*$ is the gas constant for water vapor $\approx 4.62 \times 10^6$ cm$^2$ sec$^{-2}$ deg$^{-1}$.

Unlike $\delta^*$ which is zero for $\omega \geq 0$, $\delta$ does not depend on $\omega$ since downward motion of a cloud parcel must result in dynamic evaporation so that $dW/dt < 0$ and $dr/dt > 0$. We are not free to specify arbitrarily $\delta$ since (1) and (2) must be satisfied simultaneously by (3) and (4). Ignoring variations in $\alpha$, $\beta$, and $W_1$, then (1) and (2) require that

$$\frac{dW}{dt} = 2 \beta W_1 \sqrt{c \frac{d}{dt}}$$

(12)

Since by definition

$$r = h r_s$$

then

$$\frac{dr_s}{dt} = \frac{1}{r_s} \left( \frac{dr}{dt} - h \frac{dr_s}{dt} \right)$$

(14)

The first term, the change of $h$ caused by a change in water vapor, is given by (3); the second term also depends on the fraction of mass undergoing moist adiabatic changes $\delta$, but of course does not vanish for purely dry adiabatic processes, since it changes with temperature and may be written as

$$\frac{dr_s}{dt} = r_s \delta \gamma_d + (1 - \delta) \gamma_d$$

(15)

where

$$\gamma_d(T) = p \left[ \frac{d \ln r}{dp} \right]_{th} = \gamma - 1 \geq 0$$

(16)

and it may easily be verified that

$$\Gamma = \frac{\gamma_m - \gamma_d}{\gamma} = \frac{\gamma_m - \gamma_d}{1 + \gamma_d}$$

(17)

Hence

$$\frac{dh}{dt} = \left[ (1 - h) \delta \gamma_m - (1 - \delta) h \gamma_d \right] \frac{p}{\rho}$$

(18)

Inserting (4) and (18) into (12) yields

$$\delta = \frac{\delta^* + \chi h \sqrt{c \gamma_d / \gamma_m}}{1 + \chi \sqrt{c (1 - h) + h \gamma_d / \gamma_m}}$$

(19)

where

$$\chi = 1.5 \beta W_1 / r_s$$

(20)

For $c \leq c_1$, we have that $\delta^* = 0$ so (19) and (1) uniquely define $\delta$ as a function of $c$ or $h$. Also, for $c_1 \leq c \leq c_2$ and $\omega \geq 0$, we have that $\delta^* = 0$, again uniquely defining $\delta$. On the other hand, when $\omega < 0$, we have $\delta^* = 0$ as before, and since all condensing water vapor must be precipitating when $c = c_2$, then by (4) $\delta^* = 1$. Assuming $\delta^*$ to vary linearly in this range then

$$\delta^* = \frac{c - c_1}{c_2 - c_1} = \frac{c - 1}{0.3}$$

(21)

$\geq 0$ for $c_1 \leq c \leq c_2$, $\omega < 0$.

We furthermore see that a maximum liquid water is attained for $c = c_2$, which by (2) is

$$W_z = W_1 c_2^{2/3}$$

(22)

Hence the maximum liquid water content is 50%
larger than the threshold liquid water content below which precipitation cannot occur. There is some empirical evidence [Aufm Kampe and Weickmann, 1957; Mason, 1957, p. 231] that such a maximum exists, its value depending on the type of cloud. This is reasonable since the precipitation drop size and hence \( W_1 \) should depend on the characteristic magnitude of the vertical velocities working against gravity. For stratiform clouds, with a characteristic vertical velocity of 5 cm sec\(^{-1}\), \( \rho W_2 = 0.5 \) gm m\(^{-3}\) (\( \rho \) is the air density); while for Cumulus clouds, with vertical velocities of the order of 50 cm sec\(^{-1}\), \( \rho W_2 = 5 \) gm m\(^{-3}\). Komabayasi [1957] suggests that \( \rho W_2 \) varies as the \( \frac{1}{2} \) power of the vertical velocity. We have then for stratiform clouds that \( \rho W_1 = 0.3 \) gm m\(^{-3}\).

One would expect \( \rho W_2 \) for non-convective clouds to be a maximum in mid-troposphere \((\approx 550 \) mb\) where \( \omega \) is a maximum in the large scale; this is consistent with the findings of Aufm Kampe and Weickmann [1957], who report much lower liquid water contents at the Cirrus level. However, for the purpose of estimating the variation of \( \delta \) with \( h \) in different cloud layers we use an average value of \( \rho W_1 = 0.3 \) gm m\(^{-3}\), and furthermore assume the temperature to vary as in the standard atmosphere. The corresponding values of the parameters occurring in (1), (19), (20), and (21) are given in Table 1. Figure 4 shows \( \delta \) as a function of \( h \) and \( c \) for each of the three tropospheric layers for which \( \alpha \) and \( \beta \) have been determined empirically. Figure 4 indicates that for ascending motion the maximum rate of increase of liquid water, which is proportional to \( \delta - \delta^* \), is attained just as precipitation begins. As the mean relative humidity increases beyond this point, \( \delta - \delta^* \) decreases until \( dW/dt = 0 \) at \( h = 1 \). Under continued upward motion the condensing water is directly precipitated. On the other hand for downward motion of an existing cloud the rate of conversion of liquid water to water vapor, which is proportional to \( \delta \), has a maximum at \( h = 1 \) and decreases monotonically to \( c = 0 \).

The net effect is a process analogous to that resulting from entrainment in Cumulus development, but on a smaller scale. The fact that mean relative humidities of 100\% are not often observed even from precipitating clouds must mean a significant dilution of the moist adiabatic process during precipitation. Hence even larger upward vertical velocities than calculated on the assumption of no dilution \((\delta = \delta^* = 1)\) are necessary to explain the amounts of large-scale precipitation observed.

**Table 1—Values of parameters used in constructing Figure 4**

<table>
<thead>
<tr>
<th>Layer</th>
<th>( \alpha )</th>
<th>( \beta )</th>
<th>( \rho )</th>
<th>( \gamma_1 )</th>
<th>( \gamma_2 )</th>
<th>( \gamma_m )</th>
</tr>
</thead>
<tbody>
<tr>
<td>mb</td>
<td>0.433</td>
<td>1.733</td>
<td>0.6</td>
<td>1.1</td>
<td>6.2</td>
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<td>0.9</td>
<td>3.7</td>
<td>5.0</td>
<td>2.9</td>
</tr>
<tr>
<td>(mid-)</td>
<td>2.0</td>
<td>3.33</td>
<td>1.1</td>
<td>7.9</td>
<td>4.6</td>
<td>2.4</td>
</tr>
<tr>
<td>(low)</td>
<td></td>
<td></td>
<td></td>
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**Fig. 4**—The percentage of mass undergoing moist adiabatic processes \( \delta \) as a function of relative humidity \( h \) for low, middle, and high clouds; the corresponding cloud amount \( c \) is given along the upper abscissa; solid lines are for the non-precipitating stage, dashed lines are for the precipitating stage; for \( 0 < c \leq 1.0 \), each curve is valid for \( \omega \leq 0 \); for \( 1.0 \leq c \leq 1.3 \) the solid line is valid for \( \omega \geq 0 \) and the dashed line for \( \omega < 0 \).

**References**


Smagorinsky, J., and collaborators, manuscript in preparation, 1959.


Discussion

*Mr. Jerome Naminas*—If the ultimate aim in all this is to predict in detail, at what point there may have to be a cut-off in this prediction scheme. It would have to make inferences about conditions responsible for run-away processes and various things down to some scale. That is, must we settle for a certain scale? I'd like to ask Dr. Smagorinsky if he believes there is no cut-off point and if he expects to go to the bitter end and attempt to predict weather on all scales by numerical process.

*Dr. Joseph Smagorinsky*—I would say that one can reasonably place a cut-off at the point where the statistical dynamics of the smaller scale motions are sufficiently stable and well understood. The ability to establish a threshold of turbulence permits the study of the explicit dynamics of the synoptic scale motion with adequate provision for the interaction with the scales of motion ultimately responsible for the dissipation of kinetic energy. Such a threshold of the horizontal scale appears to be of the order of 100 km. However, as I pointed out in my paper, the interaction of small scale convection with large scale motions is hardly understood.
Orographic-Convective Precipitation as Revealed by Radar

BERNICE ACKERMAN

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Abstract—Summer cloud systems in the arid and mountainous region around Tucson, Arizona, are predominantly convective in nature. Extensive radar observations of these systems have been made by the Institute of Atmospheric Physics, University of Arizona, using height-finding radar. A study of the level of formation of radar echoes, based on data collected during the summer of 1956, indicates that an all-water process, as well as one involving the ice phase, was effective in initiating precipitation. Moreover there appeared important day-to-day differences in the efficiency of the water mechanism.

Introduction—For the past four or five years the Institute of Atmospheric Physics, University of Arizona, has been investigating the cloud and precipitation characteristics of arid regions. During the course of this investigation extensive radar observations have been made using the AN/TPS-10, a height-finding radar. Radar data obtained during the summer rainy season in 1956 have been used to study precipitation features around Tucson. In particular, this report is concerned with the heights at which echo clouds first formed; the significance of such information, of course, lies in what may be learned about the mechanisms initiating precipitation.

The AN/TPS-10 radar has a 3-cm wavelength and an elliptical beam with widths of 0.7° in the vertical and 2° in the horizontal. The radar system was monitored and power levels maintained at 47 dbm and -79 dbm for transmitted and minimum detectable returned power, respectively. Automatic photography of the radar scope gave a permanent record. There was approximately a three-minute interval between successive observations at a given azimuth.

The radar, which is located on the campus of the University of Arizona, scanned the area within sixty miles of Tucson. As can be seen in Figure 1, this area is of interest because of its topography as well as its aridity. The terrain varies in elevation from 2000 to almost 10,000 ft msl. Several small but distinctive mountain ranges are in the area; most of these are fairly well defined by the 5000 ft contour.

The summer rainy season starts rather suddenly in late June or the early part of July and continues through August. During these months there are alternating dry and wet periods, the latter characterized by fairly general Cumulus activity and convective rain.

Although association between clouds and mountains is usually observed, the location of the major convective activity varies considerably from day to day. On few days are clouds associated with all of the mountain ranges, and as a corollary, a given range does not give rise to cloud systems every day. This variability is illustrated in Figures 2 and 3 in which are shown the locations of precipitation echoes on two of the days studied. (In Fig. 2 and 3 the division of the precipitation areas into groups was made for purposes of a detailed description of the echo patterns given in a more complete report of this work [Ackerman, 1959].) On only two of the seven days studied were the echo patterns similar. It is evident that processes considerably larger in scale than the cloud are factors in determining the location of cloud development, even in a region as small as the one being considered.

Analysis and discussion—Because of the time-consuming nature of the data reduction, the analysis covered only seven days of the season. The criteria used in the choice of days were sufficiently objective to insure a random sample. They were (1) that convective precipitation, as indicated by radar, has occurred; (2) that radar data be available; and (3) that the days be scattered through the summer. The resulting sample was composed of nearly 300 first echoes, that is, first appearances of echo clouds. For all of these cases the area was known to be free of an echo three minutes earlier.

In Figure 4 are shown the frequency distributions of the temperatures at the bases and tops
Fig. 1—General topographic features in the vicinity of Tucson, Arizona

Fig. 2—Plan view of echo areas on July 20, 1956; areas represent echoes detected on a single 360° azimuth scan at approximately one-hour intervals, time identification to nearest whole or half hour, mst; to simplify the presentation a single outline encloses groups of closely spaced echoes, and, occasionally, groups of echoes on consecutive hours when they occupied essentially the same region; on the base chart are shown the 3000 ft contour and elevations greater than 5000 ft msl (shaded areas); distance scale, in miles, is shown to left of center
Precipitation as revealed by radar

Fig. 3—Precipitation areas on August 24, 1956; presentation as described in Fig. 2

of the first echoes. The conversion from height to temperature was made using the regular Tucson radiosonde. Since the temperature-height relation differed little on the seven days, the histograms for the height (msl) are very similar to those shown in Figure 4.

The height at which echo clouds first formed was highly variable, with ranges of over 40°C of temperature (roughly 20,000 ft) in both bases and tops. The latter were smoothly distributed with two-thirds of the cases falling in the interval between —4 and —16°C. The modal value of —10°C is similar to that reported for the tops of initial echoes in clouds in New Mexico [Workman and Reynolds, 1949]. Almost all of the echo clouds had tops above the freezing level when they were first detected, but, as can be seen from the lower section of Figure 4, a large fraction of them (nearly 60%) extended below the freezing level.

The height distribution of the bases of first echoes (Fig. 4) is quite different from that of the tops. The bases occurred with high frequency at two levels, one above the freezing level at temperatures between —4 and —8°C, the other well below the freezing level with temperatures between +12 and +16°C. Cases for which bases were colder than —4°C were no doubt echoes from particles produced by the Bergeron ice mechanism, but at least portions of the echoes having bases at temperatures above +10°C must have been composed of particles grown without the involvement of the ice phase.

The high incidence of first echoes with bases at these very warm temperatures indicates that an all-water process must be effective in initiating precipitation in some of these Arizona clouds, if not in the entire depth indicated by the first echo, then at least in the lower portion. This conclusion is reached by reasoning similar to that used by Battan [1953] in describing the effectiveness of the all-water mechanism in Ohio; namely, that the large drops at the base of the echo cannot be explained by downward movement of large drops from the freezing level because the drop sizes or downward flow of air re-
In Figure 5 are shown the frequency distributions of the levels of echo formation on the individual days. Although there are day-to-day differences in the heights at which the tops of the initial echoes occurred they are not nearly as pronounced as the differences observed for the base heights. The total range of temperatures at which the echo bases occurred was about the same for all days but the ‘characteristic’ base height varied from one of the ‘preferred’ levels to the other. On July 24 and August 17 there was high incidence of first-echo bases at temperatures between −4 and −8°C; on August 1 and August 24, there was marked preference for first-echo bases to occur at temperatures around 14°C. The tendency for a ‘preferred’ level is less marked on July 27 and August 13, but on both days the distributions were skewed, toward the colder levels on the former day, toward the warmer ones on the latter.

Only on July 20 did the bases occur with high frequency at both levels. The echoes on this day, in particular, tend to support the thesis that precipitation may be initiated by different processes in different parts of the cloud, and that the source of variability is in the time requirements for the two processes. The formation of the echo in the lower parts of the cloud evidently lagged only slightly that in the upper portion. Over 60% of the echoes with high bases when initially detected had bases at the lower ‘preferred’ level (+12 to +16°C) three minutes later. This represented between 6000 and 10,000 ft of descent in three minutes, in many cases coincident with ascent of the echo tops.

To date, attempts to find an explanation of the variation in the effectiveness of an all-water process from cloud to cloud and from day to day have been unsuccessful. It is possible to find one that fits the observations for two or three of the days, but what is required, of course, is an explanation fitting the observations on all seven days. A consistent relationship could not be found between height of echo base and time of day or time from beginning of convection. Similarly there appeared to be no association between ‘cold’ and ‘warm’ base first echoes and topography. Nor was there any indication that proximity to older rain clouds at the time of formation made a difference. Locale and time of occurrence do not appear to be important factors.

The day-to-day differences in the characteristic height of first echo bases must reflect variation in the interval of time required for the ini-
Precipitation as Revealed by Radar

Fig. 5—Frequency distributions of the temperatures at the tops and bases of first echoes for each of the seven days studied; a few cases for which top height measurements were questionable were not included, causing the small discrepancies noted in sample sizes (N).

The data reveal that a condensation-coalescence process is effective in initiating precipitation, even in an arid mountainous region. Moreover, they indicate that precipitation may be initiated either by an all-water process in the lower reaches of a cloud or by the ice-crystal mechanism in the upper reaches or both, either simultaneously or one lagging the other. A critical parameter is the time required for these mechanisms to develop large drops. The factors deciding the time constants remain to be determined. Unexplored also is the problem of the relative efficiency as far as surface rainfall is...
concerned; that is, more or less cloud water realized at the ground when precipitation is initiated low in the cloud before or simultaneously with that in the upper regions.

Acknowledgment—This research was sponsored primarily by the Institute of Atmospheric Physics, University of Arizona, and by the Geophysics Research Directorate under Contract No. AF19(604)-1134.

Discussion

Dr. B. J. Mason—I think Miss Ackerman would agree with me that it is difficult to disentangle the two processes which may be occurring in the case where the first echo is both above and below the freezing level. I have made observations of it, and this brings me to the question, What do we mean by a first echo? A first echo is when you first see something on the radar screen, but what it means in terms of size and so forth of the particles in the cloud depends upon the sensitivity and range of the radar. So, I would like to ask Miss Ackerman if she could give some kind of figure in terms of the sensitivity of the radar or whatever it might be that she takes for a first echo. Experience in England is that the first radar echo may just straddle the freezing level; but the echoes grow explosively in a minute or two both above and below. If you catch it a minute too late, you might come to a different conclusion, because they change so quickly.

To come to the last point, the day-to-day variation, nuclei are the first things microphysicists think about but I am inclined to think that factors controlling the cloud dynamics may be more important. I wonder whether you had at the same time made any measurements of the actual humidity distributions in the air?

Miss Bernice Ackerman—The sensitivity of the radar was monitored and kept constant so we could compare the data. I made rough calculations of the minimum particle size detectable by the radar. The calculations give a diameter of around 400 microns for a range of 20 mi and an assumed water content of 1 g/m^3. Of course, the size varies with range and droplet concentration.

There was a three-minute time interval between two pictures. This could be part of the reason for the wide range of temperature (40°C) over which these first echoes appeared. I agree some of these echoes with low bases could have started just below the freezing level, but this still does not mean that they did not develop by a water mechanism, and I am not trying to separate the two types. In the lower portions of these echoes, at least, and possibly in some of the upper portions, the growth of droplets is through a water mechanism.

Regarding Dr. Mason's question on the humidity distributions, the only type of data I have for these days is that available from the regular radiosonde. This is, of course, one of the things one looks at immediately. There was essentially no difference in the vertical lapse of temperature and humidity between the days.

Mr. C. J. Todd—Did you have available any visual observation of the cloud growth rate?

Miss Ackerman—No.

Dr. W. Hitschfeld—Did you have any information about cloud densities at the time the first echo broke out; that is, the amount of water, the quantity of water per cubic meter?

Miss Ackerman—No, the only data available were the radar data.

Dr. Hitschfeld—I think that cloud-density studies, even if calculated only roughly from the tepligrams, might allow one to eliminate the uncertainty with respect to the precipitation process. If the cloud densities are high enough, it might suggest that the East process has a better chance of being active (T.W.R. East, Precipitation of convective water clouds, in Artificial Stimulation of Rain, Proceedings of First Woods Hole Conference, Pergamon Press, 1957, p. 192-201).

Dr. Donald M. Swingle—Did you make any correction for the radar beam width uncertainty?

Miss Ackerman—Not as far as the actual cited levels are concerned. I am not trying to say
these are well-defined levels in any sense of the word. They may vary by two degrees or perhaps even four degrees in temperature, because of beam-width effects.

**Dr. Swingle**—Do I understand that, in fact, you have no observation of what the temperatures actually were in the cloud?

**Miss Ackerman**—No, but the 'preferred' echo heights are very well separated.

**Dr. R. Wexler**—Did you look for any relationship between the heights at which these first echoes appeared and the maximum height which was subsequently reached or in other words, for any relation at all between the region where the echo first appeared and the subsequent maximum development?

**Miss Ackerman**—Yes, I tried to do so, using data for one day when there was a high frequency of echoes. I did not find any correlation.

**Dr. W. E. Howell**—In a situation where you obviously hope in connection with some of these experiments to have a fairly repeatable cloud condition day after day, I wonder if it is not a little disturbing to find the pattern of echo varies so much from one day to another over these different mountain ranges; and I wanted to ask if these had been related in your observations to the previous state of the ground, whether the previous occurrence of precipitation and consequent moistness of the ground affected the subsequent development on a later day.

**Miss Ackerman**—No, they were not, and outside of going into the radar data of each day, I do not know how we can really do it. In the type of region we are talking about, there is not much rain and the rain comes in the form of scattered showers. We tried to see whether there was some correlation between the amount of rainfall reported by the existing stations with any of these data, but there was none. I really did not expect there to be, in view of the scattered nature of the echo areas and the sparseness of reporting stations.

**Dr. C. L. Hosler**—Did you note any tendency of echoes when there were few echoes to reach greater heights than on the days when there were many?

**Miss Ackerman**—No, I also thought this might be a possibility.

**Mr. Jerome Namias**—A number of people seem to have suggested quite an ambitious research program for Miss Ackerman. In the geographical area concerned, as you undoubtedly know, there is great precipitation sensitivity in the summertime depending on the location of the moist tongue. The gradient of moisture on both sides of this tongue is rather intense. Small shifts of the tongue will influence the whole synoptic situation. Instead of working just with the Tucson radiosonde, I suggest plotting some isentropic charts to highlight the tongue and in this manner get an idea of the lateral mixing which could provide all sorts of variations even within six-hour periods.

**Miss Ackerman**—As I said before, a real synoptic analysis was not made, but it might be a good idea to do so.

**Dr. Mason**—I think, Miss Ackerman should not be distressed that she can not get these things to tie up. This points to the difficulty of these problems even when one is dealing with everyday Cumulus. It also points out that if one uses one tool such as radar, this is very restrictive. It only gives a very small part of the picture. We have to regard radar as one tool to be used in conjunction with a lot of others and these problems are very much more complicated than we thought at first. We have to go back and realize that Cumulus evolution is determined by the large-scale distribution of moisture, temperature, vertical motion, etc., right up the synoptic scale, and the more of these data that point this out, the better it will be for long-term research in cloud physics.

**Miss Ackerman**—Yes, I agree with you, and I am not particularly disturbed at not finding a simple explanation. We unquestionably have a complex situation. I would just like to repeat one comment made earlier by Dr. Byers; namely, the danger of taking one situation and developing a general theory from it. In evaluating the data of seven days, we find seven different situations.
Microstructure of Storms as Described by Quantitative Radar Data

Pauline M. Austin

Weather Radar Research, Massachusetts Institute of Technology, Cambridge, Massachusetts

Abstract—Instrumentation which has been developed recently presents radar echoes from precipitation in the form of range-corrected signal-intensity contours, thus making it possible to observe in a quantitative manner the smaller-scale features within the precipitation areas which appear on the normal radar-scope presentation. This paper presents some preliminary results of the analysis of such data for two types of storms: warm-front type rain in an unstable atmosphere, and showers associated with instability lines. Dimensions, durations and motions of areas of heavy rain and of individual shower cells are considered in an attempt to determine the scales of atmospheric circulations which are particularly significant in the production of precipitation.

Introduction

Studies of basic physical principles and performance of laboratory experiments have contributed greatly to our understanding of the physical processes involved in the development of precipitation particles, such as nucleation, condensation, and coalescence, and have also shown the influence upon these processes of the immediate environment of the particle. However, the manner in which such processes occur in the atmosphere and the influence of the larger scale environment in encouraging or inhibiting the growth of hydrometeors can be learned only through detailed observations of actual storms. Radar observations provide a description of the distribution of precipitation either aloft or as it reaches the ground, although until recently such information was largely qualitative. Instrumentation described by Kodaira [1957] is now available which presents radar data in the form of range-corrected signal-intensity contours, thus making it possible to observe in a quantitative manner the smaller scale features within the precipitation areas which appear on the normal radar-scope presentation. It is the purpose of this paper to describe some of these small-scale features and their behavior for two types of storms: warm-front type rain in an unstable atmosphere, and squall lines. Dimensions, durations, and motions of rain areas are considered in an attempt to determine the scales of the atmospheric circulations which are particularly significant in the production of precipitation.

The signal intensity contours are obtained with an SCR-615-B radar which employs 10-cm radiation. Therefore the measurements are not distorted by attenuation. However, the SCR-615-B radar is not sufficiently sensitive to detect light rain except at very close ranges. Therefore the data are supplemented by photographs of the PPI of the AN/CPS-9 radar maintained by the Air Force at Great Blue Hill and by hourly rainfall records from the U. S. Weather Bureau Cooperative Observer Network.

Warm Frontal Rain

General description of storms—Data are available for four storms where the rain was of the warm-front type but was not associated with a coastal storm. Only two of these storms have been analyzed in detail: October 17-18, 1957, and November 5-9, 1957. In both cases a cold front oriented approximately north-south was approaching from the west, a warm front lay to the south of the station, and the winds aloft were from the southwest. The lapse rate in the warm air ahead of the front was approximately moist adiabatic below 20,000 feet and slightly stable above that level.

The rain patterns associated with these storms are very similar. The most striking feature is a broad loose band oriented in the north-south direction which was about 250 mi in length and 40-50 mi across. The heaviest part of the precipitation was in this band and it appeared to be about one hundred miles ahead of the surface cold front. On the PPI photographs the band had a pebbly structure, especially at long ranges, indicative of many small showers within the general rain area. RHI photographs also showed the
small showers which were 15,000 to 20,000 ft in height and were about two or three miles in horizontal dimension. The echoes had a well-defined bright band near the 10,000-ft level, but the showers were so small it was difficult to tell whether the bright band actually extended through them or not.

The signal intensity contours (Fig. 1) show areas of more intense rain spaced about 30–50 mi apart along the line. These are labelled by the letters A to E in the figure. The areas A and E do not actually show because at the time they were out of range of the SCR-615-B radar. However, they were observed on previous or subsequent contour maps and indicated by the rainfall data. The approximate rainfall rates for each intensity level in Figures 1 and 2 are as follows:

<table>
<thead>
<tr>
<th>Level</th>
<th>Rainfall rate (mm/hr)</th>
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| Preceding the main band in each storm was an area of lighter rain whose internal structure could not be observed in detail by the SCR-615-B radar. In the latter part of the storm the broad band broke up into an irregular rainfall area which showed a tendency to form a series of narrow bands oriented more nearly northeast-southwest, as shown in Figure 2.

Spatial dimensions of storm areas—An attempt has been made to reconstruct the details of the storm structure by combining the information from two radars and the hourly precipitation records. The spatial scales or dimensions have
already been described and may be summarized as follows:

(1) Synoptic scale storm with low pressure center in one case in the Great Lakes region, and in the other over Hudson Bay, but in both cases with a north-south cold front well to the west of the station and a warm front to the south.

(2) Mesoscale areas of precipitation consisting of a poorly defined area of light rain containing only a few heavy showers, then a well-defined broad band containing the heaviest rain, and finally several smaller narrower bands.

(3) Areas of heavy rain within the bands whose spacing and dimensions are on the order of 30-50 mi.

(4) Individual convective showers which are about two to five miles in horizontal dimension and 15,000 to 20,000 ft in height.

Motions of storm areas—The areas of heavier rain, labelled B, C, D in Figure 1, and the individual showers as indicated by small closed contours moved in a very orderly fashion from southwest to northeast. In the two storms studied in detail most of the rain areas seemed to develop or move into radar range at one of two preferred azimuths, roughly due west or southwest. Those appearing in the west often had a slightly more westerly component to their motions than those in the southwest, so that they combined to form a north-south band (Fig. 3).

The velocities of the rain areas were in general agreement with the middle-level winds which were nearly uniform between 5000 and 18,000 ft. The large intervals in time and space between radiosonde soundings and the uncertainties in the wind measurements make it impossible to de-
MICROSTRUCTURE OF STORMS BY RADAR

Fig. 3—Regions and times of development of rain areas, November 8-9, 1957

termine the details of the upper air wind field at a particular time and place and to make a closer comparison with the motions of the rain areas.

The large band itself seemed to drift in an easterly direction at a speed slightly less than half that of the individual rain areas.

Durations of rain areas—In each of the storms the main band retained its identity as a band for a period of about three hours. Then the pattern broke up and became rather disorganized but showed a tendency to form smaller bands as illustrated in Figure 2. The areas of heavy rain within the band and the general area into which it evolved were estimated to have a lifetime of about three hours. Most of them were observed on the contour maps for only 1½ to 2½ hours. Extrapolations beyond the range of the SCR-615-B radar were based on photographs of the AN/CPS-9 scope and the hourly rainfall records.

A number of 'individual showers' as defined by small closed contour lines were tracked and their durations observed. Most of those which could be tracked easily lasted between 20 and 50 min with a few enduring for over an hour, usually the more intense ones. There were also many small areas, especially of light rain, which lasted 15 min or less and were not included in the survey.

Development of rain areas—Because of the limitations imposed by the range of detectability of the radars many of the regions of development cannot be determined with certainty. However, whether the rain areas actually developed at or near their estimated positions of origin or whether they moved into radar range at those locations, some very interesting recurrences are observed. Figure 3 shows the estimated times and appearances of ten of the heavy rainfall areas in the storm of November 8-9, 1957. They seem to ap-
pear in certain preferred regions at approximately hourly intervals. A similar recurrence was observed in the storm of October 18-19, 1957. Five rain areas appeared in northern Massachusetts along the Connecticut River valley, between the hours of 23h30m and 02h15m EST, with intervals between the appearances of 45 min to one hour. Two heavy-rain areas appeared in the vicinity of Hartford, Connecticut, at 22h45m and 23h30m EST respectively and moved along the same path about 30 mi apart.

Sufficient data have not been analyzed to permit one to assess the respective roles of local topography and larger scale storm dynamics in determining the preferred regions and recurrence intervals for the development of such rain areas. However, the fact that areas of heavy rain move along at 45-65 mi/hr and persist for periods of two to three hours suggests that local topography is not the sole factor in producing the observed patterns.

**Squall Lines and Cold-Front Bands**

*General description of storms*—During the summer of 1958 intensity contour data were taken on six squall lines or sharp cold-front bands. A study was initiated to investigate the development of these bands and, in particular, to determine the extent of the similarity in pattern details for the different storms. All except one had the characteristic that for a period of several hours they consisted of a long narrow line of intense convective cells. The one exception showed two lines, neither of which was as long or narrow as those associated with the other storms. All of the lines were either ahead or in the vicinity of a cold front and exhibited strong convective activity, although none would have been classified as a severe squall line. As in the case of warm-front storms, the data from two radars and the hourly rainfall records are being combined to reconstruct the details of the storm as well as possible. The analysis of these storms has not been completed and only a few preliminary results can be presented here.

*Spatial dimensions*—The observed length of the lines at the time of their greatest intensity and sharpness appeared to be limited by the radar range. Hence it may be concluded that they extended for at least 200 mi. The width of each line was about 20 mi, at least an order of magnitude smaller than the length. Individual convective showers whose dimensions were on the order of five to ten miles were spaced irregularly along the line about 20 to 50 mi apart. A typical pattern is shown in Figure 4. The approximate rainfall rates for each intensity level in Figure 4 are as follows:

<table>
<thead>
<tr>
<th>Level</th>
<th>1</th>
<th>2</th>
<th>3</th>
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<tbody>
<tr>
<td>Rainfall rate (mm/hr)</td>
<td>10</td>
<td>25</td>
<td>45</td>
<td>75</td>
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*Time scales*—The development of the line from the time it first appeared as a line (that is, a few scattered showers but definitely lined up) to the time of its maximum intensity and sharpness required two to three hours. For another two or three hours the line persisted and then it began to dissipate. Lifetimes of individual showers, as defined by small closed contours, varied from about 15 min to nearly three hours, but the relatively intense ones usually lasted about an hour. Since the order of magnitude of the time scales involved is the point of interest in this discussion and since the definition of ‘showers’ is dependent in a rather arbitrary manner upon the resolution, both in space and intensity, of the instrument, the observations on shower duration are presented as a qualitative estimate rather than a statistical survey.

*Development of rain areas*—The development or intensification of the line itself usually took place between 12h00m and 15h00m EST, apparently reflecting the effect of diurnal heating, although in one case maximum intensity was reached as late as 18h00m EST. Individual storms sometimes developed within the line and occasionally ahead or behind it, but, in general, there appeared to be several ‘preferred’ regions for shower development. One of these areas was along the Hudson River; on at least two days storms developed there throughout the early afternoon. They appeared on the southern end of the line so that the area of development moved southward along the river valley as the line progressed. In all of the storms the orientation of the line was approximately northeast-southwest and the individual showers moved toward the east-northeast. For some of the storms we have not yet analyzed the AN/CPS-9 data and therefore do not know whether development occurred along the Hudson Valley. A second region where many showers developed was in the mountains of New Hampshire just east of the Connecticut River Valley. On three days showers were occurring more or less randomly in that area for several hours before the line formed, and on occasion the line itself seemed to intensify as it passed over the region. A third area of development is in cen-
central Connecticut. There is a tendency for the lower end of the line to broaden and intensify in this region and to move relatively slowly so that the form and orientation of the line becomes altered as the northern end progresses more rapidly than the southern end. It is believed that the topography of each of these regions is conducive to the development of storms especially when the surface winds are more southerly than south-westerly, but the major characteristic of the pattern, the formation of a very narrow and intense line, must be dependent upon the larger scale circulation.

**Summary**

Radar observations have shown that the rain associated with warm fronts is generally widespread but exceedingly variable in intensity. Moreover, the variations in intensity do not, at first glance, appear to form any easily recognizable patterns. However, the detailed study of storms in which the synoptic situation was very similar has shown that the small-scale rain patterns also exhibited a great deal of similarity. In particular, heavy rain appeared in areas 20-30 mi in dimension, each of which contained a number of smaller convective showers. These heavy rain areas appeared to develop in certain 'preferred' regions at intervals of 45 min to one hour. They lasted for two or three hours and moved with the wind which was nearly constant between 5000 and 18,000 ft. The individual showers, as defined by small closed contours, rarely lasted more than an hour.

In the case of squall lines and cold-front bands the storms were grouped together because of the similarity of the rainfall pattern as depicted by the radar. The synoptic situations were also very similar except that some of the lines formed well ahead of the front and were called squall lines while two of them were in the immediate vicinity of the front. The latter two moved into the radar range as bands: in the case of squall lines, scattered showers occurred in the warm air ahead of the front and then became organized into a sharp line. The dimensions of the lines were over 200 mi in length and usually less than 20 mi in width at the time of their greatest intensity and sharpness. They required a period of two to three hours for development and intensifica-
Discussion

Dr. Helmut Weickmann—What is your opinion as to the mechanisms for the development and propagation of these rain areas?

Dr. Pauline M. Austin—The fact that we find areas of recurrent development suggests topographical reasons. If we compare the wind field with the areas of development, we can draw conclusions regarding the influence of convergence. On the other hand, although a heavy rain area usually develops in a particularly preferred region, it then moves along with the wind and keeps on pouring out heavy rain for about three hours, moving right across the area represented on the scope. Clearly there is some dynamic effect continuing to produce rain, so I feel sure both factors are involved; but we have not analyzed enough different cases to provide the answer to your question.

Mr. Jerome Namias—I was very fascinated by this. Of course, there are very good reasons, I think, for getting recurrence of synoptic scale events against a planetary scale background; but the amazing part of it is that there are small places in a given situation that keep getting battered by the same type of thing, and it is difficult to attempt to explain it. I do not think it comes out of the synoptic pattern itself, and as you indicate there is a possibility that the synoptic pattern sets the stage for different orographic influences. Some work done by Charles F. Brooks in Texas about 1926 consisted in studying the distribution of thunderstorms. He indicated that those areas which had thunderstorms one day would be unlikely targets the succeeding day. He explained that it was due to the fact that the ground was wet and the heating would not be so pronounced. Here is something which is highly speculative, but I think it should be considered.

The second thing is something that I think Dr. Squires is going to talk about. There is a possibility, of course, that variations of the soil and ground-cover complex affected by rain or dryness could induce variations in nuclei kind and content and thus in the concentration of droplets, which might have some effect. But the complexity of these recurrence problems suggests that we have not only to go deeper into the synoptic physics, but that we may also have to take into account other factors as well.

Dr. R. Wechsler—The motion of a line depends on: (1) the motion of the individual cells, and (2) the rate of development of new cells which later become part of a new line. In squall line situations it is frequently observed that new cells develop some 10 miles ahead of the existing line. About half an hour later, these new cells become the dominant line while the cells of the former line are dissipated. On other occasions, the new cells develop at the forward edge of the existing line evidently by a mushrooming effect. In such cases, the question arises as to how it can be determined whether a high intensity echo occurs in a previously existing cell or in a newly developing cell.

Dr. Austin—The six lines that we have were all observed in the summer of 1958. Some cells formed ahead of the squall lines and some within them. None of the storms was intensely severe. What we did in tracking individual cells was first to overlay successive maps and select a dark
was in the break hourly the understand find think that determine the still a high the area, As formed which with study an cell. topography appeared, supposed we have definitely tracking the same cell. If a break in the graph occurred, indicating an abrupt change in velocity, or if the cell disappeared, then we ceased tracking it.

Dr. C. L. Hosler—We are instituting a similar study in central Pennsylvania, and going along with the idea that much of the influence of the topography on the development of these showers is thermal, we hope to determine the thermal structure of the ground surface by infra-red radiation.

Dr. Horace R. Byers—I think the thing that strikes some of us old timers who have worked with the classical synoptic pictures is that we have thought of the cloud system in connection with the warm front as one that is oriented essentially parallel to the warm front; yet these pictures show in all cases the line is more nearly perpendicular to this.

Dr. Austin—These were peculiar warm fronts in that respect. These did have a cold front approaching so that this cold front may be what decided the north-south orientation of that band.

Dr. Tor Bergeron—I understand this was mainly a warm-front rain, and of course, I would have liked to have seen a thorough and reliable ordinary synoptic analysis of the case, first of all. Then also I wonder if you have made an analysis of the precipitation during the 24 hours in the region from all the available rainfall stations, because as a result of all my work during almost 40 years, I have found that you can do quite a lot with that network if you really treat it in the best possible way. And you will, for instance, find preferred regions where these convective cells are released. It would have been very satisfying for me personally at any rate to see such a thorough analysis of the whole situation and I wonder if there was such a one?

Dr. Austin—The cold front approached a certain region, then it slowed down and hovered in the vicinity of Albany. You can follow its progress on 6-hourly or 12-hourly basis maps, but within the few hours we were watching the precipitation on radar, it was difficult to find the exact location of the front. We are carrying the surface analysis a little farther, as far as the total rainfall was concerned. When we consider the total for the whole day, we do find regions with heavier precipitations where these preferred tracks had been, although such regions are not sharply defined. We also found a region of heavier rain down in Connecticut. The complete analysis is still in progress. The point I want to bring out primarily was the space and time scales involved—the sizes of the rain areas, the sizes of the showers, how often they recur—so we can decide whether, on whatever scale we happen to be working, we are considering a turbulent effect or organized circulation on this scale. The complete analysis is still being carried out.
Plume Formation in Thunderstorms

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Abstract—Radar data displayed in the form of precipitation maps at constant altitude above ground (CAPPI) portray the structure of storms and their anvils in a telling manner. Well-developed storms, even in the presence of severe wind shear, were observed to remain essentially vertical through their active phase. Instead of being strongly bent by the shear, parts of the storm appeared to be carried off by the wind, forming extensive plume patterns or anvils which trailed down to levels as low as 10,000 ft while evaporating. Size and shape of the plume suggested that its particles had fall speeds ranging from 0.75 to about 5 m sec⁻¹, and thus were of precipitation size.

Introduction—This paper is concerned with the motion of thunderstorms and their erosion in strong wind shears, as portrayed primarily by CAPPI-processed radar data. As will become apparent, the interpretation of the patterns, without the CAPPI (constant-altitude plan-position indication) accessory to a powerful AN/CPS-9 radar would have been more difficult or impossible. (Radar wavelength was 3.2 cm, peak power 250 kw, pulse duration 5 microseconds, PRF (pulse repetition frequency) 200 sec⁻¹. The radar is located at Montreal Airport.) A few words will suffice to explain how these constant-level records were obtained. In our procedure, the radar was used as a plan-position indicator, but with its angle of elevation being raised after every revolution in 1-degree steps up to 12°, and in 2-degree steps thereafter to 20°. The complete cycle took 3.6 min, and resulted in 17 photographic frames from which CAPPI displays were synthesized by an optical-mechanical technique [Langleben and Gaherty, 1957]. Sets of such records for heights 5, 10, 15, ... up to 40 kft (kft = 1000 ft) above ground were built up every 3.6 min, thus providing horizontal slices at these nine heights through all the showers within range. (Improved CAPPI records are now built up directly on the radar screen; in the future it will be possible to generate these displays on an electrical or magnetic store.) Figure 1 shows in vertical section how the 5, 15, 25, and 35-kft constant-altitude displays are pieced together from successive PPI beams. It may be noted that a CAPPI picture, centered at say 15 kft, really represents a layer which extends in the mean through 15 ± 0.75 kft at range 18 mi, and through 15 ± 3 kft at range 75 mi. The ranges 18 and 75 mi were minimum and maximum for the records to be discussed.

On July 1, 1956, a series of isolated well-developed air-mass thunderstorms swept over the area. Figure 2 shows a small selection of the CAPPI record. In the evening these storms gave way to wide-spread heavy rain, ahead of a cold front which passed Montreal at about midnight EST. The first of the showers, already well developed, was detected to the NW at 14h45m and decayed while still in view at 16h30m. Five subsequent showers were picked up at earlier stages of their development, and each could be followed for more than two hours before decay. Their life histories all followed the same pattern: within about 40 min of detection the echoes reached their greatest vertical extent (in excess of 40 kft), then gradually subsided, and during
Fig. 2—A selection of CAPPI records for July 1, 1956; white numbers in left column indicate heights in thousands of feet; numbers on top, EST; range of every picture, 75 mi; several storms and their plumes are coded by black numbers.
the last 30 min only weak and diffuse echo residues were detected at heights up to 20 kft. These diffuse echoes, in all cases, could be traced back to high-level echo extensions for which we used the name 'plumes,' for their orientation and development were apparently governed (like that of a smoke plume from the stack of a ship) by the motion of the source and the ambient winds.

The intense cores of the echoes, the regions in which the convection occurred and where the precipitation presumably formed, had irregular cross sections of maximum horizontal extents 18 mi at 10 kft, 15 mi at 30 kft and narrowing to less than 6 mi at 40 kft. A striking feature was the uprightness of all the storms in the face of a very strong wind shear. Except for the plumes, the radar echoes did not seem to bend with the wind, and their sections at all heights moved with the same velocity, about 45 mph, due east (from bearings ranging from 265 to 275°). Figure 3 shows sketches made for one of the thunderstorms over a period of 90 minutes. The relative movement of the centers of the storm sections at various heights is seen to be small and random, and is probably not much greater than the margin of error of the measurements. The wind hodograph is shown in Figure 4. Also shown in this figure is the echo velocity, which agrees in direction with the winds blowing at close to 4 kft and again between 8 and 9 kft; the measured winds are, however, several miles per hour slower than the echo speed. The upper of the two mentioned levels is close to 700 mb, and is representative of a height which has been called the 'steering level' of thunderstorms echoes [Liigda, 1953].

The ability of the thunderstorm to maintain itself upright in the face of the severe wind shear is remarkable. The relative wind from 22 to 28 kft was about 30 mph, and between 29 and 40 kft ranged from 40 mph to a maximum of 110 mph at 35 kft! The only apparent effect of the wind-shear was the development of the plume. In our records, plumes started at about 35 kft, and gradually worked their way downwards, eventually dissipating somewhere below 10 kft. (Only in one case did the plume reach the 5 kft level.) The motion and growth of the plumes observed were consistent with the following mechanism: at heights at which the velocity of the wind relative to the shower is great, cloud and precipitation laden air was swept out of the storm, probably from its periphery, and carried off horizontally. Gradually, the particles in this air fell into lower layers, always adapting themselves to the prevailing wind pattern. They thus trailed down through the atmosphere, in a manner which was characteristic of their fall speeds and of the wind pattern; the general motion of the plume was equal to the motion of the region where the particles were released into the wind. Comparison of the observed plume pattern with that anticipated on the basis of the wind pattern, allowed estimates of the particle fall speed (and so to some extent of their nature), and helped in pin-pointing the height at which the plume
formed. Such a comparison is attempted in the subsequent sections.

**Derivation of plume patterns from the winds**

The trail followed by particles of given fall speed continuously released into the atmosphere can conveniently be constructed by the simple graphical method developed by Douglas, Gunn, and Marshall [1957] for the derivation of the pattern formed by snow trails from generating cells. Figure 5a shows such a trail, derived from the wind hodograph of Figure 4. Each straight-line segment represents the pattern swept out by particles falling through successive 1000-ft layers. The lengths and directions of these segments of the trail are calculated according to

$$\Delta S = (W - W_c) \Delta Z/v$$

Here $W$ is the wind representative of the layer, and $W_c$ is the wind of the point of origin of the particles (the wind at 35 kft in this example); $\Delta Z$, the thickness of the layer, was taken to be 1000 ft, and $v$ represents the fall speeds of the particles in that layer. A proportional variation in the fall speed at all heights leaves the shape of the trail unchanged, but affects the scale. For $v = 1$ ft sec$^{-1}$, the over-all length of the trail as pictured here (from its origin at 35 kft to the point where it reaches 10 kft) is in excess of 200 mi. For $v = 2$, or 4 ft sec$^{-1}$, this length would be just over 100 or 50 mi respectively. The time corresponding to each segment is 1000 sec (16.7 min) for 1-ft sec$^{-1}$ particles, and 500, and 250 sec respectively for particles falling twice and four times as fast. The trail pattern as a whole must be imagined to move eastwards at 45 mph (the velocity of its origin) so that as long as the parent storm is robbed of material at 35 kft, the trail will appear to remain in contact with the storm echo. The pattern of a trail originating at any level below 35 kft, say at 30 kft, is given by the part of the trail below the point marked 30 kft.

The length and direction of plumes at given heights (as they might appear on ideal CAPPI

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**Fig. 5**—Derivation of plume patterns from the wind hodograph; (a) the trajectory of a particle descending at $v = 1$ ft sec$^{-1}$ from 35 kft through the wind field of Figure 4; trajectories of faster particles are identical, but scaled down in size proportional to $1/v$; (b) heavy dashed line joining 35 and 30 kft represents locus of positions of particles originating in storm at 35 kft, reached after descent to 30 kft; this line suggests the plume pattern at 30 kft formed by particles of fall speed 1 ft sec$^{-1}$ and up (part of the plume pattern formed by the same particles at 25 kft is also shown); (c) dashed line here suggests plume at 25 kft formed by particles released at 30 kft.
pictures) are illustrated by dashed lines. Consider the line joining the 35 and 30 kft points in Figure 5b. Its southern extremity would be formed by particles of fall speed $v = 1$ ft sec$^{-1}$ which would have traveled some 128 mi from their origin at 35 kft, requiring 5000 sec (about 84 min) to do so. Particles of twice that fall speed would reach the 30 kft level at the point marked $v = 2$, about 64 mi from their origin at 35 kft. The northern end of this plume would be formed by particles of increasingly higher fall speeds. The times required by these particles to reach their positions in the plumes also depend on their fall speeds, and three values are indicated in the diagram. The plume appearing at a height of 25 kft, and also made up of particles originating in the storm at 35 kft is similarly represented by the dashed line joining the 35 and 25 kft points. If particles from the storm enter the wind field at 30 kft their trail would be as shown in Figure 5c (solid lines). At a height of 25 kft, such particles would give rise to a plume, oriented almost exactly north-south, as shown by the dashed line.

Comparison with observations: particle fall speeds—Some of the plume observations are illustrated in the CAPPI photographs of Figure 2, and again in the schematic sketches of Figures 6, 7, and 8. Figures 6 and 7 show vertical sections through Storms 1 and 2 and their plumes. The directions of these sections were chosen along the 340°-160° plane to coincide approximately with the longest extent of the plumes. In Figure 7 the progression of the pluming storm through the field of view of the radar is also indicated by horizontal sections through the radar patterns at a height of 5 kft. (Where these horizontal sections are dashed, they were taken through the plume at 10 kft.) Figure 8 shows the development of Plume 3 as it appeared at 25 kft.

These observations are readily interpreted in terms of the preceding analysis. The predominant direction of the plumes in Figure 8 is 335°-155°, and so agrees exactly with that of the derived pattern shown in Figure 4, for particles originating at 35 kft. That this level is the principal source of particles is also borne out by an analysis of the rate of southward elongation of the plumes prior to 17h08m. This rate can be shown to be a function only of the height interval through which the particles are falling and the wind pattern they encounter en route, and is independent of particle speed. The rate observed here, which is about one mile min$^{-1}$, fits 35 kft as an origin. (For an origin at 30 kft the rate would have been only 0.44 mile min$^{-1}$.) The apparent lengths of the plumes cannot be determined with great accuracy, since the line of demarcation of storm core and plume cannot easily be drawn. But at 16h57m, the plume at 25 kft is about 32 mi long, which by direct comparison with the plume derived in Figure 5b (length 160 mi for $v = 1$ ft sec$^{-1}$) would mean that the particles in the southern tip are falling at about 5 ft sec$^{-1}$. At 17h08m, the plume is about 37 mi long and this indicates that particles falling as slowly as 4.3 ft sec$^{-1}$ have now reached the 25 kft level. By 17h20m the plume lengthened to its maximum dimension of 45 mi; minimum fall speeds are therefore 3.6 ft sec$^{-1}$. One can check the times of departure of these particles from the 35-kft level, and find that they all agree reasonably well, being between 16h15m and 16h30m. The first sign of a plume at that level appeared on the radar records at about 16h27m.

The radar record of 17h47m is the last one showing an intense core at 35 kft. This probably was therefore the last instant at which particles entered the free atmosphere at that level. From this time on, the particles already in the trail continued to fall, but the northern end of the
plumes, which at any level was formed by the fastest particles, fell faster than the rest. For lack of replenishment, the northern end of the plume at any level should therefore have disappeared faster than the rest—in agreement with observation. It is tempting to estimate the fastest particle fall speeds associated with the northern tip of the contracting plume. For the times 18h00m, 18h14m, and 18h24m such estimates are 23, 13, and 11 ft sec$^{-1}$ respectively, based on the plume contraction at these times. These estimates, though plausible, are not unique. Slower particles originating at levels lower than 35 kft could have accounted for the same observed pattern. Thus if the northern end of the plume consisted of particles falling at up to 10 ft sec$^{-1}$ which had been released at about 30 kft, the retraction of the plume at 25 kft would also proceed at the observed rate of about 0.5 mile min$^{-1}$. The supposition that these lower levels also contributed is supported to some extent by the change in direction of the plume, for a gradual clockwise rotation of the plumes was observed. All the plumes, those of Storms 2 and 4 even more notably than Plume 3, had their longest dimension in a N-S direction in their later phases. A glance at Figure 5e shows that such rotation is best accounted for by supposing that while plumes formed initially at 35 kft, in their later stages, more and more of the material in them derived from lower levels of the shower. There is no evidence in our records, however, of particles stemming from levels lower than 30 kft. Comparisons of the kind outlined were made for three of the plumes at several levels; the conclusions were essentially the same, and will now be summarized.

The storms were tall precipitation clouds, which in the face of very strong relative winds...
above 29 kft showed little tendency to bend. Instead the winds succeeded in eroding these storms by sweeping material away from them. This erosion was most pronounced at 35 kft, the level of the strongest relative wind, but occurred also to a slight extent above that level, and gradually spread down towards 30 kft. The evidence is clear that the plumes detected by the radar consisted of particles whose fall speeds were in excess of 3.6 ft sec\(^{-1}\), and are thus probably precipitation, rather than cloud, particles. The 35-kft temperature was \(-50^\circ\text{C}\). At 30 kft, it was \(-37^\circ\text{C}\), and it therefore is unlikely that any liquid material was involved. Fall speeds of precipitation particles vary with height of course. Douglas, Gunn, and Marshall [1957] used the relation

\[
u \propto \eta^{-0.5} \rho^{-0.4}
\]

where \(\eta\) is the viscosity and \(\rho\) the density of the surrounding air. On this basis the factors of Table 1 were derived. Using an average factor of 1.55 for the trajectory of the particles explicitly considered above would indicate fall speeds (corrected to 1000 mb and \(0^\circ\text{C}\) ranging from 3.6/1.55 = 2.3 to 23/1.55 = 14.9 ft sec\(^{-1}\), or in metric units from 0.75 to 4.9 m sec\(^{-1}\). Using Langleben's [1954] measurements of the fall speeds of snow crystals, the lower speeds could be associated with dendrites of melted diameter about 0.8 mm; the fast speeds (around 4 m sec\(^{-1}\)) were too high for aggregates, and probably indicate frozen rain or small hail. Using Weickman's [1955] compilation for hail particles of specific gravity 0.8, the melted diameter would be about 5 mm; for graupel particles of specific gravity 0.2, the melted diameter might be as great as 1.2 cm. It may be pointed out that scattered hail was reported at the ground, which (from its timing and location) was shown to come from the showers under study. (Hail at the ground is indicated by the arrows on the 5-kft pictures of Figure 2.)

Newton [1960] has recently discussed the pressure field surrounding a storm moving in a wind field with vertical shear. Considering the storm as a rigid structure moving with the winds appropriate to its middle levels in a typical case, he finds ahead of the storm an excess pressure near its base, and a pressure deficit near its top. On this basis, Newton concludes that enhanced lifting, and so formation of new cloud, should take place at the leading edge of the storm. Conversely, at the rear of the storm the pressure field is modified to lead to downdrafts and consequent cloud decay. These conclusions are not in agreement with the more common notion that the strongest activity is near the trailing edge of the storm. But cloud development near the leading edge and remaining separate from the rest of the storm could account for the plumes observed by us. For when this cloud reaches the layer of high winds, it would presumably be carried away from the storm in exactly the same way as the plumes described. On this model, the plume would not be the result of storm erosion, though erosion may well play an important part in maintaining the storm upright in severe wind shear. Such cloud moreover would surely compete severely with the main storm for the moisture supply which the storm draws in to an appreciable degree from its leading edge.

Yet another interpretation of plumes was given by Imai [1957] who made observations by PPI as well as by RHI (vertical) radar sections. His beautiful records of bright-band forming in the falling plume allowed him to identify the plume particles as ice, originating in the storm. On the basis of his somewhat sparser records he concluded however that the radar plume comes into being only after the decay of the convection, and that it consisted entirely of very small crystals. The only fall speed quoted is 40 cm sec\(^{-1}\).

It is noteworthy that our observations and analysis do not require the existence of cloud particles in the plume. Conceivably cloud is also swept out by the wind, but remains undetected.
by the radar. At the other extreme, particles falling faster than those mentioned may also exist in the shower at high level, and be moved by the wind, but would remain unidentified by the radar, since such particles would fall very close to, or within, the intense core of the storm itself. This point is among those borne out by Figure 9 which is a theoretical reconstruction of a plume in vertical section, the way the eye might see it. The lines sloping down to the right are the trajectories of particles of various fall speeds, all originating at 35 kft. The fall speeds are in meters per second, and are all corrected to standard temperature and pressure. To the extent that the plume has texture, e.g. unevennesses or characteristic knobs or holes, it would give the appearance of developing relative to the mother storm, in the direction of these trajectories. The dashed lines are isochrones, and indicate size and extent of the plume after 15, 30, 60, and 90 min from the incidence of pluming. The earlier stages are in excellent agreement with observation. The 90-min line would indicate a rather broader base to the plume than we observed, but it is at such later times, and in the attitudes between 25 and 15 kft, that evaporation effects, neglected here, are especially important. Especially noteworthy is the separation of the particles in the plume according to their fall speeds. For the case of snow-generating cells and the 'continuous' precipitation resulting from them, such sorting of the particles was studied in detail by Gunn and Marshall [1955].

Acknowledgments—It is a pleasure to acknowledge the interest in this work of J. S. Marshall, who as co-author of a method of analyzing snow trails, suggested that a similar approach might work for plumes. J. L. Galloway contributed a carefully documented analysis of the synoptic weather pattern of the day referred to. Though his painstaking work has not been quoted explicitly, it is he who has provided confidence in the validity of the hodograph.

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References


Discussion

Dr. H. Dessens—Was this storm with or without hail?

Dr. W. H. Witschfeld—There was hail in the plume, but very little. Hail was observed on the ground, but it was small and insignificant. We do not get much hail in Montreal.

Dr. E. Kessler—Do you think that the evaporation of parts of the plume, and therefore the cooling of the top of the layer immediately below and its destabilization might contribute to the observed structure in the base of the plume?

Dr. Witschfeld—Yes, indeed. In this connection I might mention again the work of Imai in Japan, who had RHI pictures showing some mammatus at anvil base.

Mr. C. E. Anderson—Your paper bears out some ideas that I have. I feel that by watching how a cloud dies, one can learn a lot about the dynamics of the cloud, because the manner in which it dies gives a clue as to how it has lived. You showed that you had a continuous production of particles at 35,000 ft which sent down streamers with the wind for an hour. This we have noted also in anvils in the southwest, and we could only interpret this to mean there was some type of continuous updraft within the cloud producing the plume. Sometimes we noted new cloud development in the plume away from the core of the main cloud as it moved down stream a bit.

Dr. Witschfeld—At what height are your plumes?

Mr. Anderson—These would vary, sometimes 25,000 ft, but usually higher.

Dr. Witschfeld—This, I think, is vital because as I am showing elsewhere (McGill University Report MW, July 29, 1959) if the plume is warm enough to be supercooled water, there is no reason why there should not be new generation of precipitation in it. But in the plumes I was emphasizing here, I know this was not the case because the temperatures were much, much too low.

Dr. H. Weickmann—Do you have any indications of differences of lifetimes between storms with strong wind shear and storms with little wind shear?

Dr. Witschfeld—No.

Dr. Weickmann—I am asking because by carrying away the anvil the high-level jet may alter the life history of the storm as determined by cloud physical processes. From the Byers-Braham thunderstorm model we know that the developing state is followed by the mature state and this by the dissipating state. The mature state is characterized by updrafts prevailing throughout the cloud, whereas the dissipating state is characterized by downdrafts throughout the cloud and by an increasing predominance of the ice phase in the cloud. Byers and Braham have pointed out a mechanism by which downdrafts form in the upper parts of the cloud. These downdrafts will counteract the buoyancy of the rising cloud air and may thus mark the beginning of the dissipating state. If these downdrafts are displaced away from the original cloud by the action of a strong wind shear, they may not initiate the dissipating state and the cloud may stay alive in the mature state.

Dr. Witschfeld—In spite of the fact that the system is being deprived of water all the time?

Dr. Weickmann—Yes, because it is only deprived of water which has spent already its energy for the cloud, whereas new moisture can continuously be fed in at the base, depending on the depth of the moist layer, inflow velocity, and migration velocity of the storm.

Mr. Jerome Namias—I would like to ask a question of you or the audience which has puzzled some of us for 20 years. Over the southern plains of the United States in the summer we have a condition in which an upper-level anti-cyclone develops and moisture flows around it in the form of great moist tongues. In the Great Lakes area rain falls from the warm front, as this moisture is forced over colder air, and much thunderstorm activity takes place. In many of these cases there is a very strong vertical wind shear so that the southwest surface flow from the Bermuda high is overrun by northerly components. As a result, these thunderstorms move southward in clusters. In earlier studies I showed (Bull. Amer. Met. Soc., 19, 1—14, 1938) that the moisture pumped up by these thunderstorms is carried by the wind and in some way sets off chains of thunderstorms that move in the fashion indicated. In the 1930’s we tried to explain this on the basis of radiational cooling from the cloud tops or perhaps lack of entrainment. Have you some new explanation to suggest; namely, a
physical mechanism explaining how thunderstorms form and move in the manner I described.

Dr. Hitschfeld—in the other study to which I have already referred, Report MW29, we attempt to re-interpret Dennis’ (J. Met. 11, 157–162, 1954) observations of seeding trails as plumes. Such plumes can travel considerable distances and thus can disseminate nuclei widely. It is possible that this might account for Mr. Namias observation that a favored direction of storm propagation is that of the high-level winds.
The Structure of Minute Precipitation

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Abstract—Continuous measurements with a raindrop recorder over a period of four years were used to study the unknown range of diminutive and minute amounts of precipitation. Nearly half of individual rainfalls belong to this class. Frequency distributions of amount, droplet size, and intensity in logarithmic scales are presented.

Introduction—The range of minute precipitation below 0.2 mm amount is still nearly unknown. With regard to climatological or hydrological applications this range is without interest. But in biological respects and above all in precipitation physics this range has a practical and theoretical importance. The limiting value of 0.2 mm is used because such an amount of precipitation is necessary to wet the rain gage before the first drop is recovered out of the sampling pot. The measurement of minute precipitation is only possible by continuous raindrop recordings as also suggested by Bowen and Davidson [1951] with a raindrop spectograph and, in intervals of some minutes, by Blanchard [1953, 1957] and Lamp [1958]. The Hohenpeissenberg device (Fig. 1) is of a similar principle: the raindrops fall through a $2 \times 10 = 20 \text{ cm}^2$ opening on a prepared paper band. An electric synchron-motor transports the band with a velocity of 2, 5, or 10 cm/min. After exposure the paper is dried by a resistance-wire heater below the table which guides the paper band. The role of paper has a length of 100 m, ample for a continuous record of 3½ days. Spot records of raindrops impart an instructive insight into the structure of rain, both droplet size, intensity, duration, and amount. It is laborious to measure many thousands of spot-diameters but there is no other way to this day to record all these characteristics simultaneously. For the range of minute precipitation it seems to be the only method. The evaluations related to this range cannot be measured with the normal methods of precipitation-measurement.

Frequency distribution of precipitation amounts—Comparing the records of a normal rain recorder with the raindrop recorder more than 40% of each single rainfall was not measured by the usual method. In addition to the frequency distribution of the amount of precipitation derived by the records of a Hellmann recorder, the spectrum is extended in the range of diminutive and minute amounts the maximum of which is reduced to an amount of 0.03 to 0.04 mm (Fig. 2). The dotted lines take the wetting and evaporating effect within the rain gage into account. The presentation of this frequency distribution is only possible by using a logarithmic scale of the frequency ranges as suggested for meteorological, particularly precipitation, analyses by Guss [1955], Schneider-Carius [1957] and Essenevanger [1960].

The frequency distribution is made up by part collectives the exact analysis of which would be possible with the procedure of Essenevanger [1954, 1957]. The part collective of Class 7 (1.6-2.5 mm) for instance is found on the thunderstorm showers in May and June. The instability in August favors the part collective of the range of 0.3-0.6 mm. The assumption that the minute precipitation amounts are originated only by drizzle or fine rain is not right. The most evident part collective of 0.03 mm results from instability showers in June, July, and August, while there are scarcely months of transition in this class. There are drop-showers coming out of convection cloudiness, but also the fringe areas of more heavy showers at nearby localities.

Duration—The duration of precipitation (Fig. 3) ascertained with a normal precipitation recorder is found with the frequency maximum at 50 to 200 min, those ascertained with raindrop recorder at 3 to 30 min. Therefore the frequency maximum of duration for all single rainfalls shifts to the range of 40-76 min. The relatively high quota of minute precipitation with a longer duration originates in certain weather situations: on the edge of a high pressure area, where air masses of different characteristics are adjoining, stabilizing weather in the rear of a depression and drizzle within moist-warm air together with a slight upslope circulation along the Alps.
THE STRUCTURE OF MINUTE PRECIPITATION

Fig. 1—Drop recorder for continuous sampling of raindrops; opening for exposure $2 \times 10$ cm

Fig. 2—Percentual frequency distribution of precipitation-amounts; the hatched range is verified by raindrop records

Fig. 3—Relative frequency of duration of precipitation; (a) normal precipitation recorder; (b) raindrop recorder; (c) all precipitation amounts; and (d) part of all amounts derived from raindrop records
Fig. 4—Drop sizes of minute precipitation derived from raindrop records

Fig. 5—Rainfall intensity, derived from normal precipitation records
Fig. 6—Rainfall intensity of minute precipitation, derived from raindrop records

Fig. 7—Relationship between duration (ordinate) and intensity (abscissa) of precipitation, demonstrated by isolines of equal frequency; all amounts of precipitation, derived from normal precipitation records
Drop size—The drop sizes of minute precipitation (Fig. 4) likewise demonstrate some part collectives in the curve of frequency-distribution: diameters below 0.5 mm, originating from drizzle; a more marked maximum between 0.6 and 0.9 mm originating from lability showers and a wide range from 1.0 to 2.0 mm diameter as effects of heavy showers near the station. The effect of wind shear, however, which spreads the various drops in the spectrum not only in the direction of motion of the source but also at right angles to it [Blanchard, 1957], will bring more drops of smaller size-ranges in this case. The dominating range of 0.6–0.9 mm, evident in the summer months, therefore seems to shift to smaller ranges than is the case for precipitation of higher amounts.

Intensity—The ranges of intensity shift in the same manner from higher values in case of the total collective (Fig. 5) to smaller values in case of minute precipitation (Fig. 6). The differences between the ranges of diminutive intensity and of higher ones are more conspicuous than between other factors of the structure. The minute precipitation amounts in the summer months, however, demonstrate a secondary maximum in the range of 0.0010 to 0.0030 mm/min, originating from less abundant shower precipitation.

The dependence of intensity on duration is shown on Figures 7 and 8 which illustrate clearly the ranges covered by the cases of minute precipitation. They shift to smaller values of both, duration and intensity, the angle of inclination of the mean line of this relation decreases from 42° to 35°. These ranges seem to be of interest in respect to artificial stimulation of rain. Therefore further studies of selected cases ought to clear the generating conditions of this minute precipitation.

Conclusions—At the beginning of the paper attention was called to the laborious work in evaluating the raindrop recordings. To accomplish that, the Observatory Hohenpeissenberg has designed two instruments for recording the structure of rain. The one records the duration of precipitation of any amount in minutes based on the dual principle: precipitation or no precipitation; the other records the intensity based on
the method of counting the drop flowing out of a sampling funnel each minute. I hope these means will bring further insight in the structure of precipitation, including that of more abundant precipitation such as showers and steady rain.

References


Essenwanger, O., Frequency distributions precipitation, this volume, pp. 271-279, 1960.

Discussion

(Note: Discussion of this paper is combined with that following the next paper.)
The Productiveness of Fog Precipitation in Relation to the Cloud Droplet Spectrum

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Abstract—On Mount Hohenpeissenberg in Upper Bavaria (975 m NN) the atmospheric offer of fog precipitation is measured by a cylindrical net of wires of 0.1 mm diameter, for which a nearly constant relation of deposit amount to wind velocity is found. The amount of deposited fog precipitation depends on (1) locality and exposition of the gage, and (2) weather situation. The efficiency of polar cold air, characterized by dominating small diameters of cloud droplets from 2 to 15 μ, is scanty. Increasing productiveness results, when maritime warm air masses from temperate or subtropical zones pass. The cloud droplet spectrum is then characterized by a broader range of 4 to 25 μ diameter, with a maximum frequency from 8 to 14 μ. The deposits are heaviest with amounts of 2 to 3 mm/hr when persistent cloud decks form on the windward side of the Alps. The air masses have then often degenerated by continental influence. The droplet spectrum indicates a wide range from 5 to 60 μ with a maximal frequent diameter of 12 to 18 μ.

Introduction—Cloud air streaming against an obstacle precipitates a part of cloud droplets. The deposit on trees is known as fog drip or in freezing weather is visible as rime. In flat land the fog deposits are just sufficient for wetting needles and leafage. But in mountain regions, which rise at times into the cloud space, and in coastal mist belts, where moist-warm air passes from sea to land, considerable amounts of additional water by fog drip can be expected.

Measuring method—The atmospheric potential of fog precipitation can be comparably measured with a specific fog gage as suggested by Tabata and coworkers [1953] for the research of sea fog on the coast of Hokkaido, Japan, and by the author [Grunow, 1952] for studying the productiveness of fog precipitation in mountain forests. Both types of fog collectors use the same principle, a system of wires. The Japanese pattern is that of a cylindrical wire screen; the German pattern that of a cylindrical wire net (Fig. 1). The effect of these gages is found in the theory of the dust filter derived by Albrecht. The amount \( M \) of deposit on cylinder, in case of the fog gages on each single wire, is given by

\[
M = FvWt/C
\]

It depends on the diameter and length of the wires through the factor \( F \), velocity of wind \( v \), water content of fog \( W \), time of deposit \( t \), and an efficiency factor \( C \) which is influenced by the diameter of the wire, the diameter of droplets, and the wind velocity. The most favorable re-

Fig. 1—View of the Hohenpeissenberg fog-collector; the cylindrical wire-net is mounted on a normal rain gage 10 cm in diameter
The productivity of fog precipitation is influenced by various factors. The amount of water deposit $M'$ on the wind-velocity $v$, for various wire diameters at a cloud-droplet diameter of 20 $\mu$ is shown in Figure 2.

Factors influencing the amount of deposit—The amount of deposited fog precipitation, aside from the efficiency of the wire system, depends on (1) the locality and exposure of the gage, and (2) the weather situation. The locality factor is specified by the height above sea-level, the continentality (distance from sea), and the exposure to the fog-producing air currents, especially on windward slopes. These effects are demonstrated for several mountains of Germany in Figure 3 [Grunow, 1958]. The measurements on Table Mountain (Tlb) South Africa, very critically evaluated by Nagel [1956], showed rainfall of 1940 mm and additional fog precipitation amount of nearly 3300 mm in 1954. The fog precipitation was 170% of the rainfall.

Increasing amounts of fog precipitation are dependent on the weather situation. It is not only the direction and velocity of the depositing air current, but also the origin of the operating air mass, which influences the production of fog precipitation. The heaviest deposits occur where maritime warm air-masses from temperate or
subtropical zones are in action, if a zonal circulation is predominant. The deposits are moderate or scanty, if the air masses originate in polar or arctic zones. This variable affects the value of \( W \), the water content, of our formula.

Record of cloud-droplet spectrum—A relationship between the amount of deposit and the cloud-droplet spectrum was established using the record of the droplet spectrum of cloud air for different air masses. According to the method of Dien [1947] cloud droplets were collected between two oil layers of different viscosity, the
upper being a thin film of paraffin oil placed above a base of heavier castor oil. The finding was immediately recorded by microphotography. The samples were taken under various fog situations, without any consideration of the deposited amount. Altogether 25 tests were evaluated, each consisted of an average of 12 samples with each several hundred or thousand droplets.

Types of droplet spectrum—The derived frequency distributions of droplet diameters can be classified according to the types of droplet spectrum, as seen in Figure 4. Each of these types presents obvious relations to the existing air mass and through it to the productiveness of fog precipitation. Polar cold air is characterized by dominating small diameters (Types a and b), in case of maritime origin, Type a with a range up to 10 to 12 μ and a remarkable part of diameters smaller than 2 μ; in case of continental origin Type b from 2 to 15 μ with a maximum number.

Fig. 5—Microphotographic record of cloud droplets from polar cold air; small amount of fog precipitation
Fig. 6—Microphotographic record of cloud droplets from degenerated maritime subtropic air mass; heavy amount of fog precipitation
at 8 to 9 \( \mu \). The efficiency of fog precipitation is scanty. Types c and d are maritime warm air masses from moderate (Type c) or subtropical (Type d) zones. The droplet spectrum then indicates a broader range of 4 to 25 \( \mu \) and a most frequent diameter of 8 to 14 \( \mu \). Fog precipitation becomes more and more productive. The deposits are heaviest with amounts of 2 to 3 mm/h in case of non-raining cloud decks formed on the windward side of the Alps. The air masses are often degenerated by continental influences and the droplet spectrum indicates a wide range from 5 to 60 \( \mu \) with a maximum number at 12 to 18 \( \mu \). This result conforms with investigations of Mahrous [1954], who established the increase of droplet sizes as typical evidence of degeneration in the dense coastal fog in England. In the case of a narrow spectrum with smaller diameters of droplets a meridional circulation exists whereas the broad spectrum predominantly shows up with zonal circulation. Selected cases of these two types are demonstrated, with the microphotographic records of droplets in Figures 5 and 6.

Water content of samples—The different productiveness of these types of fog precipitation classified according to the droplet spectrum is easy to understand if the water content of each sample is calculated from the product \( n \pi /6 \cdot D^3 \). With increasing droplet diameter \( D \) the water content grows rapidly even if the concentration \( n \) of these droplets is small. Integrated over the whole spectrum, the results for the samples are:

<table>
<thead>
<tr>
<th>Type</th>
<th>a</th>
<th>b</th>
<th>c</th>
<th>d</th>
<th>e</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total water content*</td>
<td>3 23 76 105 473</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Unit is \( 10^{-6} \) mm\(^3\).

Even this rough estimate indicates that the productiveness of fog deposit in case of a broad spectrum is a direct consequence of the microphysical structure of the clouds.

References


Discussion

(Releasing to the two immediately preceding papers)

Dr. C. E. Junge—Figure 3 of the paper just presented showed the increase of precipitation. This was the total increase plotted against area where there is no fog precipitation, is that right? You had figures of 200 to 300%.

Dr. J. Grunow—The figures represent only the additional fog precipitation amount deposited on the wire net of the gage at different mountain stations, without any consideration of rain or snow.

Dr. Junge—E. Eriksson, for instance, found that in Scandinavia the total amount of sea salt which is deposited from the atmosphere is approximately three times higher than can be accounted for by precipitation. The mountains in this area are often within clouds and the increase of precipitation by fog drip may be considerable and may explain the increase in sea salt deposit.

Dr. Grunow—At some places in higher altitudes, for instance, on the mountain station Wasserkuppe we have found the same result. The most interesting point where measurements were made with this type of gage, is Table Mountain in South Africa. In January 1955 the additional fog precipitation was tenfold the rain precipitation. In the annual average the unit with fog gage received four to fivefold the catch of the rain gage. According to the very critical evaluations of Nagel there was an assured excess of 1.7 fold of rain, derived only from days with fog without drizzle or rain. The best conditions for deposit are given if fresh maritime air masses are in action.

Dr. Junge—One more question: These 200 to 300% resulted from measurements with the
wire screen. Did you make any estimates of how much the precipitation is increased by fog drip in a normal forest in a location similar to yours?

Dr. Grunow—In order to test the effective deposit on natural obstacles we have made measurements with a normal rain gage and with tubs under trees which acted as a natural fog meter, and in an open area with the fog gage, all at the same time. In case of fog without rain we found a reduction factor of 0.5 to 0.8 within the forests and of 3.2 on the edge of a forest. The reduction factors thus found depend on kind, form, and size of the trees and of the density of the woods. Therefore the reduction factor is only correct for the place at which these measurements are made.

Dr. H. Weickmann—You had pine trees?

Dr. Grunow—Yes.

Dr. Weickmann—They would have a higher collection efficiency than trees with leaves.

Dr. Grunow—The pine trees stand before our door. It is true, conifers have a higher efficiency than broad leaves, but more important is the situation of a forest in relation to the fog-bearing current of air. The highest amounts will be caught at the edge of the woods, but also within the forests a deposit of fog takes place. I remember, too, the work of Hori in Japan. He has also found considerable amounts within a forest.

Dr. W. E. Howell—With reference to your first paper, I recall that some time ago, C. K. Stidd (Cube-root-normal precipitation distributions, Trans. Amer. Geophys. Union, 34, 31–35, 1953), when he was considering a possible representation of frequency distribution of rainfall amounts found zero, limited at dry stations for weekly or monthly rainfall amounts. Apparently there is a straight line relationship of the cube root of the fall amount broken down for very small amounts, and I presumed this was due to the fact that amounts less than one millimeter or at some stations less than one hundredth of an inch were not properly recorded. It occurs to me to wonder whether the measurement that you have made of the minute rainfall amounts would verify Stidd’s presumptions that the cube root relationship holds down to very close to the zero line. I think it might be of interest of investigation.

Dr. Grunow—Thank you for this reference. In our further investigations we will investigate this relationship. I think this range of minute precipitation, although the monthly amount is not modified by this with regard to climatological and hydrologic applications, is worthy of consideration with respect both for theoretical and practical purposes, especially for agricultural meteorology and for artificial stimulation of rain.

Dr. B. J. Mason—When you are considering the collection efficiency of the fog catcher, do you regard this in terms of the collection efficiency of the individual wires?

Dr. Grunow—Yes.

Dr. Mason—When much fog is coming past the cylinder, do the pores, the spaces between the wires, become filled with water?

Dr. Grunow—In winter with riming, yes, but in summer we have found that is not of significance. This can be deduced from the nearly linear relationship between the amount deposited and the velocity of the wind.

Dr. Mason—Nagel in South Africa got rather unreasonable results because when he deduced the water content of the clouds from the collection efficiency of the single wire and the velocity of the cloud air which had gone by, he was out by a factor of 5 to 10, I think. This may have been because the spaces in between the wires were being filled. Then it is no longer possible to regard the cylinder as a number of individual wires. I would not say you should go so far as to regard it as a solid cylinder, but it may have approached that.

Dr. Grunow—Nagel assumed a water content of approximately 1 g/m², and this is a very high amount. The efficiency of a wire net cylinder is higher than that of a solid cylinder, because in the latter case, with increasing wind velocity, the current lines are more conducted around the profile. But the effect is less than the theoretical factor because some spaces will be closed by water. But, more important, the efficiency factor is nearly independent of wind velocity. We can observe the behavior of the cylinder in each weather situation because we have the instruments before our door, and we found that at low wind velocity there are no more spaces between the wires filled with water than at high velocity. However, measurements with our fog gage are proposed for determination not of the water content of clouds but of the additional fog precipitation, deposited in the same manner as by natural hindrances, and for this purpose we can use any efficiency factor.

Mr. Aldaz—At Mt. Washington Observatory we do some collection for the Atomic Energy Commission. We use a frame with no more than
twenty bars, very solid and rigid, with diameters of the order of three millimeters, and we collected enormous amounts of water which reach the order of a quart, on many occasions, in about two hours. The frame is about two feet by one foot. The fog on Mt. Washington is sometimes very thick.

Dr. Weickmann—What we have just heard may have important applications to anyone who is interested in or in charge of conserving the water in regions where the mountains often reach into the base level of clouds. No trees should be cut from the tops of these mountains, because these trees are important collectors of otherwise unprecipitated water.

Dr. Bergeron—I wonder if anybody of those present has visited that wonderful place on a small mountain in Portugal named Cintra*. It has a castle on top of it. On the top, which is small and isolated, one is in a tropical rain forest. One looks down on the desert toward Lisbon. Fog drip must be the explanation for its rich vegetation.

* Serra da Cintra, rugged mountain mass, north of Lisbon. Highest peak Cruz Alta (1772 ft); castle is Palacio da Pena.—Ed.
Horizontal Distribution of Snow Crystals during the Snowfall

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Abstract—Simultaneous observations of the shape of snow crystals were carried out at 14 points in an area of about 5200 km² in the Ishikari Plain, Hokkaido, in the period from January 20 to February 28, 1959. The horizontal distribution of snow crystals and its time sequence on four days; namely, January 20, January 30, February 10, and February 16 are analyzed in detail. From these results, it was confirmed that a dependence of shape of natural snow crystals on temperature in the upper atmosphere was in reasonable agreement with that of artificial snow crystals. On the other hand, it was found that the area where the same shape of snow crystals was observed might have some relation to the isothermal lines in the upper atmosphere.

Introduction—In the recent researches on natural snow crystals, the relationship between the shape of snow crystals and meteorological conditions was studied by many workers. It was confirmed by Gold and Power [1954], Murai [1956], and Kuetten and Boucher [1958] that a dependence of shape of snow crystals on temperature in the upper atmosphere was in reasonable agreement with that found in the laboratory by Nakaya [1954, p. 249]. On the other hand, Weichmann [1957, pp. 239-241] observed the time sequence of snow-crystal forms during continuous precipitation, and suggested that the snow crystals could be used as an aerological sonde. However, there have been no studies on the horizontal distribution of snow crystals during a snowfall. The influence of the area where the same shape snow crystals are observed will be one of the important problems in the research on precipitation systems. From this standpoint, simultaneous observations of snow crystals were carried out at 14 points in an area of about 5200 km². The results from these observations will be described in this paper.

Observational procedures—Simultaneous observations of the shape of snow crystals were carried out at 14 points in an area of about 5200 km² in the Ishikari Plain, Hokkaido, in January and February, 1959. A map of Hokkaido is shown in Figure 1, in which the observation area is indicated by a square surrounding Sapporo; the location of Hokkaido is shown in the upper left corner. The topographical map of the observation area is shown in Figure 2, where the observation points or stations are numbered 1, 2, etc. Station 2 is in Sapporo, where aerological data are obtained using Rawinsonde by the Sapporo Meteorological Observatory at 09h 00m and 21h 00m JST (Japan Standard Time) every day.

The observation of snow crystals was requested of the teachers of senior or junior high schools at stations noted above. The observations at Station 2 were made by the Sapporo Meteorological Observatory, and those at Station 13 by the Iwami-zawa Weather Station. Since the observers had no experience in the observation of snow crystals, such methods as the use of microscope or replica or shadow photograph could not be employed. Therefore, observations were made with the naked eye or a magnifying glass.

The shape of snow crystals was recorded by graphic symbols of the practical classification of Nakaya [1951, p. 311], in which the snow crystals were classified into the following eight classes: (1) plates ○, (2) stellar crystals ×, (3) columns □, (4) needles ←, (5) spatial dendrites ⊗, (6) capped columns +, (7) irregular crystals Δ, and (8) graupel ±.

The period of observation was 40 days from January 20 to February 28, 1959. The time of observations was at 09h 30m, 10h 30m, 11h 30m, and 12h 30m, JST on each day that snowfall occurred. Since the aerological observation by Rawinsonde at Sapporo is carried out at 09h 00m as noted above, such times of observation are reasonable for studying the relation between the shape of snow crystals and meteorological conditions.

Besides the shape of snow crystals, it was requested that observers record the qualitative intensity of the snowfall and the occurrence of snow flakes.
Results of observation—In the 40 days of the period of observation, there were six days when snowfall was not observed at all observation points. On all other days, snowfall was observed at one or more of the stations. It can be said, therefore, that snowfall occurred in the observation area almost every day in the period of observation. Snowfall was observed at all stations on 4 days; namely, January 20 and 30, and February 10 and 16. The horizontal distribution of snow crystals and its time sequence observed in these four days will be described in the following.

On January 20, the snowfall was associated with the continental monsoon system. Sea-level and 700-mb charts at 09h 00m are shown in Figure 3. As seen in Fig. 3a, the snowfall area was the northern part of the Japan Sea Coast of Japan. The amount of precipitation on this day was 1.7 mm at Sapporo and 3.4 mm at Iwanizawa.

The horizontal distribution of snow crystals at 09h 30m, 10h 30m, 11h 30m, and 12h 30m is shown in Figure 4. In these figures, the observed shapes of snow crystals are indicated by the graphic symbols of the practical classification. The graphic symbol shown at the observation point indicates the shape of snow crystal observed most frequently; the graphic symbols in parentheses indicate those observed less frequently. In these figures, no graphic symbol is shown at some observation points. This does not mean, however, that snowfall did not occur at that place, but that the observation of snow crystals was not carried out. The open circle indicates that snowfall had stopped at the time of observation. As seen in these figures, the stellar crystals and spatial dendrites were observed widely and continuously during the period of observation. Plates were observed in the southeast region indicated by P. At 09h 30m, needles were observed in the east region indicated by N.

The sounding curve at 09h 00m is plotted in Figure 5. In the left part of this figure, the full line indicates the relative humidity with respect to water, and the dotted line indicates that with respect to ice. In the right part, the broken lines indicate the wet adiabatic lines. The wind aloft at 0.5 km intervals is shown at the left, where the long wing indicates 10 m/sec and the short one indicates 5 m/sec. These notations are also used in the figures that follow. From the
Fig. 3—Sea level and 700-mb charts, 09h 00m, January 20, 1959

Fig. 4—Horizontal distribution of snow crystals from 09h 30m to 12h 30m, January 20, 1959
Fig. 5—Sounding curve at Sapporo, 09h 00m, January 20, 1959

Fig. 6—T-s diagram for the sounding at Sapporo, 09h 00m, January 20, 1959; Region I, dendritic; II, sector and plate; III, needle; IV, scroll or cup; V, irregular needle; VI, spatial plates; and VII, column

As seen in Figure 4, it is noteworthy that the southeast region where plate crystals were observed did not change position during the period from 10h 30m to 12h 30m. This suggests that the meteorological condition suitable for the formation of snow crystals continues for several hours during a snowfall.

On January 30, the snowfall was due to the upglide motion associated with the warm front. Sea-level, 850-mb, and 700-mb charts at 09h 00m are shown in Figure 7. As seen in the sea-level chart, the rain area reached the southern end of

sounding data, the relation between the air temperature and the supersaturation with respect to ice was plotted on the T-s diagram [Nakaya, 1954, p. 249] from the study of artificial snow crystals, as shown in Figure 6. The broken line in this figure indicates a line giving the saturated vapor pressure with respect to supercooled water. On the basis of the experimental results with artificial snow crystals, it will be expected from Figure 6 that column (capped column), plate, dendritic, and scroll crystals would occur in such atmosphere aloft as shown in Figure 5. This expectation agrees quite well with the results of the actual observations as shown in Figure 4. The plates observed in the southeast region are considered to form in the warmer layer of Region II, since the plates formed in the cold layer of Region II would develop into dendritic crystals in the layer of Region I. On the other hand, the needles observed at 09h 30m in the east region cannot be explained from the sounding data, and so it will be local snowfall.

As seen in Figure 4, it is noteworthy that the southeast region where plate crystals were observed did not change position during the period from 10h 30m to 12h 30m. This suggests that the meteorological condition suitable for the formation of snow crystals continues for several hours during a snowfall.

On January 30, the snowfall was due to the upglide motion associated with the warm front. Sea-level, 850-mb, and 700-mb charts at 09h 00m are shown in Figure 7. As seen in the sea-level chart, the rain area reached the southern end of

Hokkaido, and snowfall occurred in the area northward from there. The amount of precipitation on this day was 28.5 mm at Sapporo, and 1.5 mm at Iwamizawa. The difference in the amount of precipitation between these two places can be explained as follows.

The horizontal distribution of snow crystals at

HOKKAIDO, and snowfall occurred in the area northward from there. The amount of precipitation on this day was 28.5 mm at Sapporo, and 1.5 mm at Iwamizawa. The difference in the amount of precipitation between these two places can be explained as follows.

The horizontal distribution of snow crystals at
Fig. 7—Sea-level, 850-mb, and 700-mb charts, 09h 00m, January 30, 1959

Fig. 8—Horizontal distribution of snow crystals, 09h 30m to 12h 30m, January 30, 1959
09h 30m, 10h 30m, 11h 30m, and 12h 30m was as shown in Figure 8. As seen in these figures, the peculiar characteristic of this snowfall was that needle crystals were observed in the southern half of the observation area, indicated by N. On the other hand, snowfall was not observed in the northern half of the observation area, except that plates were observed at 09h 30m and 10h 30m at some stations.

The sounding curve at 09h 00m is shown in Figure 9, and Ta-s diagram for the sounding data is shown in Figure 10. It will be expected from Figure 10 that column (capped column), plate, dendritic, scroll, and needle crystals would occur in such atmosphere aloft as shown in Figure 9. As seen in Figure 8, this expectation agrees with the result of observation in the southern half of the observation area. The predominance of the occurrence of needles can be explained from the existence of the deep layer warmer than -10°C where needles form, as seen in Figure 9.

As indicated by N and S in Figure 8, the area where needle and stellar crystals were observed is limited by an east-west line. This direction was not parallel to the direction of wind aloft, but parallel to the direction of the isothermal line in the upper air, as seen by the middle and right-hand charts of Figure 7. This seems to be a suggestive observation for further research on the system of snowfall. The reason why the snowfall in this case was limited to the southern half of the observation area cannot be explained now, but will be investigated in the further analysis.

At 09h 30m and 12h 30m, the area where stellar crystals were observed most frequently existed northeast of the area where needle crystals were predominant, as seen on the 09h 30m and 12h 30m charts of Figure 8. This situation may be explained by the drifting of snow crystals by wind aloft. As seen in the left part of Figure 9, the direction of wind aloft was SE and SSE at the altitude from 1.0 to 2.5 km, where needles formed. Therefore, the needles are considered to have drifted to the NNW by the wind aloft. On the other hand, since the direction of wind aloft
Fig. 11—Sea-level and 700-mb charts, 09h 00m, February 10, 1959

Fig. 12—Horizontal distribution of snow crystals, 09h 30m to 12h 30m, February 10, 1959
was SW at the altitude from 3.5 to 4.0 km where
dendritic crystals formed, the dendritic crystals
are considered to have drifted to the NE by
the wind aloft. On account of such difference of
the direction of drifting, it will be reasonable that
the area of stellar crystals existed to the north-
east of the area of needles.

On February 10, the snowfall was due to the
passage of the cold front. Sea-level and 700-mb
charts at 09h 00m are shown in Figure 11. The
snowfall area was the northern part of the Japan
Sea coast of Japan. The amount of precipitation
was 15.5 mm at Sapporo, and 1.7 mm at Iwami-
zawa.

The horizontal distribution of snow crystals at
09h 30m, 10h 30m, 11h 30m, and 12h 30m is
shown in Figure 12. As seen in this, the character-
istic of this snowfall was the transition from
the state of predominance of stellar crystals at 09h
30m to that of needles at 12h 30m.

The sounding curve at Sapporo at 09h 00m is
shown in Figure 13, and Ta-s diagram for those
sounding data is shown in Figure 14. It will be
expected from Figure 14 that spatial plate (spat-
ial dendrites), plate, dendritic, scroll, and needle
crystals would occur in such atmosphere. This
expectation agrees with the actual observation re-
sults shown in Figure 12, excepting the occurrence
of column crystals. The layer suitable for forma-
tion of dendritic crystals was from the altitude of
2.5 to 2.8 km; this was in concordance with
surface observation of clouds, reporting that there
was Altostratus [Kuetttner and Boucher, 1958] at
3 km at 09h 00m JST. The layer suitable for for-
mation of needles was from the altitude of 0.9
to 1.3 km, which is in concordance with the data
of surface observation, reporting that there was

![Fig. 14—Ta-s diagram for the sounding at Sapporo, 09h 00m, February 10, 1959](image)

![Fig. 13—Sounding curve at Sapporo, 09h 00m, February 10, 1959](image)

Fractonimbus at 0.9 km in the period from 09h
00m to 12h 00m.

As indicated by **X** in Figure 12, the area where
needles were observed expanded in the period
from 09h 30m to 12h 30m. At 09h 30m, the stel-
lar crystals were predominant, while needles were
observed only at Station 2 (Sapporo). At 10h
30m, the area of needles moved westward, needle
crystals being observed at Stations 8 and 13.
At 11h 30m, the area of needles expanded in the
longitudinal direction, while snowfall stopped in
the areas other than this area. At 12h 30m, the
area of needles expanded more than that at 11h
30m. The reason for this transition cannot yet be
explained. It must be noted, however, that the ex-
anding direction of the needle area was parallel
to the direction of the isothermal line in the upper
air, as seen in the right-hand chart in Figure 11.
This result was the same as that of January 30,
and is very interesting.

On February 16, the snowfall was due to the
passage of the trough in the upper atmosphere.
Sea-level and 700-mb charts at 09h 00m are
shown in Figure 15. The snowfall area was in
Hokkaido, and the rain area was over the north-
Fig. 15—Sea-level and 700-mb charts, 09h 00m, February 16, 1959

Fig. 16—Horizontal distribution of snow crystals, 09h 30m to 12h 30m, February 16, 1959
ern half of Japan. The amount of precipitation was 0.0 mm at Sapporo (Sta. 2), and 2.4 mm at Iwamizawa (Sta. 13).

The horizontal distribution of snow crystals at 09h 30m, 10h 30m, 11h 30m, and 12h 30m is shown in Figure 16. The characteristic of this snowfall was that plate and column (capped column) crystals were observed widely in the observation area.

The sounding curve at 09h 00m is shown in Figure 17, and Ta-s diagram for these sounding data is shown in Figure 18. It will be expected from Figure 18 that spatial plate, plate, and column crystals would be the main shapes in this snowfall. Besides, dendritic crystals may form in Region I, but will not be so predominant, since the supersaturation with respect to ice was 110%, that is, nearly the critical value for transition from a plate to a dendritic form. These expectations agree with the results of observations, as seen in Figure 16. The snow-crystal form recorded as irregular crystals by observers is considered to be spatial plate. From Figure 18, it may be considered that column crystals formed in the warmer layer of Region VII, but that was not the case. Since the capped column was observed at some observation points, the column should form in a layer higher than a layer suitable for dendritic or plate, that is, in a colder layer of Region VII. However, no colder layer responsible for the generation of crystals of Region VII was observed by the sounding, as seen in Figure 18. This discrepancy must have been due to the horizontal difference of supersaturation in the upper layer.

In this snowfall, the areas where the same shape of snow crystal was observed could not be

\[
\begin{align*}
\text{Fig. 18—Ta-s diagram for the sounding at} \\
\text{Sapporo, 09h 00m, February 16, 1959}
\end{align*}
\]

decided, since the mixing of snow crystals was observed at many stations.

Concluding remarks—Simultaneous observations of the shape of snow crystals were carried out at 14 points in an area of about 5200 km² in the Ishikari Plain, Hokkaido. Though this investigation was a preliminary one, it was found that areas where the same shape of snow crystals are to be observed can be detected by observations in an area as wide as in this case. However, a wider area of observation will be necessary for fully understanding the snowfall caused by the large-scale motion associated with frontal zones. Therefore, further observations in the winter of 1959–1960 will be carried out over a wider area, and by the use of more precise methods, such as plastic replica or shadow photograph.

In conclusion, the authors express their hearty thanks to C. Magone for his criticism through this work. The authors are much indebted to A. Ichikawa for his kind assistance in carrying out this work, and to the Sapporo Meteorological Observatory, to the Iwamizawa Weather Station and to the Hokkaido Scientific Education Society for carrying out the observations of snow crystals.
DISCUSSION

References


Discussion

Dr. Helmut Weickmann—How reliable are your humidity measurements?

Dr. Nakaya—These measurements are made with a hygrometer developed by Dr. Kobayashi of the Central Meteorological Institute in Tokyo. He has a ten-year experience with such measurements and succeeded in developing an instrument which is superior to other hygrometers.

Dr. B. J. Mason—If there is an isolated cloud with nothing above or below, any crystals which are collected come from just this layer. Then one can deduce its temperature and perhaps the humidity. I am doubtful of this deduction, if there is a deep cloud system where there is a large temperature range. In England, I generally observe in 90% of the cases a mixture of crystals of all sorts which come from a deep cloud layer system, and everything is mixed up and it is impossible to make any useful observations. I should like to ask Dr. Nakaya if he agrees with this, and ask him if he believes that one can only make this kind of an analysis when in fact, you have rather well defined thin cloud layers from which the crystals fall?

Dr. Nakaya—That I expected. This is a very interesting and important problem. If the clouds had three layers (Fig. 19) A, B and C, and snow crystals come down in the vertical, we must expect on the ground three types of snow; C, C plus B, C plus B plus A. If A is plate, B is dendritic, and C is needle, we can expect needle, dendrite with needles, plate with dendritic extensions with needles attached; something like that. But it was not thus. Sometimes we observed the combined type, but usually simple type of crystal corresponding to each layer of the clouds was observed. In this year's observations we found that when the upper-air condition is layered A, B and C, the upper type A or the middle type B was observed on the ground. They appear to have escaped layer B or C. This is against our simple assumption of vertical fall, and we must develop the theory to explain this phenomenon. This is a future problem, but anyhow when we observe three or four kinds of crystal types on the ground we see that three or four corresponding layers are in the upper levels of atmosphere. The only explanation at present is that the atmosphere has a cellular structure or is in the turbulent condition and is not made up of horizontal strata. In this case a certain type of crystal can come down to the ground without passing through the stratum beneath. So this is a problem of the microstructure of the atmosphere.

Dr. Weickmann—(answering Dr. Mason) In fact, your observations of various kinds of snow crystals were interesting because it indicates that you have a deep and vigorous cloud system. If you observe only one type of crystal, the cloud system underneath the generating layer is not vigorous enough to produce its own crystals, it is just able to nourish the crystals which fall through, but in other more vigorous cloud systems you may get the crystals initiated in several levels, and then you observe several types; and this seems to be the case in the systems you observed.

Mr. L. Aldaz—During the winter of 1956—
1957, on Mount Washington on the contract sponsored by Dr. Weickmann, we found as did Dr. Grunow and Dr. Nakaya, a very good correlation between the synoptic situation and the ice crystals. We made three kinds of analyses: (1) radar, (2) synoptic, and (3) ice-crystal analysis; the last I did myself. It is quite surprising how well one can determine the conditions aloft.

Now, I would like to offer a word of caution. The main problem arises not from the mixture of ice crystals, but from the Nakaya diagram in which there is a duality of identical forms coming out on both sides of the diagram. In many occasions, for example, we have the occasion of hollow needles or long columns which one does not know exactly where to place. The same thing happens with plates; only with dendrites, there is no problem.

Dr. Horace R. Byers—I notice in Dr. Nakaya’s data that the clouds seemed to be entirely below the 800 millibar surface; therefore, there are not great multiple layers involved. This is true also of the systems which produce snow on the exposed sides of the Great Lakes; therefore, you do not have there this multiplicity of crystal types. Augmenting what Dr. Weickmann said in regard to even those clouds which are very thick, it is my experience in observing them in Chicago or elsewhere, without lake or other special influences, that one almost never gets a mixture of types. It is usually one type. The type may change during the course of a storm, but at any given instant there will be usually just one type.

Dr. R. List—Regarding the comment of Mr. Aldaz, I can show you a dendrite that was found in a cloud with temperatures above about −4°C. It appears to have been grown at −15°C, but we know it was grown at temperatures higher than −4°C.

Mr. Aldaz—How do you know?

Dr. List—Our institute is on the top of a mountain, and we can see the top of the clouds very well.

Dr. C. L. Hosler—Two winters of observations of snow and activity in central Pennsylvania seldom got mixtures, and snowflakes were about 80% dendrites. We did not observe mixtures at all.

Dr. C. J. Grunow—In the cases of February 10 and 16, the behavior of columns is in contradiction with the results found in your laboratory investigations. We found the same: the conditions of growth for columns must be somewhat other than defined in the Nakaya diagram. Also the observations of Dr. Weickmann in natural Cirrus clouds suggest the beginning of growth for columns at a lower temperature. Considering results of Kobayashi, it follows that not only the temperature and the state of saturation but also the ambient vapor density is an essential factor of growth. In the Kobayashi-diagram the column dominates in a range of little ambient vapor density independently of any temperature ranges. Can the named doubtful cases in the behavior of columns be explained by this effect?

Dr. Nakaya—I admit that my diagram is now almost 20 years old and it must be revised by the advanced technique of experimentation in this field. Mr. Kobayashi is now working on this problem in the Low Temperature Institute of our University. The problem exists not only in the point of the ambient conditions but also in the point of the definition or classification of the type of crystals. The columns are the most embarrassing type. Short needle, sheath type, column with thinner wall and the ordinary column belong to the same category; growing more in the direction of principal axis and less in the direction perpendicular to it. More detailed studies must be carried out for distinguishing one type of crystal from the other belonging to this same category.
Snow Crystal Analysis as a Method of Indirect Aerology

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Abstract—Observations of form and size of snow crystals, whose conditions of growth are widely known from laboratory investigations, were compared with the synoptic situation. The qualitative analysis obtained by a general survey of all forms appearing at the same time imparts an insight to the structure of the atmospheric layers in their temporal and spatial succession. A quantitative analysis is developed to derive the thickness of layers and to reconstruct cross sections of temperature and state of saturation, based on rates of fall and rates of growth as known from laboratory investigations. The results show satisfying agreement with the cross sections derived from aerological soundings for layers up to −20°C. From measurement of numerous shadow photos, it was possible to derive frequency-sections and spectra of the distribution of sizes for some crystal forms that permit statements concerning the structure of the precipitating cloud system. From evaluation of water droplets attached to snow crystals, spectra of the diameters of droplets for different air masses were derived. Even simple observations of snow crystals from different altitudes made by the visual method show characteristic differences of form and size of the crystals. Thus the suitability of the snow crystals as aerological sonde is confirmed by many single manipulations.

Introduction—During the past decades the study of the ice-phase in the atmosphere has yielded important insights in fields of cloud physics and in the formation and release of precipitation. With this knowledge, investigations of snow crystals on the ground, particularly on mountain stations, have proven to be useful as can be seen from the work of Weickmann [1957a,b] on Mt. Hohenpeissenberg, Germany (located in the northern foreland of the Alps) and Kuettner and others [1956, 1958] on Mt. Washington, N. H. The form and size of snow crystals, whose conditions of growth have been widely known from numerous laboratory investigations, allow conclusions concerning the structure of the precipitation cloud system and upon the temperature and humidity within the upper air layers. In order to check these results under different climatic conditions and to complement the observations in the upper air layers with aerological measuring devices during the IGY, consecutive records of snow crystals by photographic methods, as shadow- macro- and microphotography, and by replica technique, were made on Mt. Hohenpeissenberg during the winter of 1957–1958 [Grunow and Haefer, 1959]. The conclusions drawn from the form analysis of snow crystals with respect to the state of the upper air layers were compared with the synoptic situation as determined by the data obtained by the usual sounding procedure.

Qualitative analysis—The manifold forms of snow crystals with their innumerable variants depend upon the atmospheric conditions, particularly upon temperature and state of saturation, that prevailed during their formation and along the path of fall. The ranges of temperature and humidity are demonstrated in the diagrams of Nakaya [1954], Aujin Kampe and others [1951], and Weickmann [1957a]. Nakaya chooses as ordinate units of the relative humidity as a measure of the supersaturation with respect to ice, while Weickmann uses the vapor-pressure difference between water and ice saturation. Recent investigations by Kobayashi [1957, 1958] on the habit of snow crystals artificially produced at low pressures show a relationship of the shape of crystals to the ambient vapor density. There- with the necessary conditions for the growth of each crystal type are known, namely, the ranges of temperature and the variation of humidity (state of supersaturation and ambient vapor pressure). It is possible to deduce from the successively grown parts of a snow particle that has fallen through the atmosphere, the conditions of the air layers during its passage. This statement about the structure of the atmospheric layers in their temporal and spatial succession we define as qualitative analysis.

Such a qualitative analysis requires a general survey of all forms appearing at the same time, and their arrangement according to a detailed
SNOW-CRYSTAL ANALYSIS AS INDIRECT AEROLOGY

Temperature °C

-85 -70 -63 -63.5 -50 -65 -65.75 -75.90 -95 -85 -104

Special forms

0700 0800 0900 1000 1100 1200 1300 1400 1500 1600 1700 1800 1900 2000 2100 2200 GMT

Fig. 1—General survey of observed snow crystals according to Weickmann, showing distribution on March 21, 1958 (Operation 27), Hohenpeissenberg, Germany.

Physical classification, which also includes their temporal variations. A method for such a representation was used by Weickmann [1957a]. An example for this method is given in Figure 1. Of the two columns of the ordinate the first contains typical forms that require only slight ice supersaturation for their growth. The second column, however, comprises forms whose growth occurs at or close to water saturation. The diagram shows two periods, where the precipitation clouds reached up into upper layers of the troposphere. A cloud layer with water saturation existed in the -15 to -20°C temperature range in agreement with the aerological cross section. During the forenoon such a layer existed also between -10 and -15°C discharging malformed plane crystals. Also the disappearance of the upper precipitation clouds during the evening hours can be seen.

Quantitative analysis—If the rate of fall and the rate of growth for each crystal type are also known, these facts in connection with the size of the grown crystal type allow calculation of the thickness of layers through which the particle fell and in which these conditions were met. With this supplementary statement the qualitative analysis is extended to a quantitative analysis. From the arrangement of these originating zones a cross section of the isotherms and isoabumes can be constructed.

For both quantities, the rate of fall and the rate of growth, Nakaya's values were used. The rate of fall proves to be nearly constant for all dimensions of plane and spatial dendrites as well as of powder snow. For crystals with droplets and even more for needles and graupels the rate of fall increases with their size. The degree to which the rate of fall depends upon the density of air, is not known and cannot be considered. The greatest influence, however, should be exerted by the vertical air transport. But the appearance of upward currents is limited to certain weather situations with a character of instability. For the computations here average values of the rate of fall...
were used. In the rate of growth of different types of crystals (such as dendrites and needles) differences exist because of variations of the vapor pressure along the path. Assuming a constant mean value for the rate of growth, as it was done here, the determination of the thickness of layers becomes somewhat doubtful, but is unavoidable. Losses of substance by evaporation in case the crystal passes through possibly existing dry layers, however, can be seen by the degenerated shape of the crystal. Those particles remain unconsidered. By picking out the largest of each type from among all crystals, with a sufficient total number of the same, it can be assumed that the size thus found presents a characteristic measure for the thickness of the layer within which the crystal originated.

With the rate of fall $V$, the rate of growth $W$, and the size of the crystal $\Delta G$, the thickness of layer $H$ is thus calculated,

$$\Delta H = \frac{(V \cdot \Delta G)}{W} (m)$$

After analysing the different snow crystals contained in the snow sample for each observation term according to this procedure, the ranges obtained for the temperature and the thickness of

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**Fig. 2**—Construction of a height cross section of temperature and humidity; March 29, 1968, 08h 34m GMT (Operation 26)
the layers are taken as basis for the construction of a height cross section. As starting-point it is necessary to have a reference value, for which the relation temperature to altitude is taken from other facts of observation, for instance, from ground observations. The construction is then carried out in two parts (Fig. 2):

1. The thickness of temperature ranges in meters is entered in an equidistant scale of temperature for each observation time (above left). If several different types of crystal appear in one observation, as it is the rule in most cases, the diagram of the stratification can be further secured by comparisons of the single-layer thicknesses determined for two or three snow particles. Thus, uncertainties of using average values for the rate of growth in relation to the state of saturation can be partly eliminated.

2. The temperature scale, derived from the analysis of the different crystals, is entered in the diagram with equidistant scale of altitude (above right). This operation is done in several steps, as explained in detail in Fig. 2, below. The statements about humidity are added at the right of the altitude-scale.

These values obtained for all observations of one operation were put in a row (Fig. 3). A cross section for temperature and humidity constructed by the snow-crystal analysis as derived according to the procedure described here is shown in Figures 4 and 5.

A comparison between the results from the snow-crystal analysis with the aerological state measured by soundings indicates (1) the accordance in the existence of layers of defined qualities, especially the state of saturation combined with a defined temperature range and the structure of precipitation clouds (qualitative analysis), and (2) the quality of a quantitative analysis, especially the question whether in spite of some necessary simplifications with respect to the use of mean values for the rate of growth the results are suitable for a fairly exact calculation of the thickness of defined layers. The results are as follows:

The growth conditions of a crystal form, known from laboratory investigations, allow conclusions regarding the temperature range and the state of saturation of the atmospheric layers in which the crystal originated. The qualitative crystal analysis from consecutive observations during snowfall shows excellent agreement between crystal types and synoptic development.

Quantitative analysis for deriving the thickness of layers and for reconstruction of cross sections, based on rates of fall and rates of growth, as measured in laboratory experiments, show satisfying agreement with the cross sections derived from aerological soundings for layers up to the −20°C level. Determination of the stratification for the upper atmospheric layers becomes more and more uncertain, as the rate of growth for columns and the upper temperature boundary for their origin seem to be doubtful.

As the agreement of the reconstructed analysis-sections with the values from aerological measurements is best near the time of sounding, it can be
concluded, that the analysis of snow crystals presents more details of the synoptic development than the aerological soundings that, because of their 12-hour intervals, often do not show these details.

Frequency analysis—The analyses of shape and size of snow crystals were also correlated with special studies on their frequency of occurrence. Frequency analyses have manifold significance in problems of cloud physics. They present information in connection with the intensity of precipitation and its dependence upon certain qualities, for instance, as carrier of atmospheric electric charges, radioactive fission products, and so on. They give also information as to the quantity of freezing nuclei contained in an air layer.

Fig. 4—Cross sections of temperature and humidity; above: derived from aerological soundings; below: derived from snow-crystal analysis, Hohenpeissenberg, March 20, 1958 (Operation 26)
of certain characteristics that did not form ice-crystals before coming into this layer. Crystals falling from a higher layer through the whole depth of a second layer may expect an approximately equal increase of their rate of fall, and their rates of growth are nearly the same. The size spectrum of these crystals or their appendices, respectively, will show only a narrow frequency culmination. If these crystals were formed in the second layer, however, they would show quite different sizes. In this case the size spectrum will be broad. The frequency curve ought to indicate the form of a gaussian distribution. In the case of a broad spectrum several frequency-cul-
minations are to be expected that correspond to different part-collectives.

Best qualified for the frequency analysis are the shadow-pictures, as they furnish a great number of single individuals in each photo. Frequencies of individual crystals were put down in a table showing for each observation the share of each form in accordance with the International Snow Classification expressed in percentage (Fig. 6). The predomination of forms that originated

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**Fig. 6**—Frequency analysis of observed snow crystals derived from shadow pictures; percentages of observed crystal types according to the International Snow Classification; Hohenpeissenberg, March 21, 1958

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**Fig. 7**—Percentage of frequency of dimensions of snow crystals; crystal group: plates, sectors, broad branches, Hohenpeissenberg, March 21, 1958 (Operation 27)
at temperatures $<-20^\circ C$ coincides with the appearance of high precipitation clouds that are verified by aerological soundings. The prevalent occurrence of supersaturation forms points to the predominating production of precipitation in lower, nearly water-saturated layers. From the alternating occurrence of certain forms the sinking or rising or a cell-like structure of the precipitation clouds can be deduced. Thus this diagram is in conformity with the general survey according to the method of Weickmann.

The variations of the size of certain crystal types with time is given in isolines of relative frequency of the diameter. An example is given for the Operation 27 in the group of plates, sectors, and broad branches (Fig. 7). Frequency of the particular forms, the most frequent diameters, width of spectrum, and size of the maximum of the spectral curve, respectively, do not run parallel to each other. A relation seems to be indicated, however, in such a manner that a small variety of forms of one type corresponds to a relatively narrow spectrum, that few appearing forms thus show nearly similar dimensions. On the other hand with a large percentage of one form the dimensions vary a great deal. The size spectra can

Fig. 8—Dimensions of single crystal forms; frequency of their occurrence expressed as percentage; Hohenpeissenberg, March 1958
be narrow as well as broad, the probability curve more pointed or flatter, and the most frequent diameters smaller or larger. With that, also the frequency curve depends upon the momentary structure and the characteristic of the layers and thus presents another means for the indirect aerological analysis.

Dimensions of crystals—Supplementary, the dimensions of all measured crystals of the same type were compiled as frequency curves, showing the dimensions of single crystal forms expressed as percentage (Fig. 8 and 9). These curves were not smoothed. Their form of a probability curve expresses that under given climatic conditions and altitude, dimensions can be expected that have characteristic sizes for the different types of crystals. Other material based on exact measurement of a sufficiently large number of individuals of crystals is only known for Sapporo, Japan, near the sea-level and, for only few forms, for Mt. Tokachi, 1060 m above sea level. The typical values found on Mt. Hohenpeissenberg, 1000 m above sea-level, mostly show smaller dimensions than the ones ascertained in Sapporo. The differences are relatively small with those forms growing in upper layers at a temperature of about below $-20^\circ$ C. But they are great for the forms
Fig. 10—Results of visual snow crystal observations; frequency of shape and size; simultaneous observations Hohenpeissenberg (975 m) and Mt. Zugspitze (2962 m) 1956–1958
(dendrites, needles) that grow within the lower layers of the troposphere, their growth being favored by the maritime climate of Hokkaido which has excessive humidity. Also the maximum size of dendritic forms found on Mt. Tokachi is 3.0 mm, still considerably higher than the value 1.8 mm of Mt. Hohenpeissenberg with nearly equal elevation but with a more continental climate.

Differences of the same kind were also found by simultaneous snow-crystal observations of Mt. Hohenpeissenberg and Mt. Zugspitze with the visual method according to the International Classification of Snow (Fig. 10). Three years of additional observations by the meteorological observers of these stations have demonstrated the suitability of this scale discussed at the Conference of Woods Hole in 1955. On Mt. Hohenpeissenberg those types prevail by number and size that need only moderately low temperatures and nearly water saturation for their growth, for instance, stellar crystals, needles, and graupel. On Mt. Zugspitze those forms dominate that originate in upper precipitation clouds, and at much lower temperatures, and are crystals of Class 7 (irregular crystals) which mostly prove to be spatial plates if viewed under the microscope. Thus even with application of simplest observation methods the snow crystal as indirect aerological sonde is capable of furnishing valuable references.

Dimension of cloud droplets—Finally a number of microphotographs of rimed crystals were used to measure the diameters of water droplets attached to snow crystals in order to obtain their size spectrum. It was found that these derivations show parallels to former investigations, with which the abundance of fog precipitation was put in relation to the spectrum of droplets of different air masses (Fig. 11). The individual frequency curves are composed of different part collectives, but the total number is not sufficient for their exact analysis, for instance, by the procedure of Esenklinger [1954]. Arctic cold air is characterized by a narrow spectrum with small diameters, warm air and degenerated air masses by a broad spectrum with larger diameters.

Conclusions—The suitability of the snow crystal as aerological sonde is found confirmed by many single manipulations and it is expected that the further evaluation of this observation ma-
terial will still give more information on the physics of cloud.

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References


Structure of Snowfall Revealed by Geographic Distribution of Snow Crystals

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Abstract—The Cloud Physics Group of Japan made observations of snow crystal forms by various methods at five points which were distributed vertically from elevations of 100 m to 1000 m at Mt. Teine, Hokkaido, through the period January 26–31, 1959. Aspirated psychrometers and rawinsondes were mainly used to measure the air temperature and humidity.

The results are summarized as follows: (1) Nakaya’s Tₜ₅-s diagram represents fairly well the growth of natural snow crystals. The crystal forms observed at the Earth’s surface are mainly affected by the temperature and humidity of air layers at altitudes lower than 2000 m. (2) It seems that sometimes dendritic snow crystals grow at a humidity very near to ice saturation, or at least at a lower humidity than water saturation. (3) The necessary conditions for the formation of large snowflakes are the existence of a thick moist atmospheric layer and of air temperature higher than −10°C.

Introduction—The relation between the snow-crystal forms and the meteorological conditions was studied by Nakaya in laboratory experiments, and his theory was proven by Gold and Power [1954], Murai [1956], and Knethner and coworkers [1958] using aerological sounding data obtained during snowfall. As for the growth of natural snow crystals, however, there are no available observational data. In order to be able to observe the rate of the growth, observations were made at several points distributed vertically.

The Cloud Physics Group of Japan made the observation of natural snow crystals by various methods at five observation points distributed vertically at Mt. Teine in January 1959 and measured the form and size of snow crystals of almost all types.

Methods employed—As one sees in Figure 1, Mt. Teine is located about eight miles northwest of Sapporo where rawinsonde soundings were carried out by the Sapporo Meteorological Observatory at 09h00m, 15h00m, and 21h00m each day during the period of the observations. Since the predominant wind direction during the period was from the northwest, the observing points are located on the windward side, therefore the atmospheric conditions measured by the aerological sounding may differ somewhat from those above the mountain except for the ease of large uniform snowfall.

The horizontal distribution and vertical distribution are shown in Figures 2 and 3. The observation points at altitudes 1023 m, 800 m, 500 m, 300 m and 100 m are called Point 1000, Point 800, Point 500, and so on, respectively, in this paper. The upper three points are close together, but the lower two points are somewhat far from the upper points; accordingly, snow crystals observed at the lower two points were sometimes not recognized as belonging to the same cloud system. Frequently no snowfall was observed at the lower points even when moderate snow showers were observed at the upper points. This discrepancy is considered to be due to orographic effects.

The observers and their work during the observation are listed in Table 1.

The work of the observer at Hokkaido University was to identify the clouds over Mt. Teine from outside of the observing area. This work was very useful, because the observers located at the mountain often could not observe the clouds by which they were surrounded. Among the methods employed in observing snow crystals, the securing of replicas and microscopic photographs were most useful. As for the measurement of humidity, Assmann’s aspirated psychrometer was most reliable.

The observations at the five points were carried out simultaneously, every ten minutes during three hours after the rawinsonde sounding times (09h00m, 15h00m, 21h00m) from January 26–31, 1959.

Results—The time cross section obtained using aerological sounding only is shown in Figure
Therefore, the data obtained by the aspirated psychrometer were taken as the condition at altitudes below 1000 m. As for levels higher than 1500 m, the data obtained by rawinsonde were accepted as they were. The conditions at the intermediate level between 1500 m and 1000 m were interpolated. The time cross section obtained thus is represented in Figure 6 in which the layer lower than 3000 m is shown. The altitudes of the five observation points are represented by horizontal short thick lines on the ordinate. The type of representative snow crystals observed at Point 1000, Point 500 and Point 100

4. Solid lines show isothermals and dashed lines represent relative humidity with respect to ice saturation. The thick solid line shows especially the isothermal of −15°C around which snow crystals grow rapidly to form dendritic type. In general, moist air existed at the −15°C level on the 26th and night of 27th when light snow showers were observed at ground level. Early in the morning of 30th a cyclone passed through Hokkaido as shown in Figure 5 and warm moist air flowed into the observing area and brought heavy snowfall. This snowfall was on a large scale and the observation points were located to the lee of the mountain, so in this case the data obtained by rawinsonde sounding are considered to be reliable. Continuous snowfall was observed at Point 1000 during the period of the observation except on the 29th, but the moist-air occurrence at the Point 1000 was limited to a few short periods. This strange phenomenon is considered to be the result of condensation of orographic ascending air, because the observed condensation level was always lower than the altitude of Point 1000 as shown on Figure 4. In addition to that, several small crystals of initial stage were almost always observed at the upper points.
at each three-hour interval is shown at the upper part of the figure schematically. The first type shown on each of the top lines means the type of snow crystal which fell most frequently. The region of humidity higher than ice saturation is shown by shaded area. Cloud bases are shown only when ascertained. From the figure it may be seen that moist atmosphere layers (hatched areas), exist near or above Point 1000 always when snowfall is observed at Point 1000. The only exception is on the night of 28th. This is important; it will therefore be discussed later.

According to Nakaya's $T_s$ diagram, a temperature of $-15^\circ{\mathrm{C}}$ is suitable for dendritic snow crystals to grow, while at a temperature of $-5^\circ{\mathrm{C}}$ snow crystals grow to needle form. From an inspection of Figure 6, note that from the 26th to 28th when the $-15^\circ{\mathrm{C}}$ isotherm existed near or above the top of the mountain, dendritic snow crystals were predominant, and on the 30th when the $-15^\circ{\mathrm{C}}$ isotherm had descended to this altitude, only thin hollow columns or needles were observed. One may understand that Nakaya's diagram agrees very well with the type of natural snow crystals. This fact leads to the conclusion that the type of snow crystals is considerably influenced by the temperature of the air layer lower than 2000 m. It is well known that in the $T_s$ diagram the region of column and plate type is distributed symmetrically around the dendritic region; that is about $-15^\circ{\mathrm{C}}$. However considering the time cross section it is possible that the snow crystal type falling naturally is mainly affected by the warmer region of two temperature regions of crystal formation. This is considered to result from the fact that at the higher level the vapor density is small as is the air density. The absolute vapor content is small even if the air of the upper level is saturated because the air is colder than that at the lower level. Ac-
Fig. 1—Time cross section during the period of the observations
Accordingly, if any crystals are born at a comparatively higher level, they do not grow rapidly until they fall into lower layers. As for the relatively dry regions above the mountain on the 30th and 31st as seen in Figure 6, it is not clear whether those regions have any connection with crystal forms observed or not.

At time Sections A(21h40m, 27th), B(21h00m, 28th), C(11h20m, 30th), D(21h30m, 30th), and E(16h10m, 31th) in Fig. 6, marked snowfalls were observed. For each time section, some descriptions should be offered in detail.

Time Section A: After 21h00m on the 27th, large snowflakes were observed (Fig. 7); snowflakes began to form at the Point 1000 level. The magnification factors of three pictures in the figure are common to all. It seems also that the snowflake formation began when snow crystals fell into the layer warmer than $-10^\circ$C [Magono, 1953]. This fall of large snowflakes was accompanied by the occurrence of a thick moist air layer as seen in Figs. 6 and 7.
Fig. 7—Growth of snowflakes 21h 30m–21h 50m, Jan. 27, 1959

Time Section B: On the 28th, snow crystals of dendritic type continued to fall. It cleared off about 15h00m but the crystals continued to fall although it was a light fall. The snow crystals were of very beautiful plane dendrites form. The vertical distribution of air temperature and humidity at 21h00m is shown in Figure 8. There existed a thick layer of −15°C near Point 1000 actually. It is notable that the air was not saturated and there were no clouds nearby; however perfect snow crystals fell and grew as shown in the left-hand part of the figure. The size distribution is shown in Figure 9 starting with a crystal diameter of 0.5 mm. It is noted that the mean diameter of the snow crystals observed at Point 500 was markedly larger than that at Point 1000. The mean diameter of snow crystals at Point 100 is about the same as that at Point 500, resulting from the fact that the air layer between Point 500 and Point 100 was dry. The tips of the branches of snow crystals observed at Point 100 were, to some extent, changed to sector form.

The observational fact that dendritic snow crystals grew in air not saturated with respect to ice may be thought to be strange. Concerning the fact, the error of the humidity measurement should be considered first. The error of the psychrometer was ±0.05°C. The error margin corresponds to about ±2% at −15°C, therefore the air whose humidity was lower than around 95% as seen in Figure 8 should be considered to be not saturated to ice. At least it is sure that the air was not saturated with respect to water.

Generally it is considered from laboratory experiments that dendritic snow crystals grow in
highly supersaturated humidity which requires
the existence of cloud droplets in air in the natu-
ral case. But if the air contains cloud droplets,
the snow crystals will collect and cannot be
called 'beautiful crystals.' It therefore appears
that beautiful plane dendritic snow crystals grow
in air of humidity lower than that assumed from
laboratory experiments, in other words, they
grow at the humidity very near to ice saturation.
In this case, the existence of cloud droplets is not
necessary.

Time Section C: It was characteristic here
that the lapse rate of the air temperature was
very small and the atmosphere was saturated
from near surface to altitude 6000 m as shown
in Figure 4. The air layer with temperatures
ranging from $-5^\circ$C to $-8^\circ$C was very thick.
This condition is suitable for the growth of thin
Fig. 10—Growth of snow crystals of thin hollow column type, 11h 20m–
11h 30m, Jan. 30, 1959; shading in column shows the temperature region of
snow crystal growth for the crystal type indicated

hollow column and needle type crystals. As assumed, it was observed as shown in Figure 10
that the snow crystals of hollow column grew moderately between Point 1000 and Point 500,
but below Point 500 the crystal type changed to needle. The layer between the lower two
points also was dry, as seen in Figure 10. During daytime on the 30th, snow crystals of thin
column or needle type were observed predominantly; also various stellar forms occurred. This
fact means that at a level colder than \(-8^\circ C\) no snow crystals formed. This may be caused
by the relatively dry region above the mountain seen in Figure 6.

Time Section D: From about 15h00m on the
30th, the temperature of the whole atmosphere
fell, and rimed dendritic snow crystals were ob-
served simultaneously with those of needle type
(Fig. 11). Snow crystals of needle type grew to
some extent and the dendrites began to form
snowflakes near the level of Point 1000 where
air temperature was about \(-8^\circ C\).

Time Section E: Near the end of the observa-
tion period, the process of the formation of
graupel of cone-type was observed in detail as
seen in Figure 12. It seems that the graupel of
cone-type originates from a rimed branch sepa-
rated from a rimed snow crystal as pointed out
by Barkow [1908].

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References


Discussion

*Dr. C. J. Grunow (communicated)*—Dr. Magono concludes from his very interesting observations, that the crystal forms observed at Earth's surface are mainly affected by the meteorological conditions within the layers lower than 2000 m. As shown by his observations of January 30, the upper layers (>2000 m) also cooperate if upper cloudiness is present. Otherwise the growth of columns and capped columns would not be explicable. It is true the rate of growth in the
upper layers is much less than in the lower layers where the ambient vapor density is higher. On Mt. Hohenpeissenberg we have observed many snowfalls where, during a long period, the columns dominate, falling doubtless from upper layers. Observations on Mt. Zugspitze (2962 m) prove that snow crystals of each type, from needle and scroll to dendrite and column, occur also at this height. Instead of a limit of altitude it would be better to use a limit of temperature; for example, $-15^\circ$C. I think the maritime climate of Hokkaido is decisive for the predominance of dendritic forms there observed.

Dr. Ch. Magono (communicated)—As for the limits in which the snow crystals of various types develop the description by temperature is better and more general than by the altitude, as you mention, but I would like to take $-15^\circ$C instead of $-20^\circ$C, when the cloudiness extends to relatively warm temperature region.

The reason is as follows. The snow-crystal form except capped columns observed at Mt. Teine always seems to be affected by the meteorological conditions of the air layer warmer than the $-15^\circ$C layer which varied from elevation 1000 m to 3500 m. This phenomenon appears to be strange, but it will be understandable if the following facts are considered.

1. The snow-crystal form is classified by the shape of its branches, so the shape of the original small portion, in other words, the portion formed in upper air layers, is usually neglected in the determination of the classification.

2. Usually the vapor density in the lower air layers is much greater than that of the higher layers perhaps owing to the maritime climate of Hokkaido, as you note.

3. Considerable numbers of small snow crystals are produced in the lower layers (warmer air layers) at the orographic precipitation.

It is true of course that the snow-crystal form observed at the high elevation is affected by the air of the upper layers.

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Fig. 12—Formation of graupel, 16h 10m–16h 20m, Jan. 31, 1959
Operation and Results of ‘Project Pluvius’

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The results of all my work on orographic and other precipitation distributions showed that the official network in no country was dense enough. So I had to construct an ‘instrument,’ which, of course, is not a laboratory instrument in the ordinary sense; but after all it is comparable to an instrument, because it is exceedingly complicated, consisting, during the summers of 1955 and 1956, of not less than a thousand rain gages of the type shown in Figure 1. Thus we had to use very cheap ones; in fact, the present cost is only one dollar for this ‘automatic rain gage Pluvius.’ We used it to build our rather complicated measuring ‘instrument’ in order to find out something more about the real distribution of precipitation. I am accustomed to having a lot of questions concerning this instrument and how it works, and about its possible errors. I am prepared to answer them in the discussion.

As to regions to select for our ‘instrument’—that is, for this special network of gages—when starting our project during the autumn of 1953, we thought that one ought to test the whole set up in three different respects. (1) How the instrument would work in the field. (2) If it were possible to get enough observers evenly distributed and working without pay. (3) We also wanted to get some knowledge as to the interior structure of different kinds of rain mechanisms, for instance, the warm-front rain and the cold-front rain, when not disturbed by orography. So we selected an area 20 by 20 miles around Uppsala that was easy to reach, and which, I thought, was so flat that there would be no appreciable orographic effects. The measurements went on for six weeks (in October and the first part of November). Unfortunately, we had only a few big rains, but it turned out that they were very suitable after all. On October 14–16, a front lay across southern Sweden with a stationary rain area and northeasterly winds on its northern side. Figure 2 shows the results of the measurements during the first 12 hr of this period. Notice that it is the night and not the day period, so there would be very little convection. When we first got the figures, I thought they did not show much, but when plotted on an appropriate map showing the regions with forest (green on the original base map), and those without, a most surprising connection appeared between the shaded areas and much precipitation, and the white areas and a minimum of it. See Figures 2 and 3 where woodland is indicated by the legend.

Now, I must spend one minute on other projects of a similar kind. You might, for instance,
think that one could use the data from the Thunderstorm Project in the United States for the same purpose. However, they had only between 50 and 60 stations, and their areas presented no marked orographic features. As early as 1953 we had 150 stations, and later, in other parts of Sweden, we had up to 500 stations. So, we could cover a much greater area, having an interesting orography, whereas the Thunderstorm Project did not have these requirements. On the other hand, evidently, the Thunderstorm Project and other similar projects had a much better instrumentation at each individual station. We could not afford that. The official measuring instrument of the Swedish Weather Bureau costs twenty times as much as the 'Pluvius,' and we had the choice between 1000 of this type and 50 of the former type. We had to choose the 'Pluvius.' Then also the data from the different American and other projects were generally not accessible to us, as you will understand.

In Figure 2 one rain maximum lies to the lee of Uppsala, and one might think of a precipitation release caused by certain nuclei produced in this town, or by the convection released through the heating effects of Uppsala; all sorts of other explanations may pass your mind. One might also think that this was just a chance distribu-
tion; but the next night, skipping the intervening 12 day hours, we got much greater amounts, but practically the same distribution (Fig. 3) with the same stationary rain, and almost the same wind direction. The wind had shifted a little more to NNE, but notice that in both cases the proportion between maximum and minimum was three to one. (Here it is about 20 to 7, on the previous map it was 7 to 2½.) During the intervening and following day-time periods there was a similar pattern, but not so regular. This may be explained by the fact that in daytime, even at this high latitude and so late in the year, there will be some convective effects. Some heat will pass through the cloud, heat the ground, and disturb the cloud sheet.

As seen from Figures 2 and 3, in this case the highest elevation of the country from the surrounding plain is about 50 m; one cannot expect an increase of the rain to the double or the three-fold only by a lifting of 50 m. With a high condensation level that would hardly give any precipitation at all. True enough, if the low layers are moist, clouds may form even over small hills. But if higher layers are cloudless, these low cloud caps will be colloidal stable and symmetric, giving practically no precipitation, since there is no upper releaser cloud. Moreover,
within a small and shallow cloud cap there is no possibility of getting warm-cloud precipitation release, because the lifetime of the droplets will be too small. Therefore, I suggest the following explanation in such cases as this. The cloud base was either low from the very beginning, or the air below the Nimbostratus base was moistened through these 100 hours of rain. Then, orographic cloud caps with more or less intense condensation, colloidally stable in themselves, formed over those regions that were higher and clad with forest, because of the great increase in friction and its stemming effect on the flow. At last, the general rain from the frontal cloud system swept down through the cloud caps and ascertained a very good release of precipitation within them.

For the next project, July–October 1954, we selected an area between the great Lake Vänern and Lake Vättern in southwestern Sweden partly shown in Figures 4 and 5, representing an interesting orography (plains, dislocations, forests, shore lines, etc.). Here we had 750 gages; in addition in 1955 we had 150 stations in the central part of the region measuring the wind direction and velocity (the latter with a cheap pressure-tube hand anemometer). The two rainfall maps chosen as examples are for July 27, 1954 (cold-front) and for August 7, 1954 (warm-front), both with a general current from south-southeast. In the cold-front case, obviously, orography has had very little effect on the distribution; the explanation evidently being that the cloud base was high, since it was summer, and the stratification was unstable. Thus, no low feeder clouds would form above the low hills or plateaus of the area, and there were, instead, convective cells moving with the direction of the gradient wind, giving streaks of precipitation parallel with that direction.

Now I may remind you of the Figure 23 of my first lecture (see T. Bergeron, “Problems and Methods of Rainfall Investigation,” p. 28, this volume, 1960), where I underlined that two main patterns may be superimposed on most precipitation distributions, owing to two main mechanisms, the convective cell mechanism, and the stationary lee-wave mechanism. Figure 4 is an example of the Case 1, and Figure 5 of Case 2 with alternating maxima and minima. There is even a minimum occurring where you would expect a maximum because of the passage of the air over the table mountain and so on.

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**Fig. 4**—Instability pattern of rainfall (cold-front with SSE wind aloft) shown by a meso-scale network of Pluvius gages in a region with plains and plateau hills between the two great lakes of southern Sweden.
Another situation from the same year, October 19 (not shown here), also with a warm-front rain and south-southeast gradient wind, showed practically the same rainfall distribution. From these examples you may see what forecasters have to cope with! They should be able to tell that, in stie such a situation, there will be heavy precipitation (25–35 mm) during 12 hours near the shore of Lake Vättern, but only a few millimeters at the shore of Lake Vänern, that is, one fifth as much. This whole area is as big as the station ring on an American weather map, at the scale of one to ten millions. So, you see what we are up against; and you may take my word that these differences are even stationary with certain wind directions.

**Discussion**

*Mr. Jerome Namias*—You described the almost local character of regional differences in precipitation. If I remember correctly you had two very low precipitation areas and also rather high zones, the high being over a forest. You mentioned friction and some other factor as causes. I wonder if you could repeat your explanation.

*Dr. Tor Bergeron*—Well, first of all you noticed in those first cases that there was a maximum to the lee of Uppsala, and people hearing of or seeing these maps have suggested that nucleation from Uppsala might have something to do with it, and also convection raised by the heating. Uppsala is rather small, 70,000 inhabitants and some industry; therefore I think the main cause will be the friction. I think these great differences which are constant would probably correspond to different kinds of precipitation efficiency: that is to say, over the plains you will have rather great evaporation of the precipitation falling from the Nimbostratus, whereas over the forest regions you will have perhaps a precipitation efficiency of nearly 100 percent and then in addition also the water content of those local cloud caps. That might explain the very great differences in the amount...
of rain compared to these small differences in space. I may also say that we are now going to have a much denser network around Uppsala than the one I showed you.

*Dr. C. W. Newton*—The hills around Stockholm are about the same as those near Uppsala, 40 to 100 m in height. While we were in Stockholm, F. H. Ludlam made time lapse movies of cloud formations and we were very much impressed when we noted that Alto cumulus clouds were very strongly influenced by these 40-m hills. I wonder if, even with such low hills, ordinary orographic lifting can be ruled out.

*Dr. Bergeron*—I did not mean to rule out that kind of lifting which goes on the theory of lee waves, observable at heights surpassing that of the obstacles by a factor of 10 to 20, as we know. I did allow for that first of all, and secondly on an obstacle with forest, 40 m high, friction would perhaps increase the effective height. I did consider those two things, but I am very thankful for the remark.
Efficiency of Natural Rain

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Abstract—The efficiency of cloud in utilizing the water made available by the updraft for precipitation may be derived from a steady state equation involving storage and horizontal advection. The efficiency is greatest in the middle of widespread rain and least in individual thunderstorms.

The microscale features are evaluated in the initiation of precipitation. Observations indicate that large drop sizes, adequate for growth by the accretion process, are present in many clouds which do not precipitate. The important factor is the product of the mean effective liquid water content and the cloud depth, which has a critical value of about 3 g m\(^{-2}\) km.

Introduction—The efficiency of the rain mechanism is of importance not only for the evaluation of artificial rain production but also for the understanding of natural rain processes. An initial analysis was made by Weickmann [1958] who showed that the influence of particle concentration on the rate of rainfall is of secondary importance. The particle size adapts itself to the available concentration. The primary factor influencing rainfall rate is the updraft. Weickmann distinguished between a releaser cloud in which there is no storage of cloud liquid water and a spender cloud in which there is storage. In the former case the water vapor made available by the updraft is deposited directly on the precipitation and no artificial increase is possible. In the latter case an increase in particle concentration could cause a temporary increase in the rainfall rate at the expense of precipitation downstream.

In the initiation of rain, probability is a factor both on the meso and microscales. On the mesoscale, probability enters in the favoring by the convergence mechanism of a more vigorous or longer lasting updraft in one Cumulus cloud over another. On the microscale, the passage of a few large drops through regions of relatively high liquid content is a probability problem. Probability also enters in the appearance of freezing nuclei in sufficient concentration in cold clouds. The question arises as to the critical conditions on both scales for the initiation and maintenance of precipitation. It is assumed here that the macro or mesoscale features largely control the cloud extent and depth (although it has been claimed that the sudden release of latent heat in the freezing of the upper portion of a Cumulus cloud can cause precipitous growth). In this paper the critical cloud depth for initiating or maintaining rain is analyzed in conjunction with the microphysical features.

The important parameter in the production of precipitation from a cloud is the product of the mean liquid water content \(L\) and the effective cloud depth \(H\) which may be defined by

\[
HL = \int \frac{EVL}{V - w}
\]

where \(V\) is the fall velocity of the precipitation, \(w\) is the updraft and \(E\) is the efficiency of catch. The integration in this equation follows the ascent and descent of a drop within the cloud. According to the equation, for equal geometric depths, warm clouds should be more effective than cold clouds in producing rain because of higher \(L\). In addition the relation between the updraft and the fall speed of the precipitation has a marked effect. For updrafts of the same order as the fall speeds effective cloud depths may far exceed geometric cloud depths; this effect is important in the production of rain or hail from shower clouds as well as the production of snow from thin Stratus clouds.

Initiation of precipitation—In clouds over the tropical ocean, the percentage of warm rain increases from near zero for a cloud depth of 5000 ft to 100% for a cloud depth of about 10,000 ft [Battan and Braham, 1956]. At intermediate depths, wind shear and the humidity of the environment probably have considerable influence on cloud liquid water content (hereafter LWC) and thus the formation of rain. The observations suggest that under the most favorable circumstances a 5000-ft cloud depth provides sufficient LWC for large cloud drops to reach raindrop size.
Assuming an efficiency of catch of 0.8, a large cloud drop released at the cloud top could attain a diameter of 1.1 mm at the base of the cloud. Over the land, the critical cloud depths are about 10,000 ft in central United States and 12,500 ft in southwestern United States. With colder cloud temperatures the maximum mean LWC is about 2.5 g m\(^{-3}\) but evaporation caused by greater turbulent mixing with an environment of lower humidity could easily cause a 50% reduction in the effective LWC. Under such conditions a drop diameter of 1.5 mm could be attained at the base of the cloud. The important factor for rain formation is the product of the mean effective LWC and the cloud depth, which the observations indicate to have a critical value of about 3 g m\(^{-3}\) km. Cloud duration of about an hour is required in order for the drops to rise from the cloud base to top and then to descend again.

It has been observed over land and ocean that cloud drop sizes in Cumulus are somewhat greater in Cumulus clouds which subsequently develop echoes (rain) than those which do not [Battan and Reitan, 1957]. For example, at concentrations of 100 per liter the drop diameters in echo-producing clouds over the ocean was 67 \(\mu\) as compared to 64 \(\mu\) in non-echo producing clouds; over the land the respective drop sizes were 62 \(\mu\) and 58 \(\mu\). However, it may easily be shown that only several meters of fall in a cloud containing 1 g m\(^{-3}\) would suffice for the 64 \(\mu\) drops to grow to 67 \(\mu\) or for the 58 \(\mu\) drops to grow to 62 \(\mu\). It is apparent, therefore, that it is not for the lack of large drops that the non-echo clouds failed to produce rain. It appears more likely that mesoscale features controlling the duration and strength of the updraft are responsible for the favoring of one cloud over another in the subsequent development of rain. As for the occurrence of rain from clouds of smaller depth over the ocean than over the land, drop size again is evidently not the deciding factor since clouds which did not develop rain over the ocean had a greater concentration of large drops than clouds which did develop rain over the land.

The critical depth for the production of precipitation via the ice phase in Cumulus clouds is smaller than that using the water phase in warm clouds. There are numerous instances of snow over the Great Lakes from Cumulus clouds less than 5000 ft in depth. Evidence from the radar indicates the existence of snow from generating cells, 2000 to 4000 ft deep and wide, with vertical velocities of the order of 1 m sec\(^{-1}\). Many instances of precipitation involving the ice phase from a cloud with top warmer than \(-12^\circ\)C suggest that when the depth and duration of the updraft is sufficient, growth of precipitation somehow occurs either by the sublimation process initiated from freezing nuclei or by the accretion process followed by the freezing of drops when they are sufficiently large.

Drizzle has been observed to fall from layer clouds about 3000 ft in depth. The important factor in the production of drizzle is the duration of the cloud, or rather the duration of the updraft, so that drops aided by turbulent motion have a path length sufficient to accrete to a size of a few hundred microns.

The efficiency of rain—The efficiency of rain production \(E\) may be defined as the per cent of water produced by the updraft that falls to the surface

\[
E_R = \frac{\int q \text{d}y}{R} \tag{2}
\]

where \(\rho\) denotes the density of the air, \(q\) is the specific humidity and \(R\) the precipitation rate. Integration is from the top to the base of the cloud.

A steady state equation for the continuity of moisture in a cloud may be written as follows

\[
\frac{\partial R}{\partial z} + w \frac{\partial y}{\partial z} = w \frac{\partial L}{\partial z} + \left( u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} \right)
\]

\[
(L + NM) + NM \left( \frac{\partial^2 \rho + \partial y}{\partial z^2} + \frac{\partial}{\partial y} \right) \tag{3}
\]

where \(R = NM (V - w)\) is the precipitation rate, \(N\) the number of drops of mass \(M\). The second term on the left-hand side is the rate at which liquid water is provided by the updraft. On the right-hand side, the first term represents storage of cloud liquid water, the second is horizontal advection (which may include evaporation due to the mixing of the cloud with drier air), and the third represents changes in the concentration of the precipitation due to the vertical wind shear (from the equation of continuity). It is readily seen that a measure of the efficiency of rain defined by (2) can be derived from this equation.

In the middle of a widespread uniform rain, where the fall speeds of the precipitation particles are large compared to the updraft, all the terms on the right-hand side of (3) may be neglected except the first. In this case it is found [Wexler
and Atlas, 1958] that storage in the form of cloud represents less than 5% of the amount of water produced by the updraft, and that the precipitation efficiency is about 95% (neglecting evaporation from the upper surface of the cloud, which should be small). For light precipitation the entire cloud above about −5°C is a releaser cloud (no liquid water or icing), although for heavy rain, storage of cloud liquid water may extend to much higher levels.

In the opposite extreme is the thunderstorm in which the precipitation efficiency was found to be about 20% [Braham, 1952]; 45% of the available moisture was evaporated in downdrafts and 35% evaporated from the cloud sides or remained in the cloud after the cessation of rain. From this analysis it is evident that most of the loss of water is dynamic so that, once the precipitation has begun, little increase can be obtained by increased particle concentration; the greatest opportunity for artificial rain production is in the initial stages of the cloud when storage of cloud liquid water is high.

From the point of view of the efficiency of rain production per unit depth of cloud, stratiform clouds are relatively inefficient since a cloud deck some 20,000 ft thick gives a rain of a few mm per hour, while much higher rainfall rates come from Cumulus of smaller depths. Most efficient from this viewpoint is the case of a Cumulus cloud imbedded in a stratiform deck. An outstanding example is the frontal precipitation of October 1, 1958. Radar photographs showed that the widespread light rain on both sides of the front originated as snow with echo top near 24,000 ft. The heavy rain with a peak of about one inch per hour in a narrow region along the front had an echo top at about 10,000 ft (Fig. 1). Since the bright band was at 12,000 ft the growth of the heavy rain occurred entirely from the water phase. The occurrence of heavy rain over land from a convective cloud mass of less than 10,000 ft thick is unusual. In addition to the smaller loss of available liquid water by evaporation, it is reasonable to suppose that drops at least the size of drizzle from the surrounding light rain were carried into the base of the cloud by the strong updrafts, which from the rainfall amounts were calculated to be about 2 m sec⁻¹ over the extent of the cloud. The presence of such drops

Fig. 1—RHI's Blue Hill CPS-9 radar, October 1, 1958; the frontal rain is at a distance of about 5 miles.
would greatly accelerate the accretion process. The gap between the echoes of the upper light rain and the heavy rain in Figure 1 indicates that the drops from the light rain did not descend into the heavy rain cloud.

Conclusion—The efficiency of cloud in initiating rain increases with cloud depth, LWC, and duration. At a critical depth of about 5000 ft for warm Cumulus, without unfavorable wind shear and mixing with the environment, sufficient liquid water content is established for cloud drops to grow to raindrop size. It is doubtful that clouds fail to rain because of the lack of large drops because observation indicates that some clouds over the land produce rain while some clouds over the sea with more numerous larger drops fail to produce rain.

Critical depths and liquid water content for precipitation formation via the ice phase are smaller than in warm clouds. For the formation of drizzle in stratiform clouds critical depths are also smaller but critical durations are probably greater.

The efficiency of cloud in utilizing available water is greatest in the middle of widespread precipitation and least in the individual thunderstorm. Rain amounts in Cumulus are greater per unit depth of cloud than in stratiform clouds. Very efficient from both viewpoints is the case of a Cumulus cloud embedded in a stratiform deck.

The author is indebted to Helmut Weickmann for suggesting the topic and for many helpful suggestions, and to Edwin Kessler III for use of the October 1, 1958 radar data.

References


Discussion

Mr. R. D. Elliott—I feel that in treating any precipitation mechanism it is necessary to integrate over the entire mechanism, and when we consider a major storm, it is necessary to consider more than a vertical column. In extending the consideration to the outer boundaries of the storm it is necessary, to put two more terms into that equation: (1) The horizontal transport which may have a relatively small magnitude in the column, but when integrated over a storm, becomes important; and (2) An evaporation term which may be very small in the particular column where all the rain is falling, but out on the boundaries of the storm and on the forward edge and possibly the top, can be large. It would appear to me that in the studies of a large scale precipitation mechanism it would be necessary to sample over the entire storm and get a complete water budget.

Dr. R. Wexler—I quite agree with you. There is no doubt that the advection term is important. In the case treated previously, I was concerned with the stratiform case in which one could reasonably leave out the horizontal advection terms. In a thunderstorm or any precipitation from a cumuliform cloud, one cannot.

Dr. B. J. Mason—I would like to take up a point concerning the time and space requirements for the growth of a raindrop, which Dr. Wexler raised. Clouds over the tropical maritime ocean of little more than a kilometer thick may produce precipitation. Over England we get coalescence rain from a cloud perhaps five thousand feet deep. In both cases, if one tries to follow the history of a raindrop, one finds that the time for which the cloud exists is a very important parameter. Assuming a uniform updraft, it is quite impossible in the space available in the cloud and in the time available to get a raindrop out. This means one must have a much more complicated picture of the vertical motion, and one way to do this is by means of successive thermals. It may be the last thermal which is very important, and indeed my theoretical analysis shows that this must be the case, if one wants to get a shower out of a cloud only one kilometer deep. If you take the exact same parameters, but a homogeneous updraft speed, nearly two
kilometers of cloud depth is needed, and a little longer time, too. Even in this very simple case, the actual pattern of the motion is important. The motion is important, and the time span is important.

Dr. H. G. Houghton—I am sure Dr. Wexler agrees with this.

Dr. J. Smagorinsky—It would seem that the question of efficiency not only arises in the transformation from cloud particles to precipitation particles, but also enters into the formation of condensed particles from water vapor. I think one rarely finds that the entire mass is cooling at the moist adiabatic rate. If one is to consider the budget for the entire water content, that is water vapor, suspended liquid water, and precipitating water, the question of the efficiency should be considered in each transformation.

Dr. Horace R. Byers—I was very glad to see Dr. Wexler give a definition of efficiency. I think this is something that has been used in too loose a sense by many people. I hope that those of us who are called upon to review papers for journals and other publications insist that the authors very carefully define what they mean by efficiency and use terminology that will not further confuse the issue.

Dr. Helmut Weickmann—As I feel Dr. Wexler's lecture has a very important bearing on the problem of rain making, I may be permitted to elaborate a little on my own thoughts on this particular subject. Figure 2 shows another expression of the water-budget equation with the terms on the left side representing the water source, and those on the right representing the water sink. Of course, the growth processes, condensation and coalescence of the precipitation particles constitute the water sink; the updraft and water storage in the cloud constitute the water sources. Note that the number of the precipitation particles is a factor on the right side, therefore, increasing their number through seeding increases only the efficiency of the sink, but unless there exists water to be depleted on the left side, no increase in the rate of precipitation will occur. This may be illustrated by a simple model for the precipitation process which Louis Batten has called the 'Weickmann plumbing model.' Figure 3 gives the conditions in a releaser cloud. Here the particles grow through sublimation; humidity is at or near ice saturation, and the liquid or solid water content is negligible. The water which is released through the updraft diffuses to the snowflakes and, if there are many, each stays small; if there are a few, each grows large. This process is illustrated by a container which has a perforated base. Each hole represents a snow crystal and the water which flows into the container supplied by the updraft mechanism runs out through the holes. If the number of the holes is increased the droplets become smaller, but the rate of rain does not increase. If a raindrop recorder could be exposed to such a type of precipitation and we would plot the number N of precipitation particles falling within a certain small period of time, their median volume diameter D, and the rate of precipitation R, then the curves for N and D should be inverse proportionally since R does not change.

In the case of the spender cloud the conditions may, however, be different (Fig. 4). Here the

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**Fig. 2**—Water budget of precipitation

**Fig. 3**—Releaser cloud, \( W = 0 \)

**Fig. 4**—Spender cloud, \( W \neq 0 \)
flux of water may be sufficiently large so that a certain water content will be maintained. In this case the liquid water content $W$ of the cloud may be compared with the height $h$ of the water level in the model container. The flux through the holes on the bottom of this container is proportional to the height of the water level, just as the drainage of cloud water by the precipitation particles is proportional to the cloud water content. Here a certain equilibrium may be reached between the influx, outflux, and the height of the water level. This equilibrium will be disturbed if the number of outflux holes is increased. More water is drained until $h$ has reached a level where equilibrium is again established. If the number of holes suddenly decreases, then water accumulates. The flux through the remaining holes increases until finally again equilibrium, with lesser but larger drops, is reached. If the droplets during this type of rain are sampled, and their number, their median volume diameter, and the rate of rain are plotted, the curves indicated in the figure could be obtained. This time, however, the curves for the number and the rate of precipitation should be proportional.

The following conclusion can be drawn regarding the efficiency of the natural rain process from plotting these curves. In the first case considered, it can be said that nature is 100% efficient. Increase of the number of particles leads only to a decrease of their size, not to an increase of the rate. In the second case, however, increase of the particle number will cause a temporary increase of the rate of rainfall, which would indicate that nature's efficiency is insufficient for a complete drainage and that additional water could be drained artificially. This water then, of course, is not available for precipitation further downwind.
The Aerosol Spectrometer and Its Application to Nuclear Condensation Studies

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Abstract—The aerosol spectrometer (A.S.) separates quantitatively airborne particles in the diameter range \(3 \mu \text{m} - 0.03 \mu \text{m}\) from the atmosphere in the form of a size-spectrum, that is, a continuous band-shaped deposit. The position of a particle thereon is indicative of its 'Stokes' diameter' while it was airborne, and independent of physical changes incurred after its separation from the suspending air. This size-classified separation results from the exposure of a *laminar*, continuous air flow to a large centrifugal field (up to 26,000 \(g\)), the flow rates vary between 3.3 and 7.4 lit/min. The size (and mass) distribution of the aerosol is derived from the typical variation of the deposit density along the spectrum, either by microscopic count (down to 0.1 \(\mu\)) or by microphotometric recording of the light scattered by the particles under reflected dark-field illumination in a special micro-analyzer.

A brief description of the instruments and the mathematical basis of the analytical procedure is presented, also its application to a "model" aerosol of polystyrene latex, consisting of equal-sized particles in various states of agglomeration. From the size definition in terms of the Stokes' diameter, a relationship between the locus of deposition of dry and hydrated hygroscopic nuclei is derived and subsequently supported experimentally for NaCl aerosols.

The A.S. has been applied to the analysis of natural and artificial aerosols in the submicron range. Samples of natural (off-shore and mountain) aerosol spectra are presented; they follow in general the pattern determined by previous authors but show a fine structure which appears to be due to traces of organic matter. The artificial generation and conditioning of NaCl aerosols and the so resulting size distribution is described, particularly the strong effect of the presence of traces of organic vapors (turpentine, pinene) during the hydration and dehydration of the salt nuclei. It is apparent that such traces prevent or delay the equilibrium of the nucleus when the humidity of its gaseous environment is altered.

Introduction

Our knowledge of the true constitution of aerosol particles (that is, nuclei with or without condensates) in the airborne state depends upon the degree by which the process of their conversion into a concentrate alters their nature.

Separation by impingement or filtration expose the surface of the particle prior to and after its precipitation to the shearing action of the passing air stream. This shear may disturb seriously the dynamic equilibria of loosely sorbed or condensed components on the particle surface, whenever present. Hence minimal interference with the physico-chemical properties of the particles prior to their separation from the suspending air is as mandatory as a reliable size-classification in the diameter range \(3 \mu \text{m} \geq d \geq 0.03 \mu\text{m}\), in order to obtain information about the size and mass distribution of natural and artificial aerosols. This is significant for at least two reasons:

1. Micro-analytic information about the constituents of unclassified precipitates can be seriously misled, as the presence of a few 1-2 \(\mu\) particles can obscure the true nature of thousands in the 0.1-\(\mu\) class because of the cubic relation between particle mass and diameter.

2. The study of the size-distribution variation with temperature, pressure, and the concentration of gaseous components (\(\text{H}_2\text{O}, \text{CO}_2\), etc.), capable of association with such particles, promises to enlarge present knowledge of the reaction kinetics of such systems.

Of particular interest in this connection appears the role of organic components of the atmosphere, contributed by industry, plants, and conceivably by the biological components of the oceans, for they may influence significantly the dynamic equilibria of nuclear condensates, suspected 50 years ago [Barus, 1907] and by numerous investigators since [Vonneugut and others, 1957].
For the last four years the efforts of this laboratory have been directed toward devising a suitable method and realizing it with a practicable instrument—the aerosol spectrometer. The principle of its operation and the methods of analysis of the resulting size spectra are briefly described in the following because an up-to-date account has not yet been published in a conveniently accessible place. This account is followed by the application of this method to natural aerosol spectra in comparison with those obtained by laboratory methods.

**Principle and Design of the Aerosol Spectrometer**

The solution of this task is based on the principle of applying high centrifugal acceleration to a continuous laminar flow of the aerosol under conditions which cause the fallout of the particles to follow Stokes law [Goetz, 1956, 1957a] at so small a velocity relative to the air flow as to render the surface shear negligible. The conflicting conditions of a low Reynolds' number \( N_R \leq 800 \) (to assure laminar flow) and a high absolute flow velocity were met by leading the aerosol through a helical channel, the latter being part of a rapidly spinning rotor. Consequently the flow rate relative to the channel walls can be small and is independent of the velocity of the rotor.

The basic element of the aerosol spectrometer (A.S.) is thus a rotor, the periphery of which forms a groove covered by a flexible foil which seals the groove to an airtight channel. After operation the foil is removed from the rotor for the investigation of the particle deposit on it.

If the time of residence in the channel is sufficient for all suspended particles to reach the outer wall, that is, the inner foil surface, the deposit thereon represents a complete 'size-spectrum' of the aerosol tested, effected by the difference of radial velocity between the large and the small particles: The former are all deposited near the entrance, the latter toward the end of the channel.

The locus of the deposit of each size class is thus determined by the geometry of the channel, the flow rate and the angular velocity of the cone, that is, by factors independently controllable. Therefore the measurement of the deposit density along the channel yields the size (and mass) distribution of the aerosol tested, independent from its fate after deposition, because the locus of each particle was determined by its 'Stokes' diameter,' while still in the airborne state.

A considerable advantage of a spinning helix is its impeller action caused by momentum transfer from the channel walls upon the gas which causes a flow in downward direction along the channels. An auxiliary pump for moving the gas through the channels is thus unnecessary.

The flow speed through the channel is controlled independently by inserting flow restrictions (jets) of various sizes at the channel exit.

As the instrument in its present form is described in detail elsewhere [Goetz, 1957b, 1958, 1959], only a brief account of principal construction and performance is given:

It consists of two separate units (Fig. 1), one is the centrifuge, shown in Figure 2, the other, a chassis with the controls, connected with the instrument by cable.

The conical rotor \( R \) (Fig. 2) carries on its surface \( 2\pi \) turns of a double-helix and represents three of the four walls of two independent, identical channels of uniform (parallelogram) cross section. The channels are sealed by an exactly fitting conical cup \( C \) which is held airtight against \( R \) by the ring nut \( N \).

At the upper end the rotor is guided by the air-lubricated 'frictionless' bearing assembly, consisting of the constituent parts \( K, L, U \).

The entire rotor assembly is contained in a heavy, removable shell \( P \), which is sealed on top by the window \( Q \).

The electric motor \( M \) (1.4 hp, 110 V a.c.), mounted in the base \( B \), supports and drives the rotor by the shaft \( S \), also the armature \( T \) of an electric tachometer with a maximal speed of (26,000 rpm).

The chassis contains a variable auto-transformer unit for the speed regulation, voltmeter and ammeter (for checking the operating conditions of the motor), the tachometer, switch, fuses etc.

To minimize turbidity prior to entry, the air is sucked through the stationary inlet tube \( I \) into the channels at a flow rate \( F \) (lit/min) and passes the channels with a relative velocity \( v_0 \) (cm/sec), and a residence time \( \tau_0 \) (sec), determined by the orifice diameter \( O \) (mm) and the rotor speed \( N \) (rpm). Because of the rotor geometry the flow is subjected to a gradually increasing centrifugal acceleration starting at \( \gamma_{\text{min}} \) (gravity units) and ending at \( \gamma_{\text{max}} \) in the outlet. The flow rates in the channels are mutually independent, hence two different orifices can be used in one test. The
dimensions of the helix yield the following

\[
F = 2.75 \times 10^{-3}N
\]

\[
v_r = 12.9F = 3.55 \times 10^{-3}N
\]

\[
\tau_0 = 3.15/F
\]

\[
N_R = 73F
\]

\[
\gamma_{\text{min}} = 5.2 \times 10^{-2}N^2
\]

\[
\gamma_{\text{max}} = 1.52 \times 10^{-1}N^2
\]

The value of \( F \) can be determined at the inlet by a flow meter which may not impart a flow resistance, while the outflow \( W \) may have to be restricted to yield the same \( F \) value at the inlet, in order to compensate for dynamic pressure conditions between rotor and housing [Goetz, 1959].

The principle of the size classification of the airborne particles in the A.S. is based on the variation of the time \( \tau \), with the diameter \( d \) of the particles, required for their radial transport to the foil across the laminar air flow in the centrifugal field. If \( (L_d) \) is the distance from the channel entry along the outer helical channel surface, where no more particles of a size greater than \( d \) are deposited, it is obvious that \( (L_d) \) must increase with decreasing \( d \) for the same values of \( O \) and \( N \). Hence polydisperse aerosols with a smallest particle size \( d \) form a deposit along \( L \) which terminates for \( (L = L_d) \), provided that \( N \) and \( O \) are such that \( (L_d < L_o) \). Similarly it follows that for the same \( d \) value \( (L_d) \) decreases with increasing \( N \) or \( \tau_0 \), also that \( (L_d) \) indicates \( d \) if the function \( d(L_d) \) is known.

Figure 3 represents the function \( d(L_d) \) for a variety of \( N \) and \( O \) values, selected for the uniform coverage of the range of \( d \) and \( L \), as obtained from calibration with monodisperse polystyrene latex aerosols of eight different sizes: they represent straight lines in logarithmic coordinates, with a slope increasing with \( N \) and decreasing with \( O \).

The size ranges \((1.2 \mu d \geq 0.08 \mu)\), indicated by solid lines, were calibrated with latex aerosols, while the hatched sections indicate the \( d \) ranges for which no latex preparations are available.

Similarly uncertain is the validity of the linear relation for \( L_d \leq 4 \) cm because of the initial disturbance of the flow after entry into the channel and at the approach to the orifices, that is, for \( L_d > 38 \) cm.

This limits the range of \( d \) for the \( N \) and \( O \) values tested to \( d_{\text{max}} \leq 3 \mu \) for \( N = 6000, O = 1.5 \), and \( d_{\text{min}} \geq 0.03 \mu \) for \( N = 24,000; O = 0.5 \). The \( d \) range accessible to analysis embraces thus the factor of about 100, and the corresponding range of particle masses of about one million.

Table 1 correlates the values of \( F, v_r, \tau_0, \gamma_{\text{min}}, \gamma_{\text{max}}, d_{\text{min}}, \) and \( d_{\text{max}} \) with the corresponding values of \( N, O \) in Figure 3, for the selection of the optimal operating conditions \( N, O \) for the
specific test requirements. The largest sampling rate is obtained for $N = 18,000$; $O = 1.5$ which corresponds to a Reynolds' number of 710.

Figure 4a shows photographs of foil deposits resulting from latex aerosols of defined diameter obtained for two values of $O$; they indicate the sharp termination of the deposit at the respective $L_d$. 
Figure 3—Particle diameter (log $d$) versus corresponding deposit length (log $L_d$) for various operating conditions $N/0$ of the aerosol spectrometer

Figure 4b demonstrates the laminarity of the flow in the channels by the linear 'shadow,' that is, absence of particles in the deposit, 'downstream' from 1 or 2 inserted thin pins, protruding from the foil into the channel at P.

Analysis of Aerosol Spectra

The analytic procedure used for the evaluation of the aerosol spectrum on a foil varies with the type of required information. This may be relative, such as the size or mass distribution of the aerosol particles, or the distribution of certain chemical or radioactive constituents over the size range, represented by the spectrum; or it may be absolute, such as the numerical or mass concentration of aerosols.

The basic considerations and the method of analytic procedure, resulting therefrom, are discussed below.

The geometry of the foil deposits—Since the Stokes' diameter of a particle determines its locus of deposition, the geometrical relationships to which these deposits are subject represent the key to all analytic procedures. These relations are as follows.

The dimensions of the foil correspond to those
Table 1—Values of operational characteristics

<table>
<thead>
<tr>
<th>N</th>
<th>O</th>
<th>ON</th>
<th>F</th>
<th>(r_r)</th>
<th>(r_o)</th>
<th>(\gamma_{\text{min}})</th>
<th>(\gamma_{\text{max}})</th>
<th>(d_{\text{max}})</th>
<th>(d_{\text{min}})</th>
</tr>
</thead>
<tbody>
<tr>
<td>rpm</td>
<td>mm</td>
<td>cm/min</td>
<td>l/min</td>
<td>cm/sec</td>
<td>sec</td>
<td>g</td>
<td>g</td>
<td>µ</td>
<td>µ</td>
</tr>
<tr>
<td>6,000</td>
<td>1.5</td>
<td>900</td>
<td>2.48</td>
<td>32</td>
<td>1.27</td>
<td>520</td>
<td>1,530</td>
<td>2.7</td>
<td>0.5</td>
</tr>
<tr>
<td>12,000</td>
<td>1.5</td>
<td>180</td>
<td>4.95</td>
<td>61</td>
<td>0.64</td>
<td>2060</td>
<td>6,100</td>
<td>2.6</td>
<td>0.38</td>
</tr>
<tr>
<td>18,000</td>
<td>1.5</td>
<td>270</td>
<td>7.40</td>
<td>97</td>
<td>0.43</td>
<td>4670</td>
<td>13,750</td>
<td>2.5</td>
<td>0.30</td>
</tr>
<tr>
<td>24,000</td>
<td>1.5</td>
<td>360</td>
<td>9.86</td>
<td>132</td>
<td>0.28</td>
<td>7280</td>
<td>24,500</td>
<td>2.2</td>
<td>0.23</td>
</tr>
</tbody>
</table>

Fig. 4—Photographs (a) typical foil deposits from latex aerosols of defined particle diameter \(d\), indicating the values of \(L_d\), corresponding to the variation with \(O\) and \(d\); (b) the same showing shadow from inserted pins \(P\), protruding from the foil into the channel, indicating laminar flow.

of the outer envelope of the rotor; that is, the outer envelope represents the surface of the frustum of a cone of the (half) angle \(\alpha\) and the radii \(r_0', r_1'\) (see Fig. 5), and thus develops on a plane the sector of an annulus with the angle \(\omega = 360^\circ \sin \alpha\), and the radii \(r_0 = r_0' \csc \alpha\), and \(r_1 = r_1' \csc \alpha\).

Similarly the locus of that fraction of the foil surface is defined which has formed the outer walls of the helical channels. The development of
Fig. 5—(upper) Vertical cross section of rotor cone indicating channels; (lower) development of the outer channel wall (represented by foil) on a plane surface indicating the configuration of the borders of channels I and II (hatched) as Archimedean spirals of equal pitch. The circles represent loci of equal channel length $L$ in cm.
a conical helix of constant pitch \( \alpha \), on a plane results in a sequence of \( v \) sectors of an Archimedean spiral, if \( v \) designates the number of complete turns of the helix. The width \( b' = b \cot \alpha \) of the channel is thus the width of a spiral band on the foil; a double helix of the same width \( b' \) develops two identical spiral bands with corresponding points shifted by \( \omega / 2 \).

Geometrically the channel deposits are defined with relation to the foil boundaries in polar coordinates \((r, \varphi)\) as follows (Fig. 5).

The radius vector \( r \) originates at the intersection of the straight edges of the foil, while \( r = r_0 \) for the circle which coincides with the start of the channels, and \( \varphi = 0 \) (arbitrarily) defined by the intersection with the upper border of the left spiral channel I. The value of \( \varphi \) increases in the flow direction, that is, antilefwise. Hence the upper border of channel I is defined by

\[
\begin{align*}
  r_n &= r_0 + q \varphi \quad \text{for} \quad r_0 \leq r_n \leq r_1; \\
  0 &\leq \varphi \leq \nu \omega
\end{align*}
\]

the lower border by

\[
\begin{align*}
  r_0 &= r_0 + q(\varphi + \delta \varphi) = r_0 + q \varphi + b'; \\
  \delta \varphi &= b'/q
\end{align*}
\]

since \( \delta \varphi = b'/q \) for

\[
\begin{align*}
  r_0 &\leq r_L \leq r_1; \\
  (-\delta \varphi) &\leq \varphi \leq (\nu \omega - \delta \varphi)
\end{align*}
\]

Channel II is defined by (1a) and (1b), if \( \varphi \) is replaced by \( \varphi' = \varphi \pm \omega / 2 \). The pitch of the spirals \( q \) is related to that of the helices on the rotor \( \varphi \), by

\[
q = q_0 \omega = q_0 / (360^\circ \sin \alpha)
\]

Numerically these dimensions are \( r_0 = 7.4 \) cm, \( r_1 = 14.5 \) cm, \( \delta \varphi = 39.4^\circ \), \( \omega = 93.5^\circ \), \( q = 2.92 \times 10^{-2} \) cm/deg, or \( q = 1.67 \) cm/radian, \( b' = 1.15 \) cm.

The distance \( L \) from the channel entry \( r = r_0 \); \( \delta \varphi = \varphi = \varphi_0 \) of any point \( r \leq r_1 \); or \( \varphi \leq \nu \omega \) along the channel wall results from the rectification of the spiral which passes through this point in terms of \( r \) as

\[
L = \frac{1}{2} \left[ \frac{R}{q} + \frac{R}{q} \ln \left( 1 + \frac{R}{q} \right) \right]
\]

for \( R = \sqrt{q^2 + r^2} \) (2)

Due to the velocity distribution within the laminar flow the actual borders of the deposit deviates from this pattern, because of the effect of the boundary layer on the horizontal channel walls. This causes \( b' \) to decrease slightly with \( L \) (Fig. 4); hence \( q \) becomes somewhat larger for the upper, and smaller for the lower deposit border than the value derived from \( q_0 \). (This evidence of the existence of a gas layer at rest with the channel walls—and thereby above the foil surface—appears to be of prime significance because it indicates absence of any disturbance of the particles after deposit.)

The loci of all points of equal \( L \) values on the spiral bands are concentric circles, the radii \( r \) of which vary with \( L \) according to (2). This results in a simple method for determining approximate \( L \) values on a foil deposit as follows. A transparent sheet is used upon which a family of concentric circles, with radii corresponding to equal intervals of \( L \) and the outline of the foil itself, are imprinted; thus, the sheet can be superimposed upon the deposit (Fig. 6). The end of the particle spectrum in each channel will then coincide with a section of the circle, which represents \( L_q \), that is, the smallest particle size noticeably present in the deposit.

For the same reason, a foil, cut along several of these circles into concentric annular sectors, yields its deposit divided into equivalent size classes available for separate analysis.

**Determination of the distribution function**—The particle size (and mass) distribution \( C(d) \) and \( M(d) \) of the foil deposit is derived from the variation of the local deposit density \( Z \) along the channels, that is, \( Z(L) \). Hence the indicating device for \( Z \) for any \( L \) value must be able to scan the entire length of the deposit along the spiral pattern, and the reference area \( A_r \) over which the individual \( Z \) values are determined has to be small.

The direct determination of \( Z \) requires microscopic counting and thus optical systems resolving each individual particle. Micro-optics involving low-angle reflected dark-field illumination (such as the Ultrapak system of Leitz with mirror condensor and 20× to 50× objective magnification) permit the identification (though not necessarily the true resolution) of scattering objects down to less than 0.1 μ diameter, provided that the foil surface does not scatter appreciably. (This latter condition has been fulfilled by the use of thin, 0.1 mm, bronze foils, chromium plated and highly polished so that all polish marks were parallel, while the azimuth fraction of the illuminating beam which the marks would scatter was eliminated.)
Microscopic counting procedures by such means have as a lower limit the statistical uncertainty, depending on the field area counted; as an upper limit the densities where the particles start to overlap and are no longer distinguishable individually. Hence the tolerance for the variation of $Z$ is about 100 fold for the 1- to 0.1 μ-diameter range for $A_e = 10^{-4}$ cm$^2$.

In spite of its independence from size, shape, and optical properties of the individual particle (also from accidental contaminations) the substitution of the counting method by surface photometry is desirable for many applications, that is, the determination of the scattering intensity $S$ from a defined small area on the channel deposit. This method introduces three variables absent in the counts: (1) the scattering power of the surface (foil) supporting the particles; (2) the variation of $S$ with the particle size $d$; and (3) the scattering capacity specific for the particle type tested (refractive index, color, albedo, etc.).

Since these variables can be controlled adequately in many cases the photometric analysis of foil deposits proved practical and led to the construction of a micro-optical foil analyzer, described in detail elsewhere (Preining and others, 1959). It represents in principle a substantially enlarged stage of a stationary microscope on which the foil is moved so that the field $A_1$, as defined by the optical system, travels accurately along a pre-selected spiral track within the borders of the deposit. A specially designed ocular divides the light beam of the objective between a sensitive photocell and an eye piece, hence renders $A_e$ simultaneously accessible for selection, focusing, and control counts. The close correlation of $S(L)$ and $Z(L)$ is thus effected as long as the constitution of the deposit is not too inhomogeneous.

The derivation of the functions $C(d)$ and $A(d)$ of polydisperse aerosols requires the distribution of a strictly monodisperse aerosol along the channel, that is, $Z_d(L)$ for $0 < L < L_d$, to be known.

The experimental production of such spectra meets, however, with certain difficulties. Although deposits resulting from aerosols made by nebulizing highly diluted suspensions of monodisperse latex particles show a sharply defined $L_d$ value (due to a defined minimum particle size $d$), such deposits are far from being monodisperse throughout. The presence of larger particle sizes is apparent from the occurrence of one or more distinct discontinuities of the deposit density at one (or more) points $L_d' < L_d$. In other words,
the aerosol contains particles with larger, discrete Stokes’ diameters representing aggregates of (2, 3, or more) individual particles (Fig. 4a). Such aggregates can be eliminated by passing the test aerosol through a cascade of two A.S.’s and operating the first under conditions of N, O so that the smallest (doublet) aggregates are eliminated just before the channel end. The aerosol leaving is thus truly monodisperse as it enters the second A.S. which is operated under standard conditions of N, O. A deposit is produced herein which has the same Ld as that without passing the first A.S., but devoid of any deposit discontinuity for L < Ld. Such strictly monodisperse spectra, when analyzed by count or photometry, result in the distributions Z(L) and S(L) in Figure 7. The distribution is not identical because of different orifices in the channels (thus different values of Ld corresponding to d = 0.50 μ). The invariance of Zd with L is strikingly evident, while Sd, the photometer indication, shows fluctuations, probably because a few dust particles or local surface impurities which are disregarded by the counting process.

Since Zd, Sd can be assumed to be constant for L < Ld, the concentration C_d, that is, the relative frequency of occurrence of particles of size d, results as C_d = (ZL)_d b', where b', the average width of the deposit (b' = 1) represents a scale factor. For S_d' as the specific scattering power of the same particle size d follows: C_d = (S_L/S_d')d.

For polydisperse deposits these relations yield

\[ Z(d) = \int_0^d \frac{C(d)}{L(d)} \cdot d \]

\[ S(d) = \int_0^d \frac{C'(d) \cdot S'(d)}{L(d)} \cdot d + S_o \]  

(3)

where S_o is the level of background scattering.

The derivative of (3) results in the size-distribution function

\[ C(d) = L(d)\delta Z(d)/\delta d \]

\[ = L(d)S'(d)^{-1}\delta S(d)/\delta d \]  

(4)

A convenient procedure is first to plot the Z(L) or S(L) record as Z(d) or S(d) by means of the pertinent L(d) characteristic (Fig. 3) and to determine from these plots the differences as ΔZ_d or ΔS_d over equidistant intervals Δd, so that ΔZ_d ~ ΔZ_d/Δd ~ δZ(d)/δd or, other things being equal, for ΔS_d, ΔZ_d is then multiplied with the corresponding L_d, and ΔS_d with L_d/S_d'; the resulting C_d values are then plotted versus d.

The relative mass-size distribution M(d) results from C(d) because of M_d = C_d × d^2 for aerosols of fairly uniform specific gravity of their components. Hence the M scale on the ordinate results from multiplying each C value with the cube of the corresponding d value.

This procedure is demonstrated graphically for the same monodisperse latex aerosol as used previously in monodisperse form in Figure 8. The resulting size distribution—typical for the monodisperse aerosol—indicates the predominance of single particles to be expected for a suitable method of aerosol generation [Silverman, 1956], but also particles corresponding to the Stokes’ diameters of aggregates 2, 3, and more particles at d = 0.64 μ, 0.73 μ, 0.81 μ, etc. at gradually lesser defined sizes. This is also to be expected because the Stokes’ diameters of higher aggregates should for statistical reasons become increasingly indefinite with larger component numbers. The frequency of single particles results as 62% of the total number, 22% for double, 9% for triple, 5% for quadruple, and 2% for still larger agglomerates.

Definition of particle size—In the foregoing it has not been necessary to distinguish between the geometrical d and the Stokes’ diameters ơ of spherical particles because the specific gravity ρ is close to unity for the latex particles used to calibrate and define d. The difference between d, and ơ becomes significant when an aerosol contains components of different densities. This case warrants brief discussion: The Stokes-Cunningham law predicts the time τ, required for the fall...
Fig. 8—Graphie derivation of the frequency-size distribution C(d); (a) micro-analyzer diagram S(L); (b) converted function S(d) with tangents; (c) graphically differentiated function \( \frac{\partial S(d)}{\partial d} = C \times \frac{S'}{L} \), transformed by the reciprocal function \( S'(d) \) (right-hand scale) into \( C(d)/L \); (d) same function (lower hatched curve, left-hand scale) converted by \( L_d(d) \) (right-hand scale) into frequency-size distribution \( C(d) \) (left-hand scale, unhatched) and mass-size distribution \( M(d) \) (hatched upper curve, right-hand scale)
in the gravity field \( a \), through a gas of the viscosity \( \eta \) (when neglecting its density) as

\[
\tau \sim \frac{\eta}{a \cdot \rho \cdot d^2 \cdot K(d)} = \frac{\eta}{a \cdot d^2} \tag{5}
\]

where the Knudsen-Cunningham factor is represented in the simplified form as \( K(d) = 1 + \alpha/d \), and \( \alpha \) represents the product of the mean free path \((\sim 8 \times 10^{-6} \text{ cm}) \) of the suspending gas and an empirical constant varying between 0.7 and 1.2 [Wasser, 1933]. Hence two particle types of \( d_1 \neq d_2, \rho_1 \neq \rho_2 \) will form the same deposit pattern \( C(d) \), for \( \delta_1 = \delta_2 \) when

\[
d_1^3 \rho_1 \left(1 + \frac{\alpha}{d_1}\right) = d_2^3 \rho_2 \left(1 + \frac{\alpha}{d_2}\right) \tag{6}
\]

and their density ratio

\[
\rho_1/\rho_2 = (d_2/d_1)^2
\]

The particle diameter for \( \rho_2 \neq 1 \) can thus be converted in good approximation from \( d_1 \sim \rho_1 = 1 \) by

\[
d_2 = d_1 \frac{\left[1 + \frac{\alpha}{d_1}\right]}{P} \left[1 + \frac{\alpha}{d_1 + \alpha P}\right]^{1/2}
\]

or as

\[
d_2 = d_1/P, \quad (P = \sqrt{\rho_2/\rho_1}) \tag{7}
\]

This relation predicts also the variation of \( \delta \) for hygroscopic nuclei (for example, NaCl) because the ratios of the densities \( \rho_2/\rho_1 \) and of the geometric diameters before and after hydration can be calculated from the solubility \( S \) in \( \text{g/cm}^3 \) water of the nuclear substance and its density. (The term ‘hydration’ is used for the almost spontaneous transition from the practically water-free crystalline state of the NaCl particle to that of a saturated solution. This transition occurs when the relative humidity \( \text{rh} \) reaches a critical value \( \text{rh}_c \) above which the solid phase becomes unstable, partly because of the increase of solubility with decreasing particle size (Ostwald-Freundlich), partly because of the decrease of vapor pressure of the resulting solution (Raoult). The volume increase (decrease of curvature of the droplet surface), caused by hydration of the crystal, acts in the same direction (Kelvin). This threshold value \( \text{rh}_c \) should increase with increasing \( d \) and asymptotically reaches a maximum (for NaCl at about 76% for \( d \geq 0.7 \mu \) [Twomey, 1953; Orr, and others, 1958a]. Hence the coexistence of two phases (solute in equilibrium with its saturated solution) is not likely to exist for any length of time in small airborne nuclei of water-soluble solids. The hydration of such nuclei is not strictly reversible when the \( \text{rh} \) is reduced below \( \text{rh}_c \), that is, a spontaneous ‘dehydration’ may be delayed for a certain length of time, while the nuclei remain in a supersaturated state out of equilibrium with the vapor pressure of their gaseous environment.)

This results in

\[
d_L/d_S = \left[1 + \rho_S/S\right]^{1/2}
\]

and

\[
\rho_L/\rho_S = (1 + S)/(\rho_S + S) \tag{8a}
\]

Hence

\[
\delta_L/\delta_S = \left[d_L \cdot \rho_L (d_L + \alpha)/d_S \cdot \rho_S (d_S + \alpha)\right]^{1/2}
\]

or if

\[
|d_L/d_S \gg |\alpha| = (d_L/d_S) \sqrt{\rho_L/\rho_S} \tag{8b}
\]

For NaCl \( S = 2.64 \times 10^{-3}, \rho_S = 2.163 \) which results in \( d_L/d_S = 2.09, \rho_L/\rho_S = 5.2 \times 10^{-1}, \delta_L/\delta_S = 1.51 \) without \( K \) factor. Its inclusion renders \( \delta_L/\delta_S \) somewhat smaller depending on the absolute values of \( d_L \) or \( d_S \). For \( \alpha = \Delta \lambda = 0.8 \times 8 \times 10^{-6} \text{ cm}, \delta_L/\delta_S \) is 1.32 for \( d_L = 0.2 \mu \), and 1.47 for \( d_L = 1 \mu \). For most experimental work the complication involved by the inclusion of \( K(d) \) appears unnecessary; instead a corrected factor \( \delta_L/\delta_S \sim 1.43 \) for 0.2 \( \mu \leq \delta_L \leq 1 \mu \) range can be used. By this relationship the variation of a known \( C(d) \) of a salt aerosol can be predicted when the nuclei undergo hydration or dehydration.

**Natural Aerosols**

The application of the A.S. to natural aerosols provides an interesting comparison with the extended previous investigations on this subject [Junge, 1958] by a principally different method, as well as an extension into the submicron range. Only a few, though apparently typical, examples are presented in Figure 9 to demonstrate the results obtainable by this method.

The selection of the sites and conditions of these field samples aimed at atmospheric conditions minimizing the chance of industrial and other pollution. The offshore marine aerosols (Fig. 9ab) were taken from a boat, anchored,
Fig. 9—Size spectra of natural aerosols; (a) taken from a boat anchored in Santa Barbara Harbor; (b) taken from a boat drifting 40 mi offshore WSW of Santa Barbara Island; (c) taken at Mt. Wilson Observatory A in heavy haze, B in fog; (d) taken from a boat anchored at the south shore of Catalina Island, A count taken immediately after sampling, B count taken after a 48-hr storage (see Table 2).

sailing, or drifting, to avoid aerosol generation by the ship itself. The samples were taken about 12 ft above the water level.

The representative of a mountain aerosol (Fig. 9c) was taken from the dome of the 100-inch telescope at Mt. Wilson Observatory.

For field work the A.S., and its auxiliaries were driven by an 800-watt a. e. generator, the instrument was also equipped with a water-cooling system to maintain constant rotor temperature during the extended test periods (1 to 3 hr.).

The analysis (micro-analyzer) was made mostly by repeated independent counts $Z$ by two observers, since the deposit density was often not sufficient for the photometric analysis. $\Delta Z$ was interpolated for equidistant $\Delta d$ and $C(d)$ derived according to (3) and (4). Figure 9c is based on photometer data, the distribution is plotted in terms of $CS'$ and of $C$ by assuming the validity of $S'(d)$ derived for latex particles of equivalent sizes. Table 2 presents the pertinent test data.

The size distribution of the marine aerosols at low wind velocity and high humidity (Fig. 9d) is strikingly similar to the mountain aerosol prior to fog formation, it is characterized by a steep maximum for $0.22 \mu \leq d_s \leq 0.25 \mu$ (or $0.30 \mu \leq d_l \leq 0.33 \mu$), and by a decline of $C(d)$ toward larger $d$, somewhat steeper than the cubic.
Table 2—Test data pertinent to Fig. 9

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>Wind (Beaufort)</th>
<th>Weather</th>
<th>Time</th>
<th>Temp.</th>
<th>Rel. Humidity</th>
<th>N 10^6</th>
<th>O</th>
</tr>
</thead>
<tbody>
<tr>
<td>July 1, 1959</td>
<td>St. Barbara Harbor shore line</td>
<td>SSW-2 on shore</td>
<td>haze</td>
<td>09-10</td>
<td>87</td>
<td>55</td>
<td>18</td>
<td>1</td>
</tr>
<tr>
<td>Aug. 22, 1959</td>
<td>About 40 mi WSW of St. Barbara Island, kelp beds</td>
<td>SW-4</td>
<td>clear</td>
<td>16-18</td>
<td>25</td>
<td>18</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Nov. 12, 1958</td>
<td>Mt. Wilson Observatory (6000 ft), 100 ft above ground at tree top level</td>
<td>A NO-1</td>
<td>heavy haze</td>
<td>12-13</td>
<td>8</td>
<td>24</td>
<td>0.75</td>
<td></td>
</tr>
<tr>
<td>Aug. 25, 1959</td>
<td>S. shore Catalina Island</td>
<td>SSW 1-2</td>
<td>haze</td>
<td>08-10</td>
<td>22</td>
<td>18</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

relation indicated by the hatched curve. Quite surprising is here the practical absence of particles for \( \delta > 0.8 \mu \). A repeatedly observed exception is Figure 9b caused by the presence of a second maximum. If the first represents a dehydrated salt aerosol with a maximum at \( d_s = 0.21 \mu \pm 0.01 \), the second a corresponding hydrated fraction at \( d_h = 0.46 \mu \pm 0.01 \), \( d_h/d_s = 2.2 \pm 0.2 \) results, identical with the value 2.09 as derived from Stokes law for this ratio. This can be interpreted as the coexistence of a dehydrated and a hydrated fraction of the same salt aerosol, either because of its different ‘age’ or because of factors delaying the evaporation of the condensate.

Figure 9d represents a comparison of two complete counts of the same size spectrum, one (A) taken immediately after the exposure, the other (B) after a 48-hr storage in an air-tight container in the presence of silica gel as deisicent. The latter count is somewhat less in the range of small \( d \) although no evidence of coagulation was noticed. (The natural aerosol spectra taken in forests, desert, and the High Sierra (since this paper was written) confirm this finding without exception. It appears thus that a fraction of the aerosol, predominantly the smallest sizes, can evaporate at a larger or lesser rate without leaving a visible trace. This would indicate an aerosol fraction consisting of volatile substances with very low vapor pressure, that is, small compared with H2O.)

In all diagrams the ordinate scale is relative, however, the counts were normalized with respect to the sampling time, in order to render the curves comparable. The total particle concentration ranged between \( 10^2 \) and \( 10^3 \) per cm³.

### Artificial Aerosols

As initially indicated, the influence of atmospheric traces of gaseous organic materials upon the nuclear condensation (or dehydration) process presents an important problem about which little appears to be known at present; here the A.S. method could be expected to yield information.

The basic effect of such traces can be postulated approximately as follows: Provided their volatility is relatively low, these molecules are likely to become associated (adsorbed) with the nuclei. At high relative humidity they will form the surface, because for components of lesser surface tension this represents the configuration of least surface energy (Gibbs theorem), and most organic materials have surface tensions much smaller than water or electrolytes. If the vapor pressure of such a surface layer is smaller than that of the condensate (for example, of the saturated NaCl solution), the rate of evaporation will be less for a droplet carrying an organic surface layer than without such [Archer and La Mer, 1954], if the relative humidity decreases. This would result in the delay or prevention of equilibrium states, such as occur in persistent fogs and haze. If, on the other hand, the nuclei were dehydrated when contacted by the organic molecules, subsequent condensation, due to an increased relative humidity, may also be delayed. If this assumption is valid, at least in principle, aerosols of hydroscopic nuclei should produce different size spectra in the A.S., depending on the presence or absence of organic traces in the gas phase, if intercepted within a relatively short time after their exposure to a change of humidity. Such tests require arti-
Fig. 10—Scheme of generator for artificial aerosols

ficial aerosol production under carefully controlled conditions of dynamic stability. The assembly and procedure developed for this purpose is described in the following.

Generation of aerosols—The aerosol is generated from suspension or solution by the calibrated nebulizer N, supplied by branch I of a compressed air system, (Fig. 10). The air passes a dense cotton filter Fi and subsequently either the tube G1 or the bypass S1, depending upon the position of the valve pair V1. G1 is lined with coiled filter paper impregnated with organic liquids of low volatility; it acts thus as source for gaseous traces to the air flow. N is submerged in a constant temperature bath, the flow rate through N is controlled by the manometer M1.

Branch II supplies dry air, desiccated by passage through the drying tube D, prior to the filter F2, while III feeds moist air, generated by dispersing the filtered air stream through porous carbon plates as fine bubbles within a thermostated water column H. Hereafter II and III join another, are filtered by the membrane filter F4, and then pass the system G2, S2, V3, identical with that in branch I. F4 assures the absence of particles in the flow when it enters the peripheral Venturi orifice, which joins the channel sections C1 - C2. The adjustment of V2 and V3 permits an accurate control of volume and humidity of the air flow when joining the aerosol in the Venturi tube.

The channel C2 guides the combined flow to the chamber C3, which contains a sensitive hygrometer (membrane type) and wet-dry thermometers T.

The A.S. A is connected by a wide tee-fitting either at the inlet or the outlet (not shown in Fig. 10) of C3. The assembly thus produces a dynamically stable aerosol supply which can be adjusted to a relative humidity varying from supersaturation to less than 10%. It also facilitates the addition of traces of volatile components prior to, or after the generation of the aerosol.

At 0.5 atm. pressure differential the flow rate through N is 5 lit/min, dispersing 40 mg/min water at T = 0° C, that is, 0.8% of the fluid capacity of N. Hence the latter is sufficient to avoid significant variations of the concentration during operation for 5 to 10 min.

For a 5% NaCl solution (0.855 n), the yield of
AEROSOL SPECTROMETER AND ITS APPLICATION

NaCl aerosol amounts thus to about 1.9 mg/min, representing a salt concentration of 380 μg/lit at the flow of 5 lit/min in C1. Since the flow addition from II and III amounted to 20 lit/min, the aerosol concentration of about 76 μg/lit results in C2, C3. A one per cent monodisperse polystyrene latex suspension of, for example, 0.3 μ particles produces thus under the same conditions an aerosol density of about 10^8 particles/lit. The reaction rate of the aerosol can be estimated from the variation of the average residence times τ1, τ2 in C1, C2, effected by changing the air volume for the same flow rates. So far two sizes of C1, representing τ1 = 2 sec and 80 sec as well as of C2 for τ2 = 5 sec and 120 sec were used, causing the variation of τ1 + τ2 of 7 sec and 200 sec at the entry into the A.S.

The hydration state of the NaCl-aerosol in C1 can in this setup be varied by the temperature T_N of the nebulizer. The equilibrium in C1 for T_N = 0° C corresponded to 35% ≤ rh ≤ 45% and thus to a fairly dehydrated aerosol prior to contact with the air flow, while for T_N = 24° C the relative humidity was 60% to 70%, and should thus result in (supersaturated) hydrated NaCl-particles. Contact with the air flow in C2 can thus either hydrate or dehydrate the nuclei, depending upon the rh in C2 [Orr and others, 1958a].

Spectra of various NaCl aerosols—The selection of size spectra of artificial aerosols (Fig. 11) from more than a hundred tests is to be considered still of exploratory character. The aim of the particular test conditions was the study of the variation of the size distribution caused by contacting either a hydrated or dehydrated (but stable) aerosol with an air stream of varying humidity, and sampling it within a few seconds thereafter. (Such conditions should be indicative of the effect by organic traces in the air stream.)
upon the condensation process.) The residence times, \( \tau_1 = 80 \) sec and \( \tau_2 = 5 \) sec, were chosen, while the nebulizer temperature was either \( T_N = 0^\circ C \) to obtain a dehydrated, or \( T_N = 24^\circ C \) for a hydrated NaCl-aerosol in \( C_1 \). The NaCl concentration in \( N \) was 3.6% (0.615 \( \mu \)), for similarity to seawater (Pacific). The organic material in these tests was turpentine as representative of the type of compound yielded by plants into the atmosphere [Vonnegut and others, 1957], also pinene and n-butanol have been used with similar effects. Control tests conducted with double distilled H$_2$O in the nebulizer showed, even after extended exposure, no traceable deposit in the A.S. and thus proved that the combination of water and organic vapor does not produce an aerosol by itself.

The operating conditions of the A.S. were identical for all spectra in Figure 11 and the same as for the natural aerosols in Figure 9a,b,d, \( N = 18,000, O = 1 \) and embracing a range of \( 3 \mu \geq \delta \geq 0.17 \mu \). Carefully cleaned (xylene), polished chromium foils were used exclusively. Since the aerosol concentration was several hundred times larger than that encountered in nature the microphotometric analysis could be used although the exposure time was but one to two minutes.

The short residence in \( C_2 \) is the apparent reason for a certain instability of the size distribution during the brief sampling period. To compensate for these, each time two spectra were taken under identical conditions and analyzed for the same 32 \( L \) values, and then differentiated by determining (\( \Delta S \)) over intervals of \( \Delta d = 0.025 \mu \). The mean of each pair of \( \Delta S/\Delta d \) was then plotted versus \( d \). Adjacent points of this derivative curve were averaged again to minimize the unavoidable inaccuracies of this semi-graphical procedure. The systematic displacement of the \( \delta \)-value so caused is (\( \delta = 0.0125 \mu \)) and was considered negligible for this type of experiment.

Each size distribution curve in Figure 11 is thus based on 74 original \( S \) determinations and 31 points derived from them. Figure 11a presents four spectra of a dehydrated aerosol \( T_N = 0^\circ C \) after contact \( \tau_2 = 5 \) sec with an air stream of a relative humidity varying between 14 and 80%. The size distribution, typical for the nebulizer under the particular operating conditions used, is closely similar to those found in nature (Fig. 9) though for a maximum at a higher value (0.30 \( \mu \geq d_5 \geq 0.27 \mu \)).

At a relative humidity of about 50% the humidity is slightly larger than in \( C_1 \), hence any effect on the aerosol, while in \( C_2 \), is unlikely except for a possible small additional adsorption of moisture on the nuclei. The curve 0-50-0° shows a distinct maximum at 0.29 \( \mu \geq d_5 \geq 0.25 \mu \) and a steady decline toward larger sizes.

If the air flow is drier (0-33-0° and 0-14-0°) than in \( C_1 \) this maximum decreases with decreasing humidity in \( C_2 \) concurrent with a slight frequency increase (as \( CS \)) for \( d_5 > 0.35 \mu \). This can be tentatively interpreted as an increased rate of evaporation of the (numerically very dense) small particles into fewer conglomerates of larger sizes, due to electric disturbances, caused by the desorption of residual moisture layers known to exist on dehydrated nuclei [Orr, and others, 1958b]. (It was assumed as a first approximation that the original nuclei \( d_5 \) form conglomerates of closest packed spheres \( d_5 \), the number \( N \) required for forming a particle of the size \( d \) is \( (\pi/6)(d_6/d_5)^3 \). A decrease of, for instance, 30% of original particles of \( d_5 \) would thus result in a 7% increase of conglomerates for \( d_5 = 2d_5 \) and only in 1.4% for \( d_5 = 3d_5 \), etc.; qualitatively compatible with the experimental indication in Figure 11a.)

The distribution is drastically altered when the relative humidity in \( C_2 \) is larger than critical, that is, when the nuclei are being hydrated (0-50-0°). This causes a general shift to larger particle sizes. The interval \( \tau_2 \), for contact with the moisture of the air stream is insufficient for the complete hydration of all nuclei, indicated by the partial remainder of the first maximum, similar to the type of marine aerosols in Figure 9b.

Figure 11b shows the effect of partial dehydration of the hydrated aerosol (0-50-24°) by brief contact with an air stream of (rh \( \sim 50% \)) in comparison with the corresponding distribution curve (0-50-0°) in Figure 11a. As is to be expected, the dehydration, while far from being complete due to the short contact time, is sufficient to cause the shift of the maximum toward smaller \( \delta \) and the indication of a maximum remaining at the \( \delta \) value of the hydrated aerosol. The variations of the distribution do thus essentially agree with the hydration pattern for hygroscopic aerosols found by other authors previously for individual nuclei.

The curves of Figure 11def show the effect of organic traces \( T \) added in \( C_2 \) with the air stream in comparison with the distribution in their absence. Figure 11e represents the hydration pattern of a dehydrated aerosol (T-50-0°)
in the presence of \( T \), which latter retards significantly the hydration of the smaller particles (\( d_s < 0.35 \mu \)) as indicated by the partial preservation of the distribution of the unaltered dehydrated aerosol (Fig. 11a, 0-50-0°) while a second maximum at \( d_L \sim 0.6 \mu \) is indicated.

The curves of Figure 11de indicate the effect of \( T \) on dehydrated nuclei which is predictably negligible as the rh in \( C_1 \) remains virtually unchanged (Fig. 11e). However, a small increase of (rh) in \( C_2 \), that is, adsorption on the nuclei, appears to increase coagulation rate (decrease of small particles and small increase of larger sizes) for \( T-50-0° \), similar to the effect of desorption from the nuclei in the absence of \( T \) in Fig. 11a.

Figure 11f compares the dehyration of a hydrated aerosol with \( T \) and without (T-50 24°). Obviously \( T \) increases sharply the number of smaller particles when contacting an increased humidity below the hydration level (rh), contrary to its effect on an already dehydrated aerosol (T-50-0°). These preliminary results leave but little doubt about the marked influence of organic traces in nuclear condensation phenomena. A detailed interpretation of these findings, in spite of their fair reproducibility, appears premature and will be postponed until more detailed data are available.

**Conclusion**

The size-distribution spectra of natural and artificial aerosols in the submicron range appear realistic because the process of separation should interfere less with the airborne habitue of the particle than that of most methods available in the past. Moreover, the method permits discrimination between hydrated and dehydrated NaCl particles, because of the relation of the Stokes' diameters, independent from the state of the nuclei after precipitation. The size-distribution spectra can thus be used to determine changes occurring among airborne nuclei by interaction with their gaseous environment, such as adsorption or desorption, hydration and dehyrdation. It appears that, whenever such changes occur, the coagulation rate of the particles is affected and that organic traces in general delay these processes, possibly in analogy to the well-known 'inhibition' of freezing nuclei [Birstein, 1957; Poppoff and Sharp, 1959].

This hypothesis satisfies also the pattern of the marine aerosol spectra. They show two maxima \( d_s \), \( d_L \) whenever a source of their generation, such as foam on wave crests [Blanchard and Woodcock, 1957], was within a few miles from the sampling site, most distinctly in the vicinity of sources of organic matter (kelp beds). At substantially larger distance from such aerosol sources, the marine as well as the mountain aerosol spectra showed one maximum only. Natural aerosols are probably rarely, if ever, free from organic traces, hence the persistence of hydrated fractions of the nuclei present under conditions of decreasing humidity could well be caused by such temporary protection against dehydration and vice versa. One has to realize that, contrary to such dispersions in the laboratory, these aerosols represent mixtures of particles airborne for largely differing time intervals available to them for attaining equilibrium states prior to precipitation in the A.S.

Independent of the validity of the above interpretation the aerosol spectra, so far available, seem to definitely indicate the significance of gaseous organic traces in the atmosphere with regard to the rate of nuclear condensation, that is, to the formation of fogs and hazes as well as to certain phases of air pollution. It thus appears that a systematic research effort which combines size-distribution studies of natural aerosols with their well defined synthetic replicas in the laboratory promises a new chapter in the understanding of our immediate atmospheric environment.

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GOETZ AND PRENING

References


Discussion

Dr. C. E. Junge—As I understand you, you evaluate the particle concentration using the microphotometer by which you measured the total scattering?

Dr. A. Goetz—We determine the surface scattering microphotometrically each time over an area of $6 \times 10^{-3}$ cm$^2$ progressively along the spiral of the deposit pattern.

Dr. Junge—In other words, you evaluate the particle concentration on the screen optically?

Dr. Goetz—Yes, that is for the artificial aerosols I described.

Dr. Junge—I would like to caution a little bit. If the same aerosol is deposited for instance in the absence and then in the presence of organic vapors, the aerosol droplets may spread on the foil surface with different contact angles. This will influence the light scattering characteristics of the individual particles which in turn would influence the analysis. The question of particle spreading is a complex one and I wonder how you took that into account.

Dr. Goetz—in the micro-analyzer which I have briefly described individual counts can be taken over an area of $10^{-4}$ cm$^2$ along the deposit with reference to a reticule in the eye piece. Deposits which are dense enough to allow reliable scattering determination are too crowded for counting, hence it is in general not possible to evaluate the same deposit in both ways. For comparison a brief and long exposure of the same aerosol are taken. This has so far been applied to a few cases only for the artificial NaCl aerosols with and without organic trace. Hence their general quantitative interpretation in terms of numbers is not yet possible. These comparisons indicated that the presence of the organic trace definitely changed the numerical size distribution as indicated qualitatively by the scattering, however, it may also have affected the specific scattering
power of the particles. The detailed relationship will require much future work.

Dr. Bernard Vonnegut—I would like to ask about the pressure drop across the centrifuge. The thought occurs to me that because of these spirals, the air that is taken through the centrifuge is experiencing a pressure change.

Dr. Goetz—The average pressure in the channels should not be different from the atmospheric because the flow restriction (the locus of the pressure drop) is at the exit of the channel. The air in the channel will, however, experience a slight compression because of the impeller action of the spinning helix prior to expansion in the jet when leaving the channel. Tests under stationary conditions indicated a maximal pressure difference of 5 cm H2O, that is, 0.5% of one atmosphere. There should exist also a radial pressure gradient, analogous to a barometric variation of the pressure between the inner and the outer channel walls. Because of the small radial depth of the channels (0.63 cm) this pressure difference should amount at 25,000 g to about 2% if the centrifugal force were constant across the channels. Since it is less at the inner than at the outer wall (about 40% at inlet decreasing to 16% at the outlet due to the conical rotor shape) this radial pressure difference should be even smaller.

Dr. Vonnegut—How nearly adiabatic is the expansion? A change of pressure of this sort would amount to a very substantial change in the temperature, were it adiabatic. If this were so, the humidity might be considerably changed.

Dr. Goetz—Because of the velocity distribution across a laminar flow a substantial fraction of the particle trajectories will pass a slow moving, almost resting gas layer prior to deposition on the outer wall. This layer must be in thermal equilibrium with the rotor. Hence we believe that the minute radial compression would be closely isothermal rather than adiabatic with respect to the equilibrium conditions of the particles.
Differences in Coalescence Tendencies in Computed Condensation Cloud Droplet Spectra

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Abstract—Four numerically computed cases of the growth of a population of cloud droplets by condensation are analyzed in terms of coalescence tendencies. Following the suggestions of Hitschfeld, 1957; Telford, 1955; and Welander, 1959, particular attention has been given to the stochastic influence on the coalescence-produced cloud-droplet spectrum. The evidence taken from the numerically computed cases strongly supports the contention that collisions between small droplets readily produce a wider spectrum than the condensation-produced spectrum. The transition time for a condensation-produced spectrum to change to a predominantly coalescence-produced spectrum depends strongly on the maximum supersaturation and resulting cloud droplet concentration produced at cloud base.

In a recently published paper [Mordy, 1959] the writer described a theoretical investigation on the effects of vertical motion and condensation nucleus spectra on cloud droplet spectra. The purpose of the present paper is to evaluate the differences in cloud droplet spectra which resulted in these cases in terms of some important factors which lead to the subsequent changing of these spectra by the coalescence of droplets.

The coalescence of droplets is mainly dependent on three factors. These factors are the size spectrum, the concentration, and the coalescence (collision times collection) efficiencies of the droplets. An equation which gives the number of contacts between two droplet sizes is frequently given as [for example, Telford, 1955]

\[ n = E N_1 N_2 \pi (r_1 + r_2)^2 \rho \]  

(1)

where

- \( E \) = collision efficiency between \( r_1 \) and \( r_2 \) radius droplets
- \( N_1 \) and \( N_2 \) are the concentrations per unit vol of the \( r_1 \) and \( r_2 \) size droplets respectively
- \( \pi (r_1 + r_2)^2 \) = the area surrounding a droplet in which a contact could be made
- \( \rho \) = relative velocity between the \( r_1 \) and \( r_2 \) droplets

Customarily the coalescence of droplets has been described by following the growth of a larger droplet as it sweeps up smaller droplets in a cloud with certain assumed characteristics of droplet size and vertical velocity [Houghton, 1950; Mason, 1957; Bowen, 1959; Liddam, 1951; Langmuir, 1948]. The growth of the larger droplet in these cases is assumed to be continuous, that is

\[ \frac{1}{\rho} \frac{dm}{dt} = 4\pi \rho^2 \frac{dr}{dt} = E \pi N_2 (r_1 + r_2)^2 \rho \frac{4\pi r^2}{3} \]  

(2)

where \( r_1 \) and \( r_2 \) are the radii of the collecting and collected droplets respectively. In fact of course it is discontinuous, each unit of growth being one cloud droplet.

If one adopts the practice of thinking of the coalescence of cloud droplets as in (2) above, then it is the initial drop size spectrum which largely determines the initial and subsequent relative velocities between the droplets. Thus droplets which are initially slightly larger, will remain larger throughout the coalescence process and there will be no tendency for smaller droplets to overtake them, except that they may follow different trajectories in the cloud [Bowen, 1950].

Such reasoning has been followed in the work of Houghton [1950], Woodcock [1952], Fournier d'Albe [1955], and others. In this line of thought Woodcock has reported measuring a correspondence between the size distributions of sea salt particles and raindrops measured in the same localities and at the same time.

As exceptions to the above treatment of this problem there have been three very interesting investigations to date [Telford, 1955; Hitschfeld, 1957; Welander, 1959], which study the importance of statistical aspects of the coalescence process.

This statistical process in its simplest form may
be illustrated by the following example. Suppose a group of 100 large droplets is falling through a cloud of randomly spaced, uniform size, small cloud droplets. Because of the random spacing of the cloud droplets, each of the large droplets will have a different collision experience with the cloud droplets and after some time will represent not a uniform-size group but a spectrum of droplets located at different levels in the cloud.

If such reasoning is applied to an initial spectrum of cloud droplet sizes instead of an initially uniform-size group, a question arises as to whether, after some time has elapsed, the resulting drop size spectrum is more dependent on this statistical process than on the initial drop size spectrum. In other words one may ask the question; do more of the large droplets present represent initially large nucleus droplets or do collisions between the smaller sizes of droplets ultimately produce more large droplets?

A complete mathematical investigation of this problem has not been made but the evidence offered by the three investigators above indicates that the statistical nature of the problem is very important.

One way of approaching this question is to examine the differences in drop sizes and drop concentrations which may result from condensation with a view toward assessing their coalescence tendencies, that is, how will $n$ (the number of first collisions) change in (1) as $r$ grows by condensation? It seems that the assumptions which are necessary to make such an appraisal are not too gross to prevent an examination of important aspects of this process.

As a model for such investigation, assume a constantly rising current of air containing a given frequency distribution of salt particles. The condensation conditions attending are then fixed by the initial conditions such as temperature, pressure, rate of rise, and the above assumptions such that, until the process of coalescence becomes important, we can describe the cloud droplet spectrum at each time and altitude above base [Mordy, 1959]. We shall then investigate, as time elapses, the characteristics of these changing spectra which are conducive to coalescence.

We shall therefore look first at the way $n$ varies with time or height if the growth of the particles is controlled only by the condensation conditions.

The fall velocity of the drops we consider is satisfactorily described by Stokes law (that is, for $r < 40\mu$) so that

$$v = \frac{2gr^2}{g} = kr^4 (k = \text{constant})$$

If the above expression is substituted in (1) and the terms rearranged the equation can be written

$$n = 2\pi N_1 N_2 dr (r_1 + r_2)^2 (r_1 - r_2)$$

(3)

Only that part of the condensation process that occurs above the zone of maximum supersaturation (where $N_1$ and $N_2$ are determined in this model) need be considered here so that $\pi N_1 N_2 dr$ may be considered as a constant. The variation of $E$ is uncertain and therefore for the moment will be included in a new variable $n' = n/E$.

If the ln $n'$ is differentiated with respect to time the equation becomes

$$\frac{1}{n'} \frac{dn'}{dt} = \frac{3}{r_1 + r_2} \frac{d(r_1 + r_2)}{dt}$$

$$+ \frac{1}{r_1 + r_2} \frac{d(r_1 - r_2)}{dt}$$

(4)

The time derivatives of the radii can now be obtained from the equation for drop growth by condensation. Here we shall for the moment consider only droplets which have grown until they are dilute enough to neglect the hygroscopic effect of the nucleus. We shall see in the discussion below that this makes very little difference in the conclusions we draw from our results. If $a$ is the ratio of the radius $r_1$ to $r_2$ or $r_1 = ar_2 (= a\bar{r},$ see (6b) below) the drop growth equations for the two drop sizes which appear in equation (4) can be written [Mordy, 1959]

$$\frac{dr_1}{dt} = \frac{a}{\bar{r}^2} (as\bar{r} - A)$$

$$\frac{dr_2}{dt} = \frac{a}{\bar{r}^2} (s\bar{r} - A)$$

(5)

where

$$s = (c - c_1)/c_1 = \text{supersaturation}$$

$$a = \frac{\alpha L D J e}{RT^4} \left[ \frac{1}{1 + \frac{\alpha L^2 D J e}{kRT^4}} \right]$$

$$= \text{const} \approx 8 \times 10^{-7} \text{ cgs at 800mb, 10°} \text{ C}$$

$$A = 2\sigma T/JL$$

= capillarity term coefficient $= 1.7 \times 10^{-8} \text{ cgs}$

$$T = \text{temp of drop}$$
\[ \sigma = \text{surface tension} \]
\[ J = \text{mech equiv. of heat} \]
\[ L = \text{latent heat of water} \]
\[ \rho = \text{density of water} \]
\[ k = \text{heat diffusion coefficient} \]
\[ \epsilon = \text{vapor pressure} \]
\[ \epsilon_s = \text{saturation vapor pressure} \]

When (5) is substituted in (4) the equation becomes
\[ \frac{1}{n^t} \frac{dn^t}{dt} = \frac{2a}{3^t} a \left[ 3^r - A + \frac{A(2a - 1)}{a(a + 1)} \right] \]  (6)

The objective now is to see how the terms in (6) vary as time proceeds. In the writer’s previous study it was shown how the bulk of the liquid water lies in the smallest one or two categories of cloud droplets in such a model (see Fig. 1), hence the mean volume radius will represent nearly the radius of droplets which contain the largest part of the liquid water. It will serve as a good index, therefore, for this discussion to consider collisions between this volume mean radius (\( \bar{r} \)) and larger or smaller droplets in the spectrum. The volume mean radius is defined by
\[ \frac{4}{3} \pi \bar{r}^3 \rho = \bar{m} \]  (6a)

where \( \bar{m} \) is mean mass of the droplets.

In the model we are considering, the liquid water is very nearly equal to a constant times the height above cloud base, which is to say
\[ N = \frac{4}{3} \pi \bar{r}^3 \rho = \frac{dx}{dz} \cdot z = \epsilon z \]  (7)

where \( dx/dz \) = the change in mixing ratio in a moist adiabatic expansion (~2 \times 10^{-5} cgs at 800mb and 10°C)
\[ z = \text{height above cloud base} \]
\[ N = \Sigma (N_1 + N_2 + \cdots) = \text{const.} \]

By differentiating, (7) can be written
\[ \frac{d\bar{m}}{dt} = 4\pi \bar{r}^3 \frac{d\bar{r}}{dt} = 4\pi \alpha (3\bar{r} - A) \rho = \frac{c_v}{N} \]  (8)

where \( z = \epsilon \bar{r} \) in the model.

The term containing \((3\bar{r} - A)\) here has been substituted from the condensation growth equation (5). If these expressions in (7) and (8) are used to replace \( \bar{r} \) and \((3\bar{r} - A)\) in (6) the equation becomes:
\[ \frac{1}{n^t} \frac{dn^t}{dt} = \left( \frac{2}{3a} \right) \left( \frac{1}{t} \right) \left[ \frac{A(2a - 1)}{a(a + 1)} \right] \]  (9)

when it is assumed that \( \rho = 1 \).

If the two equations for drop growth (5) are subtracted the following equation may be obtained in quite an analogous way
\[ \frac{1}{r_1 - \bar{r}} \frac{d(r_1 - \bar{r})}{dt} = \frac{1}{3at} \left[ 1 - \frac{A4\pi \alpha N}{ace} \right] \]  (10)

Now putting
\[ y = \frac{c_v}{4\pi \alpha N} \]

and
\[ \Delta r = r_1 - \bar{r} \]
and combining (9) and (10) the following equation is obtained.

\[ \frac{1}{n'} \frac{dn'}{dt} = -2 \frac{d\Delta r}{\Delta r} \left[ 1 + \frac{3}{1 + \frac{1}{a}} (ay - 1) \right] \]  
(11)

Note: \( c = \frac{dx}{dz} \) (see Eq. 7); \( \dot{a} \) is defined in (5).

Values of \( y \) are given in Table 1 for cases from the writer's previous condensation study. Choosing four representative values of \( y \) and computing the value of the coefficient in (11) for all radii larger than one half the mean radius one can see that a good approximation for this quantity is \( y' = 2 + \frac{6}{y} \) for the given values of \( y \). Hence the time dependence of \( a \) can be rather safely neglected in integrating (11).

Putting

\[ y' = 2 \left[ 1 + \frac{3}{1 + \frac{1}{a}} (ay - 1) \right] \approx 2 \]

when \( 60 > y > 4 \), (11) may be integrated to give

\[ n' = n_0 \left( \frac{\Delta n}{\Delta r} \right) \left( \frac{\Delta r}{\Delta r} \right)^{y'} 2 < y' < 2.75 \]  
(12)

Eq. (12) indicates that if \( \Delta r \) remains \( < \Delta n/2 \), \( n' \) will be only from 4 to 16 times greater than \( n_0 \).

It is useful now to examine the variations in \( \Delta r \) which occur as a result of condensation. Eq. (10) shows that as long as \( 1 \gg 1/\dot{a}y \) that \( \Delta n/\Delta r \) varies approximately as \( t^{-(1/3a)} \). In this range all values of \( \Delta r \) are converging. By the time that 300 sec have elapsed \( \Delta r \) has a time constant

\[ T = \left( \frac{1}{\Delta r \frac{d\Delta r}{dt}} \right)^{-1} \]

of more than 1000 sec. In fact when one includes the effect of the salt nuclei in the drops this time constant is appreciably lengthened so that \( \Delta r \) may be treated as nearly constant in the time interval of 1000 sec or so which we consider here. If variations of two in the values of \( \Delta n/\Delta r \) are allowed for, an error in the estimation of \( n' \) by a factor as large as 16 may occur. We shall see, however, that this variation is small compared to the several orders of magnitude variation in \( n' \) computed in the cases of different droplet spectra.

Eq. (12) shows effectively that the most important factors \( (n_0, y) \) which determine \( n' \) are determined very near the cloud base and are influenced by the strength of the vertical current and the nucleus distribution on which the droplets form.

If \( \Delta r \) is considered a constant then with the aid of (7), \( \dot{r} \) is known and the spectrum of condensation produced droplets can be estimated at each altitude (or liquid water content). Once the spectrum is determined an estimate can be made from Eq. (3) of the number of 'potential' first collisions \( (n' = n/E) \) per unit time between any size category and the mean radius.

Such calculations of \( n' \) are given in Tables 2-5 for four cases from the condensation study. The drop size distributions for these cases are shown in Figure 2. The calculations were made for a cloud liquid water content of 1 g/m³ which by (8)
Table 4—Case I, 100 cm/sec

<table>
<thead>
<tr>
<th>N</th>
<th>r</th>
<th>u'</th>
<th>a</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.28 \times 10^7</td>
<td>21.2</td>
<td>2710</td>
<td>0.954</td>
</tr>
<tr>
<td>6.4 \times 10^7</td>
<td>22.2</td>
<td>1950</td>
<td>0.946</td>
</tr>
<tr>
<td>3.2 \times 10^8</td>
<td>22.4</td>
<td>1195</td>
<td>0.933</td>
</tr>
<tr>
<td>1.6 \times 10^8</td>
<td>22.7</td>
<td>730</td>
<td>0.912</td>
</tr>
<tr>
<td>8 \times 10^5</td>
<td>23.2</td>
<td>436</td>
<td>0.897</td>
</tr>
<tr>
<td>4 \times 10^5</td>
<td>23.6</td>
<td>285</td>
<td>0.875</td>
</tr>
<tr>
<td>2 \times 10^5</td>
<td>21.2</td>
<td>203</td>
<td>0.840</td>
</tr>
<tr>
<td>10^5</td>
<td>25.2</td>
<td>134</td>
<td>0.809</td>
</tr>
<tr>
<td>5 \times 10^4</td>
<td>26.2</td>
<td>70</td>
<td>0.790</td>
</tr>
<tr>
<td>2 \times 10^4</td>
<td>27.3</td>
<td>18</td>
<td>0.708</td>
</tr>
<tr>
<td>3 \times 10^3</td>
<td>29.9</td>
<td>3</td>
<td>0.628</td>
</tr>
<tr>
<td>3 \times 10^2</td>
<td>33.7</td>
<td>1.8</td>
<td>0.548</td>
</tr>
</tbody>
</table>

Table 5—Case III, 5 cm/sec

<table>
<thead>
<tr>
<th>N</th>
<th>r</th>
<th>u</th>
<th>a</th>
</tr>
</thead>
<tbody>
<tr>
<td>10^7</td>
<td>24.6</td>
<td>232</td>
<td>0.943</td>
</tr>
<tr>
<td>3 \times 10^8</td>
<td>26.1</td>
<td>133</td>
<td>0.910</td>
</tr>
<tr>
<td>10^8</td>
<td>27.0</td>
<td>175</td>
<td>0.878</td>
</tr>
<tr>
<td>9 \times 10^3</td>
<td>28.0</td>
<td>197</td>
<td>0.842</td>
</tr>
<tr>
<td>7 \times 10^3</td>
<td>29.2</td>
<td>199</td>
<td>0.803</td>
</tr>
<tr>
<td>5 \times 10^3</td>
<td>30.6</td>
<td>192</td>
<td>0.743</td>
</tr>
<tr>
<td>3 \times 10^3</td>
<td>33.1</td>
<td>192</td>
<td></td>
</tr>
</tbody>
</table>

determines \( \bar{r} \). Because of the very skew nature of the droplet distribution the mean radius and the minimum radius were assumed to be the same, an assumption which could lead to an error of a few per cent in the estimate of \( u' \) at most. The four cases chosen were representative of the range of differences in nucleus spectra which resulted from different nucleus distributions and vertical velocities.

In all of these cases except one there is a markedly higher number of potential collisions among the smallest cloud droplets. The exception is the case where there is an extremely slow rate of rise, 5 cm/sec, and a dense nucleus distribution (which was taken from Woodcock's data as characteristic of the distributions of salt particles produced at cloud base by a Beaufort Force 7 wind at the sea surface).

The fact that there are a higher number of potential collisions between small droplets, however, is not enough to determine whether collisions between the smallest droplets are of comparable importance to those between the larger and smaller droplets. It must be determined whether in a reasonable time a comparable number of larger droplets are originating by combining smaller droplets as from condensation on large salt par-

![Fig. 2](image-url)

Fig. 2—The distribution of cloud droplet sizes used in the computations; these distributions are derived from Mordy [1959] and represent the extreme differences in droplet concentration and spectrum width obtained in the different assumed cases of nucleus spectra and vertical velocities; the spectra represent the sizes of droplets expected if the liquid water content is \( g/m^2 \) and is assumed constant (see text); the figures at the top give the distribution type and the vertical velocity in cm/sec.
those with a radius \( r = 2^{1/2} r = 1.26r \) in periods ranging from less than one second for the distribution Case I at 100 cm/sec to more than 1000 sec in the distribution Case III at 5 cm/sec. These values for the four cases are given in Table 6.

The values for \( T \) in Table 6 only give a comparison between the two smallest categories of droplets and the number of existing, condensation-produced droplets which have a radius roughly equal to that of the droplets produced by combining the two small droplets. Actually it would be better perhaps to compare all coalescences between the smallest categories having a radius not more than about ten per cent larger than the minimum value with the number of condensation produced large droplets. If this is done, values of \( T \) are decreased by a factor of 3 to 5.

If these computations are made for a larger liquid water content then the collisions between the small droplets become relatively more important. To see how this proceeds a useful diagram can be constructed as in Figure 3. Here the
Terms in (3) have been expressed as

\[ \ln n' = \frac{\ln n}{E} = \ln \pi N_1 N_2 \Delta r + 3 \ln (2r + \Delta r) \]  

(13)

where \( r_1 = \bar{r} \),

\[ r_2 = \bar{r} + \Delta r \]

\[ \Delta r = \text{const} \]

Since for any pair of droplet sizes \( N_1, N_2 \) is considered constant in the model, then the variation of \( \ln n' \) depends only on \( n_0' \) and \( \bar{r} \). Hence if one plots the number of collisions between two categories of droplets as given in Tables 2 to 5 it is possible to extrapolate how the relative number of collisions change with time or with different liquid water contents. As \( \Delta r/\bar{r} \) becomes small the points representing the number of collisions tend to line up more vertically as the points move along the diagonal lines (curve \( AB \) becomes curve \( A'B' \)) showing that as time elapses the relative importance of the small droplet collisions increases. The diagram also shows how important the initial droplet spectra radii are in determining the relative importance of large and small droplets, for these show up as the separation of points in the curves in the diagram. When satisfactory information about \( E \) is available as a plot of \( \ln E \) vs. \( \ln (2r + \Delta r) \) and \( \Delta r \), this will allow a more complete diagrammatic treatment.

The calculations and reasoning in this paper have been made with the assumption that the collection efficiency of the droplet was 100%. Research and theory on the collection efficiency of droplets unfortunately does not yet converge on values which can be substituted into these equations to provide more conclusive arguments. However, if one follows the reasoning of Hocking [1959], it seems that few if any collisions occur before the droplets reach radii of 18μ or more. At this point the collection efficiency rises rapidly to values exceeding unity. If these values were to be inserted in the present study an almost explosive effect on the numbers of droplets coalescing should occur at the point where the mean drop size exceeds this value, independently of whether one has a wide spectrum of cloud condensation produced droplets or not. Here the most important consideration would be the concentration of droplets per unit volume. This is largely fixed by the supersaturation conditions and nucleus concentrations at cloud base. With different droplet concentration, considerable differences in liquid water content or cloud depth are required to produce the large mean droplet size which Hocking states is necessary for coalescence.

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References


Discussion

(Note: The discussion of this paper is combined with that following the next paper.)
Computations of the Growth of Cloud Drops by Condensation Using an Electronic Digital Computer

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Abstract—The growth of cloud drops by condensation was computed for four cases involving three different assumptions concerning the manner of cooling and two different size distributions of hygroscopic nuclei. All cases show that the smallest nuclei grow only until the maximum supersaturation has been reached, and then shrink slowly, while the larger nuclei continue to grow rapidly, resulting in a gap in the size spectrum between the cloud drops and the inactivated nuclei. In the cooling models corresponding to Cumulonimbus and trade-wind Cumulus a sufficient number of large drops are formed to initiate drop growth by coalescence and warm-cloud precipitation.

List of symbols—The following symbols are used in this paper.

\( a \) Cunningham constant
\( \dot{C} \) Average molecular velocity
\( c \) Specific heat of cloud drop
\( D (D_L, D_F, D_p) \) Coefficient of molecular diffusion
\( D_T \) Thermal diffusion coefficient
\( e \) Vapor pressure
\( e_s \) Vapor pressure at surface of drop of radius \( r \)
\( e_v \) Saturation vapor pressure
\( e_p \) Ambient vapor pressure in air parcel
\( F_M \) Flux of mass of water vapor
\( F_H \) Flux of heat
\( g \) Acceleration of gravity
\( H \) Heat
\( K \) Coefficient of thermal conduction
\( k \) Boltzmann’s constant
\( L \) Latent heat of condensation
\( M \) Mass of drop
\( m_v \) Mass of water vapor molecule
\( N_0 \) Number of cloud drops per unit volume
\( N_r \) Number of drops per cm\(^3\) larger than \( r \) (cumulative frequency)
\( n \) Number of molecules (of all kinds) per unit volume
\( P \) Pressure
\( R_v \) Gas constant of water vapor
\( r \) Radius of drop
\( r_e \) Equivalent radius of nucleus
\( s \) Parameter introduced by Rooth
\( T \) Temperature (°K); \( T_r, T_e \), values of \( T \) at drop surface and in environment
\( t \) Time, seconds
\( w \) Vapor mixing ratio
\( w_L \) Liquid content
\( \alpha \) Accommodation coefficient
\( \Gamma \) Consolidated hygroscopic factor
\( \gamma \) Parameter introduced by Frisch and Collins
\( \delta \) Fractional difference between temperature of drop and ambient temperature

\( \epsilon \) Ratio of molecular weights of water vapor and dry air (0.622)
\( \eta \) Fractional frequency of cloud drops
\( \lambda \) Molecular free path
\( \mu \) Molecular viscosity of air
\( \rho \) Density of air
\( \rho_\text{n} \) Density of nucleus
\( \rho \) Density of cloud drops
\( \rho_\text{v} \) Density of water vapor; \( \rho_\text{v}, \rho_\text{v} \) values at drop surface and in distant (undisturbed) environment
\( \phi \) Drop-size frequency function
\( \sigma \) Surface tension

Introduction—It is a remarkable fact that measurement of drop-size distribution in all sorts of clouds at various locations throughout the world almost always shows a mode, or most frequent size, in the range of 5 to 10 microns radius [Dierem, 1948; Neiburger, 1949; Weickmann and Kaufman, 1953]. Methods of measurement which are able to count the very small droplets show that there are in addition a large number of sub-micron droplets and nuclei in the clouds [Eldridge, 1957]. We undertook the study we are reporting on here to see whether drop growth by condensation alone could explain the remarkable uniformity of the drop-size distributions in clouds formed under quite different conditions, and also explain the development of bimodal size distributions from the uni-modal distributions which are almost always found in nucleus counts. In addition, we were interested in seeing whether drop growth by condensation on the observed wide range of nucleus sizes would give enough large drops to initiate the growth of precipitation by coalescence.

The problem of the growth of drops by condensation involves two transport processes in the
air parcel: (1) the transport of water-vapor molecules towards the surface of the drop, and (2) at the same time the transport away from the surface of the drop of the latent heat released during condensation.

The problem has been considered successively by various investigators who took into account more and more nearly the actual conditions which prevail in the atmosphere. Thus Houghton [1933] derived the equation for drop growth neglecting curvature, hygroscopicity, and heating caused by release of latent heat; Langmuir [1944] took curvature and heat transport into account but neglected hygroscopicity; while Howell [1949] and Best [1951] took all three factors into account. Even Howell neglected some terms and treated the problem as an equilibrium problem. In addition, as pointed out by McDonald [1953], Howell made an error in his treatment of the hygroscopic effect.

The differential equation for drop growth is too complicated to integrate by analytical means. Howell resorted to numerical and graphical techniques for its solution. He treated three cases involving various nucleus distributions and rates of cooling. With the availability of the electronic digital computer it is possible to deal with the equation in a somewhat more rigorous form than that used by Howell, in addition to using the correct value for the hygroscopic factor. In addition we thought it would be desirable to try other assumptions regarding nucleus distributions and rates of cooling. As will be discussed subsequently, it turned out that the equation which Howell used is an adequate approximation, but that even with the electronic digital computer, the integration of this equation is extremely time consuming.

The equation of drop growth—The solution of transport processes in a system with nonuniform distribution of concentration of molecules and temperature must come from the kinetic theory. The theory is fully treated by Chapman and Cowling [1952]. The main development is based on the knowledge of the function giving the distribution of molecular velocities and of the effects of molecular encounters on this function, and ultimately on the solution of Boltzmann’s equation for the velocity distribution. From this solution the transport processes in the system can be determined.

In applying the theory it was necessary to make the following assumptions to reduce the complexity and thus decrease the mathematical difficulties.

(1) A binary mixture of gases is considered, water vapor and dry air, and both constituents are treated as perfect gases.

(2) The air parcel is considered a closed system with respect to mass. No exchange of matter is allowed between the parcel and its surroundings (that is, no entrainment), while exchange of energy is permitted.

(3) Droplets in the air parcel do not affect each other, so that the field around each drop is considered to have radial symmetry.

(4) Condensation takes place only on nuclei, and these nuclei behave as though all are composed of the same hygroscopic substance, taken to be NaCl.

(5) The gas that is immediately in contact with each drop is in equilibrium with the drop, so that the temperature and vapor pressure over the drop are completely determined by the properties of the drop.

In Chapman and Cowling’s treatment it is shown that the flux of one substance, in our case water vapor, through another, that is, the ‘dry’ air, is given by

\[ F_{M} = -D \nabla \rho_{M} - \frac{m \varphi}{T} \nabla T / T \] (1)

and the flux of heat is given by

\[ F_{H} = -K \nabla T + n k T (\bar{C}_{1} - \bar{C}_{2}) D_{T} / D \] (2)

The second term on the right in each of these equations is small compared with the first for the case of dilute mixtures such as water vapor in air.

For the case of spherical symmetry, the transfers of mass and heat to the surface of a drop of radius \( r \) are obtained by integrating the above equations from the drop surface to infinity, giving

\[ \frac{dH}{dt} = 4 \pi D r (\rho \varphi - \rho_{\infty}) = \frac{4 \pi D r (\varepsilon_{\infty} - \varepsilon_{r})}{T_{r} - T_{\infty}} \] (3)

\[ \frac{dH}{dt} = 4 \pi K r (T_{\infty} - T_{r}) \] (4)

The law of diffusion treats diffusion as a continuous process, implying that the individual displacements of the molecules are of infinitesimal length. When the mean free path is longer than the radius of the droplet, the question arises whether the same diffusion law can also be applied. The problem is further complicated by the fact that at the liquid-vapor boundary molecules are evaporating as well as being condensed.

Langmuir [1944], based on an equation used for calculating the rate of evaporation of sub-
stances in high vacuum, introduced a compensated diffusion coefficient of the form

$$D_L = \left[ \frac{r}{D(r + a_s)} + \frac{1}{r} \sqrt{\frac{2 \pi}{R_s T}} \right]^-1$$

to take account of these effects, where $a_s$ the Cunningham constant, is about 0.7, and $\bar{x}$ is the mean free path.

Frisch and Collins [1952], in their investigation of the growth of aerosol particles, examined the boundary condition at the surface of the droplet appropriate for the solution of the diffusion equation. A modified boundary condition was introduced which, when applied to the diffusion coefficient, gives

$$D_{FC} = D(r / (r + \gamma))$$

The parameter $\gamma$ is defined as

$$\gamma = (1 / a_s)(\bar{x} / \bar{a})$$

where $\bar{a}$ is the mean square free path of the diffusing molecules, and $a_s$ is the probability that a molecule striking the droplet is absorbed.

Rooth [1957] obtained a modified diffusion coefficient of the form

$$D_R = D_r / (r + s)$$

The thickness $s$ of the layer through which molecular diffusion acts is given by the relation

$$s = (D/a_s) \sqrt{2 \pi / R_s T}$$

When this expression for $s$ is substituted into Langmuir’s compensated diffusion coefficient, $D_L$ takes a form similar to $D_R$ or $D_{FC}$:

$$D_L = D \left[ \frac{r (1 + a_s / r)}{r + a_s (1 + a_s / r)} \right]$$

The actual value of $\gamma$ or $s$ is subject to considerable uncertainty. Anderson [1957], based on some experimental cloud chamber data by Barrett and Germain [1947], suggests a value of $\gamma$ of the order of two microns. The value of $s$ is taken by Rooth to be five microns at 10° and 1000 mb, based on the value of $a_s = 0.035$ obtained by Alty and Mackay [1935] by measurements of the rate of evaporation of drops. As Rooth points out, if the true value of $a_s$ were larger, then $\gamma$ or $s$ would be proportionally smaller and its effect would then soon approach the limit of meteorological insignificance. Because the correct values will remain uncertain until accurate data regarding the exchange of water molecules across a liquid-vapor interface are available, modification of the diffusion coefficient as well as similar modification of the thermal conductivity has not been included in this study.

The heat which must be transferred from the drop surface consists of the latent heat released by the condensation and the heat produced by friction of the drop falling through the air, less the heat stored in the drop and the energy required for increasing the surface area of the drop

$$\frac{dH}{dt} = 4 \pi p_L r^2 \frac{dr}{dt} + \frac{S}{27} \pi^2 \rho \rho_L (\rho_v - \rho_s)$$

$$- \frac{4}{3} \pi^2 \rho v \frac{dT}{dt} = S \pi r \frac{dr}{dt}$$

It is readily shown that all the other terms are very small compared to the first term on the right in (5). Neglecting them and combining (4) and (5) we obtain

$$T_r = T_s + \frac{\rho_L r dr / dt}{K} = T_s (1 + \delta)$$

where

$$\delta = \frac{(\rho_L r dr / dt)}{KT_s}$$

The vapor pressure of a small drop containing a soluble nucleus is given by

$$e_r = e_r(T_s) \left[ \exp \frac{2 \gamma}{\rho_v r T_s} \right] \left[ 1 - \frac{\Gamma_{p_r r^2}}{\rho_v r^2} \right]$$

where the second factor expresses the increase of vapor pressure due to the curvature, and the third factor the reduction due to the effect of the solute.

It is convenient to replace the drop temperature $T_s$ by the ambient temperature $T_x$ in (7). This is accomplished using (6) and the Clausius-Clapeyron relationship. The result is

$$e_r = e_r(T_x) \left[ \exp \frac{L \delta}{R T_s (1 + \delta)} \right]$$

$$\left[ \exp \frac{2 \gamma}{\rho_v r T_s (1 + \delta)} \right] \left[ 1 - \frac{\Gamma_{p_r r^2}}{\rho_v r^2} \right]$$

Substituting this expression and $M = 4 \pi r^2 \rho_v / 3$ in (3) and dropping the subscript $\alpha$ we have

$$\frac{R_p r dr}{D} = \frac{e_r(T)}{T} \left[ \exp \frac{L \delta}{R T (1 + \delta)} \right]$$

$$\left[ \exp \frac{2 \gamma}{\rho_v r T (1 + \delta)} \right] \left[ 1 - \frac{\Gamma_{p_r r^2}}{\rho_v r^2} \right]$$
For usual rates of drop growth \( \delta \) does not exceed \( 10^{-3} \) and thus may be neglected in comparison to unity. The growth equation is thus

\[
\frac{dr}{dt} = \frac{D}{\rho \cdot \tau \cdot T} \left[ e - e(T) \right] \left[ \exp \left( \frac{Lb}{R \cdot T} \right) \right] \cdot \exp \left( \frac{2\sigma}{\rho \cdot \tau \cdot R \cdot T} \right) \left[ 1 - \frac{V_{\eta}(r_{0})}{\rho \cdot \tau^{2}} \right] \tag{10}
\]

It is to be remembered that \( \frac{dr}{dt} \) occurs in \( \delta \), so that this is an implicit equation for the rate of drop growth. If the exponential terms are expanded in series and the terms higher than the second arc neglected this equation is practically identical with that used by Howell. Langmuir’s equation for small drops is obtained by omitting the last expression in the square brackets; for large drops he also omitted the second (curvature) factor. Houghton’s equation is obtained by neglect of all three factors in brackets.

To test the validity of these various approximations, \( \frac{dr}{dt} \) was computed for various-sized drops grown from various-sized salt nuclei, using (1) Houghton’s equation, (2) Langmuir’s large drop approximation, (3) Langmuir’s small drop approximation, (4) Howell’s equation, with the correct expression of the effect of solute, and (5) Eq. (9), and assuming constant temperature and relative humidity (5°C and 101%). The results are shown in Table I. They show that the neglect of the difference in temperature of the drop from that of the air leads to large errors for all sizes of drops and nuclei. The neglect of the curvature term leads to errors of less than ten per cent for drops larger than about one micron; whereas the size at which the effect of hygroscopic influence the growth rate by less than ten per cent depends on the size of the nucleus; it appears to be roughly the size for which \( (r_{0}/r)^{3} \) is about 0.0005.

Howell’s equation gave growth rates within three per cent of that given by (9) for all the cases tested.

The method of integration—So far only the growth of a single drop has been considered. The simultaneous growth of drops of different sizes is more complicated, as the distribution of sizes at each instant must be taken into account.

Let \( \phi(r)dr \) be the fraction of the drops in the size range \( r \) to \( r + dr \). Then at all times

\[
\int_{0}^{\infty} \phi(r)dr = 1 \tag{11}
\]

Suppose that at time \( t_{0} \) the fraction of the drops in the range \( r_{1} \) to \( r_{1} + \Delta r_{1} \) is \( \eta_{1} \), and that time \( t_{2} \) the \( r_{1} \)-sized drops grow to \( r_{12} \), and the \( r_{1} + \Delta r_{1} \)-sized drops grow to \( r_{12} + \Delta r_{12} \). Since the nature of the growth process is such that a given-sized drop never overtakes a larger one, at time \( t_{2} \) the fraction in the range \( r_{12} \) to \( r_{12} + \Delta r_{12} \) will still be \( \eta_{1} \), that is

\[
\int_{r_{12}}^{r_{12} + \Delta r_{12}} \phi(r)dr = \int_{r_{11}}^{r_{11} + \Delta r_{11}} \phi(r)dr = \eta_{1} = \text{constant}
\]

Thus

\[
\phi_{1}(\Delta r_{11}) = \phi_{1}(\Delta r_{12})
\]

or

\[
\phi_{12} = \phi_{1}(\Delta r_{11}/\Delta r_{12}) \tag{13}
\]

The procedure in this treatment will be to divide the initial nucleus distribution into size groups \( \Delta r_{1} \), of average frequency \( \phi_{1} \). The average frequencies \( \phi_{1} \) at each time \( t \) will be computed by (13).

An approximation of the cumulative frequency distribution at any time is arrived at by graphical interpolation. It was found that the interpolation could be best approximated by dealing with the cumulative distribution, that is, the variation of the number larger than each given size. If \( N_{0} \) is the total number per cm², the number \( N_{r} \) larger than \( r \), called the cumulative frequency, is

\[
N_{r} = N_{0} \int_{r}^{\infty} \phi(r)dr = \int_{r}^{\infty} \frac{dN}{dr}dr \tag{14}
\]

where \( dN = N \phi(r)dr \) is the number in the size range \( r \) to \( r + dr \).

The values of \( N_{r} \) were computed for each group at selected times and plotted against \( r \log - \log \) graph paper. Then a smooth curve was fitted for each time, keeping the physical process of drop growth in mind in establishing the relative positions of the various interpolated curves. The slope of the curves for \( N_{r} \) were computed to obtain the frequency per unit size interval, \( dN/dr = N \phi(r) \), which is called the differential frequency.

In interpreting the results it is important to keep in mind the distinction between the definite results of the computations and the results indicated or suggested by the curves interpolated between the computed values.

In treating cloud development in the atmosphere it is assumed that the temperature of an
Table 1—Comparison of rate of growth of drops (micron/sec) at various sizes for given initial conditions

<table>
<thead>
<tr>
<th>((n/t)^{0.5}) (micron)</th>
<th>(r) (micron)</th>
<th>(dr/dt) (micron/sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Houghton</td>
<td>Langmuir (large drop)</td>
</tr>
<tr>
<td>0.5</td>
<td>1.26</td>
<td>0.7810</td>
</tr>
<tr>
<td>0.1</td>
<td>2.51</td>
<td>0.6484</td>
</tr>
<tr>
<td>0.05</td>
<td>2.71</td>
<td>0.5428</td>
</tr>
<tr>
<td>0.01</td>
<td>4.64</td>
<td>0.3321</td>
</tr>
<tr>
<td>0.0005</td>
<td>5.85</td>
<td>0.2679</td>
</tr>
<tr>
<td>0.00005</td>
<td>10.00</td>
<td>0.1558</td>
</tr>
<tr>
<td>0.000005</td>
<td>12.60</td>
<td>0.1237</td>
</tr>
<tr>
<td>0.0000005</td>
<td>25.14</td>
<td>0.0672</td>
</tr>
<tr>
<td>0.00000005</td>
<td>27.14</td>
<td>0.0657</td>
</tr>
<tr>
<td>0.000000005</td>
<td>40.42</td>
<td>0.0306</td>
</tr>
<tr>
<td>0.0000000005</td>
<td>58.48</td>
<td>0.0266</td>
</tr>
<tr>
<td>0.00000000005</td>
<td>100.00</td>
<td>0.0160</td>
</tr>
</tbody>
</table>

For \(r_0\) equals one micron

<table>
<thead>
<tr>
<th>((n/t)^{0.5}) (micron)</th>
<th>(r) (micron)</th>
<th>(dr/dt) (micron/sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>0.126</td>
<td>7.810</td>
</tr>
<tr>
<td>0.1</td>
<td>0.251</td>
<td>6.484</td>
</tr>
<tr>
<td>0.05</td>
<td>0.271</td>
<td>5.428</td>
</tr>
<tr>
<td>0.01</td>
<td>0.464</td>
<td>3.321</td>
</tr>
<tr>
<td>0.005</td>
<td>0.585</td>
<td>2.679</td>
</tr>
<tr>
<td>0.0005</td>
<td>1.000</td>
<td>1.588</td>
</tr>
<tr>
<td>0.00005</td>
<td>1.260</td>
<td>1.237</td>
</tr>
<tr>
<td>0.000005</td>
<td>2.514</td>
<td>0.724</td>
</tr>
<tr>
<td>0.0000005</td>
<td>2.714</td>
<td>0.575</td>
</tr>
<tr>
<td>0.00000005</td>
<td>4.162</td>
<td>0.336</td>
</tr>
<tr>
<td>0.000000005</td>
<td>5.848</td>
<td>0.267</td>
</tr>
<tr>
<td>0.0000000005</td>
<td>10.000</td>
<td>0.160</td>
</tr>
</tbody>
</table>

For \(r_0\) equals one-tenth micron

<table>
<thead>
<tr>
<th>((n/t)^{0.5}) (micron)</th>
<th>(r) (micron)</th>
<th>(dr/dt) (micron/sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>0.0126</td>
<td>78.10</td>
</tr>
<tr>
<td>0.1</td>
<td>0.0251</td>
<td>64.81</td>
</tr>
<tr>
<td>0.05</td>
<td>0.0271</td>
<td>54.28</td>
</tr>
<tr>
<td>0.01</td>
<td>0.0464</td>
<td>33.21</td>
</tr>
<tr>
<td>0.005</td>
<td>0.0655</td>
<td>26.79</td>
</tr>
<tr>
<td>0.0005</td>
<td>0.0100</td>
<td>15.58</td>
</tr>
<tr>
<td>0.00005</td>
<td>0.1260</td>
<td>12.37</td>
</tr>
<tr>
<td>0.000005</td>
<td>0.2514</td>
<td>7.24</td>
</tr>
<tr>
<td>0.0000005</td>
<td>0.2714</td>
<td>5.75</td>
</tr>
<tr>
<td>0.00000005</td>
<td>0.4642</td>
<td>3.36</td>
</tr>
<tr>
<td>0.000000005</td>
<td>0.5848</td>
<td>2.67</td>
</tr>
<tr>
<td>0.0000000005</td>
<td>1.0000</td>
<td>1.60</td>
</tr>
</tbody>
</table>

air parcel containing a given amount of water vapor decreases either isobarically (fog and Stratocumulus) or by adiabatic expansion (Cumulus and Cumulonimbus). The temperature is thus a given function of time

\[ T = T(t) \quad (15) \]

The total amount of water substance must remain constant, so that as the liquid content \(w_L\) in-creases the vapor mixing ratio \(w\) decreases

\[ \rho_a w + w_L = \rho_a w + \frac{4}{3} \pi N \sum \Delta \xi \rho_a r^2 \Delta r = \frac{4}{3} \pi N \sum \rho_a r^3 \Delta r_0 = \rho_a \Delta w \quad (16) \]

where \(w_0\) is the water vapor mixing ratio before condensation begins.
The ambient vapor pressure \( e \) at the time when the drops have grown from \( r_0 \) to \( r_i \) is thus

\[
e \left( \frac{P}{e} \right) = P \left[ w_2 - \frac{4\pi N}{3\rho_o} \sum \phi_i r_i^3 \Delta r_i ight]
\]

\[
- \sum \phi_i \rho_i r_i^2 \Delta r_i)
\]

(17)

This value of \( e \) is used in the growth equation (10).

For the numerical integration of (10) on the electronic digital computer the method of Runge-Kutta was chosen. It was found that the equation was subject to great computational instability, so that very small time steps had to be used at first. A method was devised for testing when it was feasible to increase the time steps. With this procedure the machine time required was very large. It had been hoped that it would be possible to carry out computations for a number of cases in which the various parameters (size distribution and composition of nuclei, manner of cooling, etc.) were varied, but the machine time required was so large that only four cases were computed, representing three rates of cooling and two distributions of sizes of nuclei.

The nuclei distributions used—The measurement of particulate matter in the atmosphere has been carried out by various means, each capable of counting particles in a portion of the range of sizes which occur. For a long time most counts were of the range called Aiken nuclei \((<0.1 \text{ micron})\) on which condensation occurs only at high supersaturation. Recently it has become recognized that the 'large' \((0.1-1 \text{ micron})\) and 'giant' \((>1 \text{ micron})\) nuclei are the only ones which participate in cloud formation.

Size distributions in the various ranges have been summarized by Junge [1952, 1953], Gilbert [1954], Woodcock [1952, 1953], and Lodge [1955]. Their data were combined into two idealized size distributions, shown in Figure 1. The difference between the two lies in the larger concentration of large and giant nuclei in Type B distribution.

In the figures are shown both the concentrations of each size per unit size interval and the cumulative concentration of all nuclei greater than a given size. In Type A distribution there are 1030 particles per cm\(^3\) larger than 0.01 micron, 115 per cm\(^3\) larger than 0.1 micron, about 1.5 per liter larger than one micron, and much less than one particle per cubic meter larger than 10 microns. In Type B distribution the same numbers are respectively 120, 135, 50, and 14.

The models of cooling—The models of cooling were selected to simulate in a general way the conditions of formation of Stratus cloud or fog, Cumulonimbus, and trade-wind Cumulus.

For the two Stratus cases it was assumed the cooling proceeded at constant pressure at a rate of 6°C per hour. The initial temperature was taken to be 288°C, and the initial relative humidity was about 75%. For the first Stratus case the Type A distribution of nucleus sizes was assumed, and for the second, Type B. Correspondingly, these cases will be called Stratus Case A and Stratus Case B. The pressure was assumed to be 1000 mb.

For the Cumulonimbus model we assumed a distribution of vertical velocities based on the measurements of the Thunderstorm Project [Byers and Braham, 1949]. The initial temperature was taken to be 290.2°C, and the relative humidity about 75% at about 1000 mb pressure, and adiabatic cooling was assumed. Figure 2 shows the assumed variation of vertical velocity and

![Figure 1](image.png)  
**Figure 1**—Assumed distributions of sizes of nuclei (solid curves, cumulative distributions, dashed curves, differential distributions)
GROWTH OF CLOUD DROPS BY CONDENSATION

height and the corresponding changes of temperature, pressure and height with time. The Type A nucleus distribution was assumed.

The temperature difference between the cloud and the environment, computed from the vertical acceleration, is shown at the top of Figure 2. In the lowest part of the cloud the assumed acceleration corresponds to only a small fraction of one degree, and even in the upper part, in which the vertical velocity increases rapidly, it corresponds to the cloud being only about one-half degree centigrade warmer than the environment.

In the trade-wind Cumulus model the vertical velocity distribution was based on the observations by Malkus [1954]. The temperature and relative humidity at 1000 mb were assumed to be 301.7°K and about 75%, and the cooling was assumed to be adiabatic. Figure 3 shows the assumed variation of vertical velocity and the corresponding changes of temperature, pressure and height for this case. For the trade wind Cumulus case the Type B nucleus distribution was assumed.

The assumed acceleration corresponds to a larger temperature difference between cloud and environment in the lower part of the trade wind Cumulus than in the Cumulonimbus (see upper portion of Fig. 3). The deceleration near the top corresponds to the cloud becoming 1°C colder than the environment.

Properly, in computing the cooling the release of latent heat should be computed directly from the amount of liquid condensed at each step. J. E. McDonald has pointed out that in the early stages of drop growth the difference between this and the assumption of saturation adiabatic equilibrium might lead to significant differences in the cut-off point between the nuclei which grow to cloud drops and those which remain small. In a future computation it is planned to examine this point. For the computations presently being reported the decrease in environmental temperature was assumed to be that resulting from saturation adiabatic cooling.

The Stratus cases—The variation of the relative humidity during the isobaric temperature decrease (Stratus Case A) is shown in the upper portion of Figure 4, and the variation in size of the different droplet groups in the lower portion. Saturating with respect to a plane water surface is reached after about 2650 sec. Until about 2400 sec the hygroscopic nuclei all increase slowly in size; from then to the time of maximum supersaturation, about 2700 sec, all groups grow more and more rapidly; after that the degree of supersaturation decreases, and the smallest two groups of drops decrease in size, while the groups growing on nuclei of 0.1 micron or larger continue to grow rapidly. By 2800 sec all the growing groups exceed four microns, while the non-growing nuclei re-
Fig. 3—Assumed variation of vertical velocity and corresponding changes of temperature, pressure, and height with time, Trade Wind Cumulus Case; at top, temperature difference between cloud and environment.

Fig. 4—Variation of relative humidity and growth curves for various drop-size groups, Stratus Case A
main smaller than 0.2 micron. Thus the separation between the cloud droplets and the non-growing nuclei is established.

To see what this droplet growth means in terms of numbers of drops, the cumulative distribution curves are shown for various times in Figure 5. At 2400 sec there is only about one droplet per cm³ larger than one micron; by 2700 sec this number has increased to about 250, and there is already almost one per cm³ larger than four microns; and at 3000 sec there are around 300 per cm³ larger than four microns, and about ten per liter greater than 10 microns. Thus the development from no cloud to a dense cloud occurs in a very few minutes. However, even after an hour, only one per liter has grown to 20 microns, where it might be expected, by the Langmuir theory, to start colliding with other drops.

The development of the cloud, as distinct from the increasingly dense haze, is shown by the differential frequency curves in Figure 6. While the computations for the relatively small number of groups do not establish these curves uniquely, the interpolation of the cumulative frequency curves is sufficiently definite to indicate their general character. The initial nucleus distribution has a mode at about 0.03 micron. At 2700 sec there is just a slight suggestion that a second mode is developing at about two microns. By 3000 sec the mode is well established at 5.4 microns, and the gap between the cloud drops and the inactivated nuclei is conspicuous about three microns. The development of the second mode and the gap between the two modes is due to the fact that all the drops forming on nuclei above a critical size (between 0.032 and 0.1 micron) grow, while those forming on smaller nuclei do not grow or actually shrink. This results in a decrease in number, and eventually development of a frequency minimum in the vicinity of the critical size. As the cooling continues the mode moves to larger values, reaching 12 microns at 6000 sec, and the gap appears to have moved upward to include seven microns. The nature of the gap is made clear if it is noted that between one and four microns at 3000 sec there are 90 drops, while between four and seven microns there are 310. At 6000 seconds there are 50 drops between two

**Fig. 5**—Cumulative drop-size distributions after various elapsed times, Stratus Case A

**Fig. 6**—Differential drop-size distributions after various elapsed times, Stratus Case A
and eight microns, and 300 between eight and 14 microns.

The rapid development of the visual cloud is shown by Figure 7, in which the visual range computed from Koenschmieder's formula and the liquid content are shown as a function of time. The visual range drops from 5 km at 2400 sec to 50 m at 3000 sec, a change to one per cent of its earlier value in ten minutes. In the same interval the liquid content increases from $3 \times 10^{-4}$ grams per m$^3$, to 0.3 grams per m$^3$, a factor of one thousand. In the unrealistic conditions postulated (rapid cooling and no fallout) the visual range drops below ten meters and the liquid content rises almost to three grams per m$^3$. The thin curves in Figure 7 show the contributions of the various size groups to the liquid content. Almost all of the liquid content is due to the 0.1 micron and 0.32 micron groups, with the former accounting for about five-sixths and the latter about one-sixth of the total. This is because the larger nuclei are so much less frequent and the more frequent smaller ones do not grow.

The growth curves in Stratus Case B, in which the same cooling rate was used but with many more large nuclei, are shown in Figure 8. Saturating with respect to a plane water surface is reached a little later than in Stratus Case A, the peak supersaturation attained is slightly lower, and the amount of supersaturation remains lower throughout the drop growth. However, there is very little change in the rate of growth of the various-sized drop groups, so that the change in the humidity curve reflects almost entirely the additional water required by the growth of the larger number of large drops.
In Case B, in order to determine more closely the separation between the nuclei which do not grow and those which are 'activated,' an additional group was inserted, with average equivalent radius of nucleus 0.056 micron. It turned out that this group also was activated, so that the boundary between non-activation and activation for this rate of cooling lies between 0.032 and 0.056 micron.

In Figure 9 are shown the cumulative distributions for Case B. As might be expected, the main difference between these and those for Case A lies in the 'tails' of the curves, representing the largest drops. Thus at 5000 sec the number greater than four microns is 330 per cm³ compared with 300 in Case A, but the number larger than ten microns is 230 per liter, compared with ten per liter in Case A, and 2.3 per liter are greater than 20 microns. At 6000 sec about 50 per liter are greater than 20 microns in radius, compared with one per liter for Case A. Since in light rain there may be from two to 60 raindrops per liter, it will be seen that for the cooling rate assumed the Type B nucleus distribution might be expected to produce drizzle or light rain within a short time (from 5 to 50 minutes) after the cloud formed.

As in the cumulative frequency curves, the differential frequency curves for Case B (Fig. 10) show little difference from those of Case A except that the tails show the larger number of large drops. The modes are nearly the same. Similarly the visual range and total liquid content curves are almost identical (Figure 11), since the larger drops are not sufficiently numerous to affect them. The bulk of the liquid content is due to the same range of sizes as in Case A, but the subdivision of this range into an additional group emphasizes the fact that the 0.1 micron group is responsible for the bulk of the liquid content even though
crons (although only 12 per \(10^8\) cm\(^3\) larger than 20 microns) at 1800 sec; at 2400 sec the number larger than 20 microns has increased to about one per liter, and by 3600 seconds it exceeds 60 per cm\(^3\).

It should be noted that the numbers per unit volume are altered as the air parcel rises because of the expansion of the volume of the parcel with decreasing pressure as well as because of the drop growth.

Assuming that the growth by accretion would become significant when the drop radii exceed 20 microns, precipitation development by the warm-cloud process would begin at about 2400 sec (20 min after saturation) when the parcel has reached 2300 m (7500 ft), with a temperature of 12°C. The 0°C level (4750 m) is reached at 3180 sec, and the -10°C (6400 m) at 3400 sec. Thus there would be more than 15 min during which the warm process (collision and coalescence) would be active in producing raindrops before the ice-crystal process could begin, and radar echoes might be expected at about 2500 m (8000 ft).

![Fig. 9—Cumulative drop-size distributions after various elapsed times, Stratus Case B](image)

The more numerous 0.056 micron group also continues growing.

*The Cumulonimbus case*—The more rapid cooling due to the vertical velocity distribution assumed in the Cumulonimbus model results in saturation being reached much more quickly than in Stratus Case A (1180 vs 2650 sec), and a slightly higher supersaturation being attained (0.16 vs 0.14 per cent, see Figure 12). Nevertheless the separation between the activated and the non-activated groups of nuclei is the same; the 0.1 micron group grows rapidly to 100 times its original radius and continues growing, while the 0.032 micron group reaches less than eight times its original radius at the maximum supersaturation and then gradually shrinks as the degree of supersaturation goes down.

By 1760 sec (less than ten minutes after saturation) the average radius of all the growing groups exceeds ten microns, and by 3450 sec they all have grown to more than 20 microns. Thus, as shown in the cumulative distribution curves (Fig. 13), there are about 150 per cm\(^3\) larger than ten microns.

By 1760 sec (less than ten minutes after saturation) the average radius of all the growing groups exceeds ten microns, and by 3450 sec they all have grown to more than 20 microns. Thus, as shown in the cumulative distribution curves (Fig. 13), there are about 150 per cm\(^3\) larger than ten microns.

![Fig. 10—Differential drop-size distributions after various elapsed times, Stratus Case B](image)
GROWTH OF CLOUD DROPS BY CONDENSATION

These figures may be compared with the data of Bowen, Smith, and Styles and Campbell cited by Mason [1957] for non-freezing showers in southeastern Australia. Four of the six cases with radar data had echoes spreading downwards from levels between 7000 and 10,000 ft, and all of the six cases for which radar data was not available had clouds extending above 8500 ft. Mordy and Eber [1954] reported that rain occurred in the cases they studied of the orographic rainfall in the Hawaiian Islands whenever the cloud extended more than 5000 ft above the cloud base of 2000 ft. While the clouds in the cases cited were different than the thunderstorm Cumulonimbus for which the assumed vertical velocity distribution is typical, the correspondence of the height of radar echo or the thickness of precipitating clouds with that required in the model for development of drops large enough to initiate the warm cloud precipitation process suggests that the general magnitudes involved in the computation have some relationship to natural processes.

With the more rapid cooling, the cloud development occurs more rapidly in the Cumulonimbus case than in the Stratus cases. At 1200 sec (Fig. 14) the second mode in the differential frequency curves is just beginning, at about 0.5 micron; less than two minutes later it is well developed at 5.4 microns, with a conspicuous gap at about four microns. The mode moves rapidly to larger sizes exceeding ten microns after five more minutes and reaching 20 microns by 3600 sec. The rapid development of the cloud is shown likewise by the liquid content and visual range curves (Fig. 15).
The change in liquid content from $3 \times 10^{-4}$ to 0.3 grams per m$^3$ occurs in three minutes in this case, compared to ten minutes in the Stratus cases. By ten minutes after saturation the liquid content has reached 1.6 grams per m$^3$, and at 3400 sec it reached a peak value of 6.8 grams per m$^3$. The subsequent decrease is due to the expansion of the ascending air parcel, which begins to overbalance the condensation at that time. As in the Stratus cases, the 0.1 micron group is responsible for most of the liquid content.

The decrease in visual range in the Cumulonimbus is likewise more rapid, and the visual range reaches lower values, with a minimum of 7.2 m at 3300 sec.

Except insofar as it might affect the condensation process, the development of drop growth by collision and coalescence would not affect the liquid content if we assume that the parcel receives as many of the larger drops from above as fall out of it. The visual range, however, would be influenced considerably, for the removal of smaller drops would reduce the scattering area much more than the growth of the larger drops would increase it.

The Trade-Wind Cumulus Case—In the Trade-Wind Cumulus Case the large vertical velocities assumed for the lower portion of the cloud (see Fig. 3) resulted in saturation being reached earlier than in the Cumulonimbus Case; namely, at 816 sec; and a higher maximum supersaturation, 0.24 per cent at 835 sec, being attained (Fig. 16). The second maximum in relative humidity, which occurs near the level of maximum vertical velocity, has no significance in terms of activating new nuclei.

Because of the higher maximum supersaturation, not only is the additional nucleus group at 0.056 micron activated (as in the Stratus Case B), but so is the 0.032 micron group. This, together with the larger number of giant nuclei in the assumed initial distribution, results in larger num-

Fig. 12—Variation of relative humidity and growth curves for various drop-size groups, Cumulonimbus Case

Fig. 13—Cumulative drop-size distributions after various elapsed times, Cumulonimbus Case
numbers of droplets of all sizes near the cloud base in the Trade-Wind Cumulus Case than in the Cumulonimbus Case (Fig. 17). Table 2 shows the relative values at two corresponding levels above the cloud base. The modal sizes for the Trade-Wind Cumulus Case are taken from the differential frequency curves in Figure 18.

At 130 m above the cloud base the trade-wind Cumulus has more drops in all size groups larger than four microns than the Cumulonimbus. However, because of the larger number of small nuclei which are activated, the mode is smaller even at this height. By 800 m above the cloud base this greater competition for the available liquid has resulted in smaller numbers of drops larger than nine microns in the Cumulus Case than in the Cumulonimbus, and a modal size of eight microns as opposed to 11. At the top of the trade-wind cumulus the modal size is 11 microns, and there are about ten per liter greater than 20 microns in radius. Thus it would appear that also in this case precipitation initiation by the warm process could be expected, with the possibility of light showers reaching the sea surface.

The liquid content in the Trade-Wind Cumulus Case (Fig. 19) increases faster than in the Cumulonimbus Case, and the visual range decreases more rapidly. The values at corresponding heights above the cloud base are shown in Table 2. The maximum liquid content in the trade wind Cumulus, however, reaches only 2.7 grams per m², less than one-half that in the Cumulonimbus, because of the continued ascent in the latter case, but the minimum visual range is a little lower because of the larger total number of activated droplets in the Trade-Wind Cumulus Case.

Because the 0.062 micron group is activated in
Fig. 16—Variation of relative humidity and growth curves for various drop-size groups, Trade-Wind Cumulus Case

this case the very large number of nuclei in this group makes it the largest contributor to the liquid content.

Conclusions—All four cases for which computations were carried out, representing three widely differing rates of decrease of temperature and two different nucleus distributions, showed the tendency for the separation of cloud drop sizes from the inactivated nuclei. While the number of groups used in the computation is not sufficient to define it uniquely, it appears definite that a mode in the cloud drop sizes is established shortly after saturation occurs, and that this modal size increases as cooling continues. The mode is in the same general size range as in observed cloud drop distributions, but tends to move with time to larger sizes than in the observed distributions. On the other hand, there are fewer very large drops in the computed distributions than in size distri-

butions observed in Cumulus congestus and Cumulonimbus [Dien, 1948, Weickmann and aufl Kampe, 1953].

Since the initial conditions and rates of cooling for the observed cases were doubtless not the same as those assumed for the computations, the computed distributions cannot be expected to correspond closely to the observed. Nevertheless it is interesting to note that the processes neglected, namely coalescence and in the case of the Cumulonimbus the ice-crystal (three phase) process, would tend to keep the modal size small and increase the number of very large drops.

As an example of the comparison of the computed distributions with observed values, Table 3 gives the drop-size distributions observed for three cases of Stratus off the coast of California, taken by blimp (lighter-than-air ship) in the summer of 1945, and the data for the Stratus Case B computation for 3000 and 6000 sec, as taken from Figure 9. The 3000-sec data correspond in a general way to the observed cases, but even this dis-
Table 2—Comparison of relative values at corresponding levels above cloud base for the Cumulonimbus and Trade-Wind Cumulus Cases

<table>
<thead>
<tr>
<th>Item</th>
<th>Distance above cloud base</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>130 m</td>
</tr>
<tr>
<td>Cloud type</td>
<td>Ch</td>
</tr>
<tr>
<td>Time of rise, sec</td>
<td>1300</td>
</tr>
<tr>
<td>No. of drops per cm$^3$ larger than:</td>
<td></td>
</tr>
<tr>
<td>20 microns</td>
<td>9 $\times$ 10$^{-6}$</td>
</tr>
<tr>
<td>14 microns</td>
<td>1.7 $\times$ 10$^{-4}$</td>
</tr>
<tr>
<td>9 microns</td>
<td>1.5 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>4 microns</td>
<td>220</td>
</tr>
<tr>
<td>1 micrometer</td>
<td>350</td>
</tr>
<tr>
<td>Modal size, micrometer</td>
<td>5.4</td>
</tr>
<tr>
<td>Liquid content, g/m$^3$</td>
<td>0.29</td>
</tr>
<tr>
<td>Visual range, m</td>
<td>45</td>
</tr>
</tbody>
</table>

Fig. 19—Variation of liquid content and visual range (for light of wavelengths 4000 and 7000 A), Trade Wind Cumulus Case; thin lines show liquid content in various drop-size groups.

distribution has the modal radius somewhat larger than the observed distributions, and in the 6000-sec distribution the modal radius is more than twice as large as the observed modes. An explanation of the difference probably lies in the rapid rate of cooling assumed in the computations. The coastal stratus forms by slow cooling, during which the large drops may tend to settle out of the cloud. Another factor which may contribute to the small modal size in the observed distributions is the fact that the observations were made in the daytime, when the clouds were tending to dissipate.

While the initial nucleus distributions and vertical velocities used by Howell were quite different than those we used, his results are quite simi-
Table 3—Comparison of computed distributions and observed values for Stratus, number of drops per cm$^3$

<table>
<thead>
<tr>
<th>Size range η (micron)</th>
<th>Case 1</th>
<th>Case 2</th>
<th>Case 3</th>
<th>Computed (Stratus R) 3000 sec</th>
<th>6000 sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5-2.9</td>
<td>14</td>
<td>66</td>
<td>47</td>
<td>35</td>
<td>40</td>
</tr>
<tr>
<td>3.0-4.3</td>
<td>70</td>
<td>120</td>
<td>142</td>
<td>25</td>
<td>20</td>
</tr>
<tr>
<td>4.4-5.7</td>
<td>66</td>
<td>60</td>
<td>210</td>
<td>230</td>
<td>10</td>
</tr>
<tr>
<td>5.8-7.0</td>
<td>17</td>
<td>17</td>
<td>59</td>
<td>90</td>
<td>5</td>
</tr>
<tr>
<td>7.1-8.1</td>
<td>12</td>
<td>9</td>
<td>18</td>
<td>9</td>
<td>5</td>
</tr>
<tr>
<td>8.2-9.1</td>
<td>6</td>
<td>2</td>
<td>7</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>9.2-10.5</td>
<td>3</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>10.6-11.7</td>
<td>1</td>
<td>6</td>
<td>3</td>
<td>205</td>
<td></td>
</tr>
<tr>
<td>11.8-12.8</td>
<td>1</td>
<td>2</td>
<td>7</td>
<td>90</td>
<td></td>
</tr>
<tr>
<td>12.9-14.1</td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>28</td>
<td></td>
</tr>
<tr>
<td>14.2-15.0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>15.1-16.0</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

It is planned to undertake computations of the growth of drops by the combined effects of condensation and coalescence, and subsequently to take account of the three-phase process as well. For this purpose accurate collection and coalescence efficiency data are needed. At present the uncertainty, particularly with respect to the efficiency of collection of nearly equal sized small drops, would leave the validity of the results of such computations considerably in doubt. Nevertheless the computations may give a considerable insight into the precipitation process. Simplified computations have already contributed to the clarification of the problem (Houghton, 1950; East, 1957). The computations using the electronic digital computer should provide further enlightenment.

Acknowledgments—The computations were carried out on SWAC, the electronic digital computer operated under a contract with the U.S. Navy Office of Naval Research by the Numerical Analysis Research Group of the Department of Mathematics, University of California, Los Angeles. Time on the machine was provided without charge, and our gratitude to the Numerical Analysis Research Group and the Office of Naval Research is hereby acknowledged. We wish to thank the staff of the Numerical Analysis Research Group also for suggestions and assistance in preparing the problem for the computer.

References


Langmuir, I., Super-cooled water droplets in rising currents of cold saturated air, Rep. RL-223, General Electric Research Laboratory, 1941.


Discussion

(Relating to the two immediately preceding papers.)

Dr. Helmut Weickmann—Did you use in your computations nuclei distributions that had actually been measured?

Dr. M. Neiburger—This certainly would be desirable. I should have mentioned that when we started we were not very particular about the conditions we chose, because we did not realize what a big job it was for the electronic digital computer. We decided we would start with a composite using Junge's size distributions for the small ones and then some different values of Lodge or Woodcock for the big ones for a rough mean of a continental aerosol. But since there is never any individual aerosol which is actually like any of these models we used, it would make more sense to use some real ones. The trouble is that nobody has taken simultaneous measurements over the whole size range from the very small nuclei to the large ones. It is simple to incorporate measured size distributions into the program, if they are available.

Dr. Weickmann—Is it important to know the size distribution down to the Aitken range of about 10^-6 microns?

Dr. Neiburger—Yes, it affects the size in which the cutoff occurs, because there are so many of those. They do pick up water until the critical supersaturation is reached which is then determined by the larger sizes.

Dr. A. Goetz—What determines the size of the second maximum? You had it between ten and twelve microns. What parameter would have to be changed in order to shift its size, decrease it, or increase it?

Dr. Neiburger—The rate of cooling is one factor which is very important which would mean the vertical velocity in this case; others are relative numbers of large and small nuclei.

Dr. W. E. Howell—I realize that it is extremely hard to predict from the initial nuclei distribution what kind of cloud one is going to end up with. I remember I failed completely on that point. But I would like to raise the question whether the initial conditions that you have chosen do actually conform with the trade Cumulus and the Cumulonimbus. I believe you started with updraft velocity of 150 cm/sec in the condensation layer of the trade Cumulus; but going back to Dr. Malkus' paper, I find an estimate of 40 cm/sec from the flight that went exactly through the base of a trade Cumulus, and 108 cm/sec on the flight that went 130 m above the cloud base; and these were the velocities in the center of the updraft, not average across the whole base of the cloud. I think a reduction in the velocity would make a considerable change in the computation, tending toward a much smaller drop number and larger drop size for the trade Cumulus. Alternatively with the Cumulonimbus, where you have used an updraft velocity of 30 cm/sec, it seems from our experience in flying light aircraft for water seeding of clouds that 200 or 300 cm/sec in the updraft at the base of the cloud would be more realistic. Again the change would be in the direction of reversing the difference that you computed between the trade Cumulus and Cumulonimbus, giving the Cumulonimbus a much higher droplet number at least in the initial condensation stage. What happens later on, when coalescence takes over, may be different, of course.

Some of us are more interested in the liquid water content as a function of the drop size than we are in the drop number as such, for studying icing, for example. I think you can transform your graphs to show liquid water content by drawing the base line as a line of slope minus 6 on those graphs, which immediately transforms your double mode to a single mode which corresponds to the volume median drop size.

Dr. Neiburger—Taking the last point first, I do not know whether I got them into the version of the figures projected here, but we have some diagrams in which there are curves showing the
variation with time of the contribution to the liquid content by the various droplet groups in addition to the liquid content for the whole cloud. The droplet group just a little larger than the modal size is the one that contributes most to the liquid content in most cases.

Returning to the question of initial conditions, what I said previously applies as well to the vertical velocity as to the nucleus distribution. I would like to do the computations with a case in which the vertical velocities were actually observed simultaneously with observations of the nucleus size distribution, rather than the one we assumed as 'representative'. In arriving at the choice we made Mr. Chien plotted from the tables in Dr. Malkus' papers the vertical velocities for various positions with respect to the cloud base, drew some lines through them, and showed the result to me, and I said, "Let's use this line as a sort of compromise among them." Now, there must have been one or more measurements of a meter and a half or so per second, or our curve would not have been drawn so far over down at the cloud base; but it may be that most of the measurements really gave lower values, and I am inclined to agree that this may have been an unfortunate choice.

Dr. Tor Bergeron—I would like to connect the results especially of the papers by Dr. Neiburger and Mr. Mordy with certain things which are known to synopticians and those who observe clouds. Especially I would like to connect it with visibility observations that I have made during 25 summers up in the Swedish mountains, at latitude of 63 1/2° N. I am very pleased to see that gap in the size distribution, that gap between the inactivated nuclei as Dr. Neiburger calls them, and the nuclei that really grow. Because in those localities of which I spoke, from April to September in different months and different years, I had always a gap in the spectrum of visibility ranges. Visibility could be less than one or two kilometers because of fog or mist; but it could never be between two and twenty kilometers. Remember that we were rather far from the origin of the nuclei, that is, from the ocean, and from any industrial pollution also. So we had evidently very little amount of hygroscopic nuclei. Then we have had the visibility from 20 up to 500 km, the latter corresponding to practically pure air. And those visibilities, of course, occurred in the presence of inactivated nuclei; that is nuclei with sizes generally less than 0.1 micron. At those sizes, the amount of scattered light will be roughly proportional to the fourth power of the diameter. In order to get a haze by inactivated nuclei of that size whose visual range could be comparable with that of fog their number should be between $10^6$ and $10^9$ times as numerous as the activated particles. According to Dr. Neiburger's diagram, as far as I can see, they were more numerous, but only 10 to 100,000 times perhaps. This then will explain that gap in the visibility spectrum; a gap the existence of which has been absolutely impossible for me to get the continental colleagues in Europe to understand. Therefore any discussion in the International Synoptic Commission, and other places, of scales of visibility and such things was always doomed to fail, because Central Europe, England, Eastern United States, and even the coasts of the ocean, would all have rather plenty of big hygroscopic nuclei that grow with increasing relative humidity.
The Relation between Cloud Droplet Spectra and the Spectrum of Cloud Nuclei

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Abstract—Observations of cloud droplet spectra in various kinds of clouds have shown that the microstructure of clouds very largely determines the efficiency of the coalescence process in forming rain. They have also suggested that the observed differences in microstructure must be primarily explained by differences in the spectra of cloud nuclei in different air masses. This hypothesis has now been confirmed by simultaneous observations of the spectra of cloud nuclei and cloud droplets.

Introduction—In a classical paper, Houghton [1950] discussed the significance of the various processes leading to the formation of precipitation. He showed that sublimation onto ice crystals could quickly give rise to the formation of particles equal in mass to a droplet of several hundred microns diameter, and that, given a cloud droplet spectrum which was broad enough, raindrops could form by coalescence. The calculations were based on the collection efficiencies computed by Langmuir [1948] and it appeared that only those clouds which had very broad spectra and very large median droplet diameters could form rain by coalescence in a sufficiently short period of time. Telford [1955] has drawn attention to the fact that only a very small fraction of the 'most fortunate' droplets in a cloud grow to raindrops, and that consequently it is necessary to consider the statistical fluctuations in the growth time of individual drops, arising from the random and discrete nature of droplet growth by collision. In one example, he showed that one large (collector) cloud drop in a hundred thousand would experience its first ten collisions in five minutes, whereas the average large cloud drop would require 33 minutes to grow to this extent. This kind of consideration makes the production of rain by coalescence somewhat easier, but it still seems to be possible only in clouds with quite broad spectra and large average and maximum drop sizes.

Howell [1949] carried out computations of the growth of cloud droplets starting from assumed distributions of cloud nuclei and reached the conclusion that, as a result of the quadratic law of growth, condensation tended to produce rather uniform droplet spectra.

A study of the condensation phase of droplet growth [Squires, 1952], aimed specifically at finding those conditions which could bring about the formation of large droplets, led to the recognition of a kind of threshold effect: The size of the cloud nucleus has relatively little influence on the growth of a cloud droplet once it has grown through the critical activation region. Its influence is largely restricted to determining whether, with a given maximum supersaturation in the cloud-forming region, the droplet shall grow unstably or remain in the stable 'haze' stage. This simplification of the role of the nucleus made it possible to discuss the conditions leading to the formation of only a restricted number of cloud droplets, apparently a necessary and certainly a sufficient condition for the production of relatively large droplets. If the number of droplets formed is to be restricted, the nucleus spectrum must be such that only a small number of nuclei become activated; since a nascent cloud in which this has happened has only a relatively small absorptive power for water vapor, the supersaturation tends to rise to relatively high values; if, therefore, the few nuclei already activated are to continue to monopolize the supply of water, all the remaining nuclei must be appreciably smaller than they. Slow upward movement in the cloud-forming region obviously tends to keep the supersaturation low and so to help the processes tending to produce few and large droplets. However, a semi-quantitative analysis indicated that the variations in the spectra of cloud nuclei from one air mass to another were probably the dominant factor in determining the size of droplets produced by condensation; on the basis of the fact that Aitken counts are higher over the continents than over the oceans, this view appeared
to be consistent with the observed predominantly maritime occurrence of warm rain. The most significant conclusion was that observations of the spectra of both cloud droplets and cloud nuclei would be required in order to resolve these basic questions in the physics of warm rain.

*Measurements of droplet spectra*—Cloud droplet spectra have been measured by several observers, but generally without any specific reference to the problems mentioned above. Observations were begun near Sydney in 1950 and evidence quickly appeared which seemed to be consistent with the suggestions drawn from the distribution of the occurrence of warm rain and the characteristic difference in Aitken count between oceanic and continental air; there appeared to be a significant difference in microstructure between Cumuli in maritime and continental air masses. Maritime Cumuli contained both fewer and larger droplets. The picture was, however, confused by other effects: as Zaitsev [1950] found, there is a tendency for the droplet concentration to decrease upwards in clouds. Furthermore, large Cumuli tend to have smaller average droplet concentrations than small ones. It so happened that among the earliest observations, the maritime Cumuli sampled were significantly larger than the continental Cumuli, and furthermore, the sampling runs in the former, for operational reasons, had not been evenly distributed throughout the vertical extent of the clouds, but were mostly taken in their upper regions.

It seemed therefore that some or all of the apparent difference between maritime and continental Cumuli could have been due to these two factors: (1) cloud size, and (2) the height of the sampling run above cloud base. In 1953 it was found that there was a serious deficiency in the method then in use for measuring droplet diameter, although the droplet concentrations were not affected. This difficulty was not finally solved until 1954. Further observations in continental clouds in southeastern Australia and of maritime clouds off the Australian coast, as far south as latitude 42°S, and off Hawaii in latitude 19°N left no doubt that the suspected contrast between maritime and continental Cumuli was real. The median droplet concentrations found were 45 and 228 droplets cm⁻³ respectively. These results are in very good agreement with those of Batton and Reitan [1957] who found median droplet concentrations over the central United States and over the Carribbean which are very close to the medians quoted above for continental and maritime clouds respectively. The observations taken in 1954 and later yielded drop sizes as well as concentrations, and confirmed the expected result that a strong negative association exists between droplet size and concentration. Figure 1 shows typical droplet spectra found in a maritime Cumulus, and

![Fig. 1](image_url)

Fig. 1—The mean droplet spectrum at four levels in a trade-wind Cumulus off the east coast of Hawaii on October 23, 1954; cloud base was at 2200 ft, cloud top at 9500 ft; there was some light rain in parts of the cloud, and turbulence was moderate; droplet concentrations (n) per cm³ and liquid water content (w) g per m³: 3250 ft, n = 74, w = 0.50; 4250 ft, n = 51, w = 0.53; 5250 ft, n = 27, w = 0.50; and 6250 ft, n = 30, w = 0.42
Figure 2 shows those found in a continental Cumulus of similar size.

Interpretation of measurements of droplet spectra—This result was of course quite consistent with the observed distribution of warm rain. Indeed, when account was taken of other observations such as those taken on the orographic cloud of Hawaii [Squires and Warner, 1957], there were five groups of clouds, namely 'dark Stratus,' Hawaiian orographic cloud, maritime Cumuli, Cumuli in transitional air masses, and continental Cumuli, arranged here in order of increasing median droplet concentration: this proved also to be the order of increasing colloidal stability, as measured by the depth of each cloud type which normally yields rain [Squires, 1958]. The recent calculations of Hocking [1959] on collision efficiencies indicate that the presence of a few relatively large collector drops is not sufficient to permit the coalescence process to proceed efficiently; it is essential that a significant amount of liquid water should be present in the form of relatively large drops, for only large drops can be efficiently collected. The examples given in Figures 1 and 2 show that in the continental case there is practically no cloud water in the form of drops larger than 20 microns diameter, while in the maritime case, the greater part of the liquid water is in the form of drops larger than 20 or even 25 microns diameter.

The relationship between the colloidal stability of clouids and their microstructure seemed therefore fairly clearly established; it remained to be seen whether the difference in microstructure between maritime and continental Cumuli could be explained in terms of the factors controlling the initial formation of the clouds. The suspicion mentioned in the introduction, that such contrasts in microstructure could only be explained in terms of differences in the population of cloud nuclei could not be directly confirmed, since no method was available for measuring this small but important fraction of the Aitken nuclei, which becomes activated at very small supersaturations (rather less than one percent), and consequently is effective in cloud formation. The only alternative was to examine the other possible factors which could influence the microstructure of clouds to see whether any of them seemed adequate to cause the observed difference between maritime and continental Cumuli [Squires, 1958]. Thus a comparison of flight notes of turbulence experienced in the two kinds of clouds indicated that over the clouds which had been sampled, the updraft velocities were on the average much the same, so that this factor could not be responsible. Further, assuming equal updrafts, it could be deduced from observed droplet spectra that the supersaturation in the body of the cloud was higher in maritime than in continental Cumuli, despite the effect of giant sea-salt nuclei in reducing the supersaturation in the former. Thus there seemed little possibility that these giant nuclei caused the difference in microstructure by depressing the maximum value of the supersaturation and reducing the number of nuclei activated. A rather surprising confirmation of this latter point was given by simultaneous measurements of spectra of droplets and of giant sea-salt nuclei, made in southeastern Australia, 200 to 600 mi inland.

Suppose for the moment that the cloud nucleus spectrum, that is, the largest few hundred nuclei, per cm$^3$, is invariant, apart from fluctuations in the giant sea-salt nucleus content (usually less than one nucleus per cm$^3$). Then it seems possible that in air masses which are rich in giant nuclei the supersaturation may be depressed by their great absorptive power for water. This effect, if large enough, could lead to a negative correlation between sea-salt nuclei and cloud droplet concentration, such as is suggested by the continental-maritime contrast. If, as in-
Fig. 3—Scatter diagram showing a positive correlation between the concentration of cloud droplets in inland Cumuli (n) and that of giant sea salt nuclei (N), near cloud base level; corresponding points for maritime Cumuli would fall in the hatched area.

Dedicated by the theoretical analysis quoted above, the effect of the giant nuclei is not great enough to achieve this, one would expect to find no correlation between the concentrations of sea-salt nuclei and cloud droplets. In fact, observations taken to test this hypothesis showed a strong positive correlation between sea salt and cloud droplet concentration as shown in Figure 3 [Squires and Twomey, 1958]. While emphatically indicating the absence of the negative correlation, and so confirming the theoretical conclusion that the giant nuclei are not responsible for the maritime-continental contrast, these observations indicated a positive association instead of a random scatter and so raised other questions which will be discussed later.

These indirect considerations, although fairly conclusively disposing of alternative explanations, could never establish in a final manner whether variations in cloud nucleus spectra were really the major cause of variations in cloud microstructure. For this purpose, there is no substitute for direct measurement of the spectrum of cloud nuclei.

**Measurements of cloud nuclei**—The formidable difficulties in the way of this measurement need little emphasis. While it is a simple matter to produce a small expansion instead of a large one, it is not easy to be sure of the desired initial condition of saturation with the accuracy required, which was estimated from the cloud spectra data as being equivalent to a few thousandths of a degree C. Again, droplet growth at small supersaturations is naturally quite slow, so that a new problem appears: that of maintaining the supersaturation for periods of minutes. This problem is made the more difficult by the fact that even apart from the inevitable wall effects, the mere growth of the droplets tends to reduce the supersaturation.

These problems were first overcome by Twomey [1959a] using a series of diffusion chambers, the top and bottom of which were held at slightly different temperatures, so that a region of small supersaturation was established, and maintained, in the centre of the chamber. Twomey [1959a] used diffusion between a water surface and a dilute aqueous solution of hydrochloric acid to establish and maintain a small supersaturation in a closed chamber. Using this method, Twomey [1959b] established that there are large and systematic differences between the cloud nucleus populations of maritime and continental air masses, at ground level, and was able to show that a representative surface nucleus spectrum, assuming an updraft speed of 1 m sec⁻¹, would produce a cloud with a droplet concentration of around 500 cm⁻³ in continental air and of about 60 cm⁻³ in maritime air. (Observed median droplet concentrations, 228 and 45 respectively.)

This result seemed to leave little doubt that the difference in the cloud nucleus population was indeed the main cause of the difference in cloud microstructure. The fact that the computed values lay somewhat above the observed ones could be explained if, as would seem likely, the surface layers were richer in cloud nuclei than the air entering the bases of Cumuli.

Simultaneous observations were needed to resolve this question, and these were taken late in 1958, mostly about 200 mi inland from Sydney. The result, comparing observed mean droplet concentrations found on one flight with those computed from the cloud nucleus spectrum observed in air sampled below cloud base, is shown in Figure 4 for an assumed updraft speed of 1 m sec⁻¹, and in Figure 5 for a speed of 10 m sec⁻¹. As will be seen, updraft has little effect;
the agreement is slightly better for the rather more realistic assumption of 1 m sec\(^{-1}\).

The surprisingly small effect of updraft speed on the computed droplet concentration is explained by the shape of the observed spectra of critical supersaturations of cloud nuclei. In most cases these spectra could be represented by an equation of the form

\[ N = C S^k \]

where \(N\) is the number of nuclei, per unit of volume, with critical supersaturations less than \(S\), and \(C\) and \(k\) are constants. On the basis of this equation, it has been shown [Twomey, 1959b] that the computed droplet concentration just above the activation region near cloud base is proportional to \(V^{2k/1+k}\), approximately, where \(V\) is the updraft speed. The observed spectra were fitted by values of \(k\) ranging from 0.2 to 0.4. Thus the computed cloud droplet concentrations were proportional to a power of \(V\) between 0.14 and 0.25. The conclusion of some authors that updraft speed has a more considerable effect on cloud microstructure than is indicated by Figures 4 and 5 is essentially due to assuming cloud nuclei spectra which correspond to values of \(k\) much greater than those which have been found to fit measured spectra. While observed cloud nuclei spectra are occasionally found which rise steeply over a portion of their range,

- **Fig. 4**—Comparison of mean droplet concentrations observed in Cumuli with those computed from observed spectra of cloud nuclei, with an assumed updraft speed of 1 m sec\(^{-1}\); the dashed line represents exact agreement between observed and computed values.

- **Fig. 5**—Comparison of mean droplet concentrations observed in Cumuli with these computed from observed spectra of cloud nuclei, with an assumed updraft speed of 10 m sec\(^{-1}\). The dashed line represents exact agreement between observed and computed values.

the use of values of \(k\) of the order of 1 or 2 is unrealistic. In some instances the empirical law of Junge [1958] for nucleus size distribution has been invoked to deduce a supersaturation spectrum for cloud nuclei. This corresponds, for soluble particles, to \(k = 2\). However, Junge’s law was not intended to represent the supersaturation spectrum, but the size distribution of the heterogeneous natural aerosol particles down to 0.1 \(\mu\). The critical supersaturation of a nucleus is influenced by its chemical composition and surface properties, as well as by its size. Even micron-sized particles will remain unactivated at the slight supersaturations occurring in clouds if they are at all hydrophobic.

**Conclusion**—These latest observations seem to provide definite confirmation of the view that cloud microstructure is primarily determined by the spectrum of cloud nuclei. It appears that some continental surfaces at least are source regions of cloud nuclei, so that Cumuli forming in air masses which have passed over them have large droplet concentrations and consequently consist of small droplets. They are therefore relatively inefficient in releasing warm rain, which can form only by the coalescence process.

In the light of this result, a tentative explanation may be offered for the positive correlation observed between cloud droplet concentrations and sea-salt content found at places some
DISCUSSION

hundreds of miles inland. It has been observed that the most continental clouds, that is those with the highest droplet concentrations, occur during dry weather. In such weather also, the highest counts of cloud nuclei have been observed; unfortunately, simultaneous observations have not been obtained in these conditions. The salt content of an air mass moving inland is redistributed by vertical mixing during dry weather, so that the concentration in sub-cloud levels is somewhat reduced; but in wet weather the greater part of the salt is washed out. Thus when the air mass concerned has not rained since leaving the coast, there may be a tendency for both the salt and cloud droplet concentrations to be high. As regards the association between dry weather and high cloud nuclei and cloud droplet concentrations, it may be noted that dry weather would favor the formation of nuclei from fires or perhaps by the drying out of soluble salts [Twomey and McMaster, 1955].

It seems almost self evident that the release of particles into the atmosphere from the land surface would be most efficient in dry weather. For when the surface is dry, and evaporation slight, the surface temperature rises to very high values during the day, so that the surface of the soil and the air layer above it are very dry, while the lowermost layer of the atmosphere is strongly unstable.

References


Squires, P., and J. Warner, Some measurements in the orographic cloud of the island of Hawaii and in trade wind Cumuli, Tellus, 9, 475-494, 1957.


Twomey, S., Condensation nuclei at low supersaturations, pt I: The chemical diffusion method and its application to atmospheric nuclei, Geofisica pura e applicata, 1959a (in press).

Twomey, S., Condensation nuclei at low supersaturations, pt II: The supersaturation in natural clouds and the variation of cloud droplet concentration, Geofisica pura e applicata, 1959b (in press).


Discussion

Dr. C. E. Junge—This paper seems to me a very important one and I would like to congratulate the authors for their work. I would like to comment on two items:

(1) Concerning the supersaturation spectra, the reason for the difference in these spectra in continental and maritime air is the difference in the size distribution of the condensation nuclei. The relationship between supersaturation and size can be calculated on the basis of the well-known growth curves of salt droplets. This relationship is somewhat influenced by the chemical composition of the nucleus but not too much as long as a good portion of the nucleus consists of soluble salts. This relationship was recently confirmed indirectly by measurements of the growth curves of nuclei with relative humidity down to sizes of about 0.01μ radius (Ord, Hurd, Hendrix, and Junge, The Behavior of Condensation Nuclei under Changing Humidities, J. Met., 15, 240-242, 1958). The relationship between supersaturation and size can therefore be regarded as reliable and can be used to convert nuclei size spectra into supersaturation spectra.

In our measurements of the size distribution of nuclei, which were confirmed by others, we
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found a profound difference between ocean and continent. Figure 6 gives the essence of these observations. Whereas the difference in concentration is comparatively small for the giant particles larger than 1µ radius, there is an increasing lack of nuclei over the ocean when approaching 0.1µ radius. Unfortunately, there are no data below 0.1µ for the ocean, but there is some indication for a secondary increase as indicated by the part with a question mark.

Now, the nuclei between 1 and 0.1µ are the most important ones for cloud-droplet formation. In Figure 7 we converted these size distributions into supersaturation spectra and compared them with the observed supersaturation spectra of Twomey. It can be seen that the agreement is good considering the fact that the two sets of data were obtained in completely different geographical locations. The fact that our maritime curve is still lower than the corresponding values of Twomey may indicate some continental aerosol residues in the Australian maritime air where the data were obtained. Twomey’s data were taken from a manuscript submitted to the Geofísica Pura et Applicata, which Dr. Squires made available to me.

(2) The role of sulfate as a substance of condensation nuclei in areas which are normally considered completely unpolluted, and a possible difference between the southern and northern hemispheres. We found that about 50% of the atmospheric sulfur in the northern hemisphere is due to industrial activities. Since we know that sulfate is an important constituent of the condensation nuclei, it may be that during the industrial development within the last 100

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![Figure 6](image1.png)

**Fig. 6**—Size distribution of aerosols (1) over land, and (2) over the ocean; the curves give, in contrast to earlier presentation, the total number of particles larger than the indicated value of the radius (cumulative figures)

![Figure 7](image2.png)

**Fig. 7**—Supersaturation spectra. (1) calculated from Curve (1) in Fig. 1; (2) calculated from Curve (2) in Fig. 2; measurements by Twomey over Australia: (3) drought conditions; (4) continental air masses; (5) maritime air masses; the curves give the total number of particles active above the indicated supersaturation
years a gradual increase of condensation nuclei occurred which may have had some effect on cloud and precipitation physics along the lines Dr. Squires has indicated. The importance of sulfate for the composition of condensation nuclei in remote places is for instance well demonstrated by the recent finding that in Greenland the concentration of sulfate is higher by one order of magnitude than that of all the other constituents, including sea salt.

If it is true that the northern hemisphere is, as a whole, polluted to such an extent as indicated, it may be that the difference between ocean and continent with respect to nuclei concentration is more pronounced in the southern hemisphere where this pollution must be negligible, and the original natural conditions still prevail. This should result in a corresponding difference with respect to cloud behavior. I wonder if anybody has thought about this interesting problem?

Dr. P. Squires—I am very happy that there are some observations in the northern hemisphere which confirm this.

Dr. L. J. Batten—It is true that the concentrations measured by the University of Chicago were similar, but there is one thing which I wanted to ask about. As you know, in the maritime distribution there was a bimodality which was evident after averaging as well as in virtually every individual observation. There was one maximum in frequency at a radius of about 15 microns and another at about 5 microns. I wonder if there is anything in your discussion which would suggest that there should be a difference in the shape of the curves as well as in the concentration.

Dr. Squires—The methods we have used are not good below five microns. The crux of the matter, however, is how many drops share the liquid water effectively, and the ones down around four or five microns are perhaps not important.

Mr. Jerome Naiman—The problem of drought, on which I have been working for some 25 years, is very fascinating. Its incidence over the United States in summer is associated with the large-scale features such as I described earlier; namely, the presence of the upper-level continental anticyclone with its neighboring anticyclones in the Atlantic and Pacific. If the continental cell vanishes, it does so only temporarily so that it recurs persistently. This happens so frequently that a forecast rule is: "All signs fail in times of drought." Some recent studies of mine dealing with the problem of its maintenance suggest some influence of the condition of the soil below.

(Disussed in the paper Persistence of Mid-Tropospheric Circulations between Adjacent Months and Seasons, to be published in the forthcoming Rossby Memorial Volume, Stockholm, 1959.) But the only way I guessed this influence could operate was through a thermodynamic mechanism involving the availability or non-availability of the latent heat of vaporization, depending on the dryness or wetness of the underlying soil. Dryness would perpetuate the upper anticyclone through thermal heating. However, I am intrigued by the possibility implied by Dr. Squires that under very dry conditions rain might be inhibited because the clouds, once formed, contain too many nuclei and perhaps of an unfavorable kind. It could be that this is one of the feedbacks of nature to provide a sort of memory for long-term situations. Certainly if we isolate the other factors such as I mentioned, and are able to do this quantitatively, the nuclei factor should receive consideration as a feedback.

Dr. R. Wexler—Admitting the importance of the nuclei count, other factors may also cause considerable differences between land and sea clouds. The greater moisture content of the air over the sea should cause a greater liquid water content per unit depth of cloud than over the land. Entrainment of drier air into the cloud over the land may inhibit rain development. Measurements in Cumulus over both land and sea, reported by Batten and Reitan in the last Woods Hole conference, show only slight differences in cloud drop size between those which developed echoes and those which did not. Hence, the failure of rain to develop from these clouds cannot be ascribed to the lack of large drops.

Dr. Squires—In this series of measurements, the liquid content was perhaps 20 to 30% higher in the maritime clouds. This slight difference would not go far towards explaining the radically different ability of maritime and continental clouds to produce rain by the coalescence process. Some much more drastic effect, such as Hocking's calculations suggest, seems to be called for.

The differences which cause some clouds to rain while others in the same air mass do not are probably too minute to be observed. At the present stage of knowledge, we merely wish to contrast continental clouds, which are so reluctant to form rain by the coalescence process,
with maritime clouds, which are so ready to do
so.

Dr. W. E. Howell—There seems to be a real

difference here between conclusions as to the
effect of vertical velocity. Computations by
Neiburger and myself, among others, seem to
show that different nuclei spectra produce clouds
in which the difference is much less marked be-
cause of the compensating effect that, if there
are few nuclei, the supersaturation is forced
higher, and a larger proportion of them is ac-
tivated. These computations in general led to
clouds more uniform than are found in nature,
while your figures show the computed droplet
concentration more dispersive than the natural
ones. I am led to wonder whether new classes
of nuclei were brought into your computations
at close enough intervals to follow the natural
process.

Dr. Squires—The compensating effect noted
by Dr. Howell certainly exists; with a maritime
spectrum of nuclei, the supersaturation maxi-


mum is twice or three times that computed for
a continental spectrum. However, the observed
nucleus spectra show a much slower increase in
accumulated number with increasing supersatu-
ration than has usually been assumed in calcula-
tions of the condensation process. As a result,
an increase in the supersaturation consequent
upon an increase in the assumed updraft speed
does not result in so large an increase in the
number of activated nuclei. The form of the ob-
served nucleus spectrum is such that the com-
puted cloud droplet concentration increases only
as about one-fifth power of the updraft speed.
As regards the last point, the droplet concentra-
tions were estimated by an analytical method
which gives an upper and a lower bound. The
starting point was, of course, a curve fitted to
the observed nucleus spectra.

Dr. Howell—I would also like to question
further your assumption that vertical velocities
are roughly the same in maritime and continen-
tal Cumulus. Malkus has shown that maritime
Cumulus are generally rootless, without convect-
ive currents extending below the cloud base, the
energy for the convection being supplied by the
condensation within the cloud and the convective
currents becoming stronger as they rise
farther above the cloud base; whereas with con-
tinental Cumulus the convective currents origi-
nate near the ground and, especially in small
Cumulus, often decrease in intensity above the
cloud base. Even within the cloud, convective
currents in maritime Cumulus are generally cited
as being weaker than in continental Cumulus.

Finally it is my experience that the character-
istic differences between maritime and continen-
tal clouds appear even within one and the same air
mass between clouds that form over the sea and
those that form nearby over small tropical
islands, where vertical velocities are perhaps the
principal differences involved.

At Mount Washington we have found nearly
the same differences in droplet concentration be-
 tween maritime and continental trajectories that
you cite, namely, about 300 per cm$^2$ in west
winds and 60 per cm$^2$ in east winds; but mea-
urements of cloud thickness and liquid water
content seem to show that the high concen-
trations occur in clouds formed by slow warm-
front lifting. And so I do not think that vertical
velocity differences should be ruled out yet as a
major factor in determining the droplet concen-
tration in clouds.

Dr. Squires—It is probably true that up-
drafts are somewhat stronger in continental
Cumulus than in maritime clouds of the same ver-
tical extent. Our maritime clouds were, on the
average, a little deeper than the continental ones.
In any case, a comparison of subjectively clas-
sified turbulence is an extremely crude way of
comparing updraft speeds. However, because of
the slight dependence of droplet concentration
on updraft speed, this matter is not one of
great significance. Thus, if the average updraft
speed were ten times greater in continental Cu-
muli, the nucleus spectra being the same, the
ratio of droplet concentrations (continental/
maritime) should be about 1.5; the observed
ratio is about 5.

Concerning Dr. Howell’s observations of clouds
over and near tropical islands, the appearance
of a cloud and its chance of producing rain both
depend, no doubt, on the pattern of convection
below and within it, as well as on its micro-
structure. It is relevant to point out that, with
the exception of one point, all the data of Figures
4 and 5 were obtained one to two hundred miles
inland.
A Statistical Study of Cloud Droplet Growth by Condensation

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Abstract—A droplet spectrum is defined in terms of the moments of its frequency distribution, and equations are derived for the rate of change of these moments. Effects of dissolved salt are not considered; this limits the application of the theory to regions well above cloud base. It is shown that the supersaturation adjusts itself towards a quasi-steady value within a time given by $C(N\bar{r})^{-1}$ where $N$ is the number of droplets per cm$^3$, $\bar{r}$ is the average droplet radius in cm, and $C$ is a coefficient of size order one cm$^{-2}$. The basic parameter used to characterize the cloud development is the average growth rate of droplet mass. For very slow rates the droplet size spectrum has a tendency to widen, but at higher growth rates a contraction of the spectrum occurs. In the limit the linear width of the spectrum is inversely proportional to the average radius and the standardized form of the spectrum is invariant.

Introduction—The development of droplet spectra has long been a major concern of people engaged in cloud-physics research. Several investigators have constructed computation schemes for studying in detail how a droplet spectrum is formed on a specified nucleus population contained in a parcel of humid air, as the air is cooled past its point of saturation. The causal chain in such a model is depicted in Figure 1. One may divide the variables of the system into two classes, characteristic of the mesoscale and the microscale of the development. We shall refer to the mesoscale group all those properties of the bulk air, which have to do with the hydrodynamical development on the scales of convective elements, that is, pressure, temperature, vertical velocity, and content of water in liquid and in vapor state. To the microscale we refer the details of the droplet size distribution and the chemical properties of the aerosol, as well as any turbulence on scales that may interfere directly with the coalescence process.

It is evident from the work of previous investigators that, except at the very onset of the cloud formation, the rate of change of the liquid water content of the cloud is almost exactly that which corresponds to complete utilization of the available water vapor. In other words the turnover of water is very large compared to the amounts required to effect what changes in supersaturation that may occur [Howell, 1949; Mason, 1957; Mordy, 1959; Squires, 1952]. Since the warming of the bulk air by released latent heat and the total amount of supported liquid water are the only feedback effects from the condensation process on the mesoscale dynamic processes, a detailed knowledge of the microphysics is not required for the study of these scales. On the other hand the concept of complete utilization of the available water vapor provides an integral constraint on the development of the droplet spectrum, which can be formulated in terms of the mesoscale development.

If a system is governed by integral constraints, this forms an ideal basis for an attempt to study its development in some suitable statistical terms. The present paper is an example of such an approach applied to a simple system with some resemblance to what occurs in natural clouds, namely, a droplet population with negligible salt content, embedded in an air parcel subject to steady cooling.

The basic model—Consider a parcel of humid air containing a nearly uniformly distributed population of cloud droplets. Let the size distribution of the droplets be defined by a frequency distribution $n(r)$, such that $Vn(r)dr$ is the number of droplets in the size range $r$ to $r + dr$, to be found in an air sample of volume $V$. We shall denote averages taken with respect to this distribution by a bar (for example, $\bar{r}$). Let us further introduce a normalized size variable $x$, defined by

$$r = \bar{r}(1 + x)$$  \hspace{1cm} (1)

The distribution $n(r)$ is completely defined by its moments $N_v$,

$$N_v = \int_0^\infty r^v n(r) \, dr \quad (v = 0, 1, 2, \cdots) \hspace{1cm} (2)$$

Our goal is to be able to describe the develop-
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The time derivative of an average of a function of \( r \) will depend in part on the rate of growth of the individual particles, and in part on any changes in the distribution that may occur because of turbulent transfer or settling of particles, or by coalescence. We shall assume that particles are conserved in our air parcel, hence only the particle growth by condensation is considered. In that case the operations of averaging and differentiation with respect to time are inter changable in order, so for any differentiable function \( f(r) \)

\[
\frac{d}{dt} f(r) = \frac{d}{dt} \frac{f(r)}{r} \tag{3}
\]

The growth equation and the basic integral constraint—The equation governing the growth of a cloud droplet in a cloud of moderate droplet concentration can be stated in the form

\[
\frac{dr}{dt} = \frac{A}{r} \left( S - \frac{B}{r} \right) \tag{4}
\]

The definition of the different symbols is briefly discussed in Appendix I. In order to simplify the arithmetic operations \( l \) is neglected. This step may also be defended on the grounds that recent experimental work by the author (unpublished) indicates that its value is less than \( 2 \times 10^{-4} \) cm, which limits its effects to the initial stage of cloud development. So our basic equation is

\[
\frac{dr}{dt} = \frac{A}{r} \left( S - \frac{B}{r} \right) \tag{4a}
\]

Our next aim is to see how the supersaturation \( S \) is coupled to the cooling rate. The total rate of production of liquid water in a unit volume of cloud is

\[
q = \int_0^\infty n4\pi r^2 \frac{dr}{dt} \tag{5}
\]

Substitution for \( dr/dt \) from (4a) yields

\[
q = 4\pi N_0 A (S^2 - B) \tag{5a}
\]

The rate of change of the supersaturation is determined by the combined action of temperature change and water consumption. Denoting the mixing ratio of vapor to dry air by \( w \), and saturation conditions by subscript \( s \)

\[
\frac{dS}{dt} = \frac{dw}{dt} - \frac{w}{w_s} \frac{dw}{dt} = \frac{1}{w_s} \frac{dw}{dt} - \frac{w}{w_s^2} \frac{dw}{dt} \tag{6}
\]

Now \( \rho_c \frac{dw}{dt} \) where \( \rho_c \) is the density of the dry air, is clearly identical with \( q \) in Eq. (5). Furthermore \( w_s \) is a function of pressure and temperature, and since the system we consider does not exchange heat with its surroundings, it is a function of the pressure and of the total amount of condensed water. So (6) may be rewritten as

\[
\frac{dS}{dt} = \frac{-4\pi N_0 A (S^2 - B)}{\rho_c w_s} = \frac{w}{w_s^2} \frac{\partial w}{\partial p} \frac{dp}{dt} + \frac{\partial w}{\partial w} \frac{dw}{dt} \tag{6a}
\]

We observe that \( \partial w/\partial w < 0 \) and \( \partial w/\partial p > 0 \), so that (6a) is of the form

\[
\frac{dS}{dt} = -(1/\tau) [S - g(p, w)dp/dt] \tag{7}
\]

where \( \tau \) and \( g \) are positive definite functions of \( p \), \( w \), and \( \tau \), and \( dp/dt \) is prescribed by the particular mesoscale system of which our air parcel is thought to be a part. The time constant \( \tau \) is seen to have the form \( c/N_0 \hat{\varphi} \), where the coefficient \( c \) is a function of \( p \) and \( w \). Its evaluation is discussed in Appendix II. We find that its magnitude is about 1 sec/cm² for fairly typical conditions obtained in low clouds. Hence our time constant will normally be much less than one minute, while the convective processes in a cloud generally have time scales of several minutes or more. It follows
that the actual supersaturation should be very well approximated by
\[ S = g \frac{dp}{dt} \] (8)

Since the supersaturation is so sensitive to the actual rate of condensation, we may conclude that the water vapor will be condensed as fast as it is being made available by the cooling. The total rate of condensation is thus a function of the external influences on our air parcel, and essentially independent of the details of the condensation process. If we denote this rate by \( q_0 \), then from (5a)
\[ S \tau - B = q_0 / 4 \pi N \] (9)

This relation holds as long as the implied variations in \( S \) are slow in a time scale defined by the time constant \( \tau \).

The development of the droplet spectrum—The growth equation for a single drop as stated in (4a) can be transformed into a set of equations for the statistical parameters of the droplet spectrum by the following procedure. First we introduce the variables \( \tilde{r} \) and \( x \), as defined above,
\[ \frac{d}{dt}(1 + x) \frac{d\tilde{r}}{dt} + x \frac{dx}{dt} = \frac{AS}{\tilde{r}(1 + x)} \left( S - \frac{B}{\tilde{r}(1 + x)} \right) \] (10)

We shall limit ourselves to cases where \( r \leq 2\tilde{r} \), that is, the upper limit for the droplet size is twice the mean radius. In that case the fractions in (10) can be developed in geometric series
\[ (1 + x) \frac{d\tilde{r}}{dt} + x \frac{dx}{dt} = \frac{AS}{\tilde{r}} \sum_{\mu=0}^{\infty} (-x)^{\mu} - \frac{AB}{\tilde{r}^2} \sum_{\nu=0}^{\infty} (1 + \nu)(-x)^{\nu} \] (10a)

After rearrangement of terms
\[ (1 + x) \frac{d\tilde{r}}{dt} + \tilde{r} \frac{dx}{dt} = \frac{AS}{\tilde{r}} \sum_{\mu=0}^{\infty} (-x)^{\mu} - \frac{AB}{\tilde{r}^2} \sum_{\nu=0}^{\infty} (1 + \nu)(-x)^{\nu} \] (10b)

After averaging all terms in (10) yields
\[ \frac{1}{3.1} \frac{d\tilde{r}^2}{dt} = S \tau - B \] (11)
\[ + \sum_{p=2}^{\infty} (-1)^p [S \tau - (1 + \mu)B] x^{p} \]

This equation is used to eliminate \( d\tilde{r}^2/dt \) from (10b). If now the equation is multiplied through with an arbitrary power \( x^{\mu-1} \) of \( x \), and averages are taken again, then the following expression is found for the rate of change of any \( x^\mu \)
\[ \frac{\tau^\mu}{A} \frac{d\tilde{r}^\mu}{dt} = -(2S \tau - 3B) \tilde{r}^\mu \]
\[ + \sum_{p=2}^{\infty} (-1)^p [S \tau - (1 + \mu)B] x^{p} \]

It was shown above that the expression \( S \tau - B \) is a function of the enforced average rate of mass growth. We shall introduce the notation
\[ S \tau - B = \sigma B \] in (11) and (12), and so, with some transformation of the derivatives, our system of equations is
\[ \frac{\tau^\mu}{AB} \frac{d \ln \tilde{r}}{dt} = \sigma + \sum_{p=2}^{\infty} (-1)^p (\sigma - \mu) x^{\mu} \]
\[ \frac{\tau^\mu}{AB} \frac{d \ln (\tilde{r}^\mu)^{1/\mu}}{dt} = -(2\sigma - 1) \]
\[ + \sum_{p=2}^{\infty} (-1)^p (\sigma - \mu) \frac{x^{p-1} + x^{p+1}}{x^{p}} \]

This is an infinite set of equations, each containing an infinite number of terms. The possibility of deriving solutions to the system depends upon the rate of convergence of the series. Some conclusions may be drawn however, without actually going to the problem of solving the equations. It is first of all clear that the condensation process is a powerful agent for changing droplet spectra only when the droplets are small. All rates are inversely proportional to \( \tilde{r}^2 \). It is further evident that spectra of the same shape, in the sense that the set of \( x^n \) is identical between them, will develop in the same way if subject to the same forcing in terms of the specific growth rate \( \sigma \). A doubling, say, of the mean radius, will carry with it the same change in spectral shape irrespective of whether it means growth from 2 to 4, or from 20 to 40 microns. The time scale would differ by a factor 1000, however.

We shall now have to consider the actual values of the parameter \( \sigma \). Figure 2 shows how it is related to vertical velocity and total number of active particles, for the pressure 900 mb and temperature 10°C. The relation is rather insensitive to variations in \( p \) and \( T \) within the range of meteorological interest. It is obvious that values sufficiently low to change the sign of the leading terms in (13) would hardly occur. Points corre-
responding to cases studied by Howell [1949] and Moody [1950] have been entered in the diagram. It is interesting to note that Howell’s Case 3 gives a value of $\sigma = 1.5$, which brings it near to the critical value. This case, somewhat inexplicably, according to Howell, produced a very wide droplet spectrum. The condensation nuclei were considered in his computation, and they would obviously tend to widen the resulting droplet distribution, but it seems obvious that the parameter $\sigma$ should be of comparable importance also in case the dissolved salt in the droplets is considered.

The fact that the parameter $\sigma$ normally is expected to be much larger than unity lends support to the concept that (13) might be approximated by retaining only the first term on the right-hand side in each equation. The equations can then be solved, to give the relations

$$
\begin{align*}
\frac{d\sigma}{dt} &= \frac{3.333t}{4}
\end{align*}
$$

(14)

The variable $x$ was normalized with respect to $\bar{r}$. If we replace it by another variable $\xi$, defined by

$$
\xi = x/\sqrt{x^2} = (r - \bar{r})/\sqrt{(r - \bar{r})^2}
$$

then (14) gives

$$
\left(\frac{x^n(t)}{x^n(0)}\right)^{1/n} = \left(\frac{\bar{r}(t)}{\bar{r}(0)}\right)^{(n-1)/\sigma}
$$

(16)

that is the standardized distribution is invariant within this approximation, and the linear width of the spectrum, as represented by the standard deviation of $r$ from its mean, is proportional to $\bar{r}^{1+1/\sigma}$.

It may seem surprising that the capillary term in the growth equation should play the same role irrespective of the size of $\bar{r}$, but this is explained by the fact that the supersaturation is inversely proportional to $\bar{r}$, and so is the capillary term. If the ambient supersaturation is high, then the difference in growth rates between various members of the droplet population is controlled by the factor $r^{-1}$ in the growth equation. But if the supersaturation is low, then the capillary term may overcompensate the effect of this factor. The influence of the parameter $\sigma$ on the individual droplets is seen clearly if we write the growth equation (4a) in the form

$$
\frac{1}{AB} \frac{dt}{dt} = \frac{\sigma + 1}{\bar{r}^2} - \frac{1}{\bar{r}^2}
$$

(17)

Fig. 2.—The growth forcing function $\sigma$ as a function of particle number and vertical velocity; the points marked represent conditions in computations made by Howell [1949] and Moody [1959]; lines connect points representing identical nucleus spectra.

Differentiation of the right-hand side with respect to $t$ gives the result that the growth rate is largest for the droplet size $r = 2\bar{r}/(\sigma + 1)$. It is interesting to note that the condition for contraction or expansion of the linear width of a fairly narrow spectrum is identical with the condition that the radius of maximum growth rate be smaller or larger than the mean radius.

Possibilities for further development.—The model considered in this paper is very simple. The results are equally simple and straightforward. It seems possible, therefore, that the present formulation of the problem of droplet growth by condensation could be used as a starting point for more sophisticated studies of cloud development. It is my opinion, that such extended models should not attempt to treat this particular problem in greater detail, but that one should build up the more complicated system by blocks representing its different aspects, each of them being simplified in similar extent as the present study. The cloud base problem, that is, the question of finding the most economic treatment of the initial cloud formation and the question of which characteristics of the nucleus spectrum one should use, in order to describe its behavior in clouds in the most pregnant way, seems to be one that should be attacked along these lines. In the coalescence problem the obvious approach is a generalization of the ordinary ‘basic-cloud-versus-growing-raindrop’ approach by consideration of a basic cloud droplet population, described essentially with the method advanced here, and a population of droplets growing by coalescence alone. One would also like to approach the problem of whether dropl
size distributions exist that are stable in a statistical sense, like the spectral-energy distribution in statistical turbulence theory.

Acknowledgments—The approach used in this paper was inspired by the lectures on statistical methods in hydrodynamics given by Phil D. Thompson at the International Meteorological Institute in Stockholm in 1958-59. Numerous discussions about problems in cloud physics with Wendell A. Mordy, of the same Institute, provided the stimulus necessary to complete it.

APPENDIX I

The Growth Equation

The equation for droplet growth by condensation has been discussed extensively by several authors. We shall only briefly state the results, a more thorough discussion with references is found in the paper by Neiburger and Chien in this volume.

One considers the spherically symmetric case of diffusion of water vapor onto a droplet of radius $r$. The steady state provides a satisfactory approximation for natural cloud conditions. The boundary conditions are given by the vapor density in the surrounding air, taken to be valid at infinite distance from the drop, and the equilibrium value at the droplet surface, with due account taken of the capillary effect, and of the heating of the drop by the condensation. In addition one has to consider the vapor-pressure jump at the phase boundary. In the case where the exponential influences of heating and surface tension can be linearized equation (4) becomes (nuclei neglected)

$$\frac{dr}{dt} = \frac{D_p S}{1 + \frac{ML^2 D_p}{RT^2}} \left( \frac{1}{r} + \frac{2 \gamma M}{RT} \right)$$

where the symbols used are

- $D$: diffusion coefficient for water vapor in air
- $K$: heat conductivity of air
- $L$: latent heat of condensation for water
- $M$: molecular weight of water
- $R$: universal gas constant
- $S$: degree of supersaturation
- $T$: absolute temperature
- $l$: vapor pressure jump parameter
- $r$: droplet radius
- $\gamma$: surface tension of water
- $\rho_s$: vapor density at saturation

\[ \frac{1}{\tau} = \left[ 1 - \frac{\partial w_s}{\partial w} \right] \frac{4\pi N_s \gamma D}{1 + \frac{ML^2 D_p}{RT^2}} \]

If we substitute for $A$ from Appendix I, and transform the partial derivative inside the brackets, then

$$\frac{1}{\tau} = \left[ 1 - \frac{\partial w_s}{\partial T} \right] \frac{4\pi N_s \gamma D}{1 + \frac{ML^2 D_p}{RT^2}}$$

But $(\partial T/\partial w)_p = -L/c_p$ and $(\partial w/\partial T)_p = ML/RT^2$, so

$$\frac{1}{\tau} = \left[ 1 + \frac{MLw}{c_p RT^2} \right] \frac{4\pi N_s \gamma D}{1 + \frac{ML^2 D_p}{RT^2}}$$

At $\rho = 900$ mb and $T = 283^\circ K$

\[ \tau = 0.5/N_s \gamma \]

References


Discussion

Dr. M. Neiburger—I do not want to pretend that I followed the details of the computation, but I think this approach looks like the sort of thing to which we will have to turn. Now, I am not sure I see how the variations with time of the shape of the spectrum are taken into account, or how one goes back to the variations of shape from the time spectrum if the spectrum has to be char-
acterized by an increasing number of modes because of the change in shape. But I do agree, as in the case of Dr. Mason and his colleagues in their paper in the Transactions of the Faraday Society, that we have to go to some representation of the spectrum other than the complete details. I had thought of the possibility of using some sort of other representation in the computation we did, to see changes in the results these had with time.

Dr. Claus Rothe—The change in time of the spectrum shape is given by the relative changes in magnitude of the moments of the distribution. The indication here is that any such changes caused by condensation alone would be very slight. The formation of bimodal or more complicated distributions has to be attributed to mixing and differential settling, and to the partition of the nucleus spectrum into an actively growing part and a population of stable nucleus droplets. Such a partition will always occur at cloud base, but the nucleus droplets would normally be very difficult to observe because of their small size, and most observed cases of bimodality do not fall into that category. As to the number of moments that have to be considered, it would obviously have to be quite large, if an accurate description of something like the shape of a bimodal distribution is desired. But the situation might be different if we want to apply our model to the development of a physical property of the spectrum other than shape, since a property like average mass or scattering power is defined by one or a few of the moments of relatively low order.

Dr. W. Hirschfeld—I would like to comment on the particular method you chose for representing the distribution. You chose an average over number. I think this may be mathematically attractive, but physically such averages are not very attractive because the distributions we know and measure are always incomplete, notably at the small-radius end. Any ignorance there would lead to serious difficulties in evaluating a 'mean radius.' Mean radii, weighted by particle mass, surface area, or scattering power, or in fact any power of the radius higher than the first are largely free of this defect.

Dr. Rothe—The set of zero-centered moments of the distribution function, as defined here, is equivalent to the set of mean radii derived with an arbitrary power of the radius as a weighting function (see Eq. (2)). If one wants to study the time variation of a property like the scattering power, he has to make use of the proper moments of the distribution. If the physical property with which he is concerned involves the radius taken to a power equal to or higher than that involved in the basic averaging process, then he is all right, otherwise he is certainly worse off than if he had included the moments of lower order. Take, for instance, the supersaturation. It is seen to be inversely proportional to the linear mean radius. Now this fact is a consequence of the physical model used, and not of the particular method of mathematical analysis applied to it. If we have available a measured droplet spectrum, together with data on the cooling rate and other pertinent factors, then our best estimate of the ambient supersaturation in the cloud would obviously be founded on the observed mean radius, in spite of the fact that this is less well defined than any of the weighted averages suggested. So I do not go along with the direct implications of your comment, but if I may interpret it as a warning against the pitfalls of uncritical combination of mathematical method and physical reasoning, then I would support it wholeheartedly.
The Nucleation and Growth of Ice Crystals

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Abstract—Experiments to test the ice-nucleating ability of a wide variety of natural mineral dusts suggest that kaolinite is probably the main source of atmospheric ice nuclei, and that a number of silicate nuclei may be preactivated. The nucleating properties of artificial ice nuclei, mainly inorganic compounds, are discussed in relation to their solubility, crystal structure and, especially, their surface structure. It is shown that nucleation occurs preferentially at steps and other special sites on the substrate surface. A particular nucleating substance has a critical temperature above which it may act only above water saturation but below which it may act as a sublimation nucleus provided the air is supersaturated relative to ice by a well-defined critical value.

Ice crystals growing as hexagonal plates on the surface of corellite show interference colors which give a measure of their thickness. Colored growth layers may be seen spreading across the crystal surface, careful measurement of which provides new information on the mechanism of crystal growth.

The growth of snow crystals in a diffusion-cloud chamber shows that, over the temperature range 0 to −50°C, the crystal habit undergoes five changes, and these are controlled primarily by the temperature and not the supersaturation of the vapor. Their habit changes are not affected by the pressure and nature of the carrier gas, or by the presence of aerosols, as reported by Japanese workers.

Natural and artificial ice nuclei—In an attempt to discover the nature and origin of atmospheric ice nuclei, we have tested the ice-nucleating ability of various types of soil particles and mineral dusts. Of the 30 substances tested, 16, mainly silicate minerals of the clay and mica groups, were found to produce ice crystals in supercooled clouds at temperatures of −15°C or above, and of these, seven were active above −10°C (see Table 1). The most abundant of these is kaolinite with a threshold temperature of −9°C. Ten natural substances, again mainly silicates, were found to become more efficient ice nuclei having once been involved in ice-crystal formation, that is, they can be pre-activated or 'trained.' Thus, ice crystals grown on kaolinite nuclei, which are initially active at −9°C, when evaporated and warmed to near 0°C in a dry atmosphere, leave behind nuclei which are thereafter effective at −4°C. Particles of montmorillonite, another important constituent of some clays, and which are initially inactive even at −25°C, may be pre-activated to serve as ice nuclei at temperatures as high as −10°C. It is suggested that although such particles can initially form ice crystals only at Cirrus levels, when the ice crystals evaporate they will leave behind some 'trained' nuclei which may later seed lower clouds at temperatures only a few degrees below 0°C. On this hypothesis, the fact that efficient nuclei are occasionally more abundant at higher levels would not necessarily imply that they originate from outer space. Indeed, in view of our tests on particles of stony meteorites, produced both by grinding and by vaporization, which show them to be ineffective at temperatures above −17°C, it appears that atmospheric ice nuclei are predominantly of terrestrial origin, with the clay minerals, particularly kaolinite, being a major source. This work is described in greater detail by Mason and Maybank [1958].

Although a good deal of work has been carried out in different laboratories on the ice-nucleating ability of a wide variety of chemical compounds there has been little agreement in the results. Careful tests in our laboratory indicate that many of the published results are spurious because of the presence, in the air or the chemicals, of small traces of silver and free iodine, leading to the formation of silver iodide. If all such trace impurities are removed, many of the substances that have been suggested are found to be quite ineffective.

The first six substances listed as artificial nuclei in Table 1, all being practically insoluble, are active to the extent of about one particle in 109 producing an ice crystal at the indicated
NUCLEATION AND GROWTH OF ICE CRYSTALS

Table 1—Substances active as ice nuclei

<table>
<thead>
<tr>
<th>Natural nuclei</th>
<th>Artificial nuclei</th>
</tr>
</thead>
<tbody>
<tr>
<td>Substance</td>
<td>Crystal symmetry</td>
</tr>
<tr>
<td>Covellite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Vaterite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>β-Tridymite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Magnetite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Kaolinite</td>
<td>Triclinic</td>
</tr>
<tr>
<td>Glacial debris</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Hematite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Brucite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Gibbsite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Halloysite</td>
<td>Monoclinic</td>
</tr>
<tr>
<td>Volcanic ash</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Dolomite</td>
<td>Hexagonal</td>
</tr>
<tr>
<td>Biotite</td>
<td>-14</td>
</tr>
<tr>
<td>Vermiculite</td>
<td>Monoclinic</td>
</tr>
<tr>
<td>Phlogopite</td>
<td>-15</td>
</tr>
</tbody>
</table>

Fig. 1—A deposit of oriented, hexagonal ice crystals growing on special impurity sites on the surface of a single crystal of silver iodide; whole field = 300μ.
threshold temperature when introduced into a supercooled water cloud in either a diffusion- or mixing-cloud chamber. They also cause highly purified bulk water to freeze at these same temperatures. Ammonium fluoride, cadmium iodide, and iodine, being soluble in water, are inactive in a water-saturated atmosphere but produce ice crystals in an environment maintained between water and ice saturation, at the temperatures indicated. In effect, small particles of these substances act as sublimation nuclei, but on entering a water-droplet cloud, they go quickly into solution and lose their ice-nucleating ability.

In addition to the substances shown in Table 1, whose nucleating ability increases steadily as the temperature is lowered beyond the threshold value, we find a number of metallic oxides (for example, oxides of copper, cadmium, manganese, and tin) are slightly active at temperatures between -6 and -10°C but show no appreciable increase in nucleating ability at lower temperatures; again, these appear to act as sublimation rather than freezing nuclei.

Although there is a tendency for the more effective nucleators to be hexagonally symmetric crystals in which the atomic arrangement is reasonably similar to that of ice, Table 1 shows that there are a number of exceptions; but, for all the substances which are active above -15°C, it is possible to find a low-index crystal face in which the atomic spacings will differ from those in either the basal or prism faces of ice by only a few per cent. However, there is not, in general, a high correlation between the threshold nucleation temperature and the degree of misfit between the ice and the nucleus structures, indicating that nucleating ability is only partly determined by simple geometrical factors.

Full details of our work on artificial ice nuclei are given in a paper by Mason and van den Heuvel [1959].

Detailed study of the epitaxial growth of ice crystals—In an attempt to investigate the nucleating mechanism in more detail we have studied the growth of ice on well-defined faces of single crystals of various nucleating agents under carefully determined conditions of temperature and supersaturation. Oriented deposits of ice crystals have been observed on hexagonal crystals of silver iodide, lead iodide, cupric sul-
Fig. 3—A time sequence of photographs showing the orientation and growth of thin hexagonal ice plates on a single crystal of covellite, taken at about 30-sec intervals; left to right from the top.
phide, cadmium iodide, and brucite, and also on freshly-cleaved muscovite mica, mercuric iodide, iodine and calcite.

This study has revealed the great influence of the surface structure and topography of the host crystal. The ice crystals show a marked tendency to form at special sites on the surface, particularly at the edges of growth or cleavage steps. This is illustrated in Figures 1 and 2. Crystals will appear at these preferred locations under ice-supersaturations of order ten per cent but much higher supersaturations exceeding perhaps 100%, are required for nucleation on the very flat, perfect areas of the substrate surface.

Some very striking colored effects, which reveal a good deal about the detailed mechanism of ice-crystal growth, have been observed with ice crystals growing on a blue crystal of natural cupric sulphide (covelite). Figure 3 shows the crystals viewed in reflected white light. Being only a few thousand angstroms thick, the hexagonal plates show interference colors which give a measure of their thickness. Inspection of four parts of Figure 3 taken at about 30 sec intervals, reveals that some crystals grow considerably in diameter with no discernible change of thickness. This suggests that molecules arriving on the upper surface of the crystal are not assimilated but migrate over this surface and are built in at the edges.

The crystals generally thicken after meeting a cleavage step on the substrate or when they contact a neighboring crystal. This is shown very well by the line of five crystals which rapidly change color (thickness) after contact, with colored growth fronts spreading across their surfaces. These 'accidents' probably set up dislocations in the crystals from which growth fronts can emanate.

There is a marked tendency for the ice crystals to cluster along cleavage steps on the substrate. A crystal setting astride a step may be of different thickness on either side as indicated by the two-tone effects of certain crystals in Figure 3.

In order to investigate the nucleating properties of these single crystalline surfaces in more detail, careful measurements have been made, at different temperatures, of the minimum vapor supersaturations required to produce oriented deposits of ice crystals. The results for silver iodide are as follows. At temperatures above -4°C only water droplets were deposited. As the temperature was lowered from -4°C to -12°C, increasing numbers of ice crystals formed on selected sites provided that the air surpassed saturation relative to liquid water. At temperatures below -12°C, however, crystals appeared when the air was sub-saturated relative to water but supersaturated relative to ice by at least 12%. The observations suggest that between -4 and -12°C the initial deposit may have been liquid water, perhaps in droplets too small to be seen before they froze, while at temperatures below -12°C crystals may appear by sublimation direct from the vapor phase.

Very similar results have also been obtained for lead iodide, cupric sulphide, and cadmium iodide, with slightly different critical temperatures and supersaturations in each case, and also for an aerosol of silver iodide introduced into a diffusion cloud chamber in which the supersaturation could be accurately determined. These cloud-chamber experiments show that, even at temperatures above -12°C, it is not necessary for a silver iodide particle to enter a supercooled droplet in order to produce an ice crystal; it can act by adsorbing a film of liquid water. Full details of the work described in this section appear in a paper by Bryan*, Hallett, and Mason [1960].

The growth forms of snow crystals—One of our most fascinating problems, and one of great importance to the crystal physicist, concerns the remarkable variety of shapes exhibited by natural snow crystals. In order to discover the factors which influence the crystal form, and in the hope of discovering the exact nature of the controlling mechanism, we are growing artificial snow crystals under very carefully controlled conditions.

The crystals are grown on a thin fiber running vertically through the center of a diffusion cloud chamber in which the vertical gradients of temperature and supersaturation can be accurately controlled and measured. The results of many experiments covering a temperature range of 0 to -50°C and supersaturations varying from a few per cent (in the presence of a water-droplet cloud) to about 300% (in very clean, droplet-free air) consistently show that the crystal habit varies along the length of the fiber in the following manner:

<table>
<thead>
<tr>
<th>Temperature Range</th>
<th>Crystal Form</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 to -3°C</td>
<td>Thin hexagonal plates</td>
</tr>
<tr>
<td>-3 to -5°C</td>
<td>Needles</td>
</tr>
<tr>
<td>-5 to -8°C</td>
<td>Hollow prisms</td>
</tr>
<tr>
<td>-8 to -12°C</td>
<td>Hexagonal plates</td>
</tr>
<tr>
<td>-12 to -16°C</td>
<td>Dendritic crystals</td>
</tr>
<tr>
<td>-16 to -23°C</td>
<td>Plates</td>
</tr>
<tr>
<td>-23 to -50°C</td>
<td>Hollow prisms</td>
</tr>
</tbody>
</table>

This scheme is very similar to that which we obtained in earlier experiments in which crys-
DISCUSSION

Mason, a reversed take. — the...4° and the...one...a — the...the...the...one...crystal...a...this...degree has...the...H2O and D2O.

These experiments, reported by Hallett and Mason [1958], appear conclusive in showing that very large variations of supersaturation do not change the basic crystal habit as between prism and plate-like growth although, of course, the growth rates are profoundly affected. On the other hand, the supersaturation appears to govern the development of various secondary features such as the needle-like extensions of hollow prisms, the growth of spikes and sectors at the corners of hexagonal plates, and the fern-like development of the star-shaped crystals, all of which occur only if the supersaturation exceeds values which correspond roughly to saturation relative to liquid water.

The effect of suddenly changing the temperature and supersaturation of the growth form of a particular crystal could be observed simply by raising or lowering the fiber in the chamber. Whenever a crystal was thus transferred to a new environment, the continued growth assumed a new habit characteristic of the new conditions. For example, when needles grown at about -5°C were lowered in the chamber to where the temperature was between -12°C and -16°C, stars grew on their ends. In a similar manner it has been possible to produce combination forms of all the basic crystal types. Such radical changes in the crystal shape could not be produced by varying the supersaturation at constant temperature but, in some cases, were produced by only a degree or two change in temperature at constant supersaturation.

The growth habit of ice crystals is not essentially altered by growing them in hydrogen or air at reduced pressure as recently reported by Japanese workers.

The exact nature of the growth mechanism by which only a degree or two variation in temperature can completely change the crystal shape and which, furthermore, allows the habit to be reversed four times in a temperature range of only 25°C, is still something of a mystery. However, our current studies of the detailed evolution of the various crystal forms, in relation to their surface properties, are yielding some valuable clues.

References


Discussion

Dr. C. L. Hosler—How did you measure the supersaturations?

Dr. B. J. Mason—This is a very important question. In a diffusion chamber one can not measure the supersaturation if one disturbs the air. So what I tried to do and what worked out to be almost too good to be true, was to insert two parallel sheets of ice, one at a temperature level T1, and the other at a temperature level T2. Knowing the temperature profile, one can actually calculate what the supersaturation will be at any level in terms of the upper or lower temperature, and so on. One can check it because one can arrange that the supersaturation at one particular level, according to theory, would be just 100 or 98% or whatever value one wants to take. Now put in a source of nuclei, say sodium chloride; let it settle down in this region and observe optically. For instance, if it started to grow where the humidity was 78% and that coincided with the theory, one would be quite happy about the theory.
That is the way you check it. That is the way in which we essentially measure the supersaturation over this very wide range from zero to 300% relative to ice.

Mr. C. E. Anderson—If I understand, you have also examined the influence of low pressure, and have found no influence?

Dr. Mason—That is right except the crystals grow faster; the shape is completely unaltered. This is in conflict with the Japanese work.

Dr. H.-W. Georgii—Would you suggest there is a relation between the misfit and the special nucleating temperature for any given mineral?

Dr. Mason—There are really two threshold temperatures. There is the highest temperature at which ice crystals appear at water saturation, -4°C in the case of silver iodide, and -12°C, below which crystals appear in a sub-water saturated atmosphere providing the supersaturation relative to ice exceeds 12%. One can express the nucleating ability in terms of the supersaturation or of the temperature.

Dr. Georgii—Is it a specific value?

Dr. Mason—Yes, for a specific subject.

Dr. W. Hitschfeld—I would like to ask, in connection with your table of ‘activation temperatures’ whether you agree that there is no sharply defined temperature at which a given nucleus becomes active, but rather there is a range in temperature in which activation is possible with varying likelihood.

Dr. Mason—Activation temperature means the highest temperature at which one gets on ice crystal from 10,000 particles of seeding agent. If you want 1 in 100 for most of those substances, take two degrees away from those figures.

Dr. Hitschfeld—Closely associated is the ‘time of activation.’ In earlier papers from Dr. Mason’s laboratory (for example, Bigg, Proc. Phys. Soc. B, 66, 688, 1953), time was accorded an important place. But in some of the current work, one finds that not enough emphasis is always placed on it. Recent experiments at McGill University (Burkite and Gokhale, Part III of Scientific Report MW-30, Stormy Weather Group, July 1959) have clearly shown that the probability of a freezing occurrence is an approximately linear function of the time, and so the possibility arises that nuclei can become active at relatively higher temperatures if you wait long enough. This is a factor which meteorologists need to take most carefully into account when they apply the nucleation information which is becoming available.

Dr. R. Weier—You mentioned kaolinite. Is this common on the ground and in the atmosphere?

Dr. Mason—It comes in very small particles, and, in fact, in nature is hard to find in particles greater than one micron, and so it is in very finely divided form. It can not be very abundant. Most soils containing kaolinite tend to have a good vegetation cover, but obviously wherever this is disrupted it gets into the air. I would also say that volcanic ash when weathered produces kaolinite. So, I think, there is a reasonable amount in the atmosphere, but not too much. The only direct evidence that we have is that of the Japanese workers who often detect kaolinite particles in snow crystals (K. Isono, Jap. J. Geophys., 2, no. 2, 1959).

Dr. H. Weikmann—I am very happy about your paper because it is closely connected to the questions which I had posed in my letter of invitation: Can we prove that true sublimation nuclei do not exist and that AgI acts only as a freezing nucleus? There remains, however, a problem which is still unresolved: How does nature achieve the formation of ice at 0°C? So far in controlled laboratory tests this has not been achieved, not even with the best known freezing nuclei. It appears however that our experiments with freshly cleaved mica come closest. It would be very interesting if Dr. Mason would repeat those using his well controlled diffusion chamber. In our experiment breathing against the mica plate caused the formation of a very thin film of water on the mica plate recognizable only due to the formation of Newton interference rings. The crystallization of this film is easily visible. It started either at the very thin edges or at steps and irregularities in the mica plane. Crystallization occurred between 0° and -10°C wet bulb temperatures, that is, at room temperatures well above freezing! Eroneously we had assumed a practically perfect match between the geometric similarity of the cleaved plane and the base plane of an ice crystal but later we found out that the lattice structure of the cleaved plane is that of quartz and that it should nucleate at best at around -12°C. This experiment seems to indicate that we have to differentiate between the crystallization of a thin liquid film and that of bulk water, and
that perhaps here lies the answer to the astonishing performance of nature.

Dr. Mason—Yes, indeed, that is right.

Dr. U. Nakaya—My diagram was obtained in the case of mixing or convection and in your case the crystal is produced by diffusion of vapor. The experimental procedures are different, still both diagrams are quite similar, although a little different in some portions. The most characteristic type of snow crystal, the dendritic crystal, appears in the same temperature range in both diagrams. One thing to be mentioned is that the meaning of supersaturation in our case is different from the ordinary definition. In the case of mixing, usually minute fog particles are abundantly produced in the atmosphere where the crystals are made. We measured the total water content (minute drops plus vapor) by the gravimetric method, and calculated the degree of supersaturation. Our supersaturation, therefore, means the sum of water vapor and minute drops. It is very interesting that these droplets behave just like water vapor when they are very small, say, the order of one micron in diameter. Watching the process of growth of a snow crystal through a microscope, it is observed that the minute drops spread over the surface of the snow crystal without leaving any trace of the drop shape. If the drops are larger, say, several microns in diameter, they freeze in a drop shape on the surface of snow, thus giving the rimed snow crystal.

Dr. Mason—I think the mechanism responsible for the habit changes of ice crystals must be a surface phenomenon, because otherwise one might expect some difference between heavy and ordinary water. This one does not get; in fact, I can produce all shapes by adding increasing but still very small quantities of alcohol. It must be a question of the properties on the surface, and changing the surface properties. It can not be a question of the structure of the ice itself. This is one thing one has to worry about; if there is a minute amount of impurity, the experiment can come to grief.

Dr. Nakaya—we must be very careful in this point. Sometimes a very small amount of impurity changes the shape of crystal completely. We have the experience that an indetectable amount of silicone vapor changed the crystal shape completely.

Dr. Mason—This I would expect.
The Influence of Climate and Weather Elements on the Activity of Natural Freezing Nuclei

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Abstract—The paper presents results of parallel measurements of freezing nuclei and condensation nuclei. The results gained at different places on the European continent in different altitudes and under purely maritime conditions are compared. They clearly show the effect of various weather conditions and the influence of certain trace substances in the ground layer of the atmosphere on the freezing-activity of the aerosol particles.

Introduction—During the past years simultaneous investigations on condensation nuclei and ice nuclei were carried out by the author in order to gain a deeper understanding of the mechanism of heterogeneous phase transition in supercooled clouds. The experiments on condensation nuclei were performed on the basis of the results gained by Jungc [1952] on the size distribution and the composition of the atmospheric aerosols. Our studies led to the detection of certain relations between the freezing nuclei and the aerosol particles of certain size ranges.

The present paper will deal mainly with three problems:

1. What is the difference between the results gained under different weather conditions and at places with different climatic conditions?

2. In what way does the altitude at which the measurements are performed influence the activity of the freezing nuclei?

3. How is the activity of natural freezing nuclei influenced by the presence of certain trace substances and pollutants?

The freezing nucleus measurements were performed in an 85-liter mixing cloud chamber which was described in detail elsewhere [Georgii, 1956]. The following places were chosen for the investigations: (a) at Frankfurt am Main as base station, (b) on the Taunus ridge in an altitude of 800 m, 20 mi from Frankfurt (Taunus Observatory, Mt. Kleiner Feldberg), (c) on the summit of the Zugspitze at 3000 m altitude, and (d) on Valentia Island off the west coast of Ireland as maritime station. On the Irish west coast westerly winds prevail supplying a constant rate of fresh maritime aerosol particles.

Before dealing in more detail with the problems mentioned above, some earlier results of our research will be briefly summarized. The concentration of freezing nuclei active within the temperature range 0°C to −30°C shows a diurnal trend with a minimum in the early afternoon and a maximum during the night. This diurnal fluctuation runs parallel to that of the large condensation nuclei. It is most pronounced in summer and on days with calm sunny weather. The measurements taken on the Zugspitze show an inverse trend, namely, maximum in the afternoon, minimum during the night, indicating that the diurnal fluctuation of the numbers of ice nuclei is caused mainly by vertical convective mixing.

The evaluation of the parallel measurements of freezing nuclei and condensation nuclei (the latter were measured in three size ranges separately, Aitken nuclei, 'large' nuclei, and 'giant' nuclei) revealed that there definitely exists a relation between the concentration of freezing nuclei active above −30°C and the number of large condensation nuclei. This result was confirmed at all three continental stations and could be supported by direct methods, namely, by filtration of particles of definite size from the air sample to be checked on ice nuclei.

On the other hand it could be proved that the Aitken nuclei are ineffective as freezing nuclei above −30°C. A detailed description of these investigations is being published [Georgii, 1959].

Influence of weather elements—The evaluation of the daily measurements showed a considerable fluctuation of the concentration of freezing nuclei from day to day. The great number of counts made at Frankfurt permitted an analysis of the relation between the occurrence of certain weather phenomena at the time of the measurements and the concentration of freezing nuclei.
active above $-30^\circ$C. Figure 1 shows the average monthly freezing nucleus concentration compared with the concentration on such days when a certain weather situation prevailed. It can be seen in Figure 1 that the concentration of freezing nuclei is above average: (a) During and after showers as well as at the time of a strong vertical mixing. The shower activity causes a down draft of a great number of highly active freezing nuclei into the air layers close to the ground, especially of such particles which have already participated in the formation of the shower in higher levels of the atmosphere. Sometimes, such ice particles still preserve an ice-like structure on their surface which enables them to become effective once again at temperatures only slightly below the freezing point (preactivated or 'trained' nuclei). (b) During evaporation of fog droplets. Very high values of freezing nuclei were always found on days with ground fog, especially when the fog was in a state of evaporation. The activation of the nuclei may be explained in connection with drying of their surface. Possibly electric phenomena at the surface of the nuclei occur in connection with the evaporation which may also be partly responsible for the increased freezing activity [Mühlisen, 1958]. (c) When there are high wind speeds in high altitudes. The strong upwinds are mostly connected with high wind velocities near the ground thus removing the air pollution in the Rhein-Main basin.

The concentration of freezing nuclei in the ground-layer of the atmosphere is remarkably low: (a) On summer days with intensive solar radiation. The convection causes an up-current of the aerosol particles while the air sinking down by continuity reasons has a small aerosol content. (b) On days with continuous precipitation leading to a wash-out of freezing nuclei.

The influence of the different weather elements on the concentration of freezing nuclei active above $-30^\circ$C is summarized in Figure 2.

The annual average value of the 50S measurements carried out from May 1956 to May 1957 at Frankfurt was used as 100%. The other averages refer to this value subject to the above described meteorological phenomena taking place. It can be recognized that on days with evaporating fog, showers, and advection of fresh air masses in higher levels the concentration of freezing nuclei was higher than normal and amounted respectively to 138, 127 and 109% of the annual average. On the other hand the concentration on days with rain and on days with calm radiation weather reaches respectively only

![Graph showing influence of certain weather-conditions on freezing nucleus concentration at Frankfurt am Main](image-url)
Influence of weather elements on freezing nuclei—As already mentioned in the introduction of this paper the investigations were extended to three continental and one maritime location. A summarizing survey of the results is shown in Figure 3. The mean numbers of freezing nuclei, and large and Aitken nuclei at Frankfurt were set equal 100 and the average values of the measurements at the other locations were related to the Frankfurt average. Figure 3 presents an interesting result. Compared with the Frankfurt average the freezing nucleus concentration found at Mt. Kleiner Feldberg is 65%, on the Zugspitze 30% and at Valentia Island 27%. Referring however to the very active freezing nuclei agitating ice nucleation above -20°C, the picture becomes still more amazing since the absolute concentration of these freezing nuclei is higher on Mt. Kleiner Feldberg than at Frankfurt (125%). For Valentia Island we found 65% and for the Zugspitze 44%.

The distribution of the condensation nuclei shows a considerable deviation from these results. For the large nuclei (r above 0.2 μ) the numbers found at Mt. Kleiner Feldberg amount to only 18%, on the Zugspitze to only 4%, the corresponding values for the Aitken nuclei are 9% and 4% of the Frankfurt concentration. With regard to Valentia Island large nuclei could only be measured during a short period of time and are therefore not included in the summary given in Figure 3. The number of Aitken nuclei was only 2% of the Frankfurt mean number. This low number of Aitken nuclei at a maritime site is in good agreement with the findings of other investigators. The concentration is of the same order of magnitude as given in the survey by Landsberg [1938] in which the results gained at 21 different maritime locations are compiled and it is in agreement with the assumption by Mason and Moore [1954] that all Aitken nuclei found over the oceans are of continental origin. A full account of the investigations on Valentia Island had been given by Georgii and Metnieks [1958].

The vertical decrease of the particle number of the atmospheric aerosols as measured by us at the three continental stations at 100 mtrs, 800 mtrs and 3000 m altitude corresponds closely to former investigations not only as far as the trend is concerned but also with respect to the absolute concentration. In this connection attention is directed to the investigations by Weickmann [1957] on Aitken nuclei or the investigations by Drissbach [1956] on the vertical distribution of large nuclei. During 1955 and 1956 Reiter [1955] conducted measurements of Aitken nuclei on the Zugspitze to supplement his air electric measurements and found values very close to our own results.

While our investigations confirm the strong vertical decrease of condensation nuclei of all sizes they show also clearly that the number of freezing nuclei (particularly the most active) decreases much slower. This means that the properties responsible for the phase transition

![Fig. 2—Annual summary of the evaluation of the weather effect on the freezing nucleus concentration at Frankfurt am Main](image-url)
liquid to solid improve with altitude in the atmosphere. This fact can be caused by the following reasons: (1) The particles of industrial and anthropogenic origin in the ground layer of the atmosphere are poor ice nuclei. The concentration of these pollutants decreases rather rapidly with increasing altitude. (2) The high concentration of certain trace gases in the Frankfurt area causes an inactivation of the surface properties of the nuclei. In the free atmosphere the concentration of these gases is negligible. (3) Although the majority of our measurements including these at the high altitude station were carried out at outside air temperatures above the freezing level the influence of preactivated nuclei cannot be fully disregarded.

The effect of an increased activity of freezing nuclei shows up also in the phenomenon described above, namely, the occurrence of very active freezing nuclei in the ground layer after showers. In a recent paper of Kassander and others [1957] the presence of preactivated nuclei in the free atmosphere is also indicated. During flights over the southwestern United States air samples from ice clouds were taken and warmed above freezing level in a cloud chamber. Later the air samples were cooled once more and produced ice crystal concentrations of 100 per liter at only −10°C. Laboratory tests finally succeeded in conserving the efficiency of the 'trained' nuclei after they had kept above 0°C for several hours [Mason, 1959].

**Inactivation of freezing nuclei by surface reactions**—In order to clarify the problem of inactivation of freezing nuclei as a first step, parallel measurements of the concentration of certain trace gases in the atmosphere of the heavily industrialized Rhein-Main Basin were taken. Applying the methods of trace-gas analysis by Junge we sampled SO₂, NH₃, and NOₓ gases which were thought to be able to act as surface poisons. Table 1 gives a survey of the relation between the concentration of each of these gases and the activity of freezing nuclei. The latter is expressed as activity quotient, being the ratio of the number of large condensation nuclei and freezing nuclei active above −30°C. High figures of the activity-quotient stand for low 'quality' of the freezing nuclei.

The figures indicate clearly that the increasing concentration of the three gases in question deteriorates the activity of the freezing nuclei. However these parallel measurements of freezing nuclei and trace gases do not give an unequivocal answer which of the three gases is the most effective in inhibiting the ice nucleation although it seems according to Table 1 that the nuclei react rather sensitive on the ammonia.
concentration. In most cases the fluctuations of the three gases listed in Table 1 run parallel and are regulated by the actual weather situation.

In an attempt to study the poisoning effect more in detail we have started a series of tests of the freezing nuclei concentration in an air sample to which a known volume of the trace gas is being added. In addition to each ice-nuclei count a second run in a sample of untreated outside air is made to give information of the normal conditions in the atmosphere at that time. It may be mentioned that in the course of these experiments the method of counting the ice crystals in a dish containing concentrated sugar solution was applied which was first invented by Bigg [1957]. The results with this counting method agree quite well with our usual procedure of counting the ice crystals in a parallel light beam, the diameter of which can be narrowed by shutters when increasing ice-crystal numbers afford it.

Figure 4 shows the results of a first series of 20 inactivation tests using NH₃ as poisoning agent. In order to exclude the natural fluctuation of the absolute number of freezing nuclei on the different days when the samples were taken, the numbers found in normal outside air above -18°C and -21°C respectively were set equal 100. The numbers of freezing nuclei activated in an air sample containing 1500 ppm NH₃, the ammonia concentration during these experiments, is a proportion of 100. It can clearly be recognized that the presence of this amount of ammonia reduces the number of active freezing nuclei remarkably. This first series of tests applying a rather high concentration of NH₃, much higher than can be expected normally in the atmosphere, was intended to present a rough estimate on the effect of NH₃ as surface poison. An extension of these measurements with lower concentration of NH₃ and with the addition of other trace gases is under way.

The results of these experiments can be considered to be in this first stage as a confirmation of the investigations by Birstein [1954] on the inhibition of ice nucleation of silver iodide particles by certain chemicals. Testing the effect of NH₃, Birstein started with a partial pressure of 0.1 mm increasing it in stages up to 7 mm. The ice-nucleation ability decreased consequently. At a partial pressure of 7 mm no ice crystals were observed above -20°C. The most effective chemicals to inhibit ice nucleation were amines, methylated and ethylated amine. From Birstein's results the conclusion must be drawn that

<table>
<thead>
<tr>
<th>Table 1—Relation of various gases to the activity of freezing nuclei</th>
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<tr>
<td>Concentration in 10⁻³ gr/m³</td>
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<tr>
<td>-----------------------------</td>
</tr>
<tr>
<td>Ammonia</td>
</tr>
<tr>
<td>0-8</td>
</tr>
<tr>
<td>8-16</td>
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<tr>
<td>16-32</td>
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<tr>
<td>32-48</td>
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<tr>
<td>72-88</td>
</tr>
<tr>
<td>Nitrogen dioxide</td>
</tr>
<tr>
<td>0-5</td>
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<tr>
<td>5-10</td>
</tr>
<tr>
<td>10-15</td>
</tr>
<tr>
<td>15-20</td>
</tr>
<tr>
<td>Sulfur dioxide</td>
</tr>
<tr>
<td>0-160</td>
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<tr>
<td>160-320</td>
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<td>320-480</td>
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Fig. 4 — Inactivation of freezing nuclei by adsorption of NH₃ within the temperature ranges 0 to -18°C and 0 to -21°C, respectively; (a) freezing nucleus concentration in normal outside air equal 100; (b) freezing nucleus concentration after addition of 1500 ppm NH₃ to the air sample.
the inactivation of the particles is irreversible, meaning that the trace substance is not only loosely adsorbed but more firmly bound to the particle by chemisorptive forces.

The parallel measurements of condensation nuclei and freezing nuclei carried out at different places prove that besides the importance of climatic conditions the threshold temperature of ice nucleation of a given particle is not only a function of the constitution and size of the particle but is also influenced by environment conditions. The presence of certain trace substances and their concentration will effect the activation of the particle within a wide range of temperature. Preactivation of the nuclei extends the temperature range for ice formation to higher values than normal while surface poisoning will drop the threshold temperature to lower values. The factors will most certainly affect the operation of the mechanism of the rain formation process as postulated by Bergeron and Findeisen.

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References


Georgii, H.-W., and A. L. Metneks, An investigation into the properties of atmospheric freezing nuclei and sea-salt nuclei under maritime conditions at the west coast of Ireland, Geojs. Pure e Appl. 41, 150-176, 1958.


Mason, B. J. and D. J. Moore, The concentration, size distribution, and production rate of large salt nuclei over the oceans, Q. J. R. Met. Soc., 80, 583, 1954.


Discussion

Dr. Bernard Vonnegut—I am very much interested in the observation that the effectiveness of ice-forming nuclei tends to increase with convective activity suggesting that these particles may be brought down from aloft. I think this observation might shed some light on the experiences of my colleague, Charles Moore in balloon flights in thunderstorms over New Mexico. He observed on two flights that the exterior of large thunderstorm clouds was covered by a very thin mass of ice crystals that were not present in the interior. This observation, I think, suggests that air on the outside of the clouds, being pulled down from aloft, is high in nuclei content. In connection with ammonia and its effect on nuclei, it is worth mentioning Reynold's work of some years ago, in which he found ammonia had a very strong effect on silver iodide in the reverse direction; it appeared to increase the activity of silver iodide quite remarkably.

Dr. S. Birstein—With respect to inhibiting, it is gratifying to find that Dr. Georgii reports these effects. I gave a paper at the last Woods Hole Meeting, and I have another paper this afternoon on this subject. I think there is a lot more that can be said about it after that paper.

Dr. B. J. Mason—We all agree that Dr. Georgii's investigation is a very welcome one,
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and I agree with nearly everything he says; I just want to raise the problem of contamination, to see how important that is.

There is some advantage of living in a filthy atmosphere! If one takes an absolutely clean silver surface and puts it in the London air, for a few minutes, it absorbs enough iodine or sulfur to form patches of silver iodide of silver sulfide on which oriented ice crystals may be formed. If one irradiates silver iodide with ultraviolet, one finds in the first five minutes that its nucleating properties actually improve because its irradiation removes certain impurities; but thereafter its nucleating effectiveness decays logarithmically. So, this shows that impurities can work in either direction, and I think it is important to bear that in mind.

Dr. H. Weickmann—I would like to call attention to the observation regarding the increase of nucleating ability when sub-cooled droplets are in the state of evaporation. This may apply to Georgii's observations in radiation fogs as well as to Vonnegut's observation in the exterior layer of a thunderstorm cloud. Two processes seem to act: (1) the one which we mentioned before (see discussion of Dr. Mason's paper) and which seems to indicate that a thin film of water crystallizes easier than bulk water, and (2) an observation which I made in the laboratory when making ultramicroscopic studies of the melt water of cirrus crystals. Insoluble particles in the interior of these drops migrated during the evaporation, driven by Brownian motion, to the edge of the drop. Here the water surrounding the nucleus may only have the thickness of a thin film and nucleation starts.

Dr. H.-W. Georgii—The mechanism of evaporation effect is not quite clear to me yet. The increase of the concentration of freezing nuclei is also present if the fog is at temperatures above freezing without any supercooled droplets in the outside air.

Dr. C. J. Todd—On one occasion we were counting freezing nuclei and watching rain with a vertical radar, and as the rain changed from warm-cloud precipitation to ice precipitation, there was a very marked increase in the nuclei count that we measured on the ground.
Recent Observations of Freezing Nuclei Variations at Ground Level

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Abstract—Daily observations of freezing nuclei in the vicinity of Washington, D. C., during the period January 11–March 31, 1959, using a rapid-expansion technique, show indications of appreciable differences with respect to air-mass history. There was evidence of a pronounced tendency for the unusually high counts to occur in air with a recent marine fetch, while air with an extended continental trajectory characteristically showed lower concentrations of ice-forming nuclei in the subfreezing temperature ranges warmer than about \(-25^\circ C\). Some exploratory tests indicate that ocean water could be a source of freezing nuclei, as measured by the expansion technique.

Introduction—Only a few sustained series of observations of the natural variations in freezing-nuclei concentrations in the atmosphere exist. The tedious nature and other practical difficulties associated with obtaining reliable freezing nuclei observations together with uncertainties regarding current measuring techniques pose serious difficulties in this field. Present indications are that, although discrepancies of at least factors 2 to 4 in measured values may exist between different methods [Fenn and Weickmann, 1959], relative fluctuations are probably delineated by careful use of most techniques. There seems to be a general agreement that variations in the number of natural freezing nuclei per unit volume span several orders of magnitude, that increases and decreases seem fairly abrupt, and that the major anomalies may last from only a few hours to a day or so, seldom longer.

The sources and nature of the responsible natural aerosols remain in doubt, as do their connection with meteorological parameters. Rau [1954] concluded that his measurements in Bavaria showed appreciable differences with respect to air-mass types, and that the more active nuclei were of polar-marine rather than recent continental origin. An analysis of an extended series of observations at Mt. Washington, N. H. [Schaefer, 1954], suggested that the higher counts were associated with air trajectories from the western semi-arid regions of the U. S., and Isono and others [1959] reported an association between abnormal counts and the arrival of loess dust particles from the Asiatic mainland and also volcanic eruptions in Japan. Mason and Maybank [1958] found a substantial amount of laboratory evidence that various forms of siliceous materials were active nuclei. Some direct, but fragmentary, evidence of such particles at the center of snowflakes exists from electron microscopy techniques [Isono, 1955]. The effluents of certain industrial processes also appear to contain active components [Soulage, 1958]. The meteoritic-dust hypothesis advanced by Bowen [1953], although presenting seemingly insurmountable physical difficulties according to most existing astronomical and meteorological concepts, nevertheless continues to pose some interesting questions. A summary of January data [Kline and Brier, 1958] indicated some statistical support for the hypothesized singularities in freezing nuclei during this month, but anomalies in the succeeding two months at Washington, D. C. in a 1958 series of data could not be associated with any known meteor showers. There was a tendency for the abnormal counts with the mixing-chamber technique used to coincide with the intrusion of air with at least a limited marine fetch. Georgii and Methier [1958] reported evidence of parallel trends in ice-nucleation activity and the concentrations of large and giant condensation nuclei, but there was an absence of direct evidence of a maritime origin in a limited summer series of data on the coast of Ireland. On the other hand, laboratory experiments by Birstein and Anderson [1953] yielded positive indications that sea salt induced nucleation to the ice phase beginning at \(-15^\circ C\), but the role of water-soluble substances in the range of typical natural cloud nucleation temperatures remains in a somewhat controversial status.

The purpose of the following discussion is to outline the trends that appear to be emerging from a recent series of daily observations in the
Washington, D. C., area and some related test results using a rapid expansion technique. Because of existing uncertainties regarding the quantitative interpretation of measurements in this field, the data should tentatively be considered as relative rather than absolute in nature. The time factor in the activation of ice nuclei is such that the rapid-expansion technique may result in the detection of fewer nuclei than mixing-chamber or constant-temperature methods. However, the rapid-expansion technique has the extremely desirable feature of permitting replicated observations over a wide range of cloud-chamber temperatures without unreasonable time requirements. The data presented here were collected primarily to gain a little insight into the observational problem, and were obtained with a prototype model of a refrigerated expansion chamber intended for a sustained observational program at a number of sites (planned by the U. S. Weather Bureau as a joint effort with and supported in part by the National Science Foundation under Grant GI-29 with the assistance of a number of cooperating groups).

Observational procedure—The observations during the period January 11—March 31, 1959, were made about eight miles west of Washington, D. C. in an area removed from any known local sources of industrial pollution. At least two sets of data consisting of replicated temperature-spectrum runs from the threshold value (the appearance of one crystal per ten liters) to about −50°C were obtained on most days. The equipment used is similar in basic design to that described by Warner [1957], the primary difference being the addition of an electrically operated air pump for both purging and pressurization. The technique depends on the creation of a supercooled fog during an adiabatic temperature drop on expansion from a normal operating wall temperature of −10° to −12°C. Resulting ice crystals settle into a removable tray coated with a thin layer of sugar solution which is placed at the bottom of the ten liter chamber. The crystals subsequently grow to visible size in 30 to 60 sec, and the total number is then counted visually. Estimating procedures were used when the number of crystals exceeded 150 to 200 per ten liters by limiting exact counts to known fractional areas of the trays. All observations were made with the equipment located out-of-doors whenever practicable to reduce the risk of misleading results caused by possible contamination with unrepresentative inside air. Counts at −20°C (the primary "reference temperature" used here) were augmented to some extent by extra readings, particularly during variable conditions. The data from each set of observations covering a range of expansion temperatures were plotted and mean-temperature spectrum curves constructed. At least 10 to 15 separate observations were usually considered necessary to define a temperature-spectrum curve with reasonable confidence.

Results and discussion—Approximately 2950 individual measurements comprising 172 sets (mean temperature spectrum runs) were obtained during the 80-day period under consideration. The variety of weather patterns together with the abruptness with which airmass changes frequently occur during the winter months in the Washington, D. C., area offers an opportunity to evaluate the trends in freezing-nuclei observations in relation to synoptic weather features with perhaps less ambiguity than at locations farther inland or in more uniform climatic regimes.

Typically, the number of nuclei activated with decreasing temperature shows an exponential type of relationship such that a tenfold increase occurs with roughly a 5°C drop in temperature. However, as shown by several examples in Figure 1 there are considerable variations in

![Figure 1](https://example.com/figure1.png)

**Fig. 1**—Examples of freezing nuclei data obtained in the Washington, D. C., area; ordinate scale is \( \sqrt{N} + \sqrt{N+1} \), where \( N \) is the number of ice crystals.
the slope of the mean curves. A few observations have indicated maximum values in the range of \(-20^\circ\) to \(-25^\circ\)C with no appreciable increases with decreasing temperature.

Figure 2 illustrates the rather spectacular changes that were usually observed with pronounced airmass changes. In this instance the evening observations were in a well-defined maritime tropical airmass preceding the arrival of the leading edge of continental air of polar origin which dominated the area by the following morning. This example displays a feature commonly observed during the January–March period, namely, the strong tendency for air of continental fetch to have appreciably fewer freezing nuclei active at expansion temperatures warmer than roughly \(-25^\circ\)C, but higher concentrations as the \(-30^\circ\)C temperature value is approached.

Examples such as the foregoing indicated that a more detailed examination of air trajectories might profitably be undertaken. In the initial study, 12-hour trajectory estimates were derived by examination of the 6-hourly surface weather charts most nearly synoptic with the mid-period of each set of freezing nuclei observations. The stratification of the data was according to the quadrant from which the surface air arrived at Washington, D. C., using streamline flow estimates rather than geostrophic computations. This preliminary analysis indicated that the stratification could be simplified according to whether or not the air trajectory was entirely continental, of a maritime tropical nature, or was such that air moved into the Washington area from the northeastern and southeastern quadrants with a resulting high probability of a recent marine fetch. Temperature-spectrum data applicable to each of these three categories were averaged to obtain the mean threshold temperature values and ice crystal concentrations at \(-20^\circ\), \(-25^\circ\), and \(-30^\circ\)C, as shown in Figure 3. On the average, air with a recent marine trajectory contained about an order of magnitude more particles active as freezing nuclei at \(-20^\circ\)C than air with an extended continental history, while maritime tropical air with several hundred miles of overland fetch showed intermediate values. However, there were too few cases in the later category to conclude that this difference is representative. The average daily counts at \(-20^\circ\)C from January 11–March 31, 1959, and the variations between individual sets of freezing nuclei observations on each day are shown in Figure 4 with respect to airmass history. Two features in particular are evident: (1) the considerable amount of variability on many days, and (2) the strong tendency for the higher counts to occur in marine air with the most abnormal values in airmasses with a probable recent marine trajectory. Of 17 sets of observations indicating concentrations in excess of
100 crystals per ten liters at \(-20^\circ\text{C}\), all but one occurred during such airflow regimes while 67 out of 84 observations in the range of 0 to 10 crystals per ten liters were in continental air masses. Figure 5 shows a more detailed presentation of this trend. The incidence of the unusually high counts in excess of 500 per ten liters on February 9, March 14, and March 26 was in air with an over-water fetch from the northeast.

The sparsity of meteorological observations off-shore prevents a detailed examination of possible upwind factors. There was no obvious correlation with wind speed or probable extent of over-water fetch. However, there were some indications that low-level instability and widespread precipitation were associated with the freezing nuclei anomalies. If so, this may provide a clue regarding the conditions most favorable for the generation and transport of the responsible aerosols. The existence of a recent marine history alone does not appear to represent both a necessary and sufficient condition, since, as shown in Figure 5, there were several instances of low concentrations in such flow regimes.

In view of the highly suggestive nature of these empirical results, samples of ocean water were obtained from the Rehoboth, Del., coastal area. On the assumption that the bursting of bubbles at the sea surface would be the most likely mechanism for the natural production of aerosols from oceanic sources, the expansion chamber was purged and pressurized with air ingested from a few inches above the surface of agitated ocean-water samples. These tests have been conducted a number of times with similar results. Two examples are shown in Figure 6 in which the background counts were comparatively low. While there is considerable scatter in the results, there appears to be little doubt of a positive response with the expansion chamber technique. Whether the responsible particles are solely the soluble components of ocean water or substances in colloidal suspension can-
not be stated at this time. The threshold nucleation temperature of the sea-water aerosols has consistently been in the neighborhood of -14° to -16°C, in close agreement with the results of Birstein and Anderson [1953]. Trial runs with solutions of NaCl and MgCl₂ as well as other soluble substances appear to give positive responses of comparable magnitude, while carefully distilled water shows a null effect.

Concluding remarks—These results appear to offer strong circumstantial evidence, but not proof, of a marine source of aerosols active as freezing nuclei in the Washington, D. C., area. This evidence seems at variance with the experience of others in this field, and raises some provocative questions regarding the physical nature of at least some of the freezing nuclei in the atmosphere. It is of course possible that the explanation for these results may reside in the inherent nature of the rapid-expansion technique wherein soluble substances could initiate the freezing process before going into solution, but not in the slower condensation rates more representative of atmospheric processes. On the other hand, the mixing-chamber technique used in the earlier 1958 series of observations in the Washington, D. C., area showed similar tendencies for anomalous conditions to occur in air with a marine trajectory. The possibility of local sources of freezing nuclei cannot be ruled out completely. However, a series of observations at various locations within a ten mile radius of the metropolitan area on 15 separate days yielded no evidence that this might be the case.

These results plus recent work of Papée [1959] pointing toward activation phenomena in soluble substances such as NaCl which may facilitate their role as a sublimation nucleus raises some intriguing and rather crucial questions regarding the role of hygroscopic particles in atmospheric chemistry and cloud nucleation. They may have a bearing on the limited number of aircraft observations [Coons, Jones, and Gunn, 1949] indicating a tendency for the occurrence of ice crystals in clouds at warm super-freezing temperatures in marine air.

Acknowledgments—The assistance of Gilbert D. Kinzer, Glenn W. Brier, and DeVer Colson in certain experimental and analysis phases of this study is gratefully acknowledged, along with help by Thomas H. Carpenter and Frederick Van Cleef in the laborious and tedious process of maintaining a daily freezing nuclei observational program.

References


Georgi, H.-W., and A. L. Metnies, An investigation into the properties of atmospheric freezing nuclei and sea-salt nuclei under maritime conditions at the west coast of Ireland, Geophys. Pura Appl., 41, 159-176, 1958.


Rae, W., The ice-nucleus concentrations of various air masses, Met. Rundschau, 7, 205-211, 1954.
DISCUSSION

Mr. D. Blanchard—I think I would be one of the last to discourage any work related with phenomena of the air-ocean interface, but I think a word of caution might be in order. One produces aerosols by bubbling or spraying. Any way in which one produces aerosol from sea water or distilled water means this aerosol will come from the surface. This means any surface contamination on this water will end up in the aerosol, and if one has any surface active contamination, this may be what you are actually measuring. I think these experiments certainly should be carried out using very clean conditions, perhaps using artificial water.

Mr. D. B. Kline—We have tried distilled water. Carefully distilled water did not give a response. Tap water, however, did. We have tried sodium chloride, magnesium chloride, and calcium sulphate solutions and we consistently got a response.

Dr. H.-W. Georgii—In January 1957 we had very marked peaks at about the same dates at which Dr. Bowen had them. At that time, I did not yet know of Bowen’s work. We believe, as does Mr. Kline, that they are connected with the large-scale flow pattern of the atmosphere, and perhaps the change from zonal to meridional circulation that we had at this time in 1957.

Dr. Tor Bergeron—I thought our freezing nuclei measurements in January 1959 were of very little value; but I shall now be very glad to have them compared with yours.

Mr. Kline—I am very glad to establish this contact and I am very interested in the data.

Dr. Chojō Magono—A group in Japan under Dr. Isono at Tokyo University made observations like you, using different methods. They found that peaks of concentration of freezing nuclei agreed very well with volcanic eruptions in Japan. This was derived using the trajectory method.

Mr. Kline—I was interested in the report that the freezing nuclei concentration was not only correlated with volcanic eruptions, but could also be traced back to the latest dust storm in northern China (see reference to Isono and others in my paper).

Mr. Jerome Namias—In the synoptic situation for maritime flow in the Washington, D. C., area, there is usually a very strong vertical wind shear, so that fresh maritime air is seldom observed up to high elevations. It is a warm-front condition in which if the air flow is fresh from the ocean in the lowest layers, it changes to a continental flow aloft. Frequently the precipitation mechanisms develop in the higher layer, so that this type of stratification should be considered in regard to these observations.

Mr. C. E. Anderson—I wonder whether or not measuring ice nuclei at the surface is of any real value in getting an estimate of what the ice nuclei activity is likely to be aloft. Since there is not only a question of separation of trajectories with height, but also the influence of contamination at the surface.

Mr. Kline—All we can do to attempt to get around the problem is set up our network to include high altitude stations.

Dr. C. L. Hosler—What are the short-period variations in these ice-nuclei counts? The data you presented is the result, I presume, of a number of counts. How many and over what period of time? What would happen if you waited an hour longer, would this result in a greater variation from day to day?

Mr. Kline—It takes about 60 to 90 min to run through a spectrum such as is presented in Figure 3. During that period we normally have variations only within roughly a factor of two or three. Reproducibility is usually good; but we do encounter periods where there is considerable variation.

Dr. H.-W. Georgii (communicated)—In the course of our own measurements off the Irish west coast we also found a relatively high number of freezing nuclei active above —20°C. The absolute number of freezing nuclei within this temperature range was even higher than on the Zugspitze and equally high as in Frankfurt. However, at lower temperatures (between —20
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and —30°C) the freezing nucleus concentration in Valentia Island, Ireland, increased only very slowly, while at the three continental places a very sharp increase was observed. We assume that in the very clean maritime air around Valentia Island an inactivation of the aerosol particles by adsorbed trace substances does not occur while in Germany this is more or less everywhere the case because of the overall effect of air pollution. We are therefore also inclined to suggest that the relatively low concentrations of freezing nuclei which Kline observed in air masses coming from the interior of the country are also caused by a partial inactivation of the particles by pollutants which is not the case when the air is coming from the ocean.
Studies on the Effect of Chemisorbed Impurities on Heterogeneous Nucleation

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Abstract—Studies have been made on the effects of the chemisorption of ethyl amine in lead iodide insolar as they affect the nucleation of supercooled water. It has been found that concentrations on the order of one part per two million of ethyl amine in the carrier gas will decrease the temperature of nucleation by seven degrees. Higher concentrations cause an even greater temperature change. The implications of these results in terms of atmospheric phenomena are discussed.

Introduction—Previous work on the inhibition of ice nucleation by ethyl amine has shown that in a dynamic system it is possible to inhibit lead iodide nucleation of supercooled water by chemisorbing ethyl amine on the surface of the lead iodide [Birstein, 1957]. These results showed that it is possible to lower the temperature of water nucleation by 17° by passing the lead iodide nuclei over a saturator filled with ethyl amine held to a vapor pressure of approximately 0.2 mm at a standard flow rate and nuclei generator setting. Higher vapor pressures of the amine gave greater degrees of inhibition of the nucleation process (Fig. 1).

Because this work was done in a dynamic system the question arose concerning whether equilibrium conditions were reached between the nuclei and the nuclei inhibitor. The following series of experiments were run to study the reaction under equilibrium conditions to determine whether it is possible to inhibit nucleation of water droplets with traces of impurity approaching that of trace contaminants in the atmosphere.

Experimental—The apparatus used in this work consisted of our standard nuclei generator (Fig. 2) described in previous publications [Birstein and Anderson, 1955], connected to an especially designed cryostat (Fig. 3). The cryostat was essentially a block of lead in which was imbedded a copper coil. Liquid nitrogen was circulated through the coil to bring it to a desired temperature. The block was insulated with styrofoam. A Tag controller-indicator regulated the flow of liquid nitrogen through the coil. The 'Tag,' in turn, was controlled by a thermocouple set within the cryostat. Two glass tubes were located in the cryostat. One tube was filled with ethyl amine and held to a given temperature which could be interpreted in terms of the vapor pressure of the amine at that temperature. The second tube was the control and, therefore, was empty. These tubes were connected to flasks located outside of the cryostat and the flasks, in turn, were connected to the nuclei generator through a three way stopcock. All ground glass joints and stopcocks were ungreased to prevent contamination by the extremely reactive ethyl amine.

In making a run, the Tag was set for the desired cryostat temperature and the liquid nitrogen flow was started. When the ethyl amine was at the set temperature the nuclei generator was started and nuclei samples prepared with nitrogen atmosphere were collected in both flasks. The flasks were then opened to the thermostatted tubes in the cryostat and the system was allowed to come to equilibrium. Preliminary experiments to determine the equilibrium time showed that fifteen minutes was sufficient. A nuclei sample was then removed from the control flask and injected into the cold chamber to determine whether nuclei were present in the system. This was next repeated with the nuclei sample in the flask exposed to the ethyl amine vapor. The cold box temperature was lowered and nuclei injection was repeated until the temperature was reached at which ice crystals were first formed with the lead iodide sample treated with the nuclei 'poison.' These experiments were repeated until the nucleation temperature for the lead iodide had been obtained after exposure to ethyl amine vapor at pressures corresponding to the amine vapor pressure between -75° and -171°.
was cooled by circulating refrigerated methanol through a copper coil immersed in the dewar. The temperature of the cold box J and K, too, was controlled by means of a Tag controller-indicator T whose thermocouple was suspended in the cold box to sense the temperature and, therefore, control the flow of refrigerant through the valve D.

Discussion—From Figure 1, it can be seen that even in a dynamic system the effect of ethyl amine on lead iodide nucleation of supercooled water is quite marked. A comparison of the data from the dynamic system with those obtained from the static tests under known equilibrium conditions shows that there is, in general, good agreement between the two sets of data. While the static tests were quite interesting, because of the nature of the apparatus used, it was only possible to run the tests at relatively high partial pressures of ethyl amine. The system involved the passing of a nitrogen stream containing the nuclei over a saturator filled with the amine. The lowest temperature which could be reached in the saturator was approximately \(-110^\circ\). At \(-105^\circ\) the vapor pressure of the ethyl amine is on the order of 0.1 mm or approximately 0.0001 atmosphere. With the cryostat used in the static measurements, it was possible to reach a temperature approaching the boiling point of the coolant. In the case of liquid nitrogen which boils at \(-196^\circ\), the cryostat temperature could be lowered to approximately \(-180^\circ\).

The vapor pressure curve for ethyl amine is shown in Figure 5. The curve for the plot of log vapor pressure as a function of temperature can, for all practical purposes, be assumed a straight line in this range. The results of the static runs are given in Table 1. The cryostat temperature ethyl amine vapor pressure, and nucleation
temperature for the supercooled cloud are given for each series of measurements. The results, in terms of ice nucleation temperature for lead iodide as a function of cryostat temperature, are given in Figure 6 and the results in terms of ice nucleation temperature as a function of vapor pressure of ethyl amine are given in Figure 7. When one examines Figure 6, the curve for nucleation temperature as a function of cryostat temperature, it appears exponential. The plot of nucleating temperature as a function of log vapor pressure of ethyl amine, however, appears to be a straight line, indicating that the nucleating temperature bears the same relationship to vapor pressure of the ethyl amine as the vapor pressure has to the temperature of the material.

An analysis of the data obtained in the static tests shows that the lower limit of measured change is a decrease in nucleation temperature from $-6^\circ$ to $-13^\circ$ when the partial pressure of ethyl amine was approximately 0.0004 mm. This corresponds to, at one atmosphere, one part of ethyl amine per two million parts of air. Going up towards the higher partial pressures of nuclei poison, a partial pressure of 1.1 mm of amine will lower the freezing point of supercooled water to $-33^\circ$, a change in nucleating temperature of $27^\circ$. This latter pressure, however, is quite high; it corresponds to better than one part per thousand of ethyl amine and is unrealistic when one is thinking in terms of atmospheric contami-

<table>
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<th>Table 1—Results of static runs</th>
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<tr>
<td><strong>Cryostat</strong></td>
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<tr>
<td>$-81$</td>
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<tr>
<td>$-101$</td>
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<td>$-120$</td>
</tr>
<tr>
<td>$-129$</td>
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<td>$-139$</td>
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nents. One part per million of the contaminant will lower the freezing point of supercooled water droplets in the presence of lead iodide to $-15^\circ$, and one part per hundred thousand to $-20^\circ$.

**Conclusions**—While the above data were taken in the laboratory on lead iodide, an artificial nucleating material, certain conclusions can be drawn which are directly applicable to the atmosphere. The concentration of ethyl amine used in these experiments was purposely kept low in order that the lower end of the curve of partial pressures would approximate that of atmospheric pollutants. If one thinks in terms of concentrations up to one part per hundred thousand, the nucleating temperature is lowered by 14$^\circ$.

Although these data were taken on lead iodide, work done in the past [Birstein, 1957] has shown that it is possible to inhibit ice crystal formation by natural nuclei through treating the air sample with methyl amine before it is brought into a cold box. These experiments showed that with a sufficient amine concentration it is possible to prevent so-called 'spontaneous nucleation'. We were able to reach $-52^\circ$ in one experiment before ice-crystal formation took place. Ethyl amine acts in a manner similar to methyl amine but is slightly more effective because of the longer carbon chain on the molecule [Brueheman and Verhoek, 1948].

If we accept the following type of reaction as being the cause of the loss of effectiveness of the lead iodide nuclei [Biltz, 1922]

$$\text{PbI}_2 + 2\text{C}_2\text{H}_5\text{NH}_2 \rightarrow \text{Pb(C}_2\text{H}_5\text{NH})_2\text{I}_2$$

we see that this is the simple formation of coordination complexes on a metal salt. It is an extremely common type of reaction and would be expected to occur with many compounds other than amines. Hundreds of other type reactions are also possible with other particulate matter which may be responsible for ice crystal formation in the atmosphere.

While these experiments were not designed to pinpoint the exact mechanism by which ice-crystal formation may be inhibited in the atmosphere, they do show that the concentrations of industrial wastes present in an air mass are certainly sufficient to make marked changes in the ice-nuclei spectrum in the air mass.

**References**


**Discussion**

**Dr. James P. Lodge, Jr.**—Your use of very low temperatures to obtain low partial pressure of contaminants is a very difficult one. We have used the apparatus of Stephenson to obtain low concentrations by means of diffusion through tubes (A. P. Altshuler and I. R. Cohen, "The Application of Diffusion Cells to the Production of Known Amounts of Gaseous Hydrocarbons," paper presented at the 136th National Meeting, American Chemical Society, Atlantic City, September 18, 1939). In this manner we can obtain concentrations of one part per million and less in a dynamic system.

**Mr. S. J. Birstein**—I agree that this approach is a rather difficult one.

**Dr. F. W. Van Straten**—It would seem that there is something of a conflict between the results of this work and the work of Dr. Mason, because according to Dr. Mason's paper there seemed to be a preferential nucleation on, shall we say, the active lines of the host crystal. If this is true, in that case, you could not expect poisoning to be a linear function of the concentration of the poison. It should first affect the host crystal on the most efficient sites, then drop off very sharply to the point where the other surfaces of the crystal begin to nucleate.

**Dr. B. J. Mason**—I think it is important that you saw the ice crystals, in my cases, growing on the dislocation steps. It is at these that the
DISCUSSION

poison will be absorbed first. It is because the nucleation occurs only on these steps that such a small amount of poison can do the trick, because if you put a monomolecular layer of poison all around the steps, it still covers only very little of the surface. As Dr. Van Straten says, the poison affects the steps first, and the amount of material required to poison the steps is very small indeed. We certainly have evidence that the small traces of poison do, in fact, kill nucleation sites at these very steps.

Dr. Walter Hilschfeld—I think this is illustrated quite well on Figure 1. Here the relationship of the poison concentration to temperature is shown to be linear. But obviously it cannot remain linear down to zero concentration. So if one goes beyond Mr. Birstein's data to zero concentration, one knows he has to get to the temperature usually called 'activation temperature.' There is thus necessarily a non-linear region in the relationship. This is precisely what Drs. Van Straten and Mason require.

Dr. H. W. Georgii—It is right to assume from your results that the process of the poisoning is not only absorption but a chemisorption?

Mr. Birstein—On the lead iodide, I would call it reversible chemisorption. On the silicate it would probably be physical absorption of some sort. However, what it does in the activity, I could not begin to say.

Dr. Georgii—Can the poisoned particles recover their activity?

Mr. Birstein—It can recover its activity, but not easily.

Dr. Roscoe R. Graham, Jr.—Perhaps some of you did not see a paper that bears on this, although not critically, when it was published. Back in 1949 or 1950, in work we were doing at the Institute of Mining and Technology, we felt that thunder-storm electricity had its origin in freezing of water, according to Workman and Reynolds' work on freezing potentials. It had been found in the laboratory that water very slightly contaminated with ammonia (1 to \(10^2\) to \(10^6\)) gave freezing potentials which were just reversed from those of normal water. We felt we had a straight-forward experiment that was ideal. All one had to do was release quantities of ammonia into potential thunderstorms; and it would be 'flipped upside down' and simultaneously we would have proven the Workman-Reynolds effect. But it turned out when we loaded a B-17 with all the ammonia it could carry and stocked tanks of ammonia on the ground, so we could release it up underneath the thunderstorm, we were completely unable to observe any effect whatsoever on the electrical structure of this storm. We certainly were releasing quantities of ammonia into these particular storms, but Nature has a way of disposing of this material as though it were never released. This was undoubtedly real, and I would like to suggest that many times Nature does not know about many of the things we do.

Mr. Birstein—We have been doing some work in clouds inhibiting ice-crystal formation, and the results thus far have not been too great; but we seem to have gotten limited success in decreasing the ice-crystal concentration. We hope to do some work of this sort in Flagstaff in the summer of 1959.

Dr. M. Neiburger—Did you stop the rain from falling?

Mr. Birstein—This is one thing we are interested in determining.
Some Observations of Chloride-Sulfate Relationships in the Atmosphere and in Precipitation

JAMES P. LODGE

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Abstract—A series of measurements, including the determination of chloride and sulfate in particulate matter, were made at a U. S. Coast Guard weather ship midway between Honolulu and San Francisco. Even in this extremely isolated location, the sulfate concentration frequently exceeded the chloride. In the light of this finding, chloride-sulfate relationships were examined for a number of other remote stations; published data in the literature have also been studied. The chloride-sulfate regression lines were found to vary markedly in slope and intercept. There appears to be a definite relationship between chloride and sulfate concentrations; however, the nature of the relationship varies markedly with location. An interpretation is offered, based upon a three-source model.

Introduction—Among the salts circulating through the atmosphere are sizable quantities of chlorides and sulfates. Both species are of importance in precipitation physics, since they form soluble particles which are extremely effective as cloud or precipitation nuclei. Both are of interest in air pollution, since man liberates them, or their precursors, in quantity by his activities. The levels contributed by nature thus constitute a background to the urban levels measured in air-pollution studies. On the other hand, the human contribution constitutes a sort of ‘noise level’ on the natural concentrations, considered from a global geochemical viewpoint.

For several years the U. S. Public Health Service has had in progress a small study of natural levels of substances which are considered pollutants when they are found in urban areas. In collaboration with the California Department of Public Health, measurements of these natural levels were made at a number of remote coastal sites [Harrison and Lodge, 1959; Holzworth, 1959]. Subsequently two series of measurements were made during the summers of 1957 and 1958 at Weather Station November, the weather ship located approximately midway between San Francisco and Honolulu [Lodge and others, 1959]. When it was discovered that nearly all of the particulate samples from this operation contained more sulfate than chloride, a statistical study was made of chloride sulfate relationships, from both the coastal sites and from those data available in the literature.

The data—In addition to two sets of measurements from Station November which were tested for homogeneity and pooled, the California study made available air data from Southeast Farallon Island, Point Piedras Blancas, San Nicolas Island, and the top of Mount Hamilton. Also utilized were the excellent precipitation analyses of Larson and Hettick [1956] from central Illinois, and analyses of air or precipitation or both from four selected stations of the Swedish network [Egner and Eriksson, 1955, and quarterly reports by a variety of authors in Tellus]: Kinnsvika, Spitsbergen; Valenta, on the west coast of Ireland; Edinburgh, Scotland; and Bonn, West Germany.

The statistical analysis was not as complete as might be desired, either as to the depth of the treatment or as to the number of stations covered. It would be desirable to extend the analysis to all of the stations of the Swedish network and to other historic sets of air and precipitation analyses, to see if it would bear out the tentative conclusions of this paper.

The Swedish data are reported in terms of the amount of mineral matter brought down by precipitation per unit area. However, since total precipitation is also recorded, it is possible to convert the units given into concentration in the rain water. For convenience, the former figures will be referred to as ‘rainout’ data, and the latter as ‘precipitation’ data. It seems likely that rainout should be a more valid measure of the concentration of materials aloft, since it is substantially independent of the intensity of pre-
CHLORIDE-SULFATE RELATIONSHIPS IN THE ATMOSPHERE

It will, however, be in some degree dependent on the frequency of precipitation.

The findings of Byers and others [1957], among others, make it clear that particulate matter sampled near the surface of the Earth may not be representative of that aloft. It is more likely to represent local influences, and less likely to show marine influence, than samples taken at higher altitudes. Thus rainout, representing as it does the mean concentration in a layer of air some thousands of feet thick, may differ markedly in composition from surface air: it should be more representative of the atmosphere as a whole. This is especially apparent in the two sets of urban data in the present study.

The interpretation—The data, with sulfate expressed as sulfur equivalent, were fitted to best regression lines on logarithmic paper by the method of least squares. Thus all the regressions had the form

\[(CI) = a(S)^b\]  (1)

where \((CI)\) is the concentration of chloride in appropriate units, \((S)\) is the concentration of sulfate computed as sulfur in the same units, and \(a\) and \(b\) are constants. Table 1 gives the values of \(a\), \(b\), and the correlation coefficient \(r\) for all cases in which significant correlations were obtained.

A particular sample of airborne particulate matter, however taken, will contain chlorine and sulfur compounds of very local origin, together with substances which have traveled great distances. A reasonable physical model is that of the atmosphere as a great reservoir, characterized by a certain mean ratio of chlorine to sulfur. At any point within this reservoir, however, local perturbations may result in a sample composition very different from the mean.

However, from the viewpoint of a particular sampling location, it is perhaps simpler to consider the mean atmosphere and the various local influences on an equal footing. This mathematical model of the sample as the result of a number of 'sources' has the advantage for purposes of this paper of involving less prejudice concerning the over-all mechanism. Further study should be made before a mechanism can be set forth with confidence. Thus three 'sources' will be considered: the oceans, the mean atmosphere, and human activity.

The direct marine contribution contains chloride equal roughly to 21 times the mass of sulfur. Thus its composition may be represented as

\[(CI) = 21(S)\]  (2)

The total atmospheric loading has been deduced by Eriksson [1950]. From his figures, one obtains

\[(CI) = 1.43(S)\]  (3)

for the chlorine-sulfur relationship. The third component, man's own products, contributes, for practical purposes, only sulfur to the atmosphere. It may thus be taken as a relationship of the form

\[(CI) = k\]  (4)

that is, for a given station, the chloride level will be relatively constant, as fixed by that entering the area, but sulfur is indeterminate and independent of chloride. Thus the mean chlorine-sulfate relationship for any locale will reflect all three 'sources.' (It should also follow that the world-wide average of all such determinations should be identical with the above 'mean atmosphere' relationship (Eq. 3), within the accuracy of Eriksson's figures.)

Turning to the actual data, it can be seen that Kinnvika (Fig. 1) is the most truly 'maritime' station. Both air and rainout figures are for practical purposes identical in composition with sea water. The precipitation itself varies so much less than that of most stations that it may well be constant within the errors of the measurements.

Station November (Fig. 2), in contrast, is

<table>
<thead>
<tr>
<th>Site</th>
<th>a</th>
<th>b</th>
<th>r</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Champaign, Illinois</td>
<td>0.40</td>
<td>0.01</td>
<td>0.77</td>
<td>P</td>
</tr>
<tr>
<td>Mt. Hamilton, California</td>
<td>0.60</td>
<td>0.02</td>
<td>0.62</td>
<td>A</td>
</tr>
<tr>
<td>San Nicolas Island, California</td>
<td>1.22</td>
<td>0.02</td>
<td>0.57</td>
<td>A</td>
</tr>
<tr>
<td>Station November</td>
<td>1.50</td>
<td>0.30</td>
<td>0.47</td>
<td>A</td>
</tr>
<tr>
<td>Bonn, Germany</td>
<td>2.12</td>
<td>0.55</td>
<td>0.55</td>
<td>A</td>
</tr>
<tr>
<td>Valentia, Ireland</td>
<td>2.82</td>
<td>1.30</td>
<td>0.91</td>
<td>R</td>
</tr>
<tr>
<td>Point Piedras Blancas, California</td>
<td>6.20</td>
<td>1.20</td>
<td>0.71</td>
<td>A</td>
</tr>
<tr>
<td>South East Farallon Island,</td>
<td>6.24</td>
<td>1.05</td>
<td>0.57</td>
<td>A</td>
</tr>
<tr>
<td>California</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Valentia, Ireland</td>
<td>11.7</td>
<td>1.31</td>
<td>0.90</td>
<td>P</td>
</tr>
<tr>
<td>Kinnvika, Spitsbergen</td>
<td>28.6</td>
<td>0.92</td>
<td>0.92</td>
<td>A</td>
</tr>
<tr>
<td></td>
<td>61.5</td>
<td>0.51</td>
<td>0.50</td>
<td>R</td>
</tr>
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* Type of analysis: P = precipitation; R = rainout; A = air.
sources of sulfur compounds, shows the far greater scatter to be expected under these circumstances. The rainout shows a very pronounced flattening of slope as predicted by (4) above. The air regression has a greater slope which may or may not be significant, but most analyses lie farther to the right of the mean-atmosphere line than do the rainout points, indicating higher relative sulfur concentrations. This confirms the thesis that local influences are shown more strongly by surface air analyses than by rainout.

Similar effects are shown by the other stations considered. It is also worth noting that at Va-

much closer to the mean atmosphere than to seawater. Presumably this reflects the relatively calm weather of the mid-Pacific summer when little sea-spray is generated and the aerosol collected was predominantly that of the general atmosphere. Kinnvika, on the other hand, in an area of higher winds and generally less temperate climate, should show a much greater contribution of marine aerosol. Woodcock (private communication) has pointed out that a great deal of minute marine aerosol is generated when snow falls into water; this may constitute an additional source at some seasons.

Bonn (Figs. 3 and 4), with numerous local

![Fig. 1—Regression of chloride on sulfur for air, rainout, and precipitation at Kinnvika, Spitsbergen](image1)

![Fig. 2—Regression of chloride on sulfur for air at Weather Station November](image2)

![Fig. 3—Scatter diagram for rainout and precipitation at Bonn, Germany; the line is a visual fit](image3)

![Fig. 4—Regression for air at Bonn, Germany](image4)
DISCUSSION

REFERENCES


Dr. R. M. Schotland—Is there any possibility of stack gases from the ship contributing to the sulfur content of the samples taken at Station November?

Dr. Lodge—I think not. We were simultaneously measuring sulfur dioxide and found very little of it at any time.

Dr. C. E. Junge—To me it was very interesting that you found these industrial sulfur over the Pacific. I think I did not make my point very clear this morning when I spoke about the two hemispheres. Industrial sulfur, having only a lifetime of a few days, can therefore neither penetrate to the arctic areas nor into the tropical regions. So it will normally remain in middle latitudes, and was therefore found at Station November in the Pacific. The sulfur which we found in Greenland was natural, that is, we are pretty sure that the sulfur is primarily present as \( \mathrm{H_2S} \) which is oxidized and then washed out. This agrees with our findings that the residence time of the total sulfur which includes a natural component, is much longer than ten days, which is approximately the residence time of the industrial component (SO\(_2\)).

Dr. M. Neiburger—In connection with this it would be very desirable to know the lifetime of the pollutants which go into the atmosphere. How many times around the Earth will they travel before they are washed out? My frank opinion is that industrial sources still are affecting Station November even though the aerosol will have to come all the way from Europe or with easterly winds all the way around from the United States. It seems to me this is also of interest in terms of the organisms that might be being as inhibitors to the freezing nuclei. It would be very valuable if somebody could make an attack on that question.

Dr. Junge—I definitely think this material circulates around the northern hemisphere once or twice before most of it is washed out. We have now a number of residence times from various sources which are not too accurate, but they give an approximate picture. We know that the lifetimes are of the order of 30 days in the troposphere. They are shorter for some other pollutants, like sodium chloride formed from sea spray. Estimates show that they may be of the order of five to ten days. Of course, the water cycle has a ten-day turnover, so actually it looks as if most of the material which is readily washed out has approximately the same residence time.
as water. Some may remain two to three times longer depending on the kind of washout mechanism. I would say with a residence time of 10, or 20, or 30 days it can go once or twice around the world and even in the center of the Pacific you will get quite a bit of contamination from the continents.

**Dr. A. Goetz**—The analyses were all for sulfur only, not for sulfate?

**Dr. Lodge**—The samples were analyzed for sulfate, but it was computed as sulfur, the way the Swedish group reports it.

**Dr. Goetz**—The industrial SO₂ would require a fairly long time before it is converted into sulfate, whereas the ocean could probably only yield sulfate.

**Dr. Lodge**—A substantial percentage of these data were obtained from filter samples, and were not oxidized during the treatment; this would be sulfate. I am a little less certain of the Swedish data, which may include some SO₃.

**Dr. A. H. Woodcock**—Since it is fairly well established that chloride varies with the wind force over the sea, it would be interesting to know whether or not sulfur is also related to wind force.
Preliminary Results on the Aggregation of Ice Crystals

R. E. Hallgren and C. L. Hosler

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Abstract—A brief description of the experimental apparatus and procedure is given. Preliminary results of measurements of the collection efficiency of a 170-micron sphere as a function of temperature show a decrease of collection efficiency with temperature from 0.17 at −1°C to 0.04 at −20°C. The results are discussed with reference to growth of snowflakes, charge separation in thunderstorms and the bright band.

Introduction—The study of the mechanisms involved in the growth of cloud particles into precipitation particles has been one of the most important phases of cloud physics. The growth of water drops by condensation and coalescence has been treated extensively both theoretically and experimentally. Although all of the investigations are not in complete agreement and more work is needed for the smaller drops, one can make a reasonable estimate, from existing information, of the time required for a small drop to grow to the size of observed precipitation particles. In addition, the growth of ice crystals by sublimation has received considerable attention; however, the importance of the aggregation or conglomeration of ice crystals has been grossly neglected. In this case, in addition to the aerodynamic considerations, it is necessary to establish the environmental conditions in which two ice crystals will actually stick together.

The results of field experiments by Cunningham and Atlas [1953] indicated that in some storm systems the ice phase accounts for as much as two-thirds of the precipitation. It appears that a considerable amount was caused by the aggregation of ice crystals; however, it is quite difficult to separate the growth by sublimation and by aggregation in this particular experiment and to specify the conditions which are most conducive to aggregation. Several radar studies also indicate the importance of aggregation. In particular the attempts to justify the change in the echo intensity above the zero degree isotherm by sublimational growth alone seems to require unreasonable rates of growth of the ice crystals, and it appears that the aggregation of ice crystals must make an important contribution to the change in intensity of the echo. Very elaborate studies at McGill University [Douglas and others, 1957] of snow generating cells using vertically pointed radar, which gives a height-time record of snow echoes from which the terminal fall velocity of the particles could be determined, have shown that aggregates of ice crystals exist at very low temperatures. They found aggregates at temperatures as low as −34°C. In the cases they studied they found no preferred temperature at which snow cells form, but did find that the aggregation generally occurred within the first few thousand feet above a frontal surface.

Preliminary laboratory experiments carried out by Hosler and others [1957] have shown that the aggregation of ice crystals may be important at temperatures lower than had been expected, and that the temperature and the vapor pressure are extremely important in determining whether or not a collision between two ice particles will result in aggregation. They carefully manipulated two ice spheres together and measured the force required to pull them apart after being in contact for one minute. The force required to separate the spheres as a function of temperature and at ice saturation is reproduced in Figure 1. These measurements, although not realistically simulating two small ice crystals, indicate that aggregation should not be very important at ice saturation at temperatures below −25°C. Their measurements at near water saturation gave only a slight decrease with temperature in the force required to separate the spheres.

With this in mind, we have developed an apparatus in which a small ice sphere could be fixed in a flow of ice-crystal-laden air, and the growth of the sphere observed at various temperatures and vapor pressures.

1 The research leading to this paper was supported by the National Science Foundation grant G-3477.
2 Contribution No. 59-55, College of Mineral Industries.
Fig. 1—Relationship of mean force required to separate the ice spheres to temperature

Experimental apparatus—An overall view of the freezer, apparatus, and control panel is shown in Figure 2. The apparatus is mounted in a commercial freezer approximately 2 x 2 ft and 6 feet high. The temperature can be varied from room temperature to -20°C, with the temperature difference from the top to bottom of the freezer kept to approximately 1°C or less by a mixing fan located at the bottom of the freezer.

A schematic of the basic components of the apparatus is shown in Figure 3. It is, basically, a variation of the techniques used by Kinzer and Cobb [1956, 1958] in their studies on small water drops. Region A represents an inner chamber with a 4-ft³ volume in which an ice-crystal cloud is formed by means of a regulated flow of saturated vapor and subsequent seeding by dry ice. The concentration and form of ice crystals vary with temperature in the usual manner. The average diameter of the ice crystals is between 8 and 20 microns being in general larger at higher temperatures and smaller at lower temperatures. The cloud is slowly transported up to the mouth of the vertical tube (2-inch inside diameter) containing the sampling device by a negative pressure created by a pump. The crystals are then transported through the smaller tube (3/4-inch inside diameter plastic tube) in which the velocity can be varied from 0 to 100 cm/sec by varying the voltage applied to the motor of a blower which in turn varies the negative pressure created at the top of the tube by the Venturi effect. A rack gear at B permits the insertion of an ice sphere into the center of the tube. The ice sphere is mounted on the end of a fine rabbit hair or on a spider silk and is observed through a microscope. Immediately below the observational area is a sampler with which we measure the concentration of ice crystals passing through the test section. It permits a section of the tube to be shifted out of the flow, and subsequently the crystals contained in the volume precipitate onto a cover glass covered with a one-percent solution of Formvar and placed at the bottom of the shifted section of the tube. The height of the sampling section is two inches, and the cover glass is placed on a holder and then inserted into the center of the tube. Since the cover glass is located in the center of the tube, losses to the

Fig. 2—Overall view of freezer, control panel, and apparatus
walls and by impingement do not affect the accuracy of the sample. From samples taken only a few seconds apart, it appears that this method of sampling is reproducible and that the accuracy is very good. Another attempt to sample in the smaller tube immediately above the observational area has proved unsuccessful although we still have some hope of resolving the difficulties with this particular sampler.

The method currently being developed for maintaining varying degrees of supersaturation with respect to ice is to mix air at a known flow rate, temperature, and dewpoint with the ice-saturated air. The rate of flow of air is measured by means of a sensitive flowmeter. This method appears favorable for varying the vapor pressure with reference to ice saturation and water saturation, but will be limited in the sense that we will not be able to determine absolutely our position above ice saturation due to deposition on the walls of the tube and the ice crystals.

Experimental procedure—A cloud is formed within the inner chamber by a continuous source of vapor and then the cloud is slowly lifted through the larger tube. A period of ten seconds is required for the cloud to travel the length of the tube. During this time a small ice sphere is inserted into the observational area and the motors turned on so that the ice crystal cloud is carried past the ice sphere at a known velocity. The width and height of the aggregate is measured several times throughout the period of the run which in the cases reported here varied from three to six minutes. The cloud is sampled at the middle and end of the run. Immediately prior to sampling, the cloud is stopped in order to prevent impingement of the cloud on the tube. A cover glass is then removed from the plastic solution, placed on a holder and inserted at the bottom of the volume to be sampled. The section is then shifted and the run continued. In the case of the final slide of the run, the tube is not shifted, but a known volume is closed and the cover glass inserted. Several minutes are allowed for the ice crystals to precipitate onto the cover glass. The cover glass and holder are then removed and left to dry. A complete analysis for concentration, size distribution, and type of ice crystal is made from the slide. From the concentration, velocity in the tube, length of run, and the average area of the aggregate (actually the area of a circle of the diameter of the aggregate), we obtain the number of ice crystals for which collision was possible. At the end of the run the aggregate of ice crystals is melted and the diameter of the drop is again measured. From the increase in size, the mass can be determined, and using the average volume of the crystals counted, we obtain the actual number of crystals that adhered to the sphere. The volume of the crystal is determined by measuring the diameter of the ice crystal and from average measurements of the thickness obtained from a side view of the crystal through the microscope. At present a technique is being developed so that the thickness can be determined by shadowing the replicas and measurements made from electron photomicrographs. With the two, we define the ratio of crystals collected to the total number of crystals in the path of the collector.

Preliminary results—A summary of the average measurements of some of the parameters necessary to compute the ratio is shown in Table 1, along with the ratio at several temperatures. All of the measurements reported here.
TABLE I—Average values of measurements

<table>
<thead>
<tr>
<th>Temperature</th>
<th>Length of run</th>
<th>Increase in mass of ice sphere</th>
<th>Density of aggregate</th>
<th>Final diameter of aggregate</th>
<th>Crystal concentration</th>
<th>Type of crystal</th>
<th>Size of crystal</th>
<th>Ratio: crystals collected to crystals in path</th>
</tr>
</thead>
<tbody>
<tr>
<td>°C</td>
<td>sec ( \times 10^6 ) g</td>
<td>gm/cm(^3)</td>
<td>micron</td>
<td>cm(^3)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-4</td>
<td>197</td>
<td>3.2</td>
<td>0.05</td>
<td>450</td>
<td>3050</td>
<td>Needle</td>
<td>1170</td>
<td>6.3 ( \times ) 39</td>
</tr>
<tr>
<td>-8</td>
<td>147</td>
<td>2.0</td>
<td>0.04</td>
<td>435</td>
<td>8100</td>
<td>Plates</td>
<td>320</td>
<td>8.7</td>
</tr>
<tr>
<td>-12</td>
<td>175</td>
<td>5.1</td>
<td>0.03</td>
<td>740</td>
<td>14700</td>
<td>Columns Plaques</td>
<td>350</td>
<td>6.3 ( \times ) 11.5</td>
</tr>
<tr>
<td>-20</td>
<td>207</td>
<td>2.07</td>
<td>0.02</td>
<td>570</td>
<td>7300</td>
<td>Plates</td>
<td>125</td>
<td>10.5</td>
</tr>
</tbody>
</table>

were taken with a sphere with an initial diameter of 170 microns. The increase in crystal concentration with decreasing temperature accounts for the manner in which the amount of mass collected on the ice sphere and the final diameter of the aggregate vary. If reduced to a common crystal concentration and length of run, the mass collected and the final diameter is actually much less at the lower temperatures. The ratio of crystals collected to crystals in the path of the collector decreases with decreasing temperature. The decrease with temperature is quite uniform, being 0.17 at \(-4^\circ\) C and 0.04 at \(-20^\circ\) C. Whether or not the factor of four between the collection efficiency at \(-4^\circ\) C and \(-20^\circ\) C represents entirely a change in the number of ice crystals that stick after a collision is open to some question because of the method utilized in estimating the area swept by the aggregate. Actually, since the density of the aggregate is less at lower temperatures than at higher temperatures, one would expect that an ice crystal would have a better chance of actually being within the area and still not having a collision with another crystal at the lower temperatures. We are currently attempting to obtain a more realistic estimate of the actual area. However, the type of growth we observe probably is similar to that occurring in clouds since our densities are approximately the same as the density of snowflakes. Therefore, the results are satisfactory for computations of the growth of aggregates within clouds. Although we did not measure the charge accumulated by the aggregate, the studies of Reynolds and others [1957] indicate that it should be negligible since the cloud used in these particular runs contained only ice crystals and no supercooled droplets.

It is interesting to note that the area presented to the flow by the collector increases quite rapidly because the initial growth is quite open and results in a rapid decrease of the density of the aggregate. This could very well lead to a particle of sufficient size to reflect a radar signal in a rather short period of time. Observations of an aggregate growing in our chamber show extensions growing out from the sphere a considerable distance before any appreciable 'filling in' occurs. For example, in Table 1 we see that while the increase in mass was only sufficient to increase the diameter of the 170 micron sphere slightly at \(-4^\circ\) C, the final diameter of the aggregate was on the average 450 microns. Similarly, at \(-8^\circ\) C the average final diameter of the aggregate was 435 microns, at \(-12^\circ\) C the final diameter was 740 microns and at \(-20^\circ\) C the final diameter was 570 microns. Actually, the increase would be more pronounced if the concentration of ice particles had not increased with decreasing temperature. In addition, the average length of the run increased with decreasing temperature. It seems to be true when observing growth of an aggregate that the final size of the aggregate is considerably larger with larger ice crystals. In other words, the extension grows much more rapidly with the larger ice crystals (our crystals are largest at \(-4^\circ\) C) and results in a much lower density growth initially than with smaller crystals. However, it is, at least with the present information, extremely difficult to make any estimate of the time required for the particle to grow to a particular size because the relation between the length of the extension and the size of the crystals has not yet been established.

The rapid growth of the extensions could explain the recent observations of Vonnegut and Moore [1958] which indicate that the initial echo on the radar scope is in the form of an inverted cup and that the echo develops very rapidly. This, combined with their observation of a veil
of ice crystals surrounding the core of the cloud, might be explained in the following manner: Drops grow in the core of the Cumulus to a diameter that cannot reflect a radar signal, say 40 microns, and upon arriving in the layer of ice crystals surrounding the Cumulus grow quite rapidly by means of the extensions to a size that can be detected by the radar. Unfortunately, present airborne sampling techniques do not enable us to detect the presence of such small aggregates.

These extensions would only make an important contribution to the area swept out by the aggregate when the collector is rather small. It seems reasonable that there is some critical dimension after which the 'filling in' would proceed at the same rate the extensions are growing.

For some time there has been a good deal of speculation as to how much of the increase in the radar signal above the bright band can be attributed to aggregation and how much to sublimational growth of the ice crystal. With the above results, one can see that even at ice saturation considerable growth could be achieved through aggregation down to temperatures lower than had been thought.

The above results also have some important implications with reference to charge separation in clouds. The graupel process, as it has been developed by Reynolds and others [1957], requires the separation of the charge by means of frictional contact between the larger ice particle and the ice crystals in the cloud. In the experiments of Reynolds, which were carried out in an ambient temperature of between -12 and -17°C, with the growing graupel particle being three or more degrees warmer, one would expect a large number of the ice crystals actually to adhere to the graupel particle. If one assumes that the 0.17 measured at -4°C represents 100% collection after collision, then even at temperatures down to about -15°C approximately one-half of the crystals adhere after collision even at ice saturation. In addition Reynolds and others [1957] observed the maximum charge separation in a cloud near water saturation which is where Hosler and others [1957] report only a slight change with temperature in the ability for ice crystals to adhere, and therefore, places the assumption (made by Reynolds) of the amount of charge separated per collision by assuming that none of the crystals adhered, in error by a considerable amount. Our contemplated investigations should clarify this particular point.

It is interesting to speculate upon the role collection efficiencies of ice pellets in an ice or mixed cloud may play in charge separation in thunderstorms. Should the collision between a graupel or hail pellet and an ice crystal, without collection, be a necessary prerequisite to the rapid separation of charge, then a pellet falling through an all ice cloud at the lowest possible temperature would yield the largest charge separation. This is due to the observation that fewer crystals are collected at ice saturation than at higher vapor pressure, and that the lower the temperature, the fewer ice crystals that are collected. This would not agree with Reynolds and others [1957] observations, but it might well explain the increased lightning in thunderstorms where cloud seeding has been employed to transform a greater proportion of the clouds to ice [Battan and Kassander, 1960]. If Reynolds' observations are correct, then in light of our experiment, another charge separating mechanism must be operative. However, it may be that the charges generated in Reynolds' mixed clouds were due to the small amount of bounce-off that occurred and had he performed the experiment at high crystal concentrations in an all-ice crystal cloud he would have measured even more charge generation.

With our projected measurements at ice supersaturations, we also hope to explain the result of some radar studies of snow generation cells carried out at McGill [Douglas and others, 1956]. These observations indicated that the majority of the cells occurred just above a frontal surface where the uplift of the front would be most intense. Combined with the general uplift that was occurring under the situations studied, one would expect these stronger regions of uplift to be supersaturated with respect to ice, thereby making aggregation possible at all temperatures. In these studies the conditions were always such that the atmosphere was supersaturated with respect to ice, and for the several observations that indicated generating levels at very low temperatures, it appears the area was also somewhat unstable at the generating level, which would create convection and higher supersaturations with respect to ice.

References


Discussion

Dr. U. Nakaya—How do you measure these ice crystals?

Dr. C. L. Hosler—We have a man who sits there eight hours a day and measures these crystals. We have made replicas in Formvar. We have spent a year and a half trying to measure the actual number and size of the crystals in our cloud. It has been our most difficult problem to make sure we had a representative sample and to know what we had in the test section. We had inserted a collection device above and below the test section. We are finally fairly convinced we are actually measuring the population of test crystals going through our section. The process of collection of individual crystals is observed by means of a microscope attached to the test section. Their thicknesses and lengths are determined throughout the test. In addition, we have these samples collected by actually rotating out a section of the tunnel and by gravitational settling collecting them in a Formvar solution and making plastic replicas.

Dr. W. Hitzfeld—I think these are very beautiful data, but I am still surprised to find that the collection efficiencies you measured are so low.

Dr. Hosler—Do not put too much weight on these efficiencies because we computed them using an average cross-sectional area. In reality, they present weird-shaped aggregates with a lot of holes in them. Maybe one-half of this average area is actually occupied by ice and one-half is holes.

Dr. Hitzfeld—So you could double these collection efficiencies?

Dr. Hosler—Yes, now we are trying to measure the true area of these crystals, but we have not accomplished it yet.

Dr. B. J. Mason—Those values on the right in Table 1 really do not represent 'sticking' alone, because they are really partly due to aggregation and partly due to sticking. The interpretation of exactly what those figures mean will be difficult because the hydrodynamics of the problem is changing all the time depending on the shape of the crystals one is using. But, nevertheless, what you are measuring is the overall thing which meteorologically is important. Of secondary importance perhaps is whether you noticed any major difference, even in a qualitative way when you changed from needles, say, to plates.

Dr. Hosler—One observes very interesting things. The bigger the crystals are, the more quickly this aggregate assumes a greater area. Some detailed observations appear to be quite important: We were afraid, for instance, when long extentions would build out, that they would break off and fall away. Instead, they build out to a certain length and then fold right back into the aggregate. Then it merely fills in and gives a graupel-type buildup.

Dr. Helmut Weickmann—I have the impression that your experiment is a good analogy to what really happens in nature during the aggregation in a convective cloud layer. Such a layer has a little convective activity with up and down draft velocities of one-half to one meter per second. This is enough to support some of the crystals. The ensuing small-scale turbulence creates a much better chance to aggregate than if they just meet each other during a straight fall. If they go up and down, veering left and right, they meet each other much more frequently.

Dr. Hosler—in addition to this, when falling through a cloud layer a water film may form on the crystals which greatly assists the sticking.
Dr. Choji Magono—As the crystal type is very important in the formation of snowflakes; what type did you use in the experiment?

Dr. Hosler—We had plates, columns, and needles but because of the temperature range along our wind tunnel it was difficult to control this. However, we can work with plates or needles by just changing the environment we are creating.

Dr. Magono—Did you use dendritic types?

Dr. Hosler—No, we were unable to obtain them in our small chambers. Perhaps the Swiss can get dendrites. In ours, we must wait a long time before we get dendrites.

Dr. Magono—In nature we find that snowflake formation begins at an air temperature warmer than \(-10^\circ\text{C}\).

Dr. Weickmann—(communicated) Dobrowolski in his fundamental but little-known work on snow-crystal observations in the Antarctic (A. Dobrowolski, La neige et le givre—Rapports Scientifiques; Résultats du voyage du S.Y. Belgica en 1897-1898-1899; La Météorologie, 1904) gives the following table on the dependence of snowflake occurrence from the ambient temperature:

<table>
<thead>
<tr>
<th>Temperature</th>
<th>Snowflake occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td>+1.0 to (-5.0^\circ\text{C})</td>
<td>83%</td>
</tr>
<tr>
<td>(-5.1) to (-10.0^\circ\text{C})</td>
<td>9%</td>
</tr>
<tr>
<td>(\geq) (-10^\circ\text{C})</td>
<td>8%</td>
</tr>
</tbody>
</table>
Growth by Accretion in the Ice Phase

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Abstract—The growth of spherical ice particles of various densities by sublimation and accretion is considered. The less dense the particle, the greater the mass it must achieve by sublimation before accretion becomes the dominant growth mechanism. Once this stage is achieved, however, growth rates of particles of the same mass are relatively insensitive to particle density; the cloud water content exercising the major control. With low water content (0.1 gm m\(^{-2}\)) such as in stratiform clouds and in the dilute peripheral regions of cold Cumulus, the precipitation products are essentially sublimation elements rather than graupel. In four gm m\(^{-2}\) Cumulus, low-density graupel can grow to millimeter size within six minutes and to centimeter size within ten minutes, much denser particles requiring only a few minutes longer to reach the same sizes. These times are comparable with the observed elapsed times of about 15 min between the detection of the first radar echo and the first appearance of hail at the ground.

Introduction—The accretion process is important to the production of graupel and hail; there is a need for study of the initial growth of the primitive ice particle and its transformation from a sublimation to an accretion element. The appearance of graupel seems to be related to the development of electrical fields in Cumulus [Fitzgerald and Byers, 1955]. While the Thunderstorm Project [Byers and Braham, 1949, p. 48] reported very few flight encounters of hail in thunderstorms, this may have been due in part to the localization of hail within the storm; identification of graupel would likely be difficult from an aircraft. However Kuettner [1950] claims that graupel is the most frequently observed hydrometeor in thunderstorms occurring over the Zugspitze (at 10 kft), being always associated with high electric fields; large hail was found to be a rare and by no means a necessary occurrence in the 125 storms which he studied. Thus, while hail may not accompany every thunderstorm, it appears that graupel does.

For a small frozen droplet, growth proceeds mainly by sublimation until a critical size is reached, following which accretion predominates; it has usually been particles above this critical size whose growth has been studied, Ludlom's [1952] study of the development of ice particles and their role in shower initiation being a noteworthy exception. Also, the graupel problem involves densities which may be as low as 0.04, according to Magono's [1954, p. 38] data, and relatively few studies have considered densities even as low as 0.1; the present paper examines the growth of spherical particles (such as frozen droplets) of densities from 0.05 to 0.9, by sublimation and accretion occurring together.

Growth rates—Sublimational growth at a rate \((dm/dt)\), supplies heat to a growing spherical particle at a rate

\[ Q_1 = L_e (dm/dt) \]

where \(L_e\) is latent heat of sublimation, \(S\) is particle radius, \(C\) is ventilation factor, \(D\) is diffusivity of water vapor in air, \(\Delta \rho\) is excess of ambient vapor density over that at the crystal surface. Growth by accretion at a rate \((dm/dt)\) supplies further heat at a rate

\[ Q_2 = L_f (dm/dt) \]

\(L_f\) being the latent heat of fusion. Heat is lost to the environment at the rate

\[ Q_3 = 4\pi SKC \Delta T \]

where \(K\) = thermal conductivity of the environment and \(\Delta T\) = excess of temperature at the crystal face over that of the environment. In the steady state, \(Q_1 + Q_2 = Q_3\), whence

\[ \frac{\Delta \rho_a}{\Delta T_a} = \frac{K}{4\pi KL_f} \left[ 1 - \frac{L_f (dm/dt)_a}{4\pi KSC \Delta T} \right] \]

in which the subscripts \(a\) on the left hand side indicate the presence of accretion. The physical significance of \(\Delta \rho_a/\Delta T_a\) is shown in Figure 1 in an environment which is saturated with respect to water, such as would exist in the supercooled
GROWTH BY ACCRETION IN THE ICE PHASE

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water cloud in which accretional growth would occur.

Now $K/DL_d = \Delta \rho_i/\Delta T_s$, for sublimation alone [Marshall and Langleben, 1954], so that

$$\frac{\Delta \rho_i}{\Delta T_s} = \frac{\Delta \rho_g}{\Delta T_g} \left[ 1 - L_i \left( \frac{dn}{dt} \right) \left( \frac{4\pi K S C (\Delta T)^{-1}}{\rho_g} \right) \right]$$

and since $\Delta T$ is positive, the crystal being warmer than the air, it follows that $\Delta \rho_i/\Delta T_s < \Delta \rho_g/\Delta T_g$, the difference being proportional to the accretional growth rate. Thus the effect of accretion is to increase the temperature excess of the growing particle, and so to reduce its vapor density excess which in turn reduces the sublimational growth rate. The total growth rate, then, is the sum of the sublimation-only and accretion-only components, the former reduced by a term which is proportional to the latter. At 700 mb and $-10^\circ$C the reduction term is approximately three per cent of the accretional growth rate; it increases with temperature and with pressure to about six per cent at 1000 mb and 0°C. Considering all other possible sources of error in growth calculations, it appears reasonable to ignore this relatively small reduction, and to consider the total growth rate as the sum of the two components, sublimational and accretional.

Density of particle and of accreted material—The density of the particle is important in determining its terminal speed, and may of course vary during growth. Densities of hailstones are usually estimated to be about 0.7–0.8 [see, for example, Weickmann, 1955, p. 129]. Densities of graupel have been measured by Nakaya [1954, p. 115] who reports a value of 0.125, invariant with particle diameter, over a range of 1.5 to 6 mm. Magono [1954], theorizing on the terminal speed of graupel, finds agreement with observed values when the density is of that order, but presents observational data indicating densities as low as 0.04. That the density of the accreted material is dependent upon the ambient temperature and liquid water content is suggested by riming data of Clark [1948] and by the results of Melcher's [1951] experiments [see Weickmann, 1953], also by the riming data of Langmuir [1944; see Ludlam, 1958, p. 55]. Generally speaking, the warmer the temperature, the higher the cloud water content, and the greater the velocity of impact, the denser is the accreted material. Thus, for the smallest particles on which accretion is just beginning, a relatively low density of accreted material is reasonably to be expected; with increasing size and fall speed, this density should increase, and Magono's [1954, p. 38] data do indeed show a roughly linear increase in density from 0.04 to 0.17 as graupel diameter increases from 1 to 6 mm. Therefore, in considering the growth of particles, particularly in the early stages, consideration must be given to relatively low particle densities.

In the present study, growth has been treated arbitrarily at constant density. When a particle of initial mass $m_s$, and density $\sigma_s$ accretes material of density $\sigma_i$, then the particle density becomes $\sigma$ when it has grown to mass $m$ according to the equation

$$(\sigma - \sigma_s)/(\sigma_i - \sigma_s) = m/m_s$$

Thus, as the particle mass increases tenfold, the density difference between the particle and the accreted material decreases to one-tenth its initial value. It will be found (Fig. 2) that this density change can be accomplished within a few minutes.

Growth calculations—Growth rates were computed for spherical particles of densities 0.05, 0.2 and 0.9, representative of the range from Magono's [1954] least dense graupel to dense hailstones. Terminal speeds were determined on the basis of Langmuir's [1948] work. Sublimational growth rates were calculated on the basis

Fig. 1—Curves show schematically the equilibrium vapor density with respect to water $\rho_v$ and to ice $\rho_i$ as a function of temperature; at the face of an ice crystal, growing by sublimation in water-saturated air, the temperature exceeds the ambient value by $\Delta T_s$, and the growth rate is proportional to the vapor density excess $\Delta \rho_i$; accretion increases the former to $\Delta T_i$ and reduces the latter to $\Delta \rho_a$. 

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[Diagram showing relationships between various parameters]
Fig. 2—Mass of a spherical ice particle as a function of time; the coalescence curve shows the growth of a liquid droplet in 4 gm m^{-2} Cumulus; cross marks on the 4 gm m^{-2} curves indicate particle diameters of 1 cm and (on the coalescence curve) of droplet radius 50 \( \mu \); open circles indicate masses at which the sublimational and accretional growth rates are equal (in the As case, growth is predominantly sublimational within the range of the figure).

of a water-saturated environment at 700 mb and -15°C. Accretional growth rates were calculated by means of the growth equation

\[
\frac{dm}{dt} = \int_0^r \pi (S + r)^2 UEW, dr
\]

where \( S \) = particle radius, \( r \) = droplet radius, \( U \) = velocity of particle with respect to the droplet, \( E \) = collection efficiency of particle with respect to the droplet, \( W, dr \) = liquid water content contained by droplets of radius \( r \) to \( (r + dr) \). Droplet growth by condensation and coalescence has been treated at length by East [1957], whose derived cloud droplet distributions appropriate to 1 and 4 gm m^{-3} Cumulus were used in the present work; for stratiform cloud of low water content, Diem’s [1948, p. 261] distribution for Altostratus was used.

For convenience in comparing the collection characteristics of various particles in various environments, it is useful to consider an ‘effective collection efficiency’ \( E' \), the fraction of the liquid water content, contained within the swept volume, which is accreted. Thus \( E' \) is defined by the equation

\[
dm/dt = \pi S v_c EW
\]

where \( v_c \) = terminal speed of the collecting particle, and \( W \) = liquid water content. In this way \( E' \) incorporates the effects of droplet size distribution and of relative speeds. Figure 3 shows \( E' \) as a function of mass, for particle densities of 0.05, 0.2, and 0.9, in 1 and 4 gm m^{-3} Cumulus and in Altostratus. The improvement in collection as particle density increases, and as one proceeds from Altostratus to increasingly dense Cumulus, is clearly indicated, as also is the approach, for large particles, toward a maximum value. However an extension of these curves to even larger masses tends to show a decrease in \( E' \); such a decrease at large radii is evident in Langmuir’s [1948] Table 4 and in Gunn and Hitschfeld’s [1951] Figure 1.

Figure 4 shows the growth rate as a function of particle mass, obtained by the addition of the sublimational and accretional rates. The particle masses at which these two rates are just equal, when the particle density is 0.9, are indicated by the arrows; for densities of 0.2 and 0.05 the critical masses are displaced to higher values by factors of about 3 and 8, respectively. Thus the denser the particle and the higher the cloud content, the larger is the critical mass, and the radius, to which growth must proceed before accretion becomes dominant. Considering cloud of low liquid water content, it is seen that particles of mass greater than \( 10^6 \) to \( 10^7 \) \( \mu \)gm may be expected to exhibit riming. This is in general agreement with observations of (unrimed) spatial dendrites and of crystals with droplets attached; of these crystal types, the largest observed by Nakaya [1954] had masses in the range 300-500 \( \mu \)gm. Particles of masses greater than these were in the form of graupel [see Nakaya, 1954, Fig. 224].

Also shown in Figure 4 is the growth curve, by accretion only, of a 0.9-density particle in 4 gm m^{-3} cloud. This particle density is nearly enough equal to that of a liquid droplet that this curve may be used to describe droplet growth by coalescence, and to compare droplet growth with graupel growth.
Fig. 3—Effective collection efficiency $E'$ as a function of particle mass for spheres of densities 0.05, 0.2, and 0.9 in Altostratus and in Cumulus of 1 and 4 gm m$^{-3}$.

Fig. 4—Growth rate of a spherical ice particle by sublimation and accretion, as a function of its mass, for various particle densities and in various clouds; the coalescence curve shows the growth rate of a liquid droplet in 4 gm m$^{-3}$ Cumulus; particle masses (density 0.9) at which the accretional equals the sublimational growth rate in the various clouds are indicated by the arrows.
The most striking feature of these growth-rate curves is the relative insensitivity of growth rate to particle density. For any given mass, variation of density by a factor of 18 results in a variation of growth rate by a factor of only 3 or less; for a given mass, a reduction in density tends to increase the swept volume, but reduces the effective collection efficiency by an almost equivalent amount. The major control of growth rate is clearly the liquid water content. At the upper ends of the curves, the growth rate is proportional to the \(5/6\) power of the particle mass.

Curves showing particle mass as a function of time are given in Figure 2. Initial particle mass was taken to be \(10^{-2} \mu\text{g/m}^3\), but the individual curves can be used for any greater initial mass. Also included is one curve for growth by accretion only, of a particle of density 0.9 in \(4 \text{ gm m}^{-3}\) Cumulus, which may be considered to represent droplet growth by coalescence; this curve begins at droplet radius \(20 \mu\), below which growth proceeds entirely by diffusion.

Results—Houghton [1955] has compared droplet growth by coalescence in 1 gm m\(^{-3}\) cloud with the sublimational growth of ice crystals, and found the latter to be more rapid for masses up to about 10 \(\mu\text{g/m}^3\) (melted radius 130 \(\mu\)), a result in substantial agreement with the present calculations when the coalescence rate in 1 gm m\(^{-3}\) cloud is compared with the low-density sublimation rate. But the present study provides the means of comparing growth of liquid drops with growth of frozen drops by sublimation and accretion, and over a wide range of density.

Considering growth in 4 gm m\(^{-3}\) Cumulus, the frozen droplet will have the advantage over the unfrozen one for masses up to about 1 \(\mu\text{g/m}^3\) (melted radius 60 \(\mu\)) (Figs. 2 and 4). In 1 gm m\(^{-3}\) cloud the advantage persists for masses up to 10 \(\mu\text{g/m}^3\) (melted radius 130 \(\mu\)), and in Altostratus of 0.1 gm m\(^{-3}\), up to masses of at least 10 \(\mu\text{g/m}^3\) (melted radius > 600 \(\mu\)). The less dense accretion particle has a slight advantage over the more dense one (Fig. 4). If liquid drops freeze at smaller sizes than those noted above, they will continue growth more rapidly than their unfrozen neighbours, but if freezing is postponed until larger sizes are reached, the frozen particle loses most of this advantage. East [1957] shows that the initiation of the all-water coalescence precipitation process follows closely upon the appearance of 50 \(\mu\) radius droplets in 4 gm m\(^{-3}\) Cumulus. But if droplets of lesser radius should freeze before any have reached 50 \(\mu\), these frozen particles can compete vigorously with the unfrozen remainder, and precipitation growth through the ice phase is competitive with that through the all-water process. This applies only to summer Cumulus of the temperate regions, in which cloud water contents of 4 gm m\(^{-3}\) are attained above the freezing level. With higher cloud contents such as may be achieved in warmer Cumulus, the advantage of the all-ice growth process over coalescence may be greatly reduced, since accretion ‘cuts in’ at a smaller particle size the greater the cloud water content.

The effectiveness of the growing particles in removing the cloud water content is best defined as the mass collected per unit distance fallen. While Figure 4 shows the approximate equality of growth rates (within a factor of 2) at various particle densities, the less dense particle, falling more slowly, collects more cloud water per unit distance fallen. Thus, in falling through a unit depth of cloud, the less dense particle sweeps up more cloud water than a denser particle of the same mass. In the three types of cloud considered here, particles of density 0.05 may in this sense be up to six times as effective as those of density 0.9; thus low density accretional growth is the more potent cloud-consuming process.

In cloud of low water content (0.1 gm m\(^{-3}\)) growth to the stage where accretion becomes the dominant mechanism requires times in excess of half an hour, and cloud depths from 3000 to 5000 ft for low- and high-density particles, respectively. Unless cloud exists in such depth, the emergent precipitation particles will be more of the nature of sublimation elements (snow) rather than graupel; with cloud layers of reasonable thicknesses, only the lowest density graupel is likely to emerge. As cloud water content increases, and accretion is increasingly favored, the likelihood of accretion elements increases, and conditions for significant higher-density growth improve. In cold Cumulus, the dense undiluted core may be favorable for graupel growth, while the dilute cloud boundaries will yield sublimation elements, or snow; such precipitation of both graupel and snow is occasionally observed from small Cumulus during the cold seasons of the temperate regions.

In denser cloud of 4 gm m\(^{-3}\), accretion becomes predominant at masses of the order of 1 \(\mu\text{g/m}^3\), which may be achieved within the first four minutes of growth; particles of 1 cm diameter are grown within an additional twelve minutes. Ob-
servations of radar echoes and of hail occurrence indicate that, at times, hail may reach the ground within fifteen minutes of the first appearance of echo. In 4 gm m⁻³ Cumulus, coalescence will produce a 50 μ radius droplet within about ten minutes, at approximately which time the first echo will be detected [East, 1957]. Following detection, the droplets may freeze and continue growth in the ice phase, or may grow further by coalescence before freezing, depending upon the temperature.

Calculations were made (from Fig. 2) of the times required for the growth of graupel or hail of 1 cm diameter, starting with a 50 μ radius drop, and varying the size at which the liquid drop froze. Even considering drops freezing over a wide range of sizes, and subsequent growth over a wide range of particle densities, the times to reach 1 cm diameter varied only from 9 to 16 mins, the shorter time being for the low-density growth. These times are comparable to the 15-min interval observed between echo detection and appearance, at the ground, of the first hail.

The growth rates and times involved, as discussed above, are only applicable until depletion of the cloud content becomes significant, and so may be relevant only to the first ‘burst’ of hail. The shape of the primary graupel nucleus is not known, and may not be spherical. It is not known whether dense hail begins as graupel or as a frozen drop, and it may grow in supercooled rain; high density growth by collision with rain drops has not been considered, and undoubtedly should be. Reliable data on stone structure, and on the laboratory growth of various types of particles in various environments, are urgently needed to clarify these points of uncertainty.

Acknowledgments—The author is on assignment to the Stormy Weather Group, McGill University, the assistance of whose members is gratefully acknowledged. Participation in this research, and publication of the above, is with the permission of the Director, Meteorological Service of Canada.

This paper constitutes a revision and extension of work reported earlier in Stormy Weather Group Scientific Report MW-26, Growth of Precipitation Elements by Sublimation and Accretion, May 1957. The present results are considered more reliable than in the earlier work.

References

Gunn, K., and W. Hitzchield, A laboratory investigation of the coalescence between large and small water drops, J. Met., 8, 7–16, 1951.

Discussion

Dr. R. List—Have you taken into account the Nakaya-Higuchi observations of drops which are caught by the crystals, and which are rolling over their surface. I think that this may give a greater rate of growth than that by the normal sublimation which is taken into account here.
Dr. U. Nakaya—Last year we made a movie film which shows the mechanism of a small droplet captured on the surface of ice; these ice samples are polished and are exposed to the air for some time. We place a very small water droplet (one or two microns) on the surface. This droplet does not freeze to the surface of the ice but it rolls over the surface getting smaller and smaller and, finally, it vanishes. That means this droplet as a whole freezes onto the surface, but does not give a rimed crystal. It spreads out over the surface evaporating very quickly within 0.01 and 0.1 second.

Dr. Hitchenfeld—Obviously, this has not been taken into account, but I doubt that it would change Dr. Douglas' results very greatly if it had.
Frequency Distributions of Precipitation

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Abstract—Considerable difficulty is involved in the physical interpretation of the frequency distribution of precipitation. Such distributions usually do not follow the law of a gaussian distribution in a linear scale. This brings up the question for transformation to normality. The problem is rendered more difficult as the usually observed data have to be considered as truncated on the dry side.

The author suggests the use of a logarithmic scale. A frequency distribution of annual precipitation generally consists of one collective, while in monthly values and shorter amounts the mixture of the rainfall processes becomes obvious. A sample for the frequency distribution of daily amounts at Asheville, N. C., is discussed. The collective of excessive daily rain in autumn could be explained in connection with hurricanes on the east coast of the United States and the movement of extratropical cyclones through North Carolina. A two-dimensional analysis of intensity and duration for single rainfalls in Braunschweig (Germany) is also discussed. The sample demonstrates what is necessary in order to read more details from the precipitation amount either to split the rainfall data into intensity groups (such as convective and advective types) or to keep the observational interval as short as possible. Hourly data may be sufficient.

Introduction—The physical interpretation of the observed data is one of the primary topics of climatological work. If it were possible by some method to determine all physical processes of rainfall a priori, and collect the data into separate and distinct groups, then probably most of the following discussion would not be necessary. However, the basic physical processes leading to precipitation have not been completely investigated as yet and in many cases the set of observed data available consists of a complex of physical events which we call weather or climate.

This mixture exists also for other meteorological elements, for which interpretation may be easy. Considerable difficulty is involved, however, with frequency distributions of precipitation. In a linear scale they usually do not follow the law of a gaussian distribution and represent a mixture of different processes of rainfall, at least in the form of hourly, daily, monthly, etc., amounts.

The kind of distribution we may expect from daily and shorter periods of observations is well known. It shows a maximum in the class of smallest precipitation amount and decreases toward higher amounts. A typical sample is illustrated by Schneider-Carius [1954], who smoothed the frequency distribution in taking four German stations (Berlin, Bremen, Karlsruhe, and Munich) and all months of the year together (100,000 values).

This summation merely serves to demonstrate the type of frequency distribution or to derive an average rainfall probability for an area, though its applicability for climatological details is very limited.

Data in linear scale—Using a linear scale for frequency distributions of precipitation, several statistical approaches may be made. The best fit, to date, under the limitations discussed below, may be obtained by applying an incomplete gamma function though the possibility also exists to employ a hyperbola, $e^{-x}$ function (= also the limiting form of incomplete $\Gamma$ function) or the negative binomial.

The hyperbola or a form $e^{-x}$ may solve problems of rainfall probability, but furnishes practically no physical result. Wanner [1939] tried to use the negative binomial series for curve fitting in recognition of the persistence involved in meteorological elements. However, this applies a discrete function to continuous data and the author [Essenwanger, 1956a] has previously discussed the difficulties connected in arranging precipitation data in discrete form to meet the requirement of the negative binomial distribution. Besides this, the persistence for rainfall data is not a constant. This may be proved for
Table 1—Persistence factor $f$ for various amounts of precipitation, Hamburg area, Germany, winter

<table>
<thead>
<tr>
<th>Limits</th>
<th>Dry</th>
<th>Traces 0.01</th>
<th>0.02–0.03</th>
<th>0.04–0.09</th>
<th>0.10–0.40</th>
<th>&gt;0.40</th>
</tr>
</thead>
<tbody>
<tr>
<td>inch</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dry</td>
<td>1.46</td>
<td>0.89</td>
<td>0.79</td>
<td>0.68</td>
<td>0.59</td>
<td>0.44</td>
</tr>
<tr>
<td>Traces</td>
<td>0.71</td>
<td>2.09</td>
<td>1.77</td>
<td>1.31</td>
<td>1.21</td>
<td>0.94</td>
</tr>
<tr>
<td>0.01</td>
<td>0.52</td>
<td>1.17</td>
<td>1.27</td>
<td>1.28</td>
<td>1.21</td>
<td>1.32</td>
</tr>
<tr>
<td>0.02–0.03</td>
<td>0.58</td>
<td>0.81</td>
<td>1.24</td>
<td>1.33</td>
<td>1.40</td>
<td>1.53</td>
</tr>
<tr>
<td>0.04–0.09</td>
<td>0.55</td>
<td>0.72</td>
<td>0.91</td>
<td>1.06</td>
<td>1.45</td>
<td>1.72</td>
</tr>
<tr>
<td>0.10–0.40</td>
<td>0.45</td>
<td>0.68</td>
<td>0.26</td>
<td>0.71</td>
<td>1.36</td>
<td>2.16</td>
</tr>
<tr>
<td>&gt;0.40</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3.81</td>
</tr>
</tbody>
</table>

daily precipitations. Table 1 shows a persistence factor $f$, which is the observed frequency of occurrence divided by the expected value without persistence (discussed in detail by the author [Essenwanger, 1956a]).

This factor is listed for several classes of daily amounts and it is easy to see that this is not a constant for the different rainfall groups. Thus, departures between theory and observation may be expected from this result, even when it would be possible to establish an adequate discrete scale. The parameter of persistence $d$, which can be derived from this negative binomial theory may perhaps lead to some climatological relations, although in general, the result is influenced by the discrepancy mentioned above.

This leaves the incomplete $F$ function, which has been successfully applied by Berger and Thom [1949; Thom 1958, 1957ab]. This function renders good fits for amounts of approximately five days and more [Berger, 1957] depending on the proportion of near zero rainfall days, while for shorter periods the application is restricted to special collectives like the storm series [Thom, 1957ab]. Therefore, the question is still open: How should we treat precipitation data for periods shorter than five days? Before giving some answers to this problem, a few facts about transformation of scales for precipitation data should be discussed.

Transformation of scales—The author distinguishes between two sorts of transformation. In the first case one applies a pure formal mathematical function, developed to reduce the data into the desired form, such as a gaussian distribution. In the second case we derive the transformation function from the physical background. Either may be justified for particular purposes; the second one should be considered at least, as not every physical law renders a linear relationship.

By formal transformation any curve can be reduced into a normal distribution. For example, when the data follow an incomplete gamma function, one may transform them as is illustrated by the lower curve of Figure 1. The ordinate is the probability scale of a normal distribution, and the accumulated frequency has to be a straight line when it follows the gaussian law. The lower curve in Figure 1 shows the storms for Santa Barbara in February 1931–1950, [Thom, 1957a] transformed from an incomplete gamma function. The data follow a fairly straight line when we neglect about five and ten per cent at the upper and lower ends, respectively. This seems permissible.

The upper curve is constructed for observed daily precipitation sums at Asheville in September (1907–1956) in a linear scale. Suppose we are interested in computing the transformation function which reduces the curved line to a straight line. Taking the smallest possible class interval (0.01 inch) the first interval includes 12% of the total observations and for the range from 0.01 inch to 3.20 inches we would have to take more than 300 class intervals into account. By this method we would have to assume that 12% follow the same law which we derive for the other 88% of the observations (0.01 inch is the least measurable amount recorded). If we would take 0.1 inch as the first class, in order to save the laborious work and reduce the values to 30 class intervals, then we do not consider more than 40%, which is remarkably high. In practice it has been found that those small amounts do not fit well in the theoretical curve derived for the larger amount. Further discussion follows in connection with Figure 2.

As the frequency in those lower classes increases, the shorter the time period of measurement becomes, such as hourly totals, it is natural that the formal transformation is not strictly valid for the whole curve. This may be demonstrated by Figure 2. It represents the frequency distribution of daily precipitation December–February at Munich for a cubic scale. The theoretical normal distribution and the observed values agree fairly for sums $\geq 1.0$ mm (≈ 0.04 inch). However, the part less than 1.0 mm is drastically different. This is the part in the small precipitation amounts. In the accumulated frequency the part in the hatched area is compensated at the 40th percentile.
Accumulated sum of daily precipitation amount Asheville, September 1907-1956 (N = 458)

Transformed incomplete \( \Gamma \) function to normality (accumulated frequency) for Santa Barbara storms (February 1931-1950)

Fig. 1—Accumulated rainfall frequency in probability scale; abscissa scale for Santa Barbara storms at bottom, for daily precipitation amount for Asheville at top

Cubic Scale

Fig. 2—Comparison of observed frequency distribution for daily precipitation amount at Munich in December through February 1880-1950 with theoretical frequency in a cubic scale
In some instances, where the small amounts are unimportant or negligible, we may be satisfied with such an approach. For a physical interpretation usually we have to deal with both great and small amounts.

The transformation for physical reasons is based on the physical law involved in the process. Therefore, it is not based on the good fit for a part of the material but includes everything from a general point of view.

The logarithmic scale—Now return to the treatment of precipitation data of shorter time intervals. For these, the author suggests the use of a logarithmic scale which has been applied in hydrology for a long time. Wischmeier and Smith [1958] established a logarithmic law for the kinetic energy of rainfall. Schneider-Carius [1954] classified the rainfall process as autochthonous and also came to the conclusion that a logarithmic law may dominate the rainfall process. The writer [Essenwanger, 1956b] discussed some other facts pointing to a logarithmic law.

The theory for the log-normal distribution by Chow [1954] and others indicate that any logarithmic scale is suitable. Schneider-Carius [1955] and the author [Essenwanger, 1959] have published some proposed scales which proved to be convenient for meteorological data. The daily data for Munich, December through February, previously shown in a cubic scale (Fig. 2) are now arranged in a logarithmic scale (Fig. 3).

First we consider the total frequency only. The graph shows a mode at 2.51 mm. The frequency for the small amount is left open. This incomplete part on the dry side is justified. Grunow [1956] has studied this problem thoroughly at Observatory Hohenpeissenberg. To summarize his result, a large percentage of the dry days may be reclassified as days with rainfall by careful consideration of the precipitation amount, either not recorded or given under ‘traces.’ Therefore, rainfall records of shorter time periods such as daily amount and less are truncated on the dry side, which almost eliminates any type of formal mathematical transformation. The detailed analysis of the data in Figure 3 renders three partial collectives, each one a normal distribution in a logarithmic scale. Under collective the author understands the definition used in many statistical books, a random sample for homogeneous conditions of the same physical process. They are also submitted with Figure 3, while a detailed discussion of the process and problem is given below.

Frequency distributions and their analysis—In many cases meteorological events are defined

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Fig. 3—Frequency distribution of daily precipitation amount at Munich in December–February 1889–1950, logarithmic scale.
a priori and then a frequency distribution for those defined groups is established. This is an expeditious way to present climatological summations. The author feels that in cases where the various basic processes are not completely known, we should try the reverse way and analyze the total material to find out how many and which subgroups may appear. This approach is elucidated in the following discussion.

In the previous sections the author has suggested the use of the logarithmic scale. We start now the analysis with annual amount of precipitation. Figure 4 demonstrates for two different stations and climates (Asheville, N.C., and Madras, India), that for the annual amount, in general, the various rainfall processes become effaced. We obtain one collective, following one normal distribution in a logarithmic scale, except for locations in a pronounced desert climate.

In the consideration of monthly amounts, the mixture of the rainfall processes becomes obvious [Essenwanger, 1959]. This means that these frequency distributions cannot be considered any more as one collective and the problem is one of how to separate those frequencies into the individual parts and to investigate the physical basis for the partial groups. As a demonstration sample the daily precipitation amounts for Asheville may serve. First a frequency distribution (in logarithmic scale) has been arranged for each single month and then the frequencies have been split into partial collectives, each one a gaussian distribution. The author [Essenwanger, 1954, 1957] has developed an objective method for this resolution. The survey (Table 2) shows

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**Table 2—Asheville daily precipitation, 1907-1956 (50 yr); analysis of frequency distribution in logarithmic scale**

<table>
<thead>
<tr>
<th>Month</th>
<th>Excessive</th>
<th>I, Heavy</th>
<th>II, Moderate</th>
<th>III, Slight</th>
<th>Total cases</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$x_m$ N</td>
<td>$x_m$ N</td>
<td>$x_m$ N</td>
<td>$x_m$ N</td>
<td>$x_m$ N</td>
</tr>
<tr>
<td>Jan.</td>
<td>0.8 11 1.1 0.4 22 0.1 35 1.2 0.03 20 1.1 0.01 12 0.8 534</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Feb.</td>
<td>0.8 9 1.0 0.3 43 1.3 0.1 18 1.0 0.03 20 1.3 0.01 10 0.9 534</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mar.</td>
<td>0.8 17 0.8 0.3 41 1.2 0.06 29 1.5 0.01 13 1.1 602</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Apr.</td>
<td>0.6 25 1.2 0.2 38 1.3 0.05 32 1.5 0.01 15 1.2 526</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>May</td>
<td>0.5 33 1.4 0.2 20 0.8 0.06 27 1.2 0.01 18 1.1 600</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>June</td>
<td>0.5 33 1.6 0.2 26 1.5 0.05 27 1.5 0.01 12 1.2 650</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>July</td>
<td>0.3 62 1.8 0.1 21 0.9 0.05 22 1.2 0.01 16 1.1 628</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aug.</td>
<td>1.3 9 1.2 0.4 32 1.1 0.1 21 0.9 0.05 22 1.2 0.01 16 1.1 628</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sept.</td>
<td>1.3 9 1.0 0.3 54 1.7 0.03 26 1.1 0.01 13 1.1 742</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oct.</td>
<td>1.3 15 1.2 0.5 17 0.9 0.2 26 0.9 0.05 19 0.9 0.01 23 1.1 368</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nov.</td>
<td>0.6 15 1.3 0.3 30 1.4 0.06 33 1.4 0.01 22 1.1 382</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dec.</td>
<td>0.8 11 1.1 0.3 42 1.3 0.06 30 1.3 0.01 17 1.1 506</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$N$ in percent.

$x_m$ in inches.

$\sigma$ in units.
the individual groups, with parameters, where \( N \) is the occurrence in per cent of the total frequency, \( \bar{x}_m \) the mean amount in inches (for the log-normal distribution of the partial collective) and \( \sigma \) the measure of scatter (in class units, which can easily be converted into the logarithmic measure). One of the striking facts is the existence of a collective in fall with excessive rains (defined by its average daily amount of 1.3 inches). This should be worth a closer look. Therefore, the distribution of daily amounts in September has been selected here to illustrate the result of the analysis (Fig. 5). We recognize four groups of which the last group to the right is the subject of further interest. We find that the collectives overlap, but that there is also a part where mainly one collective prevails. We may now take those ranges for the limits in our study of the physical basis of the rainfall process for the excessive group. The investigation combines these groups in the months of August through October. Precipitation \( \geq 1.0 \) inch for all three months was determined as the limit where the excessive collective appears almost uninfluenced. From Table 3 it is seen that from the total of 115 cases with amount \( \geq 1.0 \) inches, 13 belong to the adjoining group (heavy daily rainfall). The 115 cases were examined using synoptic weather maps and the reason for the precipitation determined. One-third of the total sun and between 40 and 50% in August and September is caused by hurricanes which is not a surprising result. Fall is the main season for these phenomena and Asheville lies in the region of their influence. Most of the remaining parts were typical bad-weather conditions, where extratropical cyclones moved through North Carolina. Eleven cases had to be listed under miscellaneous such as other frontal or air-mass rains. From the theoretical analysis we expected that 13 values are part of the adjoining collective with another basis of rainfall. Therefore, the 11

![Figure 5](image)

**Fig. 5**—Resolution of frequency distribution of daily precipitation amount into partial components for Asheville, N. C., in September, logarithmic scale.

<table>
<thead>
<tr>
<th>Item</th>
<th>August</th>
<th>September</th>
<th>October</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \geq 1.0 ) inch</td>
<td>4</td>
<td>36</td>
<td>38</td>
<td>115</td>
</tr>
<tr>
<td>Thereof in Group I(^a) (heavy)</td>
<td>4</td>
<td>6</td>
<td>3</td>
<td>13</td>
</tr>
<tr>
<td>Caused by hurricanes</td>
<td>15</td>
<td>17</td>
<td>10</td>
<td>42</td>
</tr>
<tr>
<td>Cyclones</td>
<td>20</td>
<td>16</td>
<td>26</td>
<td>62</td>
</tr>
<tr>
<td>Others</td>
<td>6</td>
<td>3</td>
<td>2</td>
<td>11</td>
</tr>
</tbody>
</table>

\(^a\) The second line is included in the count of the first line and serves merely to compare with the last line. More details in text.
cases which do not belong to the hurricane rains or cyclonic rains are in good agreement with the theory.

The detailed investigation of the group with excessive daily rain shows clearly two reasons for the produced rain and demonstrates again the complexity involved in rainfall sums.

Refinement of analysis—The rainfall amount originally contains two parameters: intensity and duration. The recorded data are available in most instances for a predetermined constant duration such as hourly and daily sums. Very seldom a study is made which examines single rainfalls. Fortunately such a tabulation was available for Braunlage, Germany, in summertime for the period 1935-1951 with approximately 1000 values. A frequency analysis for both duration and intensity (in logarithmic scale) was made and the analysis exposed three partial collectives in both intensity and duration.

About 20% have an average duration of 3½ hr (log-normal distribution), the main part, including almost 50%, is just short of an hour, and the remaining 30% have an average of ½ hour. This may differ for other climatological regions, of course. The group division for the intensity proves to be slightly different as the parts are almost equal. One group is a little less with 0.009 mm/min, \(3.10^{-4}\) inch/min) while the other two show an average of 0.02 mm/min (10\(^{-2}\) inch/min) and 0.063 mm/min (2.5\(10^{-2}\) inch/min).

In a two-dimensional analysis, the amount should now yield nine groups. But the analysis rendered three groups only. Hence, we may conclude that some of the groups may have been combined. Figure 6 demonstrates this fact. This is exactly similar to the result discussed for the daily precipitation amounts in Asheville in the preceding section where we discovered two physical processes in one group. Hence, a refinement of the analysis would be to separate the data by intensity and then split them into duration groups or vice versa. This takes for granted the knowledge of the various kinds of intensity or duration groups. Also a preparation of the observations in the form of single continuous rainfalls would be necessary. This is very difficult, laborious and costly and there is little hope of accomplishing it.

The alternative to the above-mentioned recording of data is to split the material into intensity groups by observation which means...
classification as to advective or convective types. Dr. Bergeron, at the opening of this conference, has outlined some possible schemes.

Another way could be to keep the duration for the summation short so that we may be able to comprehend the intensity groups. Hourly sums may be close to the goal. Literally, it means to cut off parallel strips in Figure 6 which demonstrate the two dimensional analysis.

Conclusion—The author has tried to discuss some of the difficulties we have to face with commonly available sources of data for precipitation, when we are interested in the interpretation leading to the physical basis of the rainfall process. Besides the statistical techniques available for a linear scale, it could be demonstrated that the logarithmic scale may also be employed. How to study frequency distributions for investigations of the physical basis of the rainfall process has been demonstrated on the example of daily precipitation amounts in Asheville for the time of August through October. For the selected collective of excessive daily rainfall, derived by theoretical analysis, the connection to the generating process of rainfall can be explained. The result differentiates between rainfall from hurricane appearance on the southeast coast of the United States and extratropical cyclones through North Carolina as the possible reasons. This shows that collectives selected by using daily precipitation amounts may not be completely uniform in the physical origin. Therefore, the question is justified. Is the precipitation amount an adequate parameter for the interpretation of the rainfall process? A two-dimensional analysis of intensity and duration for single rainfall data in Braumage, Germany, points out the limitations we have to expect. If they can be accepted, the material in the present form is sufficient. For refinement of the analysis, however, it is necessary to do more research on rainfall intensity data which may be aided by radar analysis. We may study the classification groups of convective and advective type.

Using the duration as parameters in the presently convenient form of equal time intervals, we may find that probably the daily amount is the upper limit needed to study the physical processes involved. A shorter time interval would be preferable. Hourly precipitation values have been prepared by the U. S. Weather Bureau on a routine basis for several years at a number of stations and might be sufficient. However, much more study has to be done on this subject before final recommendations can be made.

References


Berger, G. L., Preliminary report to Technical Committee on Relation to Weather on Agriculture (NC-26), unpublished manuscript, Iowa State College, October 11, 1957.


Essenwanger, O., Linear and logarithmic scale for frequency distribution of precipitation, 1959. *Grafica, Pura e Applicata, in press.*


Thom, H. C. S., A note on the gamma distribution, *Statistical Laboratory, Iowa State College,* 1948, manuscript (manuscript of the U. S. Weather Bureau, 1958).


DISCUSSION

Discussion

Mr. C. J. Todd—I like Dr. Essenwanger's idea of looking for discontinuities in the probability distribution and using them as a lead in the search for a physical cause. Some years ago I was interested in Los Angeles hourly precipitation and probability functions, and I found out it rained nine per cent of the hours. I plotted the cumulative frequency of this nine-per-cent tail starting at 91% on linear probability paper, and got a very nice straight line.

Dr. O. Essenwanger—There are several ways to approach accumulated frequency amounts. As shown in hydrology hourly precipitation may follow a straight line in the log probability paper quite well, if they are from a homogeneous population. If we have different types of physical processes, depending on the regional climate, then it might be that this is practically a formal approach. The result is caused by the fact that one is missing the small precipitation amounts. The frequency distribution is practically truncated.

Dr. Tor Bergeron—I was very impressed to see what you could achieve just only by statistics of precipitation frequencies. From my experience with precipitation maps of all sorts, of all scales, and in many countries, I have always been looking for maps containing what I call a one-factor rain, and it has always been very difficult to find typical one-factor rains from daily precipitation measurements. We would like to have precipitation observations twice a day, as in the telegraphic reports. There is not a sufficiently large number of stations, but we would be so much better off. The 12-hour period often manages to take a one-factor rain. The 24-hour period does not. I have no figures, just general experience, and I would like to invite Dr. Essenwanger or somebody else to look into this matter. If the figures could be produced, then this could perhaps be recommended and carried through in the WMO. Although, as you know, it is generally easier for the observers to measure once a day in the morning upon arising.

Dr. Essenwanger—I can support Professor Bergeron's remark absolutely, because this is just what I wanted to say. Our daily amount is probably not sufficient for a subdivision into physical collectives. I rather would like to see very short periods like hourly amounts and then everybody who works in this field can analyze the period he likes. But this is probably not possible, at least on a world-wide basis.

Dr. Bergeron—Hourly amounts one can get in the United States, but you can not get them anywhere else. That is why I propose twelve hours.
Some Aspects of the Optics of the Rainbow and the Physics of Rain

FRIEDRICH E. VOLZ

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Abstract—It has hitherto usually been believed that the classic rainbow theories of Descartes and Airy for spherical drops completely satisfy the observations. However the well-known deviations of large falling drops from a spherical shape require changes in the rainbow angle, especially for scattering in the vertical cross section of the drop. Flattened raindrops of different radii cannot form a perceptible bow; this may explain why the interference bows are more visible at the top of the bow than towards the base, although their spacings do not change.

Preliminary investigations on the oscillations of large falling raindrops have been made. Such oscillations can result in blurring or washing out the rainbow. Rainbow intensities for measured drop size distributions have been calculated from the Airy theory. Since drops with radii larger than 0.2 to 0.5 mm are flattened and oscillating, they contribute little to the rainbow. Analysis of the rainbow can therefore only give information about the spectrum of the smaller drops.

Studies of the rainbow have found little place in modern investigations into the physics of rain; and yet this beautiful natural phenomenon is by no means as lacking in interesting problems as it may seem to be. The well-known classical Airy theory requires much modification before it can be applied to actual atmospheric conditions.

Few meteorologists may know that the supernumerary or interference bows (if they can be seen at all) gradually fade as we go from the vertex of the bow to the horizon, but without any change taking place in the displacement from the main rainbow. Also there have been only two recorded observations in the past of vibrations in the rainbow caused by thunder.

As this paper will try to demonstrate, the first phenomenon is probably connected with the deformation of any large raindrops, while the

Fig. 1—A rainbow photographed by Clarke [1920] showing how the interference bows vanish from the apex to the foot of the bow (Permission of Constable and Co., Ltd., London)
second raises the question of the oscillations of falling raindrops. Other developments in rainbow theory concern the effect of actual drop size distributions and the comparison of calculated and observed rainbow brightness. Further details may be seen in a review article [Volz, 1960].

The unusually fine photograph by Clarke [1920] (Fig. 1) illustrates well some of the important results of the Airy theory [see Pierter and Étner, 1922; van de Hulst, 1957]. The primary rainbow, and some secondary bows caused by interference, can be seen. The distance of a rainbow in monochromatic light from the antisolar point depends upon the refractive index of water and therefore slightly upon the wavelength; consequently, the rainbow shows colors in natural light. The overall width of the intensity distribution depends upon the radius of the raindrops (proportional to $\lambda/r^2$). Small drops give a broad rainbow with well separated supernumeraries. The close supernumeraries of larger drops are washed out by the divergence of the Sun’s rays. Only rain consisting of small and very uniform drops can give a rainbow such as that shown in Figure 1.

The supernumerary bow—The disappearance of supernumerary bows towards the foot of the rainbow is a long-established result [Pocci, 1863] clearly to be seen in some modern color photographs. It might be supposed that this phenomenon is caused by increasing drop size because of collisions and evaporation during fall. However, it seems that this cannot be the principal reason because the spacing between the mean bow and the supernumeraries is constant. Furthermore the calculations of Rigby and Marshall [1952] show no significant change of drop size distribution by collision with low rainfall intensities. Another possibility which must be investigated is the shape of the falling drops.

Drop shape and the rainbow—The Airy theory assumes that the raindrops are spheres, and yet we know from the experiments [Lenard,
1904; Laws, 1941; Blanchard, 1950; Magono, 1954] and the calculations [Liznar, 1914; Spilhaus, 1948; Imai, 1951 and McDonald, 1954] that falling drops flatten more and more as the size increases (Fig. 2). The path of horizontal incident light through a drop of 2.4 mm radius is shown in Figure 3. The rainbow angle has decreased to 25° from 42° for a sphere. It is also important that now some rays suffer total internal reflection while in a spherical drop only partial reflections are possible. Probably such a rainbow is much less intense than that of spherical drops.

Figure 4 shows calculated changes of rainbow angle for an elliptical cross section [Moebius, 1907] and measurements of this angle for a flattened drop as a function of the angle between the major axis of the drop and the incident light.

From these results it is clear that flattened drops with a low Sun will lead to rainbows whose radii at the apex decrease as the drop size increases. This distortion of the rainbow is important for drops greater than 0.25 to 0.5 mm radius (Fig. 2). Larger drops of different size can contribute nothing to the rainbow of small droplets. For sideways scattering, however, the drops have a circular cross section and hence the rainbow angle at the foot of the bow is the same regardless of the drop size. These statements agree well with the observations (Fig. 1).

Raindrop oscillations—Of great interest both to the physics of rain and to the theory of the rainbow is the problem of drop oscillations. Early photographic observations by Lenard (1887) show that drops falling down a tube oscillate between vertically elongated and flattened ellipsoidal shapes, in addition to the flattening mentioned earlier. The oscillations of the drops

![Fig. 5](image-url)
(which were about 6 mm in radius) were strongly damped, decaying to 1/10 amplitude in one second. Moreover, Blanchard [1950] recently found other types of oscillations in drops suspended in an upward-directed jet, which was presumably turbulent. He and Lenard roughly confirmed the theoretical law of Rayleigh [1879] (Fig. 5)

\[ v = \frac{1}{\pi} \sqrt{\frac{2\sigma}{\delta \rho^3}} \]

\((= 3.84 \pi^{-2/3} \text{ for distilled water})\)

where \(\sigma\) = surface tension, and \(\rho\) = density.

Here we may recall two earlier observations by Poey [1863] and Laine [1909]. These observations noted that the rainbow (particularly the more sensitive secondary bow) showed vibrations with each peal of thunder. It is quite likely that this indicates droplet oscillations excited by the intense sound waves.

Almost nothing is known however about the oscillations of freely falling raindrops. Earlier, Lenard [1887] and Schmidt [1913] carried out a few experiments at night but they recorded only a very small fraction of drops in oscillation.

Some preliminary results of a new investigation indicate that most of the larger raindrops are oscillating. If the falling raindrops are illuminated with a reflector bulb, and we look in the direction of the divergent beam, observations or photographs (Fig. 6) near the rainbow angle show bright traces of raindrops. Oscillating drops show traces broken by equidistant dark spaces. The distance between dark spaces \((s = v_t/v, \text{ where } v_t \text{ is the fall velocity})\) lies visually between 0.5 and 10 cm, and photographically between 2 and 10 cm. Using Rayleigh's formula this gives droplet radii between 0.4 and 1.4 mm or between 0.6 and 1.4 mm for photographic images. Analysis of a photograph taken in shower rain of medium intensity, gives a radius distribution for the oscillating drops (Fig. 7, top).

There are many physical problems about these oscillations which have yet to be solved. Are the oscillations maintained by eddies generated in
As the drop size increases, eddy frequency increases rapidly while the oscillation frequency of the drop falls. Only the questionable assumption that the proper vibrations of drops larger than 0.6 mm may excite turbulent eddies of the same frequency can lead to a source of eddying energy which may maintain the damped droplet oscillations. It is possible that small-scale atmospheric turbulence can excite the oscillations. In this case the amplitude and number of oscillating drops should be influenced by the factors which control the Austausch coefficient.

Depending upon the orientation of the oscillations, the rainbow can be affected in a way analogous to that caused by the flattening, discussed previously. With the Lenard type of oscillation (ellipsoidal with elongation and flattening in the vertical) and the Sun near the horizon, the wake of the drops, or are they excited by atmospheric turbulence which exists independently of the falling droplets? How large are the oscillations and what is their form?

The eddy frequencies behind rigid spheres falling in water were measured by Moeller [1938] for varying Reynolds' numbers ($N_R$). In Figure 5 the frequencies for raindrops are shown using the relation between Reynolds' number and drop size given by Gunn and Kinzer [1949]. In Moeller's experiments regular eddies first appear for $N_R = 400$ to 600; only in this region, when the droplet radius is near 0.6 mm the frequencies of eddies and drop oscillations are equal. This agrees with sideover slipping of drops of 0.5 mm radius observed by Gunn [1949] and with the observed lower limit of 0.4 to 0.6 mm for oscillating drops. The rainbow will not be affected in a sideways direction, but it will be broadened or missing at the top thus reinforcing the effect of droplet flattening.

Raindrop size distribution and the rainbow—Up to now relative intensities of color distribution in the rainbow have been calculated only for uniform drop size. However we know that raindrops range in size from 0.2 to 1.5 or even 3 mm. Figure 8a shows average size distributions for various rainfall intensities by Best [1950] and Figure 8b shows distributions for typical kinds of rain measured by Lamb [1958]. In the following we will take the raindrops to be spherical, thus neglecting flattening and oscillations. The results are therefore only valid for the rainbow near to the horizon.
OPTICS OF THE RAINBOW AND THE PHYSICS OF RAIN

Because the intensity of the rainbow increases with drop size (geometric scattering intensity \( \sim r^2 \), Airy intensity \( \sim r^2 \lambda^{-1} \)) the efficiency of drops for rainbow intensity is very broad and has a weak maximum near 1.0 to 0.5 mm (Fig. S a' and b').

The rainbow intensities in Figure 9 were calculated corresponding to various raindrop distributions. These rainbows have almost entirely lost the secondary Airy bows. The divergence of the sunlight would destroy the secondary bows even more.

Calculations of the absolute intensity of rainbow are shown in condensed form in Table 1. If we neglect the extinction of light by the rain itself the intensity of the rainbow should increase with the rain intensity. However the extinction of a screen of falling raindrops also increases with rainfall intensity and, as a result, rainbow brightness for normal conditions is nearly independent of the rain intensity. In green light the principal rainbow maximum is roughly three times as bright as the clear-sky brightness.

**Table 1—Extinction of rain and brightness of the rainbow**

<table>
<thead>
<tr>
<th>Rainfall rate (mm/h)</th>
<th>1</th>
<th>5</th>
<th>25</th>
</tr>
</thead>
<tbody>
<tr>
<td>Drop numbers (cm⁻² Km⁻¹)</td>
<td>34.5</td>
<td>50.0</td>
<td>205.0</td>
</tr>
<tr>
<td>Extinction coefficient (Km⁻¹)</td>
<td>0.19</td>
<td>0.45</td>
<td>1.33</td>
</tr>
<tr>
<td>Visibility range (Km)</td>
<td>21.0</td>
<td>8.7</td>
<td>3.0</td>
</tr>
<tr>
<td>Brightness of rainbow* (unit: 10⁻⁸ Sm disk brightness) without extinction</td>
<td>6.4</td>
<td>9.1</td>
<td>(38)</td>
</tr>
<tr>
<td>with extinction</td>
<td>5.1</td>
<td>5.6</td>
<td>(10.1)</td>
</tr>
</tbody>
</table>

* Rain falling from 1 km height, Sun altitude and angle of view to the horizontal 21°; no rain between Sun and observer.

**References**


MAGÓN, CH., On the shape of water drops falling in stagnant air, J. Met., 11, 77-78, 1954.

Discussion

Mr. C. E. Anderson—I am sure most of us had not been aware of the importance of the rainbow as a possible means of getting information on raindrop sizes and on oscillations of the drops.

Dr. Bernard Vonnegut—I think it is possible that raindrops and rainbows may be affected not only by shock waves but also by the sudden changes in electric field caused by lightning. The raindrop shape should be altered by an electric field, and it would be most helpful if we could use the rainbow as a tool for measuring the electric field. The usual measurements of the electric field in thunderstorms are complicated by the fact that most instruments profoundly modify the field being measured. If we could use the rain itself as an indicator of the electric field it would be an ideal measuring device.

Dr. F. Volz—But this vibration of the rainbow is not synchronized with the lightning, but rather with the thunder.

Dr. C. L. Hosler—My next-door neighbor is from Berlin and several times when there was fog he observed a variation in brightness that seemed to coincide with the 16-cycle alternating current of the electric overhead powerline. There was no visible discharge anywhere from the powerline to the ground. Can you explain this to me, so I might explain it to my neighbor?

Dr. Heinz Kasemir—This phenomenon is well known and can be observed at any railroad station where steam engines are operated and electric locomotives are fed from an overhead line. As soon as the steam from the steam locomotive enters the neighborhood of the powerline, which usually runs on a very low frequency (105 cycles per second) the steam cloud shows an oscillation in brightness in synchronization with the frequency of the powerline. This problem was studied by G. Escherich of the Technische Hochschule, Munich, and the results are published (see review of this paper by A. Schmauss in Meteorologische Zeitschrift, 57, 83–85, 1940). The steam droplets, contaminated by soot particles, form an electrical dipole and oscillate with the alternating electric field. The optical impression is that of a periodic variation in brightness.
A Possible Effect of Lightning Discharge on Precipitation Formation Process

BERNARD VOXNEGUT AND CHARLES B. MOORE

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Abstract—The electrical effects caused by lightning strokes within clouds may cause a rapid and effective drop coalescence process. It is suggested that the lightning stroke intensely electrifies cloud drops in its immediate vicinity. These drops acquire a charge opposite to that carried by the rest of the cloud and under the influence of electrical forces move rapidly away from the stroke, colliding and coalescing with the unaffected droplets.

Introduction—Atmospheric electrical measurements show that the great majority of precipitation-producing storms are accompanied by rather intense electrification even though they may not produce lightning. It follows that the precipitation-forming processes often occur in the presence of electrical activity.

It is well known that electrical effects are of great importance in determining the stability of emulsions and suspensions and there appears to be good reason to believe that atmospheric electrical activity may play a significant role in determining the stability of clouds. Various workers have approached the problem of the effect of electric fields on precipitation growth both experimentally and theoretically and in general they conclude that strong electric fields can accelerate coalescence phenomena.

With few exceptions these investigations have been confined to considerations of the effect of steady electric fields and have not dealt with the transient effects that might occur when a lightning stroke occurs. It is the purpose of this writing to offer some speculations on the nature of a lightning discharge in a cloud and its effects on the formation of precipitation.

Nature of lightning discharge within cloud—An understanding of the effects that a lightning stroke has on precipitation growth processes is obviously dependent on knowledge of the details of the discharge within the cloud. Unfortunately we presently know very little of these details, for this portion of the stroke is always screened from view. In this discussion it is therefore necessary to make some guesses concerning the nature of the stroke.

It is generally agreed that a large part of the electric charge is carried on cloud particles. Since a lightning stroke obviously serves as the mechanism for the release of the electrical energy accumulated in a cloud, it is frequently assumed that the lightning discharge neutralizes electrified precipitation and cloud particles.

Atkinson [187] has considered the effect of lightning on the coalescence process in the light of this idea, that the lightning removes the charge from the cloud droplets. He says, “The small drops which were before kept apart by mutual repulsion from being highly charged and of the same potential now coalesce and form the large drops, which being too heavy to be sustained in the atmosphere, fall.”

While it is undoubtedly true that some of the electrified particles in a cloud are neutralized by a lightning stroke, we question that this occurs generally. We prefer instead a somewhat different picture based to a large extent on the appearance of the lightning-like patterns that are results of the dielectric breakdown of electrified, methylmethacrylate plastic blocks. Gross [195] has described an interesting phenomenon which he produced by irradiating plastic with high energy electrons from an accelerator until the electric field within the plastic became sufficiently large to cause a spark quite similar to lightning. The pattern produced within the plastic by the electric discharge is shown in Figure 1. This phenomenon is in many ways analogous to a lightning stroke within the cloud, for here we have a charged region in a dielectric that produces sufficiently high electric fields to cause a spark discharge. Because the plastic is quite transparent, it is possible to see the fine details of how the discharge behaves within the highly charged region that gave rise to it.

An examination of Figure 1 shows that the
Fig. 1—Lightning-like pattern produced by dielectric breakdown in charged methylmethacrylate block

spark pattern is a treelike system that extends throughout the region of high charge density. There is no question but that the electric discharge has greatly reduced the high potential within the plastic and that from a gross point of view the charge within the block was neutralized almost instantaneously by the discharge.

However, if we look at the fine structure of the pattern with a microscope we see that it is not infinitely subdivided. The successive branches become smaller and smaller and finally terminate. We see that between the tendrils of the discharge there is transparent unaffected plastic, whose volume is considerably larger than the volume affected by the spark. Since a large fraction of the charged plastic is unaffected by the discharge it is reasonable to conclude that the spark did not neutralize a good part of the charge within the plastic but merely balanced this negative charge with an equal and opposite positive charge that was distributed in the discharge pattern.

Our picture of the nature of the lightning discharge within a cloud is based on the assumption that it is similar to what we believe takes place in the plastic. If this is true, the lightning does not neutralize the charged cloud particles but merely introduces a treelike pattern of approxi-
mately equal and opposite charge into the charged region. According to this view the lightning discharge would produce a flow of current that would greatly reduce the potential differences within the cloud but would not cause complete neutralization such as occurs when a spark jumps to a charged metal electrode.

This view of what may happen in the cloud has been arrived at largely on an extrapolation of what happens in the pastic, but it seems reasonable from another point of view. Before a lightning stroke can occur a large volume of charge, of the order of 10 or 20 coulombs must accumulate within the cloud. When the discharge begins, large voltages and high currents are available to supply the large energies required to form and maintain the intensely ionized path. As the stroke progresses and the flow of current continues, it is clear that the initial large reservoir of charge steadily decreases and along with it the potential gradients and electric currents. Finally a point is reached at which the energy required to form the ionized path is greater than the available energy. It seems likely that this happens long before the discharge proceeds to each electrified cloud particle.

Effect of discharge on precipitation particle growth.—The lightning discharge as envisioned above would suddenly introduce a large quantity of concentrated charge into the cloud that would cause neutralization on a gross scale. Following the stroke, neutralization on a much finer scale would occur by the migration of ions through the cloud. It appears that this final stage of neutralization following the stroke could result in a rapid and effective coalescence process.

The largest part of the length of the disruptive discharge pattern in the plastic is in the form of the rather fine terminal branches of the discharge. Similarly the greatest length of the lightning structure in the cloud is probably also in the form of the branch sparks, which are far smaller in diameter than the main stroke. When a discharge into a cloud takes place, we can imagine that these smaller branches penetrate into the charged region and suddenly deposit along their path a localized intense electric charge of approximately equal magnitude and opposite sign to that in the charged region. This intense charge introduced by the lightning creates an intense local electric field. Initially this charge is doubtless in the form of fast ions, and under the influence of the electric field these ions move rapidly away from the stroke.

We have no knowledge of the intensity of the local electric field in the region of the stroke just after it occurs but it is unquestionably high, perhaps approaching the dielectric breakdown strength of air. If this is so we may expect that the fast ions move outward with initial velocities of the order of 20 to as high as 100 m/sec.

The heat released in the immediate region of the stroke is probably sufficient to vaporize the cloud droplets so that initially the ions may move in clear air. After moving only a short distance away from the stroke the ions will encounter cloud droplets to which they will become attached and impart their charge. The intensity of charge acquired by the cloud droplets near the lightning stroke depends on the ion density and electric field, both of which are probably quite high. It appears likely that in the immediate vicinity of the stroke the droplets will acquire sufficient charge to raise the fields on their surface to values approaching dielectric breakdown. A rough estimate shows that if this occurs, these highly charged droplets will move outward in the local intense field with initial velocities that may be as high as 100 m/sec.

The process visualized above would be exceedingly effective in causing coalescence. The highly charged droplets would move rapidly away from the stroke out into the unaffected region of the cloud where they would collide with the other cloud particles. If we assume that the droplets in the region of the stroke are charged to values approaching dielectric breakdown this is equivalent to a space charge density of about two or three orders of magnitude greater than the average values one might expect in a thunder cloud. If this is true, the highly charged droplets produced by the lightning stroke will collide with hundreds or thousands of the oppositely and far less strongly charged cloud droplets in the surrounding cloud before their charge is neutralized. By the time the highly charged droplet has lost most of its charge by collision, its mass will probably be so large that it will have an appreciable rate of fall and gravitational forces will begin to play a part in its growth.

Discussion.—If lightning does indeed cause a rapid coalescence of the sort suggested here, one would expect to observe an increase in precipitation shortly after a lightning stroke.

It has been observed that gusts of heavy rain frequently follow a lightning stroke. [1953, p. 111] has stated, "The precipitation left the cloud base regularly at those places where
one half to two minutes earlier a cloud-to-ground stroke had come out of the cloud base."

In our observations on Mt. Withington, New Mexico we observed a similar sequence repeated six times on one day in which heavy rain with 3 mm drops appeared on the summit about two minutes after nearby cloud to ground strokes. On this occasion the cloud base was well below the summit. The time that elapses between the stroke and the appearance of the rain appears to be sufficiently large that the phenomenon could be explained on the basis of accelerated coalescence. Alternatively it is possible that the rain gushes that were observed may have been the cause of the lightning.

Measurements made at close range with a sensitive radar to determine whether or not an increase in reflectivity precedes or follows the stroke should determine which of these possible explanations is correct. If such observations show that the lightning occurs first they will indicate that the stroke is effective in promoting precipitation formation. Whether or not this happens in the way that is proposed or by some other mechanism will require sufficient detailed knowledge of the nature of the discharge and the electric fields and ion densities that it produces so that quantitative calculations can be carried out.

It is worth noting in this discussion of the possible effects of lightning in thunderstorms that a possibly related phenomenon has been observed in connection with volcanic electricity. In describing the electrical discharges that sometimes accompany volcanic eruptions, Guest [1927] states, "These dark clouds of finely divided particles of ejected matter are often brightened by vivid flashes of lightning discharging through the cloud to the rim of the crater. At such times a sudden agglomeration and precipitation of dust may follow . . . ."

Laboratory experiments are another approach that may shed some light on the possible effects of lightning in promoting particle growth. It should be possible to produce high voltage spark discharges in aerosols that simulate conditions in the thundercloud and to determine what effects occur.

Conclusions — It is concluded that the sudden redistribution of electric charge caused by lightning in a cloud may significantly accelerate the formation of precipitation. In order to evaluate this effect, radar measurements, laboratory experiments, and investigations of the lightning discharge are required.

Acknowledgment — This work was made possible by the support of the Office of Naval Research, Bureau of Aeronautics and the Atomic Energy Commission under Contract Nonr 1684(00). We wish to thank High Voltage Engineering Corp. of Burlington, Massachusetts, which organization irradiated the plastic blocks for us.

References


Discussion

(Note: Discussion of this paper is combined with that following the next paper.)
Estimates of Raindrop Collection Efficiencies in Electrified Clouds

C. B. Moore and B. Vonnegut


Abstract—Observations of thunderstorms from the summit of Mt. Withington, New Mexico, with a sensitive 3 cm radar indicate heavy rainfalls from electrified clouds a very short time after the initial detection of a radar echo. Estimates of the drop collection efficiency necessary to fit the observed time sequence give values of 200 to 500% depending on the assumption liquid water content of the cloud. These deduced values for raindrops in electrified clouds are 4 to 10 times greater than the mean collection efficiencies determined by Kinzer and Cobb for drops in the absence of an electric field. It is suggested that electrification in clouds may greatly enhance the accretion and coalescence processes and thus can be a causative factor in the formation of precipitation.

Introduction

During the summer of 1957 a sensitive 3.2-cm RHI radar was operated on the summit of Mt. Withington, New Mexico, in conjunction with a meteorological and atmospheric-electrical observatory. The primary purpose of the study was to relate the time and location of the first precipitation formed to the development of electrical activity in the stationary clouds growing over the mountain. It was found that in general the potential gradient within the cloud reversed and increased greatly in the negative direction several minutes prior to the appearance of the first radar echo. These results have been reported in detail elsewhere (Moore and others, 1958, Vonnegut and others, 1958, 1959). The subject of this paper is the short time interval between the appearance of a precipitation echo overhead and the collection of large raindrops at the mountain summit.

Experiment

We designed an experiment to extend the earlier work of Reynolds and Neill [1955] and Reynolds and Brook [1956] aimed at determining the initial precipitation of organized electrification and precipitation in thunderstorms. We did this by increasing the sensitivity of the radar detection of precipitation and by making the potential gradient measurements in the cloud as close as possible to the electrical activity.

Radar—The primary components of the APQ-13A radar we used were made available to us by Marx Brook and were those that he and Reynolds had used in their earlier work. We increased the sensitivity of the radar for the detection of initial precipitation by (1) bringing the equipment much nearer to the cloud by installing it on a mountain summit beneath the cloud being studied, (2) changing the antenna mount to give a zenith scan, and (3) carefully adjusting the radar components. New, low-noise 1N23E crystals were used so that a measured receiver sensitivity of −105 dbm was attained repeatedly. TE tubes with a recovery time of 4 microsec or less (to within 3db of full sensitivity) were used so that precipitation could be detected at ranges as small as 2 or 3 km.

During the summer of 1957 we used the standard 76-cm (30-inch) diameter parabolic antenna furnished with the set, modified only to permit a zenith scan. This scan took 3 sec and was presented as a range-height indication. In addition the plane of the scan was rotated slowly about a vertical axis, completing one-half revolution every 2 min. This arrangement permitted radar inspection of the hemisphere overhead once every 2 min.

In 1958, we replaced the original antenna with a larger parabolic reflector 152 cm (5 ft) in diameter. The beam width produced by the new antenna was 1.4°, as compared with about 3° for the smaller dish. The larger antenna was mounted to give either zenith or azimuthal scans as desired, thus permitting RHI or PPI presentation of any echoes.

Our estimates of the smallest raindrops that could be detected in these clouds during 1957 indicate that the initial precipitation echoes could be caused by raindrops with a median diameter of reflectivity no larger than 100 microns at 2 km ranges.
Fig. 1—Plot of computed $Z_{\text{min}}$, minimum detectable cloud-reflectivity factor for 3.2-cm radar, versus target range for APQ-13A sets used in 1957 and 1958.

Fig. 2—Estimates of minimum detectable median-volume drop diameter versus target range for APQ-13A radar sets used in 1957 and 1958 on Mt. Withington, New Mexico.

In 1958 the radar sensitivity was determined by use of 10-cm spherical metallic reflectors supported on captive balloons 6 or more km distant. From these measurements, estimates were obtained for the minimum detectable cloud reflectivity and for the minimum detectable median-volume cloud-drop diameter as a function of distance. Plots of these estimates are shown in Figures 1 and 2. The pertinent data on the APQ-13A radars used are given in Table 1.

The curves of Figure 1 were derived from calibrations of the radar with metallic reflectors carried on captive balloons. Assuming that the path to the minimum detectable target lies through cloud, we applied appropriate corrections for cloud-droplet attenuation. These curves are based on the relationships summarized in Mason [1957]. The lower limit on the usable range is imposed by the time required for recovery of the TR tube after the transmitter pulse. The curves of Figure 2 were estimated from the data in Figure 1, by use of relationships derived by Atlas [1954]. For this we assumed a cloud liquid-water content of 0.5 gm m$^{-3}$. We feel that these estimates of minimum drop size detectable are of the right magnitude, for with the 152-cm-diameter parabolic antenna, we repeatedly obtained properly shaped echoes from nonprecipitating Stratocumulus clouds.

Other instrumentation—We increased the sensitivity of the detection of the initial electrification by supplementing measurements made on the summit with potential-gradient measurements made within and above the cloud. Radiosondes modified to measure the vertical component of the potential gradient were suspended in the cloud from a tethered balloon. Most of the radiosonde field measurements were made with apparatus employing three radioactive probes. (These will be described in detail elsewhere.) To supplement the data obtained with this apparatus, several tethered balloon flights were also made with a newly designed passive electric field meter devised by Roy Hendrick of Cornell Aeronautical Laboratory. The operation of this field meter was difficult because of the conditions within the cloud. Nevertheless, we obtained several comparative measurements with this instrument and the potential-gradient measuring radiosondes employing radioactive probes. Since the measurements with the two instruments were in substantial agreement, we gained some confidence in the radioactive probes when used alone.

Measurements directly above the top of the cloud were made with a P-38 aircraft equipped for measuring the component of the potential gradient normal to the plane of the aircraft wings. (The participation of James Cook, owner and pilot of the airplane, was made possible through the co-operation of the United States
Weather Bureau and the Office of Naval Research.) This airplane was also used to obtain time-lapse motion pictures of the cloud tops and to make temperature and humidity soundings. To determine the nature of the precipitation, the airplane made several flights through the cloud according to radioced instructions from the radar observer.

**Observations**

During a typical summer day in the mountains of New Mexico, the early morning sky is usually cloudless. Under the influence of solar heating small Cumulus clouds begin to form over the mountains and after a few hours these clouds grow in size and become thunderstorms. Figure 3 shows how the cloud grows with time and the sequence of cloud electrification and radar echo formation on August 13, 1957. On this day an isothermal stable layer about 200 m thick at about 7 km limited the early convection; thus, several episodes of cloud activity were studied. The sequence seemed to be as follows: A new cell rose at about 4 m sec⁻¹. When the top rose beyond 6 km or so, negative charge concentrations appeared in the lower portion of the cloud. In two minutes or so, precipitation was detected by radar as a cup echo above the upper radiosonde, which still reported fair-weather polarity. The precipitation echo spread, and the vigorous updraft ceased. The electrification then vanished. In five minutes heavy rain and small hail fell to the surface. In 15 minutes or so the rainfall diminished, a new updraft appeared, and the entire sequence was repeated.

In situations where the wind speeds aloft were low, we found that quite commonly the first radar echo in a developing Cumulus is in the form of a hollow inverted cup that changed in appearance quite rapidly. Initially the cup echo is distinctly hollow. However, in a matter of several minutes it increases in reflectivity and fills in. As the echo grows and moves downward, virga makes its appearance beneath the cloud and on occasions when the cloud is directly overhead rain falls on the mountain summit.

Cook in his instrumented P-38 airplane played an important part in our study by making observations directly over the tops of the growing cloud. On several occasions we were lucky enough to observe simultaneously an initial, hollow echo and the echo of his airplane as he flew a few feet above the top of the visual cloud. When this happened, we determined that he was of the order of 500 to 700 m above the initial echo and established that in these cases the initial echo was well below the cloud top. Although we made several attempts we were unsuccessful in directing Cook into the initial hollow echo. Passes made through some of these clouds during later stages in their development showed that the precipitation was liquid water even when it was considerably above the melting level.

The phenomenon of the first precipitation echo in the form of an inverted cup was observed over quite a temperature range. In some cases the temperature at the top of the echo was as low as –17°C, while in other cases the tops of the echo were below the freezing level.

**Rapid appearance of rain from clouds in New Mexico**—One of the most striking features of the thunderstorms in New Mexico is the rapidity with which the clouds frequently develop and the great speed with which electrification, rain, and lightning make their appearance (see Fig. 4 and 5). As we have pointed out in earlier work, the available evidence suggests that the first rain is formed by a coalescence process. It is therefore of some interest to make estimates of the collection efficiencies of the drop growth process in these clouds.

**Table 1—Characteristics of the APQ-13A radars used on Mt. Washington**

<table>
<thead>
<tr>
<th>Item</th>
<th>1957</th>
<th>1958</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnetron frequency</td>
<td>9375</td>
<td>9363.8</td>
</tr>
<tr>
<td>Nominal wavelength</td>
<td>3.2</td>
<td>3.2</td>
</tr>
<tr>
<td>Peak power</td>
<td>40 kw</td>
<td>40 kw</td>
</tr>
<tr>
<td>Receiver sensitivity</td>
<td>–105</td>
<td>–105</td>
</tr>
<tr>
<td>Ratio: Measured system sensitivity/Computed theoretical sensitivity</td>
<td>1/30</td>
<td>1/3</td>
</tr>
<tr>
<td>Antenna diameter</td>
<td>76 cm</td>
<td>152 cm</td>
</tr>
<tr>
<td>Beam width to 1/2 power points</td>
<td>3°</td>
<td>1.4°</td>
</tr>
<tr>
<td>Pulse duration</td>
<td>1/2 μsec</td>
<td>0.65 μsec</td>
</tr>
<tr>
<td>Pulse repetition rate</td>
<td>nominal</td>
<td>850 sec⁻¹</td>
</tr>
<tr>
<td>TR tube recovery time</td>
<td>about 4 μsec</td>
<td>about 3 sec⁻¹</td>
</tr>
</tbody>
</table>
Fig. 3—Development of convection and electrification August 13, 1957
grown to the size shown. With the formation of rain, the vigorous updraft in the cloud ceased, and the electrification relaxed back to fair-weather values. The top of the cloud turned to ice and disintegrated while heavy rain fell to the ground. A new convective cell was formed overhead at 10h 11m MST, and negative charge concentrations again appeared within the base of the cloud. A new hollow echo (see Fig. 4) appeared in this cell at 10h 15m MST. With the development of the precipitation echo, the updraft in the cell ceased, and the electrification disappeared. Starting at 10h 20m MST, torrential rain and small hail fell to the mountain: no further electrical activity occurred until new convection appeared. A new cell 4 km to the south produced the first lightning at 10h 41m MST.

Regarding Figure 5, at 11h 16m MST the first echo appeared almost overhead, at a slant range of about 2 km and just below the freezing level. It had the cross section of an inverted hollow cup. The threshold median drop size for detection may have been less than 100 microns with the high radar sensitivity at this short range. The radiosonde echo was to one side of the first precipitation echo. The first rain arrived at the summit as 3-mm drops at 11h 20m MST. The rainfall became torrential by 11h 24m MST. From several considerations, the initial rain was probably formed by a coalescence mechanism. Close examination of the original film showed that until 11h 45m MST there was no 'bright band' in the precipitation echo at the melting level, where it first appeared on the outskirts of the echo.

The data of Table 2 and the Appendix were obtained when organized electrification in clouds was observed to precede closely the detection of a precipitation echo overhead [Moore and others, 1958] followed by a burst of rain falling to the summit.

**Computation of Apparent Raindrop Collection Efficiencies in Electrified Clouds**

With these data and the usual simplified raindrop growth model we can compute apparent collection efficiency, liquid water content products [suggested by Atlas, 1955] for these clouds assuming that each raindrop is independent of the others. According to this model, the growth of a falling raindrop by accretion is a function of its horizontal cross-sectional area, the distance it falls relative to the cloud, the cloud’s liquid water content, and the raindrop collection efficiency.

\[ dM = \rho_{\text{water}} dV = cLAdz = cLAdt \]  

where
### Table 2—Data on two electrified clouds

<table>
<thead>
<tr>
<th>Item</th>
<th>August 13, 1957</th>
<th>August 16, 1957</th>
<th>Initial echo</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Initial echo</td>
<td>Second echo</td>
<td>Initial echo</td>
</tr>
<tr>
<td></td>
<td>09h 55m MST</td>
<td>10h 12m MST</td>
<td>11h 14.5m MST</td>
</tr>
<tr>
<td></td>
<td>10h 00m MST</td>
<td>10h 15m MST</td>
<td>11h 16.5m MST</td>
</tr>
<tr>
<td></td>
<td>70μ</td>
<td>70μ</td>
<td>70μ</td>
</tr>
<tr>
<td></td>
<td>4.3 km</td>
<td>4.5 km</td>
<td>4.5 km</td>
</tr>
<tr>
<td></td>
<td>5.4 km</td>
<td>5.4 km</td>
<td>5.2 km</td>
</tr>
<tr>
<td></td>
<td>3.7 km</td>
<td>3.7 km</td>
<td>3.5 km</td>
</tr>
<tr>
<td></td>
<td>10h 03m MST</td>
<td>10h 17m MST</td>
<td>11h 20.3m MST</td>
</tr>
<tr>
<td></td>
<td>3.0mm</td>
<td>3.2mm</td>
<td></td>
</tr>
</tbody>
</table>

\[ M = \text{drop mass} \]
\[ V = \text{drop volume} \]
\[ c = \text{collection efficiency} \]
\[ L = \text{cloud liquid-water content} \]
\[ A = \text{drop cross-sectional area} \]
\[ v = \text{drop velocity} \]
\[ t = \text{time} \]
\[ \rho = \text{density} \]

\[ g = \text{gravitational acceleration} \]
\[ C_o = \text{drag coefficient of resistance for spheres which may vary between 0.30 and 0.21 in the size range between a 0.5-mm and a 3-mm diameter.} \]

For our calculation we take the minimum value for \( C_o \), which will give the smaller value of collection efficiency.

Substituting for \( v \) and clearing we get

\[ cL \frac{dt}{dt} = 2 \left[ \frac{3 \rho_o \rho_w}{g} C_o (0.9 h - 0.01) \right]^{1/2} \frac{db}{b^{1/2}} \quad (4) \]

and integrating

\[ cL = 4 \left[ \frac{3 \rho_o \rho_w}{g} C_o (0.9 h - 0.01) \right]^{1/2} \left( \frac{b^{1/2}_{\text{final}} - b^{1/2}_{\text{initial}}}{\Delta t} \right) \quad (5) \]

Now \( r \), the radius of the equivalent spherical drop, equals \( b^{1/2} \) by definition. From 

\[ \rho_{\text{water}} = 1.0 \text{ gm cm}^{-3} \]
\[ \rho_{\text{air}} = 0.79 \times 10^{-3} \text{ gm cm}^{-3} \text{ at a 4-km altitude} \]

we obtain

\[ cL = 2.5 \times 10^{-3} \frac{\left( b^{1/2}_{\text{final}} - b^{1/2}_{\text{initial}} \right)}{\Delta t} \quad (6) \]

at a 4-km altitude

From (3), the fall velocity at 4 km of 3-mm
(equivalent diameter) raindrops is computed to be about 11 m sec\(^{-1}\). On subtracting the minimum fall time through the clear air beneath the cloud from the observed total time interval, we estimate that the maximum fall time within the cloud is 140 sec and 65 sec for two echoes on August 13, 1957, and 180 sec on August 16, 1957 for the lowest part of the echo. Evaluating (6) for these days from the minimum radius detectable to the observed size on collection, for the maximum time of fall within the cloud, we get

\[ cL = 5.5 \text{ gm m}^{-2}, \text{ August 13, 1957, 10h 00m MST (3-mm drops estimated)} \]

\[ cL = 12.5 \text{ gm m}^{-2}, \text{ August 13, 1957, 10h 15m MST (second echo)} \]

\[ cL = 4.6 \text{ gm m}^{-2}, \text{ August 16, 1957, 11h 16m MST} \]

We let \( r_{\text{min}} = 50 \text{ microns for this computation.} \) The median cloud-drop diameter for reflectivity was estimated to be about 70 microns.

**Discussion of Calculations**

From the soundings for these days, the maximum liquid water content possible (assuming no dilution) from the cloud base (at about 3.5 km) to the lowest portion of the first echo (at 4.5 km) is about 2.0 gm m\(^{-2}\) before the first precipitation is formed. This calculation then suggests that drops in these clouds may have net collection efficiencies in excess of 200% after the appearance of a high potential gradient within the cloud. Use of the more realistic mean liquid water content in the layer from the cloud base to the initial echo leads to values in excess of 400% for the computed collection efficiencies.

It should be noted the simplifying assumptions have been made that no dry air is entrained into the cloud and that evaporation beneath the cloud has not decreased the raindrop radius. Our estimated values of the collection efficiency would be still higher had these factors been taken into account.

A solution of (6) for various assumed initial drop diameters is given below using the frequently observed 3-mm drop diameter as the final size and three minutes as the drop fall time within the cloud. The computed reflectivities \( Z \) which would arise from these initial sizes are shown in Table 3 for raindrop liquid water concentration of 1 gm m\(^{-2}\).

From this it can be seen that the computed high value of the \( cL \) product is not a critical function of the initial drop size chosen but the reflectivity and hence the detectability of the cloud is very sensitive to drop size. Therefore, even if the initial drop size used in the calculation were in error by a factor of 10 (so that the threshold of detection for the radar was 1/1000th as sensitive as we believe it is) the computed collection efficiency would only be decreased to unity. We believe that the sensitivity of the radar approached the design performance so that we place some credence in these indicated high values of raindrop collection efficiency.

Consider these results from another point of view. Under these conditions in New Mexico it has been observed that the rate of rainfall became "torrential" (50 mm hr\(^{-1}\) or greater) two minutes or so after the first drops arrived at the summit and only four to eight minutes after the echo was first detected. High rates of rainfall so soon after detection of the echo indicate that these first echoes did not rise from a low concentration of very large drops but from many drops in the cloud. It follows from the low initial reflectivity of the echo at detection (and the subsequent rapid increase in intensity) that these first drops must have been small in size but grew very rapidly.

These deduced collection efficiencies are quite remarkable if they are to be believed. The observations of which we are quite sure are the initial altitudes and the repeated short elapsed-time intervals between the detection of a radar echo in these electrified clouds and the collection of 3-mm raindrops on the mountain summit in a gush of rain. If our estimates of the threshold drop size for detection were in error by a factor of 10 (so that we could not detect raindrops less than 1 mm in diameter at a 2-km

---

**Table 3**—Computed collection efficiency—water content products required and cloud reflectivity as a function of the assumed initial drop diameter

<table>
<thead>
<tr>
<th>Assumed initial drop diameter</th>
<th>Required ( cL )</th>
<th>( Z ) at detection</th>
</tr>
</thead>
<tbody>
<tr>
<td>mm</td>
<td>gm m(^{-2})</td>
<td>m(^{2})m(^{-1})</td>
</tr>
<tr>
<td>0.05</td>
<td>4.5</td>
<td>1</td>
</tr>
<tr>
<td>0.1</td>
<td>4.4</td>
<td>2.5</td>
</tr>
<tr>
<td>0.3</td>
<td>3.8</td>
<td>750</td>
</tr>
<tr>
<td>0.5</td>
<td>3.2</td>
<td>350</td>
</tr>
<tr>
<td>1.0</td>
<td>2.2</td>
<td>2200</td>
</tr>
<tr>
<td>1.5</td>
<td>1.6</td>
<td>4000</td>
</tr>
<tr>
<td>2.0</td>
<td>1.1</td>
<td>20000</td>
</tr>
</tbody>
</table>

---
range overhead), the computed collection efficiency of unity required would still be greater by a factor of 2 than the collection efficiencies observed in the laboratory. Kinzer and Cobb [1956] made laboratory observations of the collection efficiency of water drops supported by an upward flow of a cloud of known properties, in the absence of any external applied electric field. The collection efficiencies they reported are shown in Figure 6. It can be seen that for drops 0.1 mm to 3.0 mm in diameter falling through a cloud of small droplets, the mean collection efficiency is less than 50%.

To a first approximation, a raindrop falling through a cloud of water droplets has the opportunity of coalescing only with those cloud droplets lying at least partly within the volume that it sweeps out. If, however, we consider the finite size of the small droplets it is apparent that the rain drop might coalesce with small drops just tangent to its path. Accordingly to be rigorous the effective radius for collision should be that of the raindrop plus that of the cloud droplet. For simplicity this consideration has not been included in (2); instead, a solution was made for the apparent collection efficiency in natural clouds as a coefficient for comparison with collection efficiencies determined in a laboratory cloud by Kinzer and Cobb [1956] who used the same simplifying assumption for

\[ C = \left( \frac{dV}{dt} \right) \left( \frac{4}{L_r} \right) \]

When the growing drop becomes much larger than the droplets that are captured, the correction for the size of the droplets becomes very small, much less than the high collection 'coefficients' indicated.

**Discussion of Coalescence**

An observer of the convective clouds in New Mexico soon notices that rain falls abruptly from these clouds as though a valve had just been opened or some trigger mechanism had been activated. Here the initial rainfall does not start slowly indicating slow droplet growth or an orderly process of rain formation. A cloud may float around producing no rain for an hour (or for as little as 15 min) then, closely associated with a burst of convection, there is an abrupt change in the cloud, with vigorous production of rain in a transient gush.

The calculations above were an effort to determine the collection coefficients necessary for initial raindrops in these clouds to fit the observed time sequence. The effective values found for the collection of actual raindrops are appreciably greater than collection efficiencies experimentally determined in a laboratory. There are several possible explanations for the surprising fact that our observations give collection efficiencies greater than unity. It may be that unwarranted assumptions have been made in the model of drop growth that we and others have used.

This model probably is a fairly good description of the growth process when a large drop falls through a cloud of small droplets. When large numbers of big drops have formed and the precipitation process is well underway, more rapid drop growth can also occur by interactions between large drops. These interactions are more complicated and probably not well described by the model.

It is possible that the high collection efficiencies we obtain may arise from large drop interactions that are not properly accounted for by the model. We discount this possibility, however, for our observations concern the first drops to fall from the cloud during the beginning of the precipitation process. At the time of our observations the concentration of raindrops was quite low (possibly 0.3 drops m⁻³), and the concentration of rain water was but 0.005 that of the cloud-liquid water content. It is doubtful that interactions between big drops are important during this period and it appears to us that the assumptions of the model are justified.

We believe that the high collection efficiencies indicated by our observations may arise from the
action of the electric fields we observed to appear shortly before the formation of rain [Moore and others, 1958].

From our observations, the appearance in a cloud of electrical activity above some threshold of perhaps 30 V cm⁻¹ seems to destroy the colloidal stability of the cloud. Precipitation echoes appear very shortly after the electrification within the cloud; these echoes increase rapidly in intensity and extent despite the subsequent disappearance of the electrical activity. A gush of rain falls from the cloud for several minutes then ceases slowly. We then observe only drizzle type rain until there is another episode of electrical activity whereupon the sequence repeats itself.

Effects of electric field on colliding water drops—Both laboratory and theoretical investigations indicate that electrical fields may have a very appreciable effect in speeding up the coalescence process in clouds. Laboratory experiments performed by Fuchs [1856], Plateau [1873], and Rayleigh [1879] have shown that even a rather weak field will cause colliding water drops to coalesce instead of bouncing off each other.

In his well known book on surface tension phenomena, Boys [1911] discusses the action of electricity to cause coalescence of two impinging jets of water. He wrote, "A piece of scaling wax rubbed on (a) coat is electrified. . . . The scaling wax acts electrically on the different water drops, causing them to attract one another, feebly, it is true but with sufficient power where they meet to make them break through the air film between them and join. To show that this is not imaginary, I have now in front of the lantern two fountains of clean water coming from separate bottles, and you can see that they bounce apart."

"To show that they really do bounce, I have colored the water in the two bottles differently. The sealing wax is now in my hand; I shall retire to the other side of the room, and the instant (the wax) appears the jets of water coalesce . . . These two bouncing jets provide one of the most delicate tests for the presence of electricity that exist. You are now able to understand the first experiment. The separate drops which bounced away from one another and scattered in all directions, are unable to bounce when the sealing wax is held up, because of its electrical action. They therefore unite, and the result is, that instead of a great number of little drops falling all over—great drops, such as you see in a thunderstorm, fall on top of one another. There can be no doubt that for this reason the drops of rain in a thunderstorm are so large. This experiment and its explanation are due to Lord Rayleigh."

Electrical effects can be expected to increase not only coalescence (by preventing bounce-off) but also the frequency of collisions between droplets. From theoretical considerations, Sartor [1957] calculated that a field of 30 V cm⁻¹ might produce ‘collision efficiencies’ of 400% for neutral raindrops falling through a cloud of 10-micron droplets, and that larger fields in thunderstorms could produce ‘collision efficiencies’ of 10,000% or more. The electrostatically induced increase in collision efficiency arises from the force of attraction between the dipoles induced on the water drops in an electric field. This force varies with the square of the external field, the radius of the drops, and in a complex manner with the distance of drop separation.

Sartor noted that this effect applies to all droplets, charged or uncharged, and that most of the droplets affected are those for which the computed efficiencies are very small. He concluded that “the electrostatic field, even when very small, plays an important role in the initiation and growth of precipitation, at least in warm or supercooled clouds and that the electrical manifestations of the thunderstorm are not merely its by-products but form an integral part of the precipitation mechanism.”

Formation of rain by coalescence—The rapid growth in the intensity of initial radar echoes in convective clouds by coalescence has been noticed by Battan [1953] and other observers. The phenomenon is worthy of further study so that coalescence mechanisms in clouds may be evaluated more properly; from computations with existing models, these mechanisms seem too slow to account for the observed production of precipitation.

For example, Houghton [1951] (who early suggested that coalescence and accretion should be important precipitation forming mechanisms) calculated the time required for the growth of a raindrop by coalescence (using collection efficiencies computed by Langmuir) and found that a 100-micron-diameter drop falling through a cloud of 20-micron drops (with a liquid water content of 1 gm m⁻³) would require 24 min to grow to a diameter of 500 microns, and another 7 min to grow to a diameter of 1 mm. (By Houghton’s calculations, an initial period of 92
minutes would be required for a 30-micron diameter drop to grow (by accretion and coalescence under the same conditions) from this size to a 100-micron diameter.) It can be seen that there is a considerable discrepancy between these computations and the period of 3 min or so elapsed from the time a radar echo overhead is detected by a sensitive radar and 3-mm drops are collected at the mountain summit.

A further discrepancy is that initial echoes were observed several times in vigorous clouds 20 min or less after the clouds first formed and rain fell from these clouds a few minutes later. From the appearance of these echoes it would seem that they formed by coalescence mechanisms, since a bright cup echo was observed, with no change in intensity at the melting level and some of these echoes first appeared below the level of the 0°C isotherm. Application of collection-efficiency computations based on aerodynamic and geometric assumptions would indicate that the drops need a much longer period from the time the cloud is formed to grow by coalescence to the size at which radar echoes might be detected (possibly more than 100 min) or to fall from the cloud as rain (possibly 30 min more).

Our observations of appreciable electric fields within the clouds during this period of rapid drop growth suggest that electrical enhancement of coalescence may actually occur. The disappearance of the electrification within the cloud as the rain developed and as vigorous convection was observed to cease makes it difficult to support an argument that the precipitation causes the electrification. The characteristic cup shape of many initial echoes [Vonnegut and others, 1958] may be a significant connection between the coalescence and the distribution of electric fields within a convective cloud. The cup shape may arise from a coalescence process and may depict the distribution of a region of high potential gradient within these clouds.

**CONCLUSION**

Our observations of the rapid appearance of rain from clouds in New Mexico support the ideas that electrical effects appreciably accelerate the coalescence process.

We have examined our data for days when the first rain formed overhead and find that in the more vigorous clouds of New Mexico, rain echoes sometimes appear (within electrified clouds) in as little as 19 min after the cloud is first formed. In these Cumuli the rate of increase of cloud reflectivity is quite remarkable; sometimes 3-mm-diameter raindrops fall in as little as 3 min after an initial echo appears. On computing effective collection efficiencies for drops in these cases we obtain values several times unity. On the other hand, computations with present coalescence models of the time required for un-electrified drops to grow from 0.1 mm to 1 mm diameter suggest that about 30 min would be required for such a size change and that two hours may be required from the time the cloud appears until rain is formed.

Since both the formation of an echo and the arrival of rain (sometimes apparently formed by coalescence) from the Cumuli in New Mexico occur much too rapidly to be described properly by present coalescence ideas, the possible effects of the observed cloud electrification must be considered. It appears desirable to carry out further studies that will provide a detailed picture of the precipitation growth process from which more defensible collection efficiencies can be computed.

Another experimental approach that may shed light on the coalescence process is to determine whether high precipitation growth rates such as we observe ever occur in clouds that have not become electrified. The examination of raindrop growth rates in warm clouds should be of extreme interest.

**Acknowledgments**—This work was made possible by the support of the Office of Naval Research, Bureau of Aeronautics, The Atomic Energy Commission and The National Science Foundation under Contract Nonr 1684(00).

**References**


Boys C. V., Soap bubbles, 1911 (reprinted, Dover Publ., New York, pp. 75-77).

Fechs, A., Verhandlungen des Vereins für Naturkunde zu Posenburg, first issue, 1856.


RAINDROP COLLECTION EFFICIENCIES IN ELECTRIFIED CLOUDS


REYNOLDS, S. E., AND M. BROOK, Correlation of the initial electric field and the radar echo in thunderstorms, *J. Met.*, 13, 376-380, 1956.


APPENDIX

Data describing the first raindrops collected on August 16, 1957 between 11 h 20 m and 11 h 22 m MST

Collection surface: Whatman #1 filter paper (31 cm dia) treated with methylene blue dye (see Fig. 7).

(a) Drop data: 30 drops collected in 120 sec

(b) Distribution of drop sizes:

<table>
<thead>
<tr>
<th>Size (mm)</th>
<th>Number of Drops</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.8</td>
<td>2</td>
</tr>
<tr>
<td>3.6 to 3.7</td>
<td>2</td>
</tr>
<tr>
<td>3.4 to 3.5</td>
<td>4</td>
</tr>
<tr>
<td>3.2 to 3.3</td>
<td>4</td>
</tr>
<tr>
<td>3.0 to 3.1</td>
<td>2</td>
</tr>
<tr>
<td>2.8 to 2.9</td>
<td>1</td>
</tr>
<tr>
<td>2.6 to 2.7</td>
<td>1</td>
</tr>
<tr>
<td>2.4 to 2.5</td>
<td>1</td>
</tr>
<tr>
<td>2.2</td>
<td>1</td>
</tr>
</tbody>
</table>

(c) Drop concentration: The rain collection area was 760 cm² and the approximate distance fallen in two minutes was about 1.2 km, so that the 30 drops came from an air volume of about 90 m³. The mean drop concentration was 0.3 drop m⁻³.

(d) Rainfall rate: The water contained in the 30 drops amounted to 0.5 cm³ over a 760 cm² area. This is a rainfall rate of 0.2 mm hr⁻¹.

(e) Liquid water concentration in the first rain: 5 × 10⁻³ gm m⁻³ (i.e., the mean liquid water content in the lower portion of the cloud is enriched by 1 part in 200 by this first rain).

Fig. 7—Photograph of filter paper stained by first raindrops, August 16, 1959
(Discussion of the two immediately preceding papers.)

Dr. B. J. Mason—The first thing that impressed me when Dr. Vonnegut showed his field records, was that his field variations fit very well the thunderstorm model of C. T. R. Wilson. If the thunderstorm is some distance away, then the electrical field at the point of observation is certainly controlled by the positive cloud dipole and one gets a positive field. Lightning discharges may cause excursions of negative sign for short periods. But if the thunderstorm approaches the point of observation, the influence of the negative charge in the base of the cloud becomes predominant, and the field changes to the negative sign. The small positive charge pocket in the base of the cloud, in passing directly over the observation point, might again cause a short positive excursion. I find nothing really very unusual about these field records since they are exactly what one would expect.

Now the other important point is whether or not one can deduce the charge of precipitation in the cloud from precipitation charge measurements at the ground. A large concentration of negative drops in the cloud will cause a strong negative field which in turn produces point discharge at the ground. The point discharge will send a stream of positive ions up to the cloud, which the negative raindrops, falling to the ground, have to pass. By this passage the raindrops may pick up enough positive ions to reverse their charge. This is the commonly accepted explanation for the 'mirror image effect,' which recognizes that the sign of the electric field and the sign of the precipitation current are inverted. So I would be very very reluctant to make any firm deductions about the precipitation charge in the cloud from measurements on the ground.

Dr. B. Vonnegut—I agree with Dr. Mason that because of point discharge it is difficult to ascertain from ground measurements what charge is carried by the rain within the cloud. However, I pointed out that occasionally we had

*At the conference Dr. Vonnegut presented a condensed and slightly different version of the two immediately preceding papers. It is because of that reason that some of the discussion remarks do not seem to have an immediate bearing to the papers as published. As the discussions are, however, of general interest and importance, they have not been omitted. Ed.
tive charges below the line. The rate of rain is shown by the dashed line.

Dr. Vonnegut—I think this is a very interesting observation. I am curious to know the magnitude of the electric fields involved. Were these high enough so that point discharge was occurring or not?

Dr. Magono—Observations were made on ordinary rainfall not thunderstorms, so that the field was about a thousand volt per meter.

Dr. Vonnegut—This value of the field is on the border line so it is difficult to know whether or not point discharge was taking place.

Dr. H. Kasemir (communicated)—I think I can give an explanation for the mirror effect and all the questions connected with it, which are brought up in the paper and discussion. Observations such as reported here where the mirror effect has been observed in the presence of low fields apparently require that we abandon the old theory, as outlined by Dr. Mason, that the original precipitation charge is identical with the negative charge in the base of the cloud. We postulate a charging mechanism whereby the precipitation falling through the lower part of the cloud assumes a positive charge leaving negative ions behind. These negative ions attach themselves to the cloud elements and thus they lose their ability to wander away. They are responsible for the negative charge in the base of the cloud but not the precipitation. As long as the precipitation does not reach the ground, the net charge of the cloud elements and of the precipitation are the same amount but of opposite sign. The electric field generated by the positive precipitation charge will almost be cancelled by the reverse electric field of the negative cloud charge. But as soon as the raindrops reach the ground, they get discharged and their charge is withdrawn from the picture. From there on the net charge of the precipitation remains at a constant value while the cloud charge continues to accumulate. A few minutes after the first raindrop hits the ground the cloud charge becomes so predominant that the field reverses its sign and becomes negative while the rain charge remains positive. The negative field increases rapidly until the discharging induction current balances the charging effect of the precipitation. If for some reason the precipitation charge reverses its sign and becomes negative then positive charge is left behind in the cloud and after a short transient period the field also changes to positive values and we observe the mirror effect, that is, that field and precipitation charge usually have the opposite sign. This picture explains all of the observed facts: (1) that we have the mirror effect with and without point-discharge; (2) that the mirror effect exists even in the cloudbase itself, as the measurement of Dr. Vonnegut and others (Kvettnin on Zugspitz, Israel and Kasemir on Jungfrau) have shown; (3) that we observe the mirror effect also, if the cloudbase is several kilometers above the ground.

There are cases, where precipitation charge and field have the same sign. Also the precipitation may lose some of the charge on its way to the ground passing through ion clouds of opposite sign. But we have to consider these as secondary effects while the basic mechanism works as outlined above.

Dr. Vonnegut (communicated)—While Dr. Kasemir's suggestion may explain the mirror ef-
fect under the conditions when the electric field is quite low. I question that the process he describes is usually the dominant one. He ascribes the field changes primarily to the discharge of electrified rain when it falls on the ground. It appears to me that a serious objection to this idea is the general fact that the current density resulting from the falling rain is usually only a fraction of that measured for point discharge. Accordingly, since field changes are the result of current flow, it would appear that the precipitation current plays a secondary role. Another objection to this postulated mechanism is our observation that the changes of the electric field associated with the falling of precipitation often begin a minute or more before any precipitation has reached the ground.

Dr. James P. Lodge (communicated)—Dr. Vonnegut has noted, as numerous previous investigators have done, that an energetic lightning discharge is frequently followed by a marked increase in rainfall intensity. He associates this as, respectively, cause and effect. Is it not more likely that the cause and effect are instead reversed, that in fact the lightning discharge simply used the descending rain sheet as a low resistance path to the ground so that the lightning flash merely appears to precede the rain because the final air gap is bridged while the rain is still a few hundred meters above the ground? It seems to me that the increased rain usually arrives only seconds after the lightning flash, which is much too short a time for increased rain to be generated as a result of the lightning stroke and to fall all the way from the cloud.

Dr. Vonnegut (communicated)—The gush of rain frequently observed after a lightning stroke is probably quite a complicated phenomenon that may arise from several different causes. The alternative suggestion proposed by Dr. Lodge seems quite reasonable and may indeed account for some observations of this kind. I believe there may be some question concerning the validity of Dr. Lodge's main premise that the lightning uses the descending rain sheet as a low resistance path. My own observations of thunderstorms lead me to doubt that lightning prefers the rain sheet. I have frequently been struck by the fact that the lightning often descends from the cloud base through clear air where no rain is falling even though a heavy rain sheet is nearby. The lightning appears to have little affinity for the rain and seldom either follows a path through or terminates in the rain sheet. Quite possibly Dr. Lodge may have in mind a situation that is somewhat different from the sort we have discussed, for the rain gush that we observed followed the lightning by a minute or two instead of only seconds after the lightning flash, as Lodge describes.
The Mechanism of Hail Formation

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Abstract—Critical considerations of the Ludlam model for producing large hailstones are presented. In connection with the results of the work on turbulence by Kolmogorov, von Karman and Heisenberg, it is questionable whether the procedure of how the water content of supercooled droplets is transported onto neighboring ice particles occurs according to the laws of molecular diffusion. The significance of the Bergeron-Findeisen mechanism for producing precipitation is mentioned and referred to the newest experimental finding on the structure of hailstones.

In a recent article Ludlam [1958] goes very thoroughly into the problem of hail formation and comes to the conclusion that the formation of large hailstones is due exclusively to droplets present near the base of Cumulus clouds and having a radius of 20–30 μ, particularly when these are very sparsely scattered with an incidence of less than 10⁻⁷/cm². These particles, which lie at the large end of the drop spectrum, are therefore to be regarded as the initial particle or embryo of the large hailstones. Ludlam bases his observation on a model of the way showers are produced, which in all essentials corresponds to the all-water precipitation process of the kind most recently employed to explain the origin of warm rain where the solid phase of the water does not appear at all. Since the fundamental features of all such models are derived from theoretical considerations, idealizing isolated aspects of a natural process and therefore representing only an approximation to reality, there is always the danger that conclusions may be drawn on which too much reliance is placed. And precisely the inference already referred to, namely, that large hailstones owe their origin entirely and only to the sparse presence of relatively large droplets around the cloud base, should properly be treated with some scepticism.

By introducing thermals into the pattern of air currents, Ludlam has overcome the difficulties which arise in explaining the production of showers by means of the air-particle theory and the related concept of air columns existing inside Cumulus clouds in the form of a steady updraft. At the same time it also becomes possible to understand the discrepancy between the values for temperature and liquid-water content measured inside the clouds, and the values calculated in accordance with the parcel theory as an adiabatic process. (The origin of these vortex motions Ludlam sees principally in locally conditioned increases in temperature or in instability accompanying an atmospheric front.) Cloud particles of initial radius 20–30 μ, may, if they are favorably placed at the start, grow to such an extent while they are within the thermal (reaching a radius perhaps of 150 μ) and gather such a speed of fall (1–2 m/sec), that they escape the general erosion of the thermal which sets in when it reaches the top of the cloud; and they are then caught up in the course of falling by the following thermal, and continue to grow into precipitation particles or drops of water. In this way the particles of precipitation have time to pick up the necessary amount of water despite the usual deficiency (in fact) in the fluid-water content by contrast with the adiabatic values, yet without having to climb to any considerable height and thereby run the risk of reaching the −40°C level where spontaneous glaciation prevents any further growth.

In the mechanism of shower formation which has just been described, the initial particles are already of such a size that growth can in practice only take place through coalescence with other particles, while the condensation of water vapor on the particle is only of secondary importance. (We shall return later to the diffusion of water vapor in the Bergeron-Findeisen process.) For this mechanism of shower formation it is unimportant whether and where the particle, during its growth into a particle of precipitation, undergoes freezing, which may be initiated by the presence of some foreign ice particles or ice-forming nuclei. Nor is it of any consequence in this connection whether the
cloud particles which have coalesced and been caught up are supercooled or not. If there is no glaciation, then we are dealing purely with a shower forming as warm rain (by the all-water process). If on the other hand there is glaciation, then the particles of precipitation, which have now solidified, can sometimes grow into huge hailstones through continued aggregation with supercooled droplets.

The fact that it makes no difference in this mechanism whether the coalescing droplets are supercooled or not, has led understandably to less importance being attached than formerly to the Bergeron-Findeisen process of rain formation. There may, however, be grounds for asking, whether the disparagement of the Bergeron-Findeisen process has not perhaps gone too far. At all events we should at least check whether, if glaciation were to take place in the growing particle later, after an initially pure all-water process, this would not bring about an increase in the amount of precipitation produced.

Firstly, here are a few basic considerations. The introduction of thermals into the general movement of upward currents of air within a Cumulus cloud is merely a first approximation to suggest the turbulent character of atmospheric movements. The fundamental works of Kolmogorov [1941] and the theoretical reflections of von Karman [1948] and Heisenberg [1948], which follow on Taylor’s [1922] enquiries into statistical-isotropic turbulence, have led to an appreciable clarification of the inner structure of turbulence. What is above all relevant to our present discussion is that a successful demonstration has been made, on a statistical basis, to show that a law of energy distribution exists, which in the case of meteorology is valid for an extremely wide spectral range of vortex elements. It is in accordance with this law that the energy contained in the largest vortex elements in the atmosphere, which may be several hundred meters across, is successively transferred to others of smaller dimensions until it finally disappears in vortices of almost molecular size. Two questions are to be raised in this connection.

(1) When Ludlam uses a mathematical description of the amount of aggregation to calculate the growth of hailstones, relying entirely on Langmuir’s [1948; Langmuir and Blodgett, 1948] concept of collection efficiency, which in practice is equated to unity, is sufficient allowance made for the turbulent character of the air stream, particularly in view of the fact that during ice formation additional heat is liberated?

(2) In the mathematical account of how the water content of supercooled droplets is transported onto neighboring ice particles, is it enough to take into consideration only the molecular diffusion process? In discussing the problem already mentioned of the effect which the Bergeron-Findeisen mechanism has on the production of precipitation, we shall have to pay particular attention to this second point. With regard, finally, to the admissibility of a statistical approach to turbulence phenomena, it should be pointed out that thereby a variation is tacitly included in the product of the growth processes, and this may reveal itself in a considerable multiplicity of size and shape in the precipitation particles.

It is generally acknowledged that, if the major part of the supercooled droplets in a cloud were to freeze, this would impair the further growth of a solid particle of precipitation through coalescence and stop the development of hailstones. Ludlam has calculated that for this the density of ice-forming nuclei would have to be 10/cm², and Weickmann [1953] arrives at a similar value by an analogous argument. It is a high figure for the density of icing nuclei and it could never be realized artificially, for instance, through seeding the atmosphere, by any economically feasible methods.

Ludlam’s calculation of the nucleus density necessary to bring about a sufficient glaciation of the supercooled parts of a cloud to prevent coalescence, is based, as I have suggested previously, on the same theoretical ideas that underlie the Bergeron-Findeisen mechanism. And here the transference of the water from the supercooled droplets onto the precipitation particles, which have themselves solidified into ice under the influence of ice-forming nuclei, occurs according to the laws of molecular diffusion. Since the freezing process releases considerable quantities of heat, however, and thereby initiates pronounced turbulent intermingling, the following experimental enquiry is suggested: May not a closer approximation to reality be reached if, in the expression for the movement of water from the fluid droplet to the ice particle, the constant D of molecular diffusion (D = 0.45 cm²/sec in Ludlam’s calculation) is replaced by a constant more in keeping with turbulent diffusion? The conception of this constant (which cannot, for
obvious reasons, be the same) may be justified on the basis of Prandtl's considerations aimed at a better quantitative description of the exchange of ponderable matter brought about by turbulence. It may be remarked in passing that Sutton [1947] has made successful use of this concept in deriving his formulas for the spread of atmospheric dust. By adopting such a procedure, a density would be obtained for ice-forming nuclei, which would turn out to be markedly smaller in order of magnitude than that given by Ludlam. This density could clearly often be realized in nature at temperatures less than \(-10^\circ\text{C}\), and there would also be no great difficulty in producing it artificially.

These suggestions make it evidently necessary to say, that the possibilities of precipitation being produced according to the Bergeron-Findeisen process should not be rated so low as has sometimes been done more recently. This view is supported ultimately also by the Final Report of the U.S. Advisory Committee on Weather Control which states that it has been proved statistically with a satisfactory degree of certainty, that in orographic conditions an increase in rainfall amounting to between 10 and 20\% may be expected from artificial seeding with silver iodide by means of ground generators. Also the results of the first two experimental years of the Swiss Hail Suppression Project, in which a randomized ordinance has been consistently adhered to, likewise give indications for believing that a surprisingly large increase in precipitation can be effected by seeding; a report on this is, however, being presented in a separate paper. Dessens [1958] points out in a short article, in which he takes up the ideas propounded by Ludlam, that the effects known to be produced in clouds by seeding with ice-forming nuclei could never be observed in clouds seeded with condensation nuclei. And in a work published recently Schaefer and Dietrich [1959] have also confirmed in an impressive manner the influence on supercooled clouds by seeding with silver iodide particles. Finally, the excellent experiments designed by Schaefer must be recalled, in which, within a few seconds after dropping dry-ice particles through supercooled clouds, centimeter-broad dark bands appear along the sides of the threads of small ice crystals, after approximately another 30 sec the whole picture becomes blurred as turbulence sets in.

These comments on the significance of the Bergeron-Findeisen mechanism have been put forward less to provide a more solid foundation for hail-prevention experiments by silver iodide seeding than to indicate from the point of view also of these experiments what kind of problems are involved in constructing any theoretical model to represent some part only of a weather process. It is important to stress here with particular clarity that only experiment can provide the ultimate answer. This is no less true when it is a question of clarifying the processes which lead to the formation of hail, than in the practical field of hail prevention.

Recently List [1958] has published two articles on investigations into the structure of different types of graupels and of large hailstones collected at different places in Switzerland in the years 1953 to 1957. The structural analyses have been carried out by means of techniques for producing very fine cross sections, which have been developed to a degree of great proficiency at the Swiss Federal Institute for Snow and Avalanche Research, Weisshüflikoch-Davos. This Institute houses also one of the research centers of the Federal Commission for the Study of Hail Formation and Prevention and it is here that for the last few months the Swiss Research Hail Tunnel has been operating, about which a separate report is presented as the next paper in this volume by the supervisor of the tunnel experiments.

On the basis of pictures which have been made of the structure of actual hailstones, List arrives at the view that the hailstones which were examined and had dimensions of several centimeters, all possessed in their original form as growth centers (to use the author's own term) some type of rime or frost graupel, which may well in certain cases have come about through snow crystals entering a graupel phase. A particularly pertinent example is offered by the hailstone 57.7 (see Fig. 1 and 2), of which structural photographs are reproduced. There can be no doubt that this argument is entirely correct in this case, and above all his claim that the initial graupel cannot have developed from a drop of water which subsequently froze. Thus we are able to make a statement, which is certainly confirmed by experiment, but contradicts Ludlam's suppositions regarding the growth of hailstones, as we described them at the beginning of this paper. Figure 3 is a picture of a fine cross section of a graupel which has reached such
an advanced stage of development that the original particle, namely, a snow crystal, can now only just be recognized.

In conclusion, a remarkable finding should be mentioned, which has already resulted from investigations made with the Hail Research Wind Tunnel. It appears that large ice particles, which contain imprisoned within themselves considerable quantities of still liquid water, can be produced relatively easily by coalescence with sub-cooled droplets. This immediately raises the question, whether the bigger natural hailstones could not also carry enclosures of water. This possibility would allow us to understand how growth may take place in a considerably shorter time than is generally assumed nowadays. In order to find out, investigations would have to be carried out immediately after the hailstones had fallen, and might therefore prove especially difficult. For there can be no question of pre-
serving the hailstones by any form of refrigeration, since this would lead to a change in structure.

References


Discussion

(Note: Discussion of this paper is combined with those of the two following papers at the end of the second following paper.)
Design and Operation of the Swiss Hail Tunnel

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Abstract—An acceptable theory on hail formation should be based on the general physical conditions prevailing within thunderstorm clouds and in particular on the processes of nucleation and growth of ice particles. An attempt has been made to approach this problem from the experimental side by means of a vertical wind tunnel offering a wide range of air conditions. In this paper the Swiss Hail Tunnel as it has been built in the Laboratories of the Federal Snow and Avalanche Research Institute, Weissfluhjoch, Davos is described and discussed. It is shown that in the tunnel stationary and variable conditions can be reproduced similar to those to be expected in a natural hail-producing atmosphere.

Introduction—The hail tunnel was designed and built in order to grow hailstones experimentally in the laboratory. For it was felt that comparative observations of these with natural hailstones would enable conclusions to be drawn as to the conditions in which hail forms in nature. Experience gained over a number of years at the Swiss Federal Snow and Avalanche Research Institute in investigating thin sections, when applied to soft hail (grapel) and hailstones [List and de Quervain 1953, List 1958ab], showed that here too the key to any explanation of how these natural ice particles arise must lie in an interpretation of their physical structure. It may be expected that specific structural zones in iced-up particles of precipitation are a direct consequence of specific growth conditions in a cloud; that the structure of a particle is essentially the product of the conditions in which it originates.

The experiment has thus to clarify the connection between conditions of growth and the resultant ice build-up on a nucleus particle. The hail tunnel is the plant which permits this experimental aim to be carried out. It consists basically of a wind tunnel which is vertical in the section where measurements are made, has a closed circuit and an adjustable climate.

When formulating the problem it must be clear that, as a result of the many parameters of the experiment, it is not possible in every case to classify a certain ice formation under a particular set of growth conditions. This is especially true of different growth phases built up in time at the same place. The final conclusions are also complicated by the multiplicity of forms which occur in natural hailstones [List, 1959a].

Before designing a hail tunnel, it is important to elucidate the extent to which the conditions concerning temperature, humidity, air speed, air pressure, impurities in the air, electrical effects, etc., such as may be found in a hailcloud, can in fact be imitated in the experiment. Enquiry into these factors gives a positive answer on all counts. It would in itself be desirable to reproduce variations in pressure, but here the financial expense is disproportionately high, and so experiments have to be designed at constant pressure and the results transformed by suitable rules of similarity [List, 1959b]. In addition it was decided when planning our wind tunnel not to try to check and influence the electrical conditions, on the preliminary assumption that they are only of secondary importance. The possibility still remains, however, of elaborating the plant in this respect or of otherwise improving its capacity.

The construction of the hail tunnel—The basic construction of the hail tunnel can best be seen in Figure 1. The blower a generates the necessary air-speed relative to the object b suspended or floating in the measuring section. (The direction of the air stream is counterclockwise). In the position c next to the blower is the air cooler or vaporizer. Ammonia vaporizes in its ribbed tube system, is compressed in the refrigerating compressor d (or Fig. 2) and is then cooled and condensed in the condenser e by means of air from the atmosphere (since the plant stands at 2665 m above sea level there is no supply of cooling water). A heater f with a capacity of 0 to 19 kw, which can be engaged immediately at any level, makes possible rapid periodic changes of temperature as well as the eventual warming of the ex-
Fig. 1—General diagram of the hail tunnel; (a) blower; (b) test object; (c) air-cooler or vaporizer; (d) refrigerating compressor; (e) ammonia condenser; (f) heater; (g) electrostatic filter; (h) point where humidity and ice-forming nuclei are injected; (i) section where the flow achieves homogeneity; (k) measuring section; (l) control panel; (m) liquid separator; (n) alcohol-water cooler for cooling the cold compressor

The electrostatic filter g, which has particular advantages with regard to ice-forming nuclei, serves to purify the air. (The special characteristics of the electrostatic filter have been discussed, for instance, by List and de Quervain [1956].) At the point h, humidity is continuously added to the air which since leaving the blower has been dried, cooled, and cleaned. It is a matter of choice whether the humidity is introduced via a steam generator (Fig. 3), an air-water compressor (Fig. 2), or a rotor atomizer (Fig. 4). By varying the method of producing the droplets their size can be altered to quite extreme limits. At the same point in the tunnel h, nucleating substances are added (Fig. 4), after having been vaporized in a high-temperature oven. The next vertical section of the tunnel i, which follows the humidity injection point, allows the air current to become less turbulent and the temperature of the droplets to adjust to the temperature of the air about them.

The air which has been climatically conditioned in this way now flows into the actual measuring section k, where its ice-forming capacity is examined through the growth of a test object. Various measuring probes are also installed in the measuring section, and the test object can be illuminated and observed at any time through plexiglass windows. In this connection it should be noted that the measuring section has in prin-
The remaining parts of the tunnel serve only to allow the air to expand slowly and to bring the circulation back to the blower.

The whole plant is supervised and run from the control panel (marked as 1 in Figure 1). Here the object under investigation can be directly observed at the same time that all the values which can be regulated are set on the panel. All the important values, whether of temperature or some other electrical scale, are simultaneously recorded on three compensation recorders. The apparatus for regulating temperature is housed independently; the gages for measuring pressure and humidity are likewise located close by the measuring section (Fig. 5).

From this description it will be clear how, in the part of the tunnel below the measuring section, a cloud is continuously produced more or less subcooled according to need; it does not circulate more than once, however, as drops of water are deposited at the latest in the air cooler and any particles of ice on the electro-filter.

The whole tunnel is constructed of separate components, which can easily be taken apart and changed. The main substance from which it is built is a cellular, slightly spongy plastic called Polystiroi; it combines lightness of weight ($\gamma = 20 \, \text{kg/m}^3$) with insulating properties similar to those of cork, and has by comparison the advantage of an even lower thermal capacity relative to the unit of volume. These characteristics all guarantee minimal losses of heat with minimal ther-
mal inertia. External surfaces are additionally protected by a coating of araldite, which preserves a certain elasticity even at very low temperatures.

The Performance of the Hail Tunnel

During stable operation—The operation of the tunnel is very complex, in particular with regard to the regulation of the temperature. The various factors which have to be observed for the general case may therefore be listed as follows:

1. The basic temperature of an individual experiment is given by the lowest temperature in the tunnel, that of the air-cooler: $t_F$.

2. On the way to the measuring section the temperature of the air is raised as a result of cold losses by conduction to the outside and also of indirect heat generated by the blower: values which depend on the temperature of the tunnel itself and on the air speed.

3. The tunnel temperature is further influenced by the excess energy injected through the medium of the humidity.

(4) Ice-forming nuclei may also influence the temperature according to their activity in transforming energy, as freezing processes take place in the subcooled cloud, before it reaches the measuring section.

The sum of these influences determines the actual temperature in the measuring section, which is the effective temperature of the experiment, $t_M$. To regulate it by means of the compressor acting on the vaporizer temperature is extremely difficult and quite impossible where there is any sudden alteration in humidity. The procedure adopted was therefore to produce in each experiment always more cold in the vaporizer than was actually needed. The surplus cold is then destroyed by the heating element, whose output can be altered almost instantaneously; so that a maximum temperature adjustment can be achieved by hand control, or alternatively be regulated automatically.

The efficiency of the tunnel with regard to temperature and humidity input is a function of the vaporizer temperature as well as of the air-speed, which helps to determine the amount of the losses from the system. The effective performance available for a particular experiment can be ascertained directly, by observing the following procedure: the refrigerating compressor is allowed to run at full capacity while the air-speed and heat-setting are constant. With time the stable vaporizer-temperature $t_V$ develops, and once this is arrived at it provides an operational point which describes the connection between vaporizer temperature, effective refrigerating capacity (which equals heating capacity), and air-speed. The group of curves in Figure 6 were given by the sum of such measurements. They show

Fig. 5—The measuring section of the tunnel, the gages for pressure and the control panel

Fig. 6—The effective cold output $L_K$ available for the experiment, at different air speeds $v_M$, as a function of the vaporizer temperature $t_V$.
The value \( Q'_{VM} \) can be measured experimentally, so long as the value \( Q''_{VM} \) is not involved. Corresponding values are entered in Figure 7 as mean values for the air speeds in question. This presentation shows the connections in accordance with Eq. (1); it is used especially when the operational point for the compressor, that is to say, an appropriate vaporizer temperature has to be determined on the basis of a desired measuring-section temperature \( t_M \), a given energy value \( Q'_{VM} \) present in the tunnel as a result, for example, of humidity, and a certain air speed \( v_M \). In practice, one starts with the energy value \( Q'_{VM} \) and finds the point of intersection with the line representing the chosen speed (Fig. 7). What is then given is principally the degree to which the measuring-section temperature diverges from the vaporizer temperature. Additional choice of a \( t_M \) value allows the operational point in the \((t_M, t_v)\) diagram to be fixed and the vaporizer temperature \( t_v \) to be read off which is appropriate to the particular experiment.

Before an experiment calculated in this way can be practically carried out, it is necessary to establish whether the assumed amount of added energy can in fact be produced for the resultant vaporizer temperature. This information is given in Figure 6.

Figure 8 shows the maximum performance of the tunnel as a function of the air speed with the parameter \( t_M \), while Figure 9 gives similar correlations for the maximum water injection. (It was assumed here that the added water has a temperature of \(+10^\circ C\).)

The effect of the heating apparatus provides other possibilities beyond improving our control over the tunnel. It enables the water content of the air in the measuring section to be substanc-
tially reduced for evaporation measurements, with the vaporizer elements then serving as an air dryer. The moisture content in the measuring section corresponds in this case to the moisture content of the atmosphere at saturation point relative to the vaporizer temperature as compared with ice. Suitable diagrams could easily be worked out from the data given earlier.

The behavior of the tunnel during unstable operation—In nature relatively rapid changes in the conditions of growth play an important part in the way hail forms, as is seen from hailstones having a layered, or 'onion-coat,' structure. In order to clarify the possibility of effecting some equivalent imitation, the flexibility of the atmospheric conditions in the measuring section of the tunnel was also investigated. The greatest obstacle in the way of rapid alterations in the operational conditions, and particularly in temperature, lies in the large mass of the air cooler. As a result of its large heat capacity (corresponding to 1200 kg of iron) the quickest changes of temperature that can be achieved in either direction are in the order of 1.5 to 2°C/min. At first, these values appeared to be inadequate. They can, however, be very considerably improved by the use of the 19-kw heater, although the temperature changes thus produced are always upwards, making the tunnel air warmer. In creating periodic variations the basic temperature \( T_f \) is arranged so that it remains constant, with the cooling compressor set at a fixed output-level. The maximum rise in temperature with the heating switched fully on is accordingly dependent only on the air speed, and it goes up as the volume of air put through decreases. The rapidity with which this rise in temperature shows up in the measuring section is dependent on the heat inertia of the tunnel between the heater and the measuring section. Figure 10 shows these relationships, observed for various air speeds as a function of time with maximum heating. If, moreover, the refrigerating compressor is stopped at the moment when the heating is switched on, increases in temperature result which are on an average 20 to 40% higher than the values shown in Figure 10.

Since we have been able to observe that heating takes as long as cooling, we are in a position to estimate the double periodic amplitude \( 2A \) (which equals the maximum temperature increase) as a function of the period time; (this is shown in Figure 11). These periodically induced changes of temperature are, however, subject to some limitation from the capacity of the refrigerating plant in the lower temperature ranges. Here care must be taken that the heat setting for the lowest temperature wanted does not exceed the requisite cold output, otherwise the compressor cannot adjust itself to this operational point. Similar

![Diagram](image-url)

**Fig. 10**—The maximum increase of temperature \( \Delta T_M \) which can be produced in the measuring section by switching the heater full on, expressed as a function of time and for various air speeds \( v_M \) (showing the heat inertia of the hail tunnel at a constant vaporizer temperature \( t_v \))

![Diagram](image-url)

**Fig. 11**—Behavior of the hail tunnel under periodic changes of temperature: maximum double temperature amplitude \( 2A \) as a function of the air speed \( v_M \) and for different period times \( T_A \)

![Diagram](image-url)

**Fig. 9**—The maximum possible water injection \( w_{in} \) as a function of the measuring-section temperature \( T_M \) and at various air speeds \( v_M \)
considerations are called for when extra humidity is injected.

From these particulars it may be seen how the cold output which is too high during stable operation makes possible large periodic temperature changes during unstable operation of the tunnel.—To what extent, however, the variations brought about in our experimental atmosphere may suffice to simulate the growth of natural hailstones, is a question which can only be properly answered by experiment and appropriate comparison.

Problems of Measurement

That subcooled water clouds may be produced in the way we have shown, does not, of course, solve the problem of how the relevant conditions are to be measured. For measuring the temperature alone a great variety of methods is therefore being tested and used according to need; and as the state of research shall dictate, new types of measurement can be tried out and new measuring apparatus built into the tunnel.

Prospects

The performance of a plant has also to be judged from the point of view of how far it broadens our knowledge of known phenomena or helps us to discover new instances of regularity and law. In the latter connection the first experiments with particularly high water contents (up to 20 g/m³) have indicated the following interesting effects:

(1) It can easily be shown that a close relationship must prevail between the conditions in which hailstones arise and their shape. But since different growth phases become overlaid the connections are sometimes noticeably complicated.

(2) A number of authors have assumed, when postulating a hail theory, that no more water can accumulate on a hailstone than can become frozen as a result of the warmth generated by freezing escaping into the surrounding air. The surplus water would then be carried away by the air current. These assumptions have proved to be quite arbitrary, since this surplus water can to a large extent be incorporated into the ice structure of a hailstone. This fact suggests a basic revision in the way the origin of large hailstones is explained. At the same time it also becomes obvious that nature produces hailstones of density greater than that of ice [List, 1959c].

The hail tunnel was planned and built under close cooperation with the firm Sulzer AG, Winterthur, as part of the research program of the Swiss Commission for the Study of Hail Formation and Prevention. I should like to express my particular thanks to my distinguished Principal, M. de Quervain, under whose helpful guidance I have been able to design, build and run the wind tunnel. The plant was paid for by the "Schweiz. Nationalfonds."

References


Discussion

(Note: Discussion of this paper is combined with those of the preceding and following papers at the end of the following paper.)
Growth and Structure of Graupel and Hailstones

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Abstract—Interpretation of the structures of air bubbles and single ice crystals makes it possible to show that the growth of hailstones on graupel and small hail particles follows certain laws of symmetry. For the present a number of general qualitative statements concerning corresponding glaciation conditions can be made, which explain these facts.

INTRODUCTION

The elucidation of the structure of natural ice particles is fundamental for an understanding of their growth. This principle takes precedence over any theoretical consideration, since in the case of atmospheric freezing processes, only observation of the natural ice particles covers the full range of variation in shape, structure, and growth potential.

An attempt is here made to give some coherent order to all the facts which have thus far been established from graupel and hailstones at Weissfluhjoch. Of course, so far it has not been possible that every type occurring in nature has been taken into account. Further observations or equally intensive parallel studies in other places will perhaps give more complete results.

In particular what will be shown is the way in which the growth of graupel or hailstones is to be characterized and by what structural factors this suggests. For specialized information reference may be made to the more detailed publications [List, 1958a].

THE THIN-SECTION TECHNIQUE

The structural analysis of graupel and hailstones is based primarily on the thin-section technique as it was developed by de Quervain [1950] for snow. His method enables, in particular, layers 0.3-0.4 mm thick to be saved out of ice particles of low density (0.1-0.7 g/m³), without the ice structure being disturbed. The layer obtained in this way provides us with knowledge of the amount of air contained in the ice and of the arrangement of the individual air bubbles. The use of polarized light shows up the crystalline structure of the ice and indicates the arrangement and size of the individual single crystals.

The structural picture is strongly influenced, of course, by the zone and direction in which the section is taken from the graupel or hailstone. Normally the plane investigated should run through the growth center, the original nucleus of the particle, the growth directions of the individual crystals being contained in this plane. This is called a main section; other sectional planes may be taken for special examination.

THE FUNDAMENTALS OF GROWTH IN ATMOSPHERIC ICE PARTICLES

Crystallographic considerations—The crystallographic main axis, the c-axis of the individual ice crystallites, is as a rule approximately vertical to their direction of growth. This results from the speed of growth being dependent on direction; this determines the selection of starting points for new unit crystals. A further consequence of this characteristic is that the unit crystals, growing out symmetrically from the center like rays, exhibit a pyramidal or truncated pyramid form. Lenthwise sections are therefore generally triangular or trapezoidal, while sections taken at right angles to the direction of growth are rather polygonal. With practice all intermediate cuts can also be recognized distinctly. The arrangement of the crystallites yields a certain symmetry for which conclusions can be drawn as to the manner of growth. It is, for instance, possible in every hailstone to determine the center of growth, the oldest part of the whole particle, on the basis of the form and arrangement of the single crystals. So long as this center comprises the growth directions of the first generation of single crystals, it is regarded as symmetry-center I. It continues to occupy this role so long as the conditions of growth, the type of glaciation, the shape of the particle, and its aerodynamics all remain
practically constant. If the falling particle, however, should change its relative direction of fall, begin to rotate or be arrested in its initial rotation, then normally a new center of symmetry (II) is formed and a second generation of unit crystals become oriented about this other point. Frequently different generations of this kind are interrupted by a so-called intermediary phase which consists of relatively much smaller crystals.

A single hailstone may exhibit a third or even more centers of symmetry. Figure 1 gives a particular example of a hailstone core having three centers of symmetry. Changes of symmetry, equivalent to changes of growing conditions, are happening over a whole intermediate surface of a growing hailstone and can be recognized later in the thin sections. This fact helps us to conclude from the variations of symmetry to earlier shapes which the hailstone has passed through. In the case of hailstone 57.15 (Fig. 1 and 2), for instance, we can recognize a conical structure which turned into an ellipsoid (symmetry-center I). Continued growth led to a rotary ellipsoid (symmetry-center II) and afterwards reached the final shape of a triaxial ellipsoid compressed along the line of the smallest axis (symmetry-center III).

The causes leading to the formation of new crystallites are unknown. The assumption that there is some connection with the incorporation of additional freezing nuclei [List, 1955a,b] cannot be maintained, as these would have to be active at approximately 0°C. The whole mechanism follows, however, an apparently strict pattern, as can be seen from the structural pictures, and further research should throw light on its causative principle.

Bubble structures—Analogous conclusions may be drawn from the arrangement of air bubbles contained in the ice: bubbles arranged in lines indicate the direction of growth; an inner shell of bubbles indicates an earlier form of the hailstone. Zones of equal density may, with certain exceptions, be taken to indicate zones where the same growth conditions prevailed. Figure 2 shows the same section from hailstone 57.15 as Figure 1 but seen this time by translucent light. From this picture it is clear that the intermediary phase I–II has resulted from a change in the conditions of growth, causing a shell of bubbles to form partially around the growth zone I. As
a result the particle developed different aerodynamic behavior and this led to the establishment of a new center of symmetry.

It may be remarked here that growth in one place may be the result of two different phases, the one 'overlaying' the other. This occurs when a primary ice formation produces a loose ice framework, air capillaries of which become filled with liquid water at a later stage; this then freezes either entirely or in part.

It is this 'overlaying' process which gives us a generally better picture of the original center than of the later phases, contrary to the views of Dessens [1959]. As an example, Figure 1 may again be referred to. We know that normally a graupel acts as the basic particle of a hailstone. Hailstone 57.15 indicates by the conical symmetry of the initial particle that the graupel was also conical. Since, however, the shape of this central zone is ellipsoidal, this means that the graupel has absorbed slow-freezing water into its cohesive system of caverns. This has brought about its transformation into a small hail particle with the surface tension of the wet surface leading to a roundish shape. Such overlaidings cannot be so well recognized in later phases.

**Fig. 2**—Thin section through hailstone 57.15 under translucent light; real length of the figure, 3.0 cm

**Fig. 3**—Characteristics of the growth of a hailstone

GROWTH FROM ICE-FORMING NUCLEUS TO HAILSTONE

On the basis of observations made between 1933 and 1959 on graupel and hailstones the
diagram given in Figure 3 can be used to represent the various stages in the development of a hailstone. On the left the type of growth is given, whether by sublimation or accretion of drops, and on the right the main growth phase. This may consist either of an increase in volume or of an increase in particle density. Figure 3 also shows whether the accreted drops freeze slowly or quickly.

While there is no need to give a definition of freezing nuclei and ice crystals, the other particles may be characterized as follows:

Graupel (soft hail)—Appearance: Graupel are white opaque particles, resembling sectors of spheres, roundish or irregular in shape, and with a diameter rarely exceeding 7 mm. Origin: Graupel originate through the accretion of cloud drops on an initial ice crystal which has grown by sublimation. Water caught up in this way freezes relatively quickly at the point of contact.

Small Hail—Appearance: Small hail particles are particles having generally rounded surfaces, but of conical, spherical, or irregular shape. They appear white and opaque in parts; certain zones of their surface are glassy or wet. Their diameters rarely exceed 7 mm. Origin: Small hail originates through accretion of water drops on graupel. Freezing takes place slowly, so that the floating water has time to spread out and penetrate the air capillaries on the original graupel. It may freeze entirely or in part.

Hailstones—Appearance: Hailstones are transparent or partially opaque particles with a diameter of about 5 to 150 mm or more. The number of shapes which are met with is extremely large; the structure is most frequently 'onion-skin' with alternating clear and opaque (white) layers. Larger hailstones often contain liquid water in a system of connected caverns. Origin: Hailstones originate from small hail through ac-
cretion of water drops which then freeze partially or entirely.

(Small hail particles are, in fact, essentially graupel which have already reached a higher stage of density. To call them small hail makes sense only according to this manner of characterization, for there is otherwise no possible means of distinguishing small hail from hailstones.)

Figures 4-7 show thin sections of a graupel, a small hail particle (these are wet as a rule and freeze together before examination, with the result shown in Fig. 5), and of a hailstone.

With regard to Figure 3 note that in all cases observed there was no indication that graupel arise from frozen water drops. This possibility may, however, exist in nature: this would necessitate a corresponding modification of the model.

The color pictures, Figures 8-11, should give an impression of the real appearance of a thin section under polarized light. The range of variation in color is given by the thickness of the ice, where the brown tone sections correspond to 0.3 mm; the other ones are thicker. As the limits of the single crystals are much sharper the thinner the section, it is better to cut the ice slices as thin as possible, perhaps contrary to the aesthetic point of view.

The comparison with black and white reproductions of thin sections under polarized light (see Fig. 7 and 11 which were taken of the same thin section) show that the impression of the arrangement of the single ice crystals and the symmetries is more evident in the color pictures.

**Concluding Comments**

The information which can be inferred from analysis of bubble arrangements and crystalline structure is of crucial significance but only qualitative. Our improved knowledge should, on the one hand, prevent us from carrying out calculations concerning layer formations on the basis of entirely untenable assumptions [Mason, 1958], and on the other hand encourage research which will yield quantitative results. It is to this end that investigations are to be carried out with the Swiss Hail Tunnel, which aims at explaining natural ice structure by creating similar structures by artificial means.

 Whereas a year ago it seemed possible to understand growth behavior perfectly with the help of structural investigations, laboratory
Fig. 6—Thin section of hailstone 59.1 under translucent light; diameter, 4.0 cm

Fig. 7—Thin section of hailstone 59.1 under polarized light; diameter, 4.0 cm
Fig. 8—Thin section through hailstone 57.50 under polarized light; diameter, 5.0 cm
Fig. 9—Center of hailstone 57.A under polarized light; diameter of the center, 2.2 cm; large single crystals surrounded by smaller ones
Fig. 10—Thin section of hailstone 57.10, under polarized light; diameter, 4.1 cm
Fig. 11—Thin section of hailstone 59.1, under polarized light; diameter, 4.0 cm
measurements have established that not only are there growth phases which in places become overlaid, but that many times a water layer forms that is stabilized by a subsequent ice framework; it is this which, together with the action of capillary forces, is responsible for maintaining the fluid phase where it is [List, 1959]. This phase, which is apparently of such prime importance, means that even the usual and seemingly reasonable classification that rapid growth produces opaque ice layers must be given up. Very rapid particle increase may be connected with a two-phase ice-water addition, with the water involved only able to freeze at a later stage, when it forms a frequently clear zone in conjunction with the original ice framework. A technique for establishing such distinctions after the event has not yet been found, although the requisite comparative experiments are in progress.

References

Dessens, H., La Gréle, Association d’études des moyens de lutte contre les fléaux atmosphériques, no. 7, pp. 3-17, 1959.


Discussion

(This discussion relates to the three immediately preceding papers.)

Dr. C. L. Hosler—Do the thin sections eliminate the possibility that a large portion of the growth is due to the collection of the ice crystals rather than entirely to supercooled water?

Mr. R. List—We have not yet made experiments which could show very clearly that that would not be the case. But our experiments show that it is not necessary to have snow crystals which can aggregate.

Dr. Hosler—My picture would be one of aggregation of crystals; then filling in of the spaces by water of the supercooled droplets.

Mr. List—Only when the surface is wet can ice particles aggregate. When the particles, big or small, are dry they bounce off.

Dr. Choji Magono—The pictures shown by you are actual hail stones?

Mr. R. List—Yes, and the center of natural hailstones can be recognized in 80% of the cases as granulated.

Mr. D. Blanchard—I would like to take exception to a statement Dr. List made about liquid water being shed from the hail pellet itself. You said that the liquid water is itself absorbed into the hail stones or is shed. At the first meeting here at Woods Hole, I presented a paper (The Supercooling, Freezing, and Melting of Giant Waterdrops at Terminal Velocity in Air, Artificial Stimulation of Rain, pp. 233-249) reporting on some experiments I did. One day, I tried the reverse process, taking an ice sphere, suspending it in the wind tunnel, to see what happens when the ice melts. When the melting occurred, a water ring would form around the center of the sphere. Of course, eventually water will be shed, but first there will be a horizontal ring around the center.

Mr. List—Horizontal rings can be observed when the rotating axis of the growing stone is stable. That means the form of hailstones depends on the icing conditions. When the water freezes slowly, it has time to soak into the interior and finally to form corresponding to the aerodynamic pressure distribution a ring around the rim of the stone. Then one finally arrives at plates which can rotate about the vertical axis. One can see this rotation also in the frozen water droplets. They are arranged around the vertical axis.

From the floor—Do we know enough about what a frozen water drop looks like when it is freezing? I wonder whether your analysis would indicate whether the center of the stone was formed by the particle which Ludlam talks about, that is, by freezing of a water drop having a radius something of the order of 30 to 50 microns. If this froze because of a violent collision of two such drop sizes, or because of the crystallization of a single water drop or if the center was formed from the conglomeration of a number of frozen particles which resulted from
growth caused by sublimation, I wonder whether from an examination of the frozen center of hailstones one could differentiate these processes.

Mr. List—I think I can answer that last question, but not in an absolute manner. Since our institute lies 8000 ft above sea level, we can observe these graupel stages 20 to 30 times a year. We find not many graupel with what is probably a frozen droplet in the center, but in about 40 or 50% we are sure that the center is an originating snow crystal. I would say it is not necessary to have frozen droplets.

Dr. Helmut Weickmann—How much did the tunnel cost?

Prof. R. Sänger—It cost about 600,000 Swiss francs or $200,000.
Hailstorm Structure Viewed from 32,000 Feet

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Abstract—Photographs taken near Cheyenne, Wyoming, on a high-altitude flight from the cloud-studies aircraft of AFRC show hail thrown out of the side of a large organized thunderstorm. A small vortex is visible at one edge of the hail shaft. The relation of the hail region to other regions of the cloud is clearly evident. The cloud and hail patterns, revealed after mapping the clouds from the photographs, suggest a horizontal cyclonic circulation of the whole storm.

While studying the effects of low topography on the cloud patterns east of the Denver area, aerial photographs were taken of a well-developed large hailstorm. This paper will discuss these photographs and the measurements made from them.

On the afternoon of September 11, 1958, the AFRC C-130 Hercules aircraft, equipped with cloud-probing and large mapping cameras (T-11’s) flew along a SSE-NNW line some 80 mi east of the Front Range of the Rockies. Small showers and thunderstorms had produced a general chaotic cloud mass over the Rockies. Figure 1 shows the tops of these clouds as they descend and evaporate in the westerlies flowing down from the mountains. The photograph was taken looking toward the west. A west-east line of cumulus is visible at a lower altitude. These mountain thunderstorms have not developed into self-propagating storms; presumably at least on this day they cannot survive passage through the downdraft region east of the Front Range. In contrast to the poorly organized thunderstorms over the mountains, large well-organized hailstorms were photographed east of our flight path (Figs. 2 and 3). These storms were self-propagating, lasting until after midnight. Mountain wake circulations may be of importance here; that is, the large storms first develop in and under a region of general upward motion in the first downstream wave beyond the mountain. Once started over flat terrain, these storms, on drifting eastward, can apparently become organized enough to outweigh the effects of the farther downstream portion of the mountain wave.

The surface and upper-air conditions can be described as follows. A thin layer of moist hot air was moving at moderate velocity from the south and southeast over the region just east and north of Denver and Cheyenne. Temperatures reached into the upper eighties (all surface readings in °F) in this region and dew points were recorded at Akron, Colorado, in the lower fifties, and at Scotts Bluff, Nebraska in the lower sixties. The surface air at Denver, however, was part of the direct flow down off the mountains as the dew point fell in the afternoon to the upper thirties, the temperature reached the upper eighties, the wind was light and variable. Cheyenne recorded light southeast winds with surface temperatures in the lower eighties, dew points near fifty. The vertical thermal and wind structure of the atmosphere is shown on a skew T, log P diagram, (Fig. 4). The features to note are the following: (1) The thinness of the surface layer of southeast wind with slowly increasing westerly wind above, and the strong west-northwest winds above the isothermal layer at 35,000 to 35,000 ft. (2) The difference in the moisture structure over Denver and over Scotts Bluff; the air at Denver below 23,000 ft is relatively homogeneous with a nearly constant potential temperature and small range in moisture content, suggesting again that this air is well mixed and comes from the higher terrain to the west. The air at Scotts Bluff, on the other hand, is moist in the lower layers and very dry above 15,000 ft. (3) The isothermal layers, a minor one at 22,000 ft, a major one at 33,000 to 35,000 ft. This latter layer is perhaps a remnant of a more northerly tropopause. The sharp tropical tropopause appears at 51,000 ft.

The hailstorms to be discussed were, it appears, surrounded by an atmosphere close to that measured by the Scotts Bluff radiosonde. The maximum temperatures during the day in the storm area reached as high as 91°F (33°C). A
Fig. 1—Right T-11 Photo 122, looking west

Fig. 2—Right T-11 Photo 91, looking east northeast
more representative temperature sounding for mid-day (radiosonde released one to two hours after time of maximum temperature) would lie in between the Denver and Scotts Bluff curves in the lower layers. A dry adiabat drawn from the point of intercept of the storm cloud base (15,000 ft MSL, 600 mb by measurement from the photographs) and the Scotts Bluff sounding reaches the surface at a temperature of 34°C. The moist adiabat from the cloud base intercept crosses through the isothermal layer at 34,000 ft (266 mb).

With the structure of the ambient atmosphere in mind, a closer look will now be taken of the large storm masses that developed during the afternoon of the 11th. In the upper middle part of Figure 2, a large storm is just beginning to produce rain at the ground. Almost the whole perimeter of the cloud shadow can be seen, indicating only a few hydrometeors below cloud base. This storm has grown, however, to above 40,000 ft and a large volume of cloud and other hydrometeors have been blown downstream in the strong upper winds to form a large anvil with maximum horizontal spreading at the upper isothermal layer (33,000–35,000 ft). Another similar storm appears in the upper left but is largely obscured in this picture by lower, nearer Cumulus.

A third storm, the one to be studied here in detail, appears on the right side of Figure 2. A more complete view of the western end of this storm, (Fig. 5), shows several interesting features. (The dark vertical line in the upper part of the picture is one blade of the aircraft propeller.) Figure 6 should be referred to when studying Figure 5. Figure 6 was traced from photographs with the use of a Canadian grid technique and by triangulation from several pictures. Note that because of the varying distances from the flight path, this tracing is divided into three parts with three different scales. The upper and left parts are from Photo 90, the lower right from Photo 89.

The features of interest will be discussed from the top down. The portion of the cloud above flight and cirrostratus level reaches some 10,000 ft into stable air; individual cloud elements (bubbles) do not greatly outrun their neighbors, giving the whole cloud a rather rounded top.
lack of distinct individual turrets at the top suggests that the storm is characterized by a large region of fairly continuous upward velocity with a narrow velocity spectrum, in contrast to many Cumulus systems where the velocity spectrum is wide and individual turret production is a major feature. Much of the cloud material that reaches these heights gains momentum from the ambient air and spreads downstream in the anvil. This feature is best seen in the young cloud in the upper middle of Figure 2. The Cirrostratus layer at flight level (Figs. 5 and 6), may be caused by the upwind spreading of that portion of the upper cloud material that has lost its upward momentum, is relatively cold, and descends against the ambient wind down to the isothermal layer. More likely, however, this upstream Cirrostratus is formed by an uplift of the ambient air flowing around the storm. The horizontal, thin protrusions on the north side of the storm, pictured best in Figure 3 at about 27,000 ft and 22,000 ft, may be caused by this blocking action forcing the ambient air to ascend. That portion of the protrusion at 22,000 ft, visible on Figures 5 and 6, appears rather to be caused by motions within the storm. Mammatus formations mark the lower portion of this protrusion, suggesting that small precipitation particles and clouds are flowing down and out of the main cloud and evaporating into the ambient air.

The most interesting feature shown on Figure 5 and sketched on Figure 6 is the great fall of hydrometeors into the sunshine below the marked protrusion. The bright portion of this fallout is presumably a region of snow, snow pellets, and hail (S + A). The dimmer portion, rain and hail (R + A). The brightest portion in these two regions may be shafts and bunches of falling heavy hail. The level at which most of the small frozen particles have melted is 5000 to 12,000 ft or at +20 to +8°C ambient air temperature; the more relevant local air and wet bulb temperature in the precipitation is unknown.

To the right of the precipitation is a large
Fig. 5—Right T-11 Photo 89, looking east northeast

Fig. 6—Two vertical cross sections west end of hailstorm parallel to flight path; scaled from Photos 89 and 90
region of Cumulus and Cumulus congestus seemingly leading into the storm, the cells growing larger the nearer to the storm mass they are. These clouds and their relation to the main storm are also seen in Figure 3. At the juncture of the western edge of these clouds with the area of precipitation a dark sloping line is visible, bordered by two bright lines. This configuration is assumed to be the result of a vortex circulation, the precipitation particles being concentrated on the perimeter of the vortex. This apparent vortex was short-lived; it is not visible on Photo 88 (taken one minute previous); and it appears about double the size but very faintly on Photo No. 90 (taken one minute later). One may imagine that the Cumulus cloud material on the right is streaming into the storm while the precipitation is flowing out and that the region of

Fig. 7—Enlargement of part of Photo 89
strong shear between these flows spawns eddies, some of which develop sufficiently to show up as a long vortex such as the one visible on Figure 5. Figure 7 is an enlargement of the region in which the vortex is visible.

Close inspection of the ground shading in Figures 2 and 5 reveals the pattern of rain coverage from the nearby storm. The darker ground leading to the left of the storm is the ground wet by the heavy rain and hail. The spots of white on the upper side of this dark area are patches of hail left by the storm. According to W. B. Beckwith (private communication) hail on the ground that falls with little rain will show well from the air but when accompanied by heavy rain it will be washed into the vegetation and gullies and will not show. Hail, therefore, could have fallen throughout the precipitation area but only show on the ground where it separated from the other hydrometeors, descending along a different trajectory because of its greater fall velocity.

Many of the features discussed above are mapped on Figure 8 through the use of the Canadian grid technique on single photographs and by triangulation using several photographs. The similarity in size, shape, and spacing of these storms is striking. The anvils appear to spread off in the down shear direction, extending for about 100 mi. The motion of the storm mass itself has been measured as along the direction of the anvil and at the low velocity of 21 knots, or a speed equivalent to the speed of the ambient air (Scotts Bluff 300 mb) at the cloud base. This measurement was made utilizing all the time fixes indicated except the 1345 radar position. The radar positions were obtained from W. B. Beckwith (private communication). Additional fixes were obtained from surface observations at Cheyenne, noted on Figure 8, and Akron, Colorado, where the storm was observed passing north of the town at 17 h00m. The aircraft photograph also gives a fix at 15 h00m. The fix used from the photographs was placed five miles east of the indicated vortex position. The speeds calculated from the six fixes were in close agreement,
being $21 \pm 3$ knots. The radar positions at 13h45m are not consistent with the other points.

Note should be taken of the narrowness of the precipitation area in comparison with the size of the main cloud mass. The main mass of the storm passed over Cheyenne Airport at 13h00m but only a trace of rain fell in seven minutes. Marble size hail was reported three miles south of the airport. The hail region outside the main cloud mass is indicated on Figure 8. As mentioned previously, it is suggested that this is a region of outflowing air and the building cumulus on the southwest side is a region of inflow relative to the slow storm movement. One might imagine that, besides the obvious vertical motions taking place, there may be a consistent horizontal half spiral circulation, inflow rising, hydrometeors growing rapidly, part of the flow gaining eastward momentum and turning downstream to form the anvil, the other part turning westward and descending with the heavier hydrometeors. In an earlier stage, most of the cloud material would flow out downstream in the anvil, (see mid-cloud in Fig. 2), showers would fall from the main mass between stronger updraft cells, creating a low level downdraft region which might induce a steady westerly branch of the flow in the upper cloud (particularly of the more protected volume of vertical rising air), creating a steady supply of hydrometeors and therefore a steady downdraft.

Rotary motion, cyclonic, of the thunder and hailstorms in the Denver area have been noted by W. B. Beekwith (private communication) and in other areas in studies by Fujita [1960].

In this paper a great deal has been suggested concerning the air circulations in a western hailstorm from the analysis of several still photographs taken from a high altitude. The actual circulations in these storms await the time when flow measurements in and around these storms are taken with aircraft or by some safer means. The intent here was to point out what can be gained from a high-altitude view of this type of storm in areas where the surrounding atmosphere is dry and largely void of other obscuring clouds. The intent is also to point out that some large Cumulonimbus may not consist entirely of a series of cellular elements in various stages of their life cycle, but may to some extent consist of a steady horizontal and vertical circulation system.

Reference


Discussion

(Note: Discussion of this paper is combined with that following the next paper.)
Severe Hailstorms Are Associated with Very Strong Winds between 6,000 and 12,000 Meters

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Abstract—Both physical and statistical studies indicate that the factor which transforms thunderstorms into severe hailstorms is a very strong horizontal wind at levels between 6,000 and 12,000 m. A structure for the hailstorm is suggested, in which the 'chimney' updraft of the Cumulonimbus is considered as forming a link between two main sources of energy: in the lower part, the instability and the latent heat of condensation and of freezing; in the upper part, the kinetic energy of the strong wind.

In any given region, on certain days thunderstorms cause a very light or moderate damage by hail; on other days the damage is widespread and greatly increased. Furthermore there are regions of the world where thunderstorms are very frequent and violent but where hail is extremely rare. We have tried here to delineate a factor or factors which transform an ordinary thunderstorm into a destructive hailstorm.

The influence of atmospheric aerosol particles has not been demonstrated—Following a suggestion of Ludlam (1951), we have investigated if the formation of hail is not hindered by an abundance of giant hygroscopic nuclei in the air. Some trials at hail prevention were undertaken in 1951 using rockets which took crystals of sodium chloride up to cloud base. Neither the results of these trials nor air sampling have indicated that there is any correlation between hail formation and the concentration of giant nuclei in the air.

I have observed on several occasions that destructive hail occurs more frequently when the lower atmosphere is very clear; if these observations are confirmed, they will mean that hail forms less easily when there is a high concentration of condensation nuclei. Remember that the concentration of giant nuclei does not necessarily vary in the same manner as the small condensation nuclei. Soulage (1958) has studied the variations of the ice-forming power of the air during two thunderstorm seasons. He has not found any strong diminution of the ice-forming power during destructive hailstorms. These measurements, made at ground level, have so far failed to reveal any significant correlation between hailstorms and the concentration in ice-forming nuclei.

Finally I call your attention to the fact that hailstorms are extremely rare in the Congo Basin as observed at Lukolela (1°S, 17°E). The observations at ground level made by Soulage [1956] indicate, at least during the dry season, there is an almost total absence or at most a very low concentration of freezing nuclei in the area. Furthermore, the photogrammetric measurements of clouds according to a method described [H. Dessens, 1953] show that in the rainy season the effective freezing level in the clouds is observed at approximately −35°C [J. Dessens, 1959]. This tends to show that there are very few freezing nuclei active in the clouds in that area. These data therefore contradict rather than support the hypothesis that destructive hail is connected with an absence or low concentration of ice-forming nuclei.

A strong vertical development appears to be a necessary but not a sufficient condition for the formation of hailstorms—Since 1949 we have measured the altitude of Cumulonimbus tops in the region north of the Pyrenees by photogrammetry, using a base line of 8-km length. Measurements are possible only when the Cumulonimbus are from 30 to 120 km distant from the observation stations and if the sky is clear in the neighborhood of the stations. For the 14 thunderstorm days without heavy hail, the maximum height reached by the Cumulonimbus on each day, as follows: 6,000 to 9,000 m, four days; 9,000 to 12,000 m, eight days; and 13,000–13,200 m two days.

We have been able to make similar observations on days of destructive hail on only three occasions. The corresponding summits' heights were: 12,500 m, 14,000 m, and 14,500 m.

The two storms which reached 13,000 m without giving hail were formed over peaks of the
Pyreness and their anvils were only slightly displaced by the wind. These results are generally in agreement with those observed by Donaldson [1958], using radar, if one assumes that the top of the radar echo coincides with the top of the anvil.

Nevertheless, one cannot conclude that a strong vertical development constitutes the determining factor in hailstorm formation. In the Congo Basin a vertical development up to 16,000 m occurs without a single hailstone reaching the ground. This happens in spite of a vertical temperature structure very similar to that which is observed in temperate regions on those days of heavy hail, and also, in spite of a degree of instability and of humidity equally favorable for hail formation.

The transformation of a thunderstorm into a hailstorm is determined by the wind velocity at high levels—The only fundamental difference between hailstorms of temperate regions and thunderstorms of the Congo Basin lies in the horizontal wind speed at high levels. These observations have led us to make a statistical analysis of the thunderstorms of southwest France. In Figure 1, the full line represents the frequency distribution of maximum tropospheric wind velocity for 38 days on which the damage by hail, over the region, is estimated not less than 10 million francs. The wind data are obtained from radiosonde and the value for each day is that of the maximum velocity observed between the ground level and the tropopause, at whatever level this occurs.

In order to emphasize the significance of this diagram, we have included also (broken line) the frequency distribution of wind speeds for an equal number of days during the same season of the year (April to September) and the same period (1951-55), these 38 control days were days without significant damage due to hail, although thunderstorms did occur over the region.

A comparison of these two frequency distributions shows clearly the correlation between the wind velocity at upper levels and the degree of damage caused by hail. It is probable that the correlation between wind speed and the intensity of the hailstorms is actually closer than indicated on Figure 1. In effect the correlation is probably somewhat reduced by the fact that the radiosonde observations are often made at places distant both in space (that is to say, up to 250 km) and in time (up to 12 hr) from the hailstorms. It is worth mentioning in addition, that although on five days, winds of more than 90 km/h were observed without wide-spread destructive hail. On four of these days, heavy hailstorms did in fact occur but the damage done was not very great because they were very limited in extent. Finally, to sum up, on a total of 76 days with Cumulonimbus activity, there is only one case in which the maximum wind-speed exceeded 90 km/h without being accompanied by destructive hail.

In conclusion, it appears that the presence of a jet stream, or at least a very strong wind at upper levels, is the factor which determines whether or not a thunderstorm situation will transform itself into a heavy destructive hailstorm.

A suggested structure for the hailstorm—In air masses with relatively little variation of horizontal wind speed with height, the ascending currents in Cumulonimbus are approximately vertical. The air in the 'chimney' updraft takes its energy from the instability of the atmosphere, perhaps increased locally by differential heating of the soil and from the latent heat of condensation and of freezing; this energy is liberated mainly in the lower part of the ascending air column. Therefore chimneys are fed from the low end and receive no important energy at high levels; furthermore, if air is not removed from the top, then the 'chimney' may fail to draw. Chimneys therefore tend to be short lived, depending, as they do in this case, on the hazards of local conditions and particularly on the nature of the terrain over which they travel, under the influence of the

![Figure 1](image-url)

**Fig. 1**—Frequency distribution of wind speeds at high levels for 38 days with heavy destructive hailstorms (full line) and for 38 control days with thunderstorm but without significant damage by hail (broken line)
prevailing wind. In the cloudy edifice, a chimney is born, develops, and quickly dies to give place to another. Such sporadic activity is not favorable to the formation of large hailstones [Wieckmann, 1953].

On the contrary, under conditions of strong windshear, as in Figure 2, the chimney is inclined to the vertical. On the right side of Figure 2 is included a chimney of very restricted transversal dimensions, in accordance with the observations of Wieckmann [1951]. It is certain that in a case such as this, a part of the air rising in the chimney is entrained horizontally at the top of the chimney at a speed of the order of 80 m/s. Therefore I envisage a coupling taking place above 6000 m between the updraft in the chimney and the very strong horizontal current and that this current tends to stabilize and prolong the life of the chimney, which now receives, in addition to the energy of thermodynamic origin at low levels, kinetic energy of the strong wind at high levels. The chimney may, in fact, be considered as forming the link between these two main sources of energy; the lifetime of an individual chimney appears to be prolonged under these conditions for periods of more than 30 min sufficient to make possible the formation of very large hailstones.

Arguments in favor of the suggested structure—In addition to the data which led me to suggest this structure of the hailstorm, there is the following argument in its favor. It is the correlation, which has been observed, between the maximum size of hailstones and the maximum wind velocity between 6,000 and 13,000 m.

Figure 3 shows the correlation between these two parameters for seven hailstorms which occurred in the region within a 200 km radius of Bordeaux. The curves showing the variation of terminal velocity of hailstones with height and with diameter are taken from Wieckmann. The fact that the crosses representing observations lie fairly close to the curve corresponding to 11,000 m, shows that the maximum velocity of the updraft tends to approach that of the horizontal wind as measured outside the storm-cloud. Thus the maximum size of the hailstones would appear to depend on the velocity of the horizontal wind.

It is because of this that a hailstone cannot remain indefinitely within the ascending chimney when this is inclined to the vertical, and therefore we can obtain from this picture of the hailstorm a forecast of a maximum size which
the hailstones may reach. The value of such a forecast lies in the fact that in general the upper air winds have already been present for several hours at least before convective activity becomes important.

This last paragraph and Figure 3 represent only a preliminary version of the suggested hailstorm structure. It is to be hoped that the different studies at present being pursued in various parts of the world will enable us in the near future, to reach some more definite conclusion as to the value of this suggestion. On the other hand there is need for a theoretical explanation of the fact that the strong horizontal wind at high levels tends to accelerate and stabilize the updraft in Cumulonimbus-clouds.

References


(Discussion pertains to the two immediately preceding papers.)

Dr. Helmut Weickmann—I call attention to the fact that another plausible explanation exists as to why there occurs little or no hail in the tropics. It has been found (Fawrbrush and Miller, Bul. Amer. Met. Soc., 34, 235-244, 1953) that no hail falls from hailstorms which have a wet-bulb freezing level above 11,000 ft, because then even the biggest stones melt before they reach the ground. I suspect that this is the case in these tropical hailstorms.

Dr. H. Dessens—I think that the greater height of the 0°C level cannot explain the scarcity of the hail in the equatorial regions. In France, in summer, the 0°C level is near 3500 m, and the top of the hailstorms near 14,000 m; at this top, the temperature is about —55°C.

In the Congo Basin, the 0°C level is near 4500 m, and the top of the thunderstorm near 17,000 m; at this top, the temperature is about —80°C.

In both of these regions, during the heavy showers, the temperature near the ground is of the same order: +17°C.

Thus, the difference in the 0°C level is of the order of 1000 m. The time of fall of big hailstones through this layer is shorter than one minute. This time seems inadequate to melt the hailstones noticeably, the more so as the mean temperature between the 0°C level and the top of the cumulonimbus is much lower near the equator than in the temperate regions.

Mr. Jerome Namias—In the United States one of the atmospheric characteristics looked for in the prediction of severe storms or tornadoes is the existence of a mid-tropospheric jet. No one has yet given a satisfactory explanation for this association. Nevertheless, it is a pretty good indicator, provided the lower thermodynamic air-mass structure is conducive. I would like to suggest two points which may have something to do with this. Springtime is the most favorable for tornadic activity. Then we frequently have conditions which lead to the development of deep layers of very dissimilar air masses—very cold air from the Pacific over-riding moist and warm, tropical gulf air. This differential flow often leads to convective instability and, simultaneously, the presence of jets which are found in the middle troposphere. Perhaps there is something about the circulation in and around these jets that might encourage convection; this would be enhanced by the thermodynamic instability.
Dr. J. Smagorinsky—It might be possible to explain the convective instability underneath a jet as a gravitational shearing instability.

Dr. Weickmann (communicated)—I may make an attempt to explain this phenomenon from purely cloud-physics reasons. We know from the Byers-Braham thunderstorm model that a storm's life span can be subdivided into a developing stage, a mature stage, and a dissipating stage. The dissipating stage begins with the development of a downdraft in the rain area which slowly gains momentum and extends upward in altitude. Downdrafts also develop aloft and within the cloud because of entrainment of drier outside air.

In conditions of great instability the updraft velocities in the inner core of the storm may be sufficiently high as to prevent even larger particles from descending against the updraft. They too end up in the anvil where the updraft finally ceases. The particles, however, continue to grow until they are large enough to descend against the powerful updraft and to initiate the downdraft mechanism. The hailstones thus formed have a conical shape in the ideal case. Suppose now the existence of a strong jet in the upper levels. It will displace the anvil downwind from the mother cloud, so that large particles cannot fall back into it and cannot stimulate the formation of a downdraft which would lead eventually to the dissipating stage. Of course, from the downwind spreading anvil large particles will fall into other up-coming clouds and may form hailstones in these. The lifetime of the original mother cloud depends now only on the correct adjustment of external factors, such as inflow characteristics, migration velocity, and air consumption in the updraft. The increased lifetime of this updraft may be a decisive factor for tornado formation as it will permit the encounter with an already existing micro-low, or it will even assist in the organization of the correct low-level cyclonic circulation. Then the tornado may form.

Dr. W. E. Howell—Of course, in the tropics there are very seldom high-level winds of velocities that are anywhere comparable with middle latitude high-wind velocities. However, on the Pampa de Junín, the high plain in central Peru, there is often a strong easterly wind at high levels across the main range of the Andes while the Pampa lies, protected, between two ranges. Here the air is relatively slow moving and in fact often undergoes a reverse circulation so that there is a west wind near the ground beneath the east wind aloft. On this Pampa it is not unusual to have rather strong showers of graupel. Of course, since the elevation of the ground is 14,000 ft, there is much less depth of cloud in which hail could grow, but we do have the first phase that Mr. List has shown.

Dr. E. Kessler—Referring to this problem in a purely intuitive way, I have been struck also, as many others have, with the association of thunderstorms and high winds aloft, and their occurrence together in the spring. In Corpus Christi, Texas, where I used to live, numerous severe thunderstorms occur in the spring; later, in the summer they are not as severe. I have thought at times, that the explanation for this may lie in the transport of heat away from the region of active convection by the strong upper-level winds, which are weak or absent in summer. The unstable thermal stratification which is associated with the development of the springtime storms may then persist locally for a comparatively long time, with a correspondingly long time available for an organized, strong circulation to develop in the lower troposphere.

Dr. D. Swingle—I just have a quick question for Dr. Cunningham. Do you have proof this is the same storm that made hail near Cheyenne?

Dr. R. M. Cunningham—I do not know exactly what you mean by proof, but the track of the storm is right for that conclusion, shall I say.

Dr. Swingle—You do not have the continuity that actually ties it back?

Dr. Cunningham—Radar showed it coming from that area, but I do not believe we have the radar echo right over Cheyenne.

Mr. Alan Fuller—I would like to ask if the hail was falling through the zone of strong shear from the region of high winds into the lower region where the wind is not so strong?

Dr. Cunningham—The visible hail coming out of the side of the cloud was all lower than the strong winds. The strong winds are up in the Cumulus part of the cloud and the hail was down where we only had a wind of 20 knots. Of course, I do not know where it came from inside.

Dr. Tor Bergeron—Concerning Dr. Dessens' paper, I must confess that I on the whole agree with the first remark made by Dr. Weickmann. On the other hand, it is not my intention to deny the possible effect of the shear. But I want to remind you of the fact that in the United States
you have the maximal thunderstorm frequency in summer with one maximum over Florida and the other over New Mexico, the first being characterized by practically no hail, but by much rain and low cloud base. In the other maximum you have rather little precipitation reaching the ground, with the cloud base more than 3500 m ab. s.-l. I think, on average the cloud base is generally higher than Mount Sandia at Albuquerque, and yet hail reaches the valley, which is 1500 m ab. s.-l. My explanation, when I tell my students about these things, is that in Florida the hailstones exist probably high up in the cloud, but they melt before reaching the ground because there is constant condensation on them all the time, almost down to the final melting. In the New Mexico region they fall through dry air. They may evaporate, but they do not melt because they keep cool. Those principles would probably, on the whole, apply to all the storms in the tropics, and I thought it was generally recognized that there was no hail within the equatorial region. The really devastating hail is in central Europe, in Switzerland, and certain parts of the United States, but not in Florida. And then as to the shear and the high winds aloft, in Florida there are no such high winds aloft in summer. The jet may reach down to the northern part of New Mexico, which has a higher latitude. So jet and hail may have a common basis, but there need not be any direct connection between them.

Dr. C. J. Todd—F. H. Ludlam pointed out (Nabila, 1, p. 1, 1958) that the fall velocity of hailstones of radius of one centimeter or larger at 0°C is so fast that they should reach the ground with little loss of radius even in hot weather.
Morphology of Thunderstorms and Hailstorms as Affected by Vertical Wind Shear

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Abstract—Within convective clouds imbedded in a current with pronounced vertical shear, horizontal velocities are considerably different (1 to 10 m/sec or more) from environment winds, because of intense vertical transfer of momentum inside the cloud. Pressure fields induced by relative motions tend to promote new cloud growth on the downshear flank of a large convective system containing both updrafts and downdrafts. Physical analysis of the forces, with the aid of experimental analogy, indicates that the induced pressure field is quantitatively capable of triggering convection and that it significantly augments vertical accelerations in the newly formed cells. Possible influences on growth and distribution of large hail relative to the main body of a storm, are discussed with examples.

Introduction

Although theories for hail formation differ in important details, it is generally accepted that the vigor of updrafts (or successions of 'bubbles') is a major factor in hailstone growth, since this determines the possible effective length of the path swept out by a hailstone [Ludlam, 1958]. It is thus natural that the degree of instability serves as an excellent predictor for maximum hail size.

Air-mass structure, however, does not by itself determine the character of the convection. In air masses having much the same thermodynamic structure, small and chaotically-distributed, or very large organized thunderstorm systems may occur. Particularly in the latter, the distribution of severe convective phenomena tends to be strongly asymmetrical.

In the outline for the Woods Hole Conference, H. Weickmann posed the questions: "What do we know of hail formation; is it due to a very intensive or persistent updraft?" and "What is the significance of the zones of high winds (jet streams) which appear to be a typical feature of hailstorms?" This paper is designed to suggest how these two questions are related, for certain types of convective storms.

This analysis will concern only the effects of vertical shear on the structure of a cloud system, without regard to antecedent conditions for its formation. The latter, as is well known, is strongly influenced by the temperature and moisture advection and by large-scale weak but persistent vertical motions [Folks, 1951], which are most pronounced in the jet-stream region.

Interactions Between In-Cloud and Ambient Winds

As background for comments on the hail question, it is necessary to review some results of an earlier investigation [Newton and Newton, 1959] which treats, in greater detail, the interactions between convective cloud systems and the wind fields in which they are imbedded.

Rainfall and radar observations show that organized convective systems, even relatively solid squall lines, are composed of distinct agglomerations of thunderstorm cells, having average widths in the range 15 to 50 km. Such an aggregate, henceforth called a 'rainstorm,' is the entity to be discussed here.

In a typical organized convective situation where storms are found in the warm sector or ahead of a middle-latitude cyclone, the wind veers with height. Figure 1 shows a rainstorm in such a wind field, $V_w$ and $V_L$ being the environment winds in upper and lower levels. Large rainstorms are made up of alternate up- and downdrafts which exchange horizontal momentum between upper and lower levels. Complete mixing would result in mean in-cloud velocities as shown by the dashed arrows in the figure.

As indicated by the double-shafted arrows, relative motions would exist between ambient winds and in-cloud air. Evidently the right flank (with respect to motion of existing cloud) is most favored for new cloud growth, due to relative inflow of moist air on that flank in lower levels.

Several authors [Desai and Mal, 1938; Humphreys, 1940, pp. 353 and 350; Newton, 1950; C. S. Weather Bureau, 1949] have suggested that
triggering of new convection by cold air flowing outward from beneath a thunderstorm is enhanced if the air brought down in downdrafts has initially high horizontal momentum. In the following, an attempt will be made to analyze the broad-scale aspects of this process in terms of the physical forces at work.

**Induced pressure field and new cloud growth**

At a given level in Figure 1, the rainstorm may be regarded as an obstacle in motion relative to the ambient winds. If \( \mathbf{V}_E \) is the velocity in the undisturbed environment (for example, \( \mathbf{V}_c \) or \( \mathbf{V}_L \) in Fig. 1), and \( \mathbf{V}_r \) the mean in-cloud velocity, the basic relative motion is \( \mathbf{V}_R = \mathbf{V}_E - \mathbf{V}_r \). Hydrodynamic pressure is defined as

\[
P = p - p_h
\]

being the departure from the hydrostatic pressure in the undisturbed environment. Aerodynamics experiments [Goldstein, 1938] show that, in the case of a circular cylindrical obstacle, a positive pressure \( P = \rho \mathbf{V}_R^2/2 \) is induced on the upwind side (with respect to \( \mathbf{V}_R \)), and negative pressures slightly larger on the lateral and downwind sides.

In Fig. 1, the sign of \( P \) is indicated on the right flank at upper and lower levels; the signs are opposite on the left flank. Evidently one effect of this pressure distribution is to tilt the cloud along the direction of the vertical shear.

Substitution of \( p \) from (1) into the equation for vertical acceleration gives

\[
\frac{\partial w}{\partial t} = g \left( \frac{\Delta T}{T} - \frac{P}{\rho} + \frac{\partial P}{\partial p} \right)
\]

Here \( \Delta T \) is the excess of temperature relative to the undisturbed environment. On the right flank, the second term (not always negligible) is small compared with the third, when averaged through the depth of the cloud.

In general, for 'triggering' convective instability, a certain amount of lifting is required to reach a state of free convection (positive thermal buoyancy). During lifting, the air becomes temporarily cooler than in its undisturbed state. Eq. (2) shows that lifting can be accomplished by the hydrodynamic pressure force, if there is a strong enough decrease of \( P \) with height.

On the right flank of the rainstorm in Figure 1, the induced pressure field favors such lifting. Thus, the same pressure field that acts to shear the cloud away from the vertical, tends to promote new convective cloud growth on the downshear side.

Radar observations [U. S. Weather Bureau, 1949; Light, 1956] show that nonpropagating thunderstorm cells move very nearly in the direction of the mean wind in the cloud layer (or the 700-mb wind). If growth on the right flank predominates, this should show up as a tendency for rainstorms to deviate toward right of the winds. In case studies using hourly rainfall data [Newton and Katz, 1958] this was consistently observed, rainstorms moving on the average about 25° (10-15 knots vector velocity) to right of the mean wind, in situations where the wind veered markedly with height.

**Quantitative Estimates**

*Relative motions between cloud and environment*—Because the induced pressure field depends on the relative motions, it is essential to establish the probable magnitude of \( \mathbf{V}_R \) to see whether the process described above can be quantitatively significant. Pressure forces acting horizontally across the cloud (Fig. 1) tend to accelerate it, at
a given level, from high to low induced pressure, and thus to decrease \( V_R \). The force acting on a unit slice is \( D \Delta P, \Delta P \) being the difference between upstream and downstream pressures, averaged across the cloud diameter \( D \). The acceleration is given by this force, divided by the mass of the slice \((\pi i D^4/4)\), being
\[
\frac{dV_c}{dt} = \frac{4 \Delta P}{\pi D}
\]
(3)

In a coordinate system following the cloud, if advective transfer of momentum across storm boundaries is neglected
\[
\frac{\partial V_c}{\partial t} = \frac{\partial V_c}{\partial t} - \frac{\partial V_c}{\partial z}
\]

On substitution from (3),
\[
\frac{\partial V_c}{\partial t} = \frac{4 \Delta P}{\pi D} - \frac{\partial V_c}{\partial z}
\]
(4)

The in-cloud air can be maintained at a constant mean velocity different from that in the environment, so long as the sum of the two terms on the right is zero.

With strong shear typical of spring squall-line situations, it is found [Newton and Newton, 1959] that for large rainstorms \((D = 30 \, \text{km})\), with modest mean vertical motions \((\bar{w} = 1-3 \, \text{m/sec})\), balance between the terms in (4) can exist if the in-cloud shear is about \( 1/3 \) that in the environment (the difference between \( \partial V_c/\partial z \) and \( \partial V_c/\partial z \) determines \( V_R \) and thus \( \Delta P \), at upper and lower levels). In the cases studied, it turned out that the estimated difference between mean in-cloud and ambient wind velocities was around 10 m/sec in lower and upper parts of the cloud.

This proposition may be illustrated by use of observations provided by Malkus and Rome [1954]. Figure 2 shows a comparison of horizontal velocities of cloud towers, with winds at a nearby station. Cloud turrets penetrating a strong shear layer and entering the subtropical jet stream, had speeds 9-15 m/sec lower than the environment winds.

From data tabulated by Malkus and Rome, the following mean values are estimated for several towers in the layer 10-12 km: \( D = 8 \, \text{km}, \bar{w} = 6 \, \text{m/sec}, V_R = 12 \, \text{m/sec}, \rho = 0.4 \times 10^{-3} \, \text{gm/cm}^3, \partial V_c/\partial z = 3.5 \times 10^{-3} \, \text{sec}^{-1} \). At the appropriate Reynolds number, \( \Delta P \) is about 0.4 mb. Substitution of these values gives, for the first term on the right in (4), 1.6 cm/sec², and for the second term, -2.1 cm/sec².

Considering the crudity of our estimates from their data, this is good agreement, showing that vertical transfer of momentum can effectively offset the tendency for the cloud to be accelerated by outside forces. Furthermore, the close agreement suggests that the analogy between a vigorous cloud tower and a rigid obstacle (for which the laboratory measurements of \( P \) are valid) is not a bad one.

The expression for form drag used by Malkus and Scorer [1955] is physically similar to that used here. Analogous to (3), their expression would be \( dV_c/dt = KV^2 \). The experimental values quoted above give \( K \approx 1/(3R) \), \( R \) being the cloud radius. This value for the drag coefficient is several times smaller than that derived by Malkus and Scorer for cloud bubbles.

**Vertical pressure-gradient forces**—Maximum values of \( P \) on the upstream and downstream sides of an obstacle being about \( \rho V_e^2/2 \), relative motions of the order 10 m/sec give a total vertical decrement of \( P \), in a case like Figure 1, of about 1 mb.

In a typical large thunderstorm, there is a radial outflow in lower levels of order 10 m/sec, due to thermodynamic processes within the storm. When this is superimposed on the mean relative motions caused by momentum transfer, the total relative motion between 'in-cloud' and ambient winds on the right flank is double the amount estimated above. The vertical decrement of pressure can then be 1.5 to 2.5 mb. With \( \Delta P = 2 \, \text{mb} \) in a layer 500 mb deep, the vertical acceleration given by the last term of (2) is 4 cm/sec², comparable with the buoyancy acceleration if \( \Delta T = 1^\circ \text{C} \) averaged through the whole layer.

**Significance for triggering new convection**—Particularly because low-level outdrafts decay rapidly with height, most of the hydrodynamic pressure differential tends to be concentrated in the lowest 200 to 300 mb or less. According to (2), a 200-300-mb layer can be lifted until it is 2-3°C cooler than the undisturbed environment, if \( \Delta P = 2 \, \text{mb} \). In a typical mT air mass, the amount of lifting involved is enough to set off the existing potential instability.

Lesser lifting, such as provided by the downdraft outflow alone, can suffice to trigger new convection when the air mass is very unstable in lower levels, such as in mid-afternoon. The estimates of \( V_R \) and \( P \) above are characteristic
of spring situations wherein the vertical shear is strong. In summer, these values would be on the average considerably smaller; on the other hand, the mechanical lifting required is often correspondingly less. The significance of momentum transfer lies in the fact that it augments the relative motion due to outflow alone, and that the induced pressures are proportional to \( V_w^2 \). Thus, if a relative motion of only 5 m/sec because of momentum transfer with modest vertical shear is added to a relative motion of 10 m/sec due to the outflow field of a thunderstorm, the kinetic energy of relative motion, and the potential for lifting, is still more than doubled.

Effects of storm size—Equating the right-hand terms in (4) and noting that \( \Delta P \) is proportional to \( V_w^2 \), it is seen that for steady motion of the cloud

\[
V_w^2 = \left( \frac{\partial V_w}{\partial z} \right) D
\]

Since \( \partial V_w / \partial z = \partial V_E / \partial z - \partial V_c / \partial z \), for a given environment shear small \( \partial V_c / \partial z \) is concomitant with large \( V_w \) at upper and lower cloud levels. Proportionality (5) then states that for a given intensity of vertical motion and of vertical shear, large relative motions and induced pressures are enhanced by large storm diameter.

Thus propagation arising from shear is most favored when a rainstorm is of large size. This suggests a selective growth of large storms at the expense of small ones, which can be more readily sheared off without propagation of new cells on
their boundaries. This is consistent with the fact that a few large rainstorms tend to predominate (e.g., in squall lines) when the vertical shear is strong.

On the lower end of the scale, small trade Cumuli, as shown by Malkus [1949], acquire measurable but not great velocities relative to ambient winds. Induced pressures at their boundaries are small and probably have no appreciable effect on cloud growth. For this reason, clouds of the scale treated by Malkus do not propagate in the manner described above. Rather, they appear to grow on the upshear side because [Storer and Ludlam, 1953] successive towers rising from the same base appear upwind from older towers which have been carried off by the wind. Ackerman [1956] has shown observationally that shear inhibits production of rain in tropical Cumuli, but that rain may occur with proportionately larger shear when the buoyancy is increased.

**Vertical Shear and Hail**

The above discussion leads to the following conclusions possibly applicable to hail occurrence:

(a) With strong vertical shear and pronounced veering of wind with height, growth of new convection is most favored on the right flank of a rainstorm. Since young cells have most intense vertical motions [Byers and Braham, 1948], large hail should tend to occur predominantly on that flank. The hail track should have restricted width compared with the track of the rainstorm as a whole.

(b) The vertical pressure gradient induced by cloud-environment interactions favors, in the newly growing cells, upward accelerations stronger than provided by buoyancy forces associated with temperature anomaly alone.

(c) An additional factor favoring localized strong vertical motions with large hail growth, is that new updrafts growing on the downshear flank are sheltered by the main body of the rainstorm itself (cf. Fig. 1), from the decreased buoyancy which would result from entrainment of dry air in upper levels.

(d) The above effects should be most evident in vigorous rainstorms of large horizontal extent, and probably not noticeable in small storms.

**Examples**

Heavy hail damage occurs in a variety of circumstances, often in warm sectors or ahead of warm fronts, but more commonly behind cold fronts, in the geographical regions of most frequent occurrence [Harrison and Beckwith, 1951; Douglas and Hitchens, 1958]. Since we have been unable to find any detailed descriptions of large storms behind cold fronts, attention will be confined to two warm-sector situations where the shear was of the type shown in Figure 1.

There is little evidence of asymmetry of the kind described above, in the Alberta hailstorms described in great detail by Douglas and Hitchens [1958]. On the whole, those cases involved storms of small diameter, mostly in weak shear (or with vector shear nearly along the mean wind direction). In one example wherein there was significant veering of wind with height, the principal hail fall occurred with several cells forming successively toward right of the individual cell paths.

Figure 3 shows a set of observations collected by Harrison [1952], giving the precipitation distribution in the neighborhood of an incipient tornado (stippled track) at Fort Wayne, Indiana, near 19h30m CST on April 28, 1951. As shown by Figure 4, the heavy hail was concentrated on the south edge of a large rainstorm. This storm was located a short distance ahead of the cold

![Fig. 3—Hail and rain distribution, Fort Wayne, Indiana, near 19h30m CST April 28, 1951 [Harrison, 1952]](image-url)
front of a wave cyclone, with SSW-SW winds in the lowest few thousand feet, and winds from W in the upper troposphere. Although no upper-wind observation is available for Fort Wayne, the vertical shear vectors at surrounding stations (Fig. 5) clearly suggest that the hail occurred on the downshear flank of the rainstorm. Successive rainfall maps showed movement of the rain area as a whole toward ESE, somewhat to right of the upper winds.

The second example is shown in Figure 6 from Hamilton [1958], who gives a detailed description of the use of radar in tracking the storm and in identifying the hail. First echo appeared at A at 14h40m CST; hail was first apparent at B. Large hail (up to 2-inch diameter, some larger) and high winds occurred intermittently along a narrow track 130 mi long, ending near Bremond prob-

ably around 18h30m CST. A tornado occurred at Midlothian (C). The track of large hail was confined to the right side of the storm, which produced small hail along with the extensive heavy rain.

It is not possible to reconstruct the exact conditions in the neighborhood of the storm, but as far as can be determined, the Fort Worth winds of 12h00m CST (inset, Fig. 6) are nearly representative. By 18h00m CST, winds were W to WNW in lower levels. Hamilton quotes a radar report stating that echo movements (individual cells?) were from west, an agreement with the mean winds in the cloud layer in Figure 6, inset.

The Fort Worth hodograph, as well as the 850- and 300-mb wind charts (Fig. 7) shows strong veering with elevation. According to the earlier conclusions, pronounced building of new storms should be expected on the SSE side of the rainstorm at a given time, with marked deviation of the storm track to right of the mean winds. The movement of the rainstorm is thus qualitatively in agreement with expectations, although the deviation to right is stronger than might be expected.

Without a detailed analysis on the scale of the storm itself, it is not possible to give a reason for this excessive deviation. In an earlier study [Newton and Katz, 1958] it was found that rainstorms moved on the average 7 knots slower than, and 25° to right of, the 850-500 mb mean winds in the cases concerned. Individual storms, however, appeared to deviate up to 20-30 knots vector velocity from this mean behavior.

Such aberrations are to be expected in a phenomenon so complicated as a large thunderstorm, where many different processes are at work. The above discussion is obviously oversimplified; for example, a low-level outflow, assumed to radiate uniformly outward from storm center, has been added to a mean in-cloud velocity based on the ambient winds, in arriving at the relative velocity at a given point on the storm boundary [Newton and Newton, 1959].

Very likely other hydrodynamic effects are present which modify the simple conclusions here. Byers [1942] was the first to give a systematic description of the movement of thunderstorm paths to right of the winds. His observations suggested a cyclonic rotation within individual storms. Byers explained the storm movements on the basis of the rotor principle (a counterclockwise-rotating cylinder propels itself to right of the current in which it is imbedded).

On the basis of about 30 synoptic situations
EFFECTS OF VERTICAL WIND ON STORMS

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Fig. 6—Path of heavy hail- and rainstorm near Fort Worth, Texas, April 21, 1958, from Hamilton (1958); (inset) vector storm velocity compared with winds aloft.

analyzed, the hypothesis presented above appears to account for the gross behavior of rainstorms. That it does not satisfactorily account for all details is shown by the fact that in the second case above, as well as in several published radar observations of tornadoes, the most intense phenomena apparently occurred somewhat upwind, rather than on the direct right flanks of the storms concerned. Understanding of such phenomena must come eventually from increasingly detailed observations, interpreted not only in terms of ordinary thermodynamic processes, but also in the light of the interactions between storms and their environments.

Further remarks on other studies—At the time of writing the above, I was not aware of highly pertinent remarks in some other studies. Dessens [1959] (who offers a different explanation) states that “the existence of a jet stream or at least of a very strong wind aloft is the sole factor that we have been able to isolate as very probably responsible for the formation of destructive hail”; the relation to high winds aloft was also brought out by Wichmann [1951]. These studies emphasize that instability by itself does not completely account for large hail formation. Small or medium

Fig. 7—Winds in Fort Worth vicinity at 30,000 and 5000 ft, 12h00m CST April 21, 1958

A • First echo
B •
FORT WORTH (Radar)
MIDLOTHIAN
WACO (Radar)
BREMOND

0 10 20 30 40 50 MILES

N

25 M/s
26 M/s
22 M/s
FWH

30,000 FT

E

SURFACE FRONT

1530 C

L

FWH

5,000 FT

Hevy rain, small hail

Damaging winds, large hail

0.5 1 1.5 km
850 mb
3 5
50 knots
300 mb
7 km
50 knots

1 2 3 5

Fig. 6—Path of heavy hail- and rainstorm near Fort Worth, Texas, April 21, 1958, from Hamilton (1958); (inset) vector storm velocity compared with winds aloft.

analyzed, the hypothesis presented above appears to account for the gross behavior of rainstorms. That it does not satisfactorily account for all details is shown by the fact that in the second case above, as well as in several published radar observations of tornadoes, the most intense phenomena apparently occurred somewhat upwind, rather than on the direct right flanks of the storms concerned. Understanding of such phenomena must come eventually from increasingly detailed observations, interpreted not only in terms of ordinary thermodynamic processes, but also in the light of the interactions between storms and their environments.

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B •
FORT WORTH (Radar)
MIDLOTHIAN
WACO (Radar)
BREMOND

0 10 20 30 40 50 MILES

N

25 M/s
26 M/s
22 M/s
FWH

30,000 FT

E

SURFACE FRONT

1530 C

L

FWH

5,000 FT

Hevy rain, small hail

Damaging winds, large hail

0.5 1 1.5 km
850 mb
3 5
50 knots
300 mb
7 km
50 knots

1 2 3 5

Fig. 6—Path of heavy hail- and rainstorm near Fort Worth, Texas, April 21, 1958, from Hamilton (1958); (inset) vector storm velocity compared with winds aloft.
hail is common, but large hail is apparently favored by asymmetric storm structure, with concentration of the most violent convection in a restricted part of the storm as a whole. 

Weickmann [1953, pp. 108-109] has described a process which fits in very well with the notions above. He emphasizes that the persistence of an updraft column is augmented by the migration velocity of the storm. Weickmann likens a convective storm to a snowplow, sweeping up the unstable layer ahead of it in lower levels and throwing the air out in upper levels. The amount of air swept up, and ascending in the updraft, is augmented when the storm is impelled to move along through the air mass.

References
Ackerman, B., Buoyancy and precipitation in tropical Cumuli, J. Met., 13, 302-310, 1956.
Dessens, H., Severe hailstorms are associated with very intensive wind between 6000 and 12,000 meters, this publication, 333-336, 1960.
Harrison, H. T., Notes on certain tornado and squall line features, United Air Lines Met. Circ. 36, 21 pp., 1952.
Humphreys, W. J., Physics of the air, McGraw-Hill Book Co., 1940.
Newton, C. W., Structure and mechanism of the prefrontal squall line, J. Met., 7, 210-222, 1950.

Discussion
Dr. Roscoe R. Brahim, Jr.—I have had the advantage of having heard this lecture before, and I think you have left out in the interest of time, one point that needs to be brought out in the discussion of the model; namely, the consequence of the dynamic dam created by the vertical transport of momentum and of the hydrostatic pressures being built up on one side. This would lead to the development of new clouds, as Dr. Newton suggests, off to the right; and in fact I have seen some of his analyses of hourly rainfalls tracing out this kind of storm system. These hourly rain patterns move off to the right from the path of the storm as analyzed from radar records.

Mr. W. Boynton Beckwith—I would like to cast a vote for Dr. Newton's theories on the basis of radar. Although I would not like to generalize on all thunderstorm echoes, I would say three-fourths of the thunderstorms' echoes observed tend to bear out your ideas. Incidentally, we probably have in our ten years of record, ample rain-
fall and hail records which would give you added material to work on this theory.

Dr. Helmut Weickmann—In connection with your paper I would like to recall Finley’s famous analysis of 600 tornadoes. He found that they usually developed on the right side of the path of a big thunderstorm complex, and this would tend to agree with your model, since you also have the main cloud formation on the right side with respect to the path.

Mr. R. J. Donaldson, Jr.—I would like to add a little observational radar data to the theory. There is some evidence (see my paper in this volume) which indicates that the larger the hail size in thunderstorms, the more the hail is displaced to the right of the high-intensity echo core.

W. Boynton Beckwith

Meteorology Department, United Air Lines, Inc., Denver, Colorado

Abstract—Data collected during 225 hail days over the past ten years are investigated. The synoptic and thermodynamic conditions associated with hail development are analyzed and comparisons are made between storms of this area and those of the Middle West and Alberta. Upper wind distribution and the position of the tropopause are related to the development and growth of hailstorm-producing cells.

Introduction—Since 1949 a cooperative hail reporting network in and around the city of Denver has made possible the collection of sufficient detailed hailstorm data to make the beginnings of a climatology of this troublesome hydrometeor. Organized first to learn more about the true areal frequencies of hailstorms in this vicinity, and to establish new forecasting techniques on the basis of the various parameters that can be measured with a micro network, the project has been a continuing one since then. This paper is an extension of earlier studies [Harrison and Beckwith, 1950; Beckwith, 1956; Douglas and Beckwith, 1955] with respect to hailstorm characteristics and some thermodynamic relationships. Comparisons between the Colorado hailstorms and those of the Midwest and Canada are made on the basis of studies by Stout and others [1953], Douglas and Hitzfeld [1958], Faubush and Miller [1953] and others. Since 1953 an AVQ-10 radar has been used to monitor with a PPI display, some of the heavier hailstorm developments in the Denver area. Some results of this phase of the investigation have been published earlier [Harrison and Post, 1954; Beckwith, 1956]. The relation of the network to the Front Range of the Rockies is shown in Figure 1. The area covered by the cooperative network is shown in Figure 2. Site of the United Air Lines radar corresponds very closely to the position labeled USWB. The actual number of reporting stations shown in Figure 2 has varied from 12 in 1949 to 30 in 1951 to an average of 50 each year since that time.

Time, space, and frequency variations—It has frequently been pointed out that a more correct picture of the incidence of hail for a given area is afforded by network reporting than by point reporting. Hail frequency for the Denver area is shown in Table 1 and is expressed as a ratio to the official point reporting of the Weather Bureau. The Denver network encompasses about 150 sq mi. Even in such a small area as many as five hailstorms have been reported in one day. The 225 hail days shown in the ten-year period represent about 300 individual hailstorms.

Many earlier studies have expressed the frequency of hail occurrences in ratio to thunderstorm days. This is a useful relation for point summaries, but when area figures are summarized some bias is introduced if one reporting station is the basis for thunderstorm day counts. The hail-thunderstorm ratio of Table 2 has been subjected to this bias. Also, the year-to-year ratio has varied widely during the past ten years, a function of the change in hail patterns from one year to the next. An opinion expressed very often by forecasters concerned with the problems of the hail belt as well as by transport pilots frequenting this region is that nearly every thunderstorm developing off the Rocky Mountains contains hail in some stage of its development.

A relationship that has proven more valid than the hail-thunderstorm ratio during the past ten years is shown in Figure 3. For the period April through October, precipitation is almost entirely from shower activity, the exceptions being frontal and upslope developments which are of some importance in April and October. Since this precipitation is shower-generated, a good indicator for hail activity would be expected. The precipitation for 1957 which falls away sharply from an otherwise good scattering, reflects the two very wet months of April and May of this year which were characterized by a late heavy snow and a series of cloudbursts.

The sharp rise in hail activity in the Denver area in late spring is well illustrated in Figure 4. On the average, activity falls off more gradually
Fig. 1—Topographical features of area surrounding site of Denver hail study.

Fig. 2—Hail reporting station in Denver area; Denver rawinsonde release point is just east of U. S. Weather Bureau station.
after June when the frontal contributions to shower development decrease in potency. The investigations by Douglas and Hitschfeld [1958] in Alberta and Stout and others [1959] in Illinois suggest that peaking of hail activity is somewhat later in the season in Canada and at about the same time in Illinois. However, it should be noted from Figure 4 that peaking of hail days is spread quite broadly on a year-to-year analysis.

Time of initial fall of hail in the network portrayed in Figure 5 shows a peaking of about a third of the reports between 14h 00m and 16h 00 MST. Two-thirds of the time, hail fall commences in the five-hour period between 13h 00m and 18h 00 MST. This pattern demonstrates the powerful effects of the thermals generated in the mountains to the west of the network. If the broad scale upslope conditions which are associated with half of the hail developments in this study were the major lifting force in triggering hail generating thunderstorms, we might look for a flatter distribution curve. It is well recognized by forecasters, for example, that the general upslope circulation which produces nearly all of the cold weather snows, is equally effective at night as during the time of day when thermal effects are felt.

**Hail size**—The frequency distribution of hailstone size is portrayed in Figure 6. Hailstones of one-inch diameter can inflict considerable damage not only at the ground, but also to aircraft in flight. It is to be noted that an average of seven reports of stones of this size or larger are collected in the Denver network each year. Of more importance is the consideration of the greater frequency of occurrence of these larger stones at aircraft operating altitudes before the effect of melt or sublimation has taken over in the fall from point of origin to the surface.

Geographically, hail size frequency at the northern end of the hail belt in Canada bears a close relation to the figures, according to data compiled by Douglas and Beckwith [1958]. In the Middle West, the smaller size stones appear with greater frequency, judging from figures published by Stout and others [1959]. The seasonal change in hail size expectancy is shown in Figure 7, depicting Denver and Alberta [Douglas and Beckwith, 1958] experience. It will be noted in general that in months of relatively low levels of activity, maximum stone size is small.

Another relationship between hail size and activity level which has been borne out in the

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**Table 1**—Area frequency versus point frequency of hail

<table>
<thead>
<tr>
<th>Year</th>
<th>Network hail days</th>
<th>Official hail days</th>
<th>Area to point ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>1940</td>
<td>33</td>
<td>4</td>
<td>8:1</td>
</tr>
<tr>
<td>1950</td>
<td>23</td>
<td>3</td>
<td>8:1</td>
</tr>
<tr>
<td>1951</td>
<td>34</td>
<td>9</td>
<td>4:1</td>
</tr>
<tr>
<td>1952</td>
<td>16</td>
<td>2</td>
<td>8:1</td>
</tr>
<tr>
<td>1953</td>
<td>21</td>
<td>7</td>
<td>3:1</td>
</tr>
<tr>
<td>1954</td>
<td>9</td>
<td>4</td>
<td>2:1</td>
</tr>
<tr>
<td>1955</td>
<td>26</td>
<td>9</td>
<td>3:1</td>
</tr>
<tr>
<td>1956</td>
<td>13</td>
<td>3</td>
<td>4:1</td>
</tr>
<tr>
<td>1957</td>
<td>25</td>
<td>6</td>
<td>4:1</td>
</tr>
<tr>
<td>1958</td>
<td>25</td>
<td>4</td>
<td>6:1</td>
</tr>
<tr>
<td>Totals</td>
<td>225</td>
<td>51</td>
<td>4:3:1</td>
</tr>
</tbody>
</table>

* Ten year average.

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**Table 2**—Hail-thunderstorm ratio

<table>
<thead>
<tr>
<th>Item</th>
<th>Total hail days</th>
<th>Total thunderstorm days</th>
<th>Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>10-year unofficial area</td>
<td>225</td>
<td>391</td>
<td>1:1.7</td>
</tr>
<tr>
<td>10-year official point</td>
<td>51</td>
<td>391</td>
<td>1:8</td>
</tr>
<tr>
<td>50-year official point</td>
<td>4</td>
<td>43</td>
<td>1:11</td>
</tr>
</tbody>
</table>
Denver area, is an obvious feature of individual hailstorms crossing the network during weak hail years. During 1952, 1954, and 1956, years which were relatively inactive, hail-wise (see Table 1), patterns of fall were marked by spotty occurrences instead of well defined or extensive swaths.

**Synoptic and RAOB patterns**—The 1956-1958 performance of hailstorms in the Denver area bear out the facts summarized in our earlier study [Beckwith, 1956] as follows:

1. About half of the hail developments were associated with cold frontal passages within the previous 6 to 36 hrs.

2. Wind direction on hail days at 500 mb favors WSW.

3. In only four per cent of the hail cases were cold lows found in or near Colorado.

4. Crossings of jet streams within 100 mi of the network occurred on about 15% of the hail days.

5. Surface temperature just prior to onset of hail averaged 71°F. Values ranged for the ten years from 40°F to 98°F.

6. The average surface dew point before pre-
that surface dew points for Middle West storms are nearly always in the 60's and 70's.

A preliminary survey of the Denver RAOBs for 18 hail days selected at random during 1956, 1957, and 1958, shows a temperature and moisture relationship in the vertical as illustrated in the mean sounding of Figure 9 and in the data below:

- Freezing level: 590 mb
- Wet bulb freezing level: 637 mb
- Temperature at LCL: 5°C
- (lifting condensation level): 134 mb*

*Approximately 45,000 ft

In comparing these figures with the mean hailstorm soundings computed by Fawbush and Miller [1953], a similarity is found from 700 mb to 400 mb for stone diameters between one and two inches. The greatest discrepancy lies in the 5000-ft layer above the surface. The High Plains moisture distribution is marked by a low humidity at the surface, increasing gradually to about 500 mb and decreasing gradually above this level. The inversion above the moist layer noted by them as a pattern for stones of one inch diameter or larger is also missing in the mean Denver sounding. However, the 0°C wet-bulb temperature is reached at about 7500 ft above the terrain, again in agreement with a consistent parameter for the larger stones. Extending the
Fig. 9—Mean sounding for 18 hail days in 1956, 1957, and 1958; solid line is temperature distribution, broken line is dew point curve, and dash-dot line is the parcel adiabat; mean tropopause height is at 134 mb (approximately 45,000 ft).

Investigation to the tropopause, it is found that the mean lapse rate of the 18 soundings is steep enough to produce a positive energy area from the LCL to about 35,000 ft.

Soundings published by Douglas and Beckwith [1958] for some of the larger hail situations occurring in Alberta in 1957 are mainly in agreement with this Denver mean except for the tropopause altitude which is lower in Canada in accordance with the normal latitudinal difference.

Although no firm conclusions may be drawn on this limited sampling, it appears that on days with hail, the height of the tropopause at Denver is, on the average, lower than on non-hail days.

Other work—In the attempt to develop new methods of analysis of hailstorms and in reviewing the work of others in this field, one cannot overlook the strong suggestion that the make-up of thunderstorms in general and hailstorms in particular, involves an important ingredient not recognized today. Whether or not this involves space charges, for example, or unknowns in cloud nuclei is something for the cloud physicists to determine.

As of today, the hailstorm remains as a problem for the farmer and for the property owner. In aviation, hail is less of a hazard now, thanks to airborne radar. But the forecaster must continue to be alert for these developments as aircraft now flying near the tropopause appear to be exposed to an even higher proportion of large hailstones than experience up to now would dictate.

Acknowledgments—The author again wishes to express his gratitude to the corps of unofficial observers who have cooperated since 1949 in the reporting of hailstorms in the Denver area, and without whose help this project would not have been possible.

References


Discussion

(Note: Discussion of this paper is combined with those of the two following papers at the end of the second following paper.)
Some Behavior Patterns of New England Hailstorms

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AND

CHARLES REEVE SHACKFORD

Allied Research Associates, Inc., Boston, Massachusetts

Abstract—Radar measurements of New England thunderstorms have been combined with reports furnished by cooperative observers during three years. Echo tops in storms releasing hail of less than 3/4-inch diameter were higher, colder, and penetrated the tropopause more often than tops of rain thunderstorms. The differences are even more striking for storms with large hail (3/4-inch or larger) compared with the other two categories. Extreme tropopause penetrations of 10,000 to 15,000 ft occurred on five days, of which all but one were tornado days.

A study was made of the histories of 20 hailstorms. Ten of them dropped hail for long periods of time (hail repeaters), the others for less than 20 min and at only one or two locations. All cases of severe damaging winds and tornadoes occurred with the hail repeaters. Within a wide scatter, both echo tops and maximum intensities in the hail repeaters attained higher peaks and remained high for longer periods of time than in the single hail producers. Hail locations in all storms exhibited a slight tendency to appear in the right, rear quadrant of the storm, with the larger hail sizes located to the right of the smaller hail, facing downstream.

Echo areas of various intensities as a function of height were measured in two hailstorms, one just getting under way and the other a well-developed producer of several tornadoes. Computations are made of hail-mass concentration versus height at various times in the two storms, assuming the radar echo is scattered from 1-cm ice spheres. Mean 1-cm hail concentrations in the tornado storm, averaged over the total echo area at the height of the most intense echo, varied from 0.3 to 1.3 g/m², but the maximum concentrations in the echo core ranged from 9 to 170 g/m², subject to a possible overestimate by a factor of 5 due to an unresolved radar calibration error. During two observations of echo tops penetrating the tropopause, the echo volume above the tropopause increased rapidly with time, suggesting a progressive modification of the lower stratosphere above the storm.

Hailstone information furnished by cooperative observers is summarized. Median values are: maximum diameter, 7 mm; max/min size ratio in a hailfall, 2; number concentration, 0.1/m³; hailfall duration, 3-4 minutes; hailfall started four minutes after heavy rain began. About 75% of hail shapes were divided between spheres and oblates.

Introduction—Thunderstorms have been studied in southern New England during 1956 through 1958 by a combination of CPS-9 radar observations taken from Blue Hill, Milton, Massachusetts, and storm reports received from a network of cooperating observers. This paper is a description of some of the characteristics of hailstorms observed during the three years of the investigation.

The CPS-9 radar has a 1° conical beam which can be set at any desired elevation angle. A series of antenna rotations about a vertical axis (PPI scan) with successive increases in elevation angle causes the beam to sweep out a volume defined by a cone of revolution with vertex centered at the radar position. All parts of thunderstorms (except those within 25 mi) can be observed conveniently by this means. The height of an echo volume is given by its range and elevation angle. The heights of echoes within 25 mi were usually measured by means of an RHI scan (a rapid vertical slice from near-zenith to horizon, at fixed azimuth). Echo intensity can be determined by means of a calibrated receiver gain control. Er-
rors in echo height and intensity measurement have been discussed by Donaldson [1959].

**Echo top heights and temperatures**—All thunderstorm echo top heights measured during 1956 and 1957, plus a few from 1958, were classified according to the largest hail size reported within ten minutes before or after the radar measurements (±15 minutes for the 1956 data) and within ten miles of the location of the echo top. The heights were grouped into three precipitation categories: no hail (rain only), hail diameter less than ½ inch, and hail diameter equal to or greater than ½ inch. If an echo top measurement was associated with several reports of rain without hail and only one report of one-inch hail, it was put into the larger hail category.

Echo top temperatures were read off the Albany, New York, radiosonde nearest in time to the echo height measurement. The tropopause height was also determined, where possible, selecting the height above the 300-mb level where the temperature lapse rate first became less than 2°C/1000 ft. The distance by which the echo tops exceeded or fell short of the tropopause was recorded. For echo tops penetrating the tropopause, an estimate was made of the negative energy required for penetration by measuring the area bounded by the sounding, the dry adiabat rising from the tropopause (very nearly equivalent to the moist adiabat at stratospheric pressures), and the pressure surface at echo top.

The results are shown in Figures 1 to 4, which show the cumulative frequency distributions of echo tops associated with each of the three precipitation categories with respect to height, temperature, penetration above (or below) the tropopause, and negative energy required for stratospheric penetration. Table 1 summarizes the upper quartile, median, and lower quartile values (in descending order) of these four parameters for the three precipitation categories.

The differences between the ‘all rain’ category and the larger hail category are quite marked, even considering the several sources of error. These errors include the uncertainty as to hail occurrence and maximum hail size because of holes in the observing network almost certainly larger than significant variations in the character of thunderstorm precipitation. Thus, the proximity of the ‘all rain’ and smaller hail size curves on all four diagrams may be partly attributable to unobserved (or unreported) small hail that falls
near an echo top measurement thought to be in the 'no hail' category.

Another error is inherent in the timing of the ground events relative to the echo-top determination. Douglas [1959] associated echo tops with the appearance of hail or rain 20 min later at the ground, allowing time for the precipitation to fall. The timing used in the present study is a rough attempt to relate hail production to convective forces that result in the increase or maintenance of echo-top height. The optimum timing relationship probably lies somewhere between those adopted by the two investigations.

Other errors are randomly distributed. They include echo-top height measurement, with a mean error estimated to be about ±2000 ft; uncertainties in a few of the tropopause height determinations; and deviations in time and space (up to six hours and 150 mi) of the Albany radiosonde from conditions at an echo-top measurement. Hopefully, the sum of these random errors should tend to zero in a distribution of many cases.

The few special cases of extreme penetration are interesting. Echo tops pushed 10,000 to 15,000 ft above the tropopause on five days during the three years studied (two complete years and part of a third). Tornadoes occurred on four of these five days. The maximum penetrations of the particular tornado storms were 8, 10, 13, and 15 k ft. On the fifth day widespread hail occurred with extremely severe electrical storms (45 houses were struck by lightning in one town alone); the maximum penetration on this day, 12 k ft, was related with the fall of one-inch hailstones. The negative energies overcome by the tornado storms ranged from $1.6 \times 10^8$ to $3.0 \times 10^9$ ergs/gm, or 2½ to 5 times the median value for hailstorms with $\frac{3}{4}$-inch or larger hail. The severe lightning storm with one-inch hail had to overcome the greatest negative energy found during the entire study, $3.4 \times 10^9$ ergs/gm. These extreme cases

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**Fig. 3**—Cumulative frequency distribution of thunderstorm echo top penetrations above and below the tropopause.

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**Fig. 4**—Cumulative frequency distribution of negative energy required for thunderstorm echo top penetration above the tropopause.
Table 1—Thunderstorm echo top characteristics

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Maximum hail size</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>None (all rain)</td>
</tr>
<tr>
<td>Height (K ft)</td>
<td>42</td>
</tr>
<tr>
<td></td>
<td>37</td>
</tr>
<tr>
<td></td>
<td>30</td>
</tr>
<tr>
<td>Temperature (°C)</td>
<td>-60</td>
</tr>
<tr>
<td></td>
<td>-51</td>
</tr>
<tr>
<td></td>
<td>-38</td>
</tr>
<tr>
<td>Height above (+) or below (-) tropopause (K ft)</td>
<td>+2</td>
</tr>
<tr>
<td></td>
<td>-4</td>
</tr>
<tr>
<td></td>
<td>-11</td>
</tr>
<tr>
<td>Negative energy required for tropopause penetration (ergs/gm of air)</td>
<td>$9 \times 10^3$</td>
</tr>
<tr>
<td></td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>0</td>
</tr>
</tbody>
</table>

approach the conditions listed by Vonnegut and Moore [1958] for their category of 'giant' thunderstorms. Their simplified assumption that the updraft speed at the tropopause equals 20 m/sec for each kilometer of penetration shows good agreement with a sample calculation. The required updrafts are incredibly high for these few extremely penetrative storms. However, it is suggested later (under Hailstorm echo areas) that the first penetration of the tropopause initiates a modification of the lower stratosphere, and hence the unreasonably high requirement for updraft speeds is considerably reduced.

Composite hailstorm histories—Twenty hailstorms depositing hail one-half inch or larger in diameter were selected from case studies conducted during the three years for examination of their common features, if any. A storm is defined as a unique radar echo, in most cases entirely separate from other echoes at maximum radar receiver gain and 0° antenna elevation angle. Occasionally, the weaker, lower parts of a storm will merge with another storm, but the boundary becomes clear at slightly reduced gain settings and especially at moderate heights (for example, above the melting level). A storm may have one to five or six well-defined echo cores which become visible from the surrounding background as the receiver gain is lowered. The cores, which are very likely related to convective cells, generally have a much shorter lifetime than the total storm duration of two to six hours.

Ten of the storms produced hail during a single period of less than 20 min, reported in every case but two from a single location. The others, named 'hail repeaters,' dropped hail for periods of up to 35 min continuously and up to 140 min intermittently. This classification of storm types may be analogous to that found by Douglas and Hitchens [1958] in Alberta hailstorms. They found that the hail usually fell in short bursts but on several occasions there were continuous hail falls of 1½ hr in a swath 30 mi long. If the New England hail frequencies were increased by closer observer spacing (comparable to the tight Alberta network) and by accounting for overall regional differences in hail probability, some of our hail repeater storms might fill in to form continuous hail swaths.

One further fact emerged: the single-hail producers included no instance of severe wind damage (uprooting or fracturing of trees, structural damage to buildings, etc.). However, of the ten hail repeater storms, severe wind damage was reported in seven of them, including three storms with several tornadoes each and another one with two funnel clouds aloft. Also, the maximum hail size was only ½ inch in four of the ten single-hail producers; hail of ½ inch diameter or larger was reported in all of the hail repeater storms. Incidentally, the maximum hail size in the repeaters appeared 10 to 100 min after the time of first hail, with a median separation of 40 min.

In the analysis that follows, all times are given with respect to the first appearance of hail. This may be a dangerous thing to do because of the probable incompleteness of hail reports. Of course, the storm classification scheme is subject to the same criticism. However, for all its uncertainty, it seemed better to relate events to time
of first hail than to time of first echo, since many first echoes were not observed at all in our schedule of radar observations and only nine of them could be stated within an accuracy of five minutes and 13 within 11 min.

Figure 5 depicts the hail and severe damaging windstorm events, including tornadoes, during the time following first hail in the hail repeaters. Note the tendency toward a 100- to 120-min periodicity in tornado activity and occurrence of large hail. This is an integrated picture; the seven individual storms in which such changes are observed show a periodicity of 30 to 70 min with, at most, three periods apparent.

Figure 6 shows the time variation of maximum echo height in all 20 storms. The first echoes are arbitrarily started at a height of 15,000 ft with a maximum rate of height increase of 2000 ft/min. There is a wide variation in time between first echo and first hail (14 to more than 162 min), in maximum height reached by an echo (27,000 to 56,000 ft), and in total height interval covered by the echo tops (always greater than 20,000 ft). Wide variety in echo-top behavior is the rule. Nevertheless, some general trends superimposed on these large variations are brought out by Figure 7, which shows the median and maximum echo-top history of the two storm groups.

The echo tops of hail repeaters differ from those of the single-hail producers in the following particulars: They are rising when the first hail falls, they maintain extreme heights for greater periods of time, they attain a higher over-all maximum, and they have a longer lifetime. These characteristics are similar in certain respects to those discovered by Douglas and Hitzeffeld in their sustained hail producers.

The most intense radar echo found anywhere in the storm was observed at various times during the life cycle of 17 of the hailstorms, in the manner described by Donaldson [1958]. The height of the most intense part of these echoes ranged from near the surface up to 31,000 ft, with half of the observations centered between 13,000 and 24,000 ft. At least part of the reason for the surprisingly high altitude of the most intense radar echo in hailstorms is due to the high attenuation in heavy rain of the 3.2-cm waves radiated by the CPS-9 radar. Attenuation by ice is much lower than by the same mass of rain. Other possible causes include high particle concentrations accumulated near the maximum of a non-steady-state updraft.
BEHAVIOR PATTERNS OF NEW ENGLAND HAILSTORMS

Fig. 6—The trends of echo top height in 20 hailstorms, related to time before or after first hail; the numbers identify the various storms; storms 1, 7, and 19 each produced several tornadoes.

Fig. 7—The trends of median and maximum echo top height for the two groups of hailstorms, related to time before or after first hail.

and the attrition of falling hailstone size by melting and break-up.

The maximum echo intensities are expressed as the common logarithm of the maximum value of the equivalent radar reflectivity factor $Z$. The equivalent $Z$ is defined as $\Sigma ND^6$ (or volume concentration of the sixth power of drop diameters, in mm$^6$/m$^3$) of small spherical water droplets which would return the same radar echo intensity, at the same range, as the measured echo intensity. The water droplets must be small enough to permit the use of the Rayleigh scattering approximation, or about 2 mm or smaller in diameter for a radar wavelength of 3.2 cm. Further discussion of the meteorological interpretation of equivalent $Z$ is given by Donaldson [1958].

Hail returns a considerably lower echo power than would be indicated by computing the $Z$ of
an assortment of hailstones, for three reasons: (1) the dielectric backscattering factor is lower in ice of unit density than in water by a factor of 4 or 5; (2) hailstone densities may be as low as 0.5 and, of course, are always less than the density of water; and (3) for 3.2-cm radar practically all observable hailstones are outside the Rayleigh scattering region and begin the transition through the complicated Mie region toward geometrical scattering which is approximately proportional to the square instead of the sixth power of particle diameter.

Figure 8 shows the time variation of log $Z_{\text{max}}$ in the 17 storms in which this was measured. The right-hand ordinate gives the maximum ice content, assuming the maximum echo is scattered from unit density ice spheres of 1-cm diameter. Note the rapid growth in Storm 10, in which $Z_{\text{max}}$ measurements were made near the time of first echo. The maximum and median $Z_{\text{max}}$ trends for the two classes of storms are illustrated in Figure 9. Similar to the echo top situation, the hail repeaters attain a higher maximum and at a later time.

Radar echo maximum heights and maximum intensities appear to be conveniently observable indicators of the convective activity within a storm. Although the two parameters are loosely correlated, they reflect somewhat different aspects of the convective situation, the heights indicating more about updraft speeds and persistence, and the intensities perhaps more about moisture supply and the efficiency of converting the available moisture to large hailstones. A convective index, the product of $H_{\text{max}}$ and log $Z_{\text{max}}$, was plotted for the median trends of the two storm groups (Fig. 10). $H_{\text{max}}$ is expressed in k ft and $Z_{\text{max}}$ in mm$^3$/m$^3$. The median convective index change with time was also plotted for the three tornado-producing storms, with each tornado storm maximum plotted separately. The diagram also shows the range of convective index calculated from the medians and quartiles of two populations of hailstorms and rain-thunder-storms (the $H_{\text{max}}$ population includes the $Z_{\text{max}}$ population and many more cases besides).

Figure 10 is intended to be provocative rather than definitive. Several general features are worth mentioning: (1) the more severe the storm class, the higher and later the peak of convective index; (2) the close association in time (within the scale of time resolution in the original observations) between the maximum convective index in a tornado storm and the time of the most destructive tornado; and (3) the rapid rise but slow
Fig. 9—The trends of median and maximum log $Z_{\text{max}}$ for the two groups of hailstorms, related to time before or after first hail.

**CONVECTIVE INDEX:** $H_{\text{max}} (K \, \text{ft}) \times \log Z_{\text{max}}$

Fig. 10—Median convective index trends in hailstorms with less than 20 minutes of hail, hail repeater storms, and tornado-producing hailstorms; the convective index maximum and time of most destructive tornado are shown separately for each tornado storm; all times are before or after first hail; at left, medians and quartiles of a large number of rain-thunderstorms and hailstorms are plotted, independent of timing.

decay of convective index for the median hail repeater.

The position of the leading edge of hail was determined relative to the low-altitude echo core in 25 cases where both the timing and location of the hail occurrence were considered accurate within limits of one minute and one mile. The position of the low-altitude echo core was interpolated from the available measurements, generally taken two to five times during an hour. The relative hail locations, plotted in Figure 11, seem to have quite a scatter, but closer examination shows that half of all hail and three-fourths of hail larger than 1/2-inch falls within three miles of the echo core.

There were many hail occurrences for which
Fig. 11—Position of leading edge of hail relative to low-altitude echo core

Table 2—Mean hail locations relative to low altitude echo core

<table>
<thead>
<tr>
<th>Maximum hail diameter</th>
<th>Normal to echo core path</th>
<th>Along echo core path</th>
</tr>
</thead>
<tbody>
<tr>
<td>D ≤ 1/4</td>
<td>1.2 mi to left (facing downstream)</td>
<td>1.7 mi behind</td>
</tr>
<tr>
<td>1/4 &lt; D ≤ 1/2</td>
<td>0 mi (on path)</td>
<td>1.2 mi ahead</td>
</tr>
<tr>
<td>D &gt; 1/2</td>
<td>0.9 mi to right</td>
<td>0.4 mi behind</td>
</tr>
<tr>
<td>All hail</td>
<td>0.48 mi to right (84 cases)</td>
<td>0.16 mi ahead (25 cases)</td>
</tr>
</tbody>
</table>

the timing was in doubt but the location exact; for these cases the hail position could be determined normal to the path of the echo core but not along the path. There was similar scatter in position, but in over half of these cases the hail fell two miles or less to the right or left of the echo core path, or exactly on the path. Table 2 summarizes the mean hail locations for various hail sizes. It shows a progressive tendency for the larger hail to be located toward the right-hand side of the core path, in support of the model proposed by Newton [1959]. Since these locations are for the forward edge of the hailfall, which has a median duration of three to four minutes (see later under Characteristics of New England hail), the hail path tends to lag the echo core. Thus, the right rear quadrant of the storm (with respect to the echo core) shows a weak tendency to claim the greater part of hail.

Hailstorm echo areas—The echo areas at various receiver-gain settings and elevation angles were traced in two of the storms, one in which a first echo was detected (Storm 10) and the other, Storm 19, a producer of several tornadoes. Figure 12 is an example for Storm 10, for three gain settings.

The areas enclosed by various values of equivalent reflectivity factor Z, as a function of height, were computed for both storms at different times in their history. The areal heights were obtained simply from the height of the echo core at each antenna elevation angle. Similarly, the range of the echo core at each height was adopted as an average range for purposes of computing the
equivalent $Z$ of all of the areas at the several gain settings. As many as eight gain settings were used to minimize interpolation errors.

Some of the results are shown in Figures 13 and 14. The solid curves in Figure 13 relate to the areas of Figure 12; the observation started at 15h15m and ended at 16h22m. The dotted curve is the first echo, measured between 15h35m and 15h40m. The growth both in area and in intensity is almost explosive. Note the two regions of maximum intensity in both observations. The lower one rose at the rate of 550 ft/min, the upper one at 620 ft/min, assuming the corresponding regions to be related. The lower maximum rose through the 0°C level at approximately 13,000 ft. The upper maximum rose to a level where the temperature was $-39°C$.

Two of the five area-height measurements of...
the tornado storm are depicted in Figure 14. The areas covered by this well-developed storm are much larger than those found in the previous case. During the development of the storm the maximum equivalent Z aloft increased, while the areas covering the lesser values of equivalent Z increased markedly near the surface but decreased somewhat in the upper part of the storm. The areas were used to estimate the precipitated water of hailstone size as a function of height in the storm. For the purpose of this computation, all the precipitated water that contributes appreciably to the radar echo was assumed to consist of ice spheres of unit density and diameter 1 cm. From Ryde’s [1946] scattering curve for ice, a concentration of 1/m³ of 1-cm ice would give an equivalent Z of 10^6 mm²/m² for a radar wavelength of 3.2 cm. (An equal concentration of 7-mm water drops would have a slightly larger equivalent Z.) An integration of area by Z at various altitudes in the storm was interpreted in terms of grams of ice per meter thickness of the storm. The results are plotted in Figure 15 for the five times during which areas were obtained during the life cycle of the tornado-producing hailstorm of July 11, 1958. (The areas of two of these times were shown in the previous figure.) The early echo (but not first echo) case of July 30, 1957 is included for comparison; the areas of this storm were the solid lines of Figure 13.

Figure 15 shows a large development at all altitudes in the mass content of large particles during the half hour preceding the first tornado, which touched down at about 15h30m. During the next two hours there is little further development except a redistribution of some mass from high altitudes to medium altitudes.

The mean concentration of large particles in this storm, averaged over the total area of the radar echo at the altitude containing the greatest mass, ranged from about 0.3 to 1.3 g/m³. (In the ‘early echo’ storm, July 30, 1957, 15h45m+, the highest mean concentration was only 0.017 g/m³.) These values, somewhat lower than those mentioned by Weickmann [1953], suggest that the large (1 cm) particles, which give the strongest radar echo, contribute only a small fraction of the total water substance. On the other hand, the extremely high values of radar reflectivity in the small echo cores aloft indicate mass concentrations of 9, 19, 35, 170, and 125 g/m³ in the five measurements. Thus, the distribution of large particles is extremely spotty, with a surprisingly large concentration in the echo core, falling off to low concentrations and probably smaller maximum sizes as the storm periphery is approached.

A word of caution is advised in interpretation of these water-content estimates. First, all calibrations of weather radars reveal a systematic departure of the echo intensity by a factor of 2 to 5, approximately, below the value expected on theoretical grounds. Corrections have been made for this factor. If the factor is somehow related to the calibration scheme but is not operative in meteorological observations, then the water concentrations have been considerably overestimated. Secondly, attenuation by thunderstorm rain and water-coated hail, which is not accounted for here, leads to an underestimate of the water concentrations by an unknown amount. Finally, the assumption adopted regarding particle size and state has a marked effect on the water concentration capable of giving the same intensity of radar echo. For example, the same radar echo

![Fig. 15](image_url) - The relationship of total precipitated water mass per unit height to height, during five measurements in the tornado storm of July 11, 1958 and one in the growing hailstorm of July 30, 1957.
attributed to 1-cm hail would be returned by three times the mass concentration of 3-mm water drops, about 1/3 the mass concentration of 4-cm hail, and somewhere between 1/3 and 1/4 of the mass concentration of a 1-cm particle composed of a mixture of ice and water in equal parts.

With these limitations in mind, the total mass of precipitation, assuming it to consist entirely of 1-cm hail, was added up for the two storms under consideration and is listed in Table 3.

Echo areas near and above the tropopause were measured with reasonable accuracy at two times in the tornado storm of July 11, 1958. In both cases the areas above the tropopause decreased with height, at first in a linear fashion and then more rapidly near the echo top. However, the second measurement, about one-half hour later than the first, showed much larger echo areas penetrating above the tropopause level. The echo volume in the stratosphere increased more than five times, and the energy required for penetration increased by more than a factor of three (see Table 4). The penetration energies were computed in two steps because the specific negative energy from the tropopause at 38,000 ft up to 44,000 ft was relatively small, but increased rapidly above 44,000 ft in a strong temperature inversion.

Malkus [1959] has demonstrated the critical role of element size in the penetration of Cumulonimbus towers; the lower entrainment rate for the large element size inhibits the dilution of the 'protected core.' The echo areas of this tornado storm at the tropopause are much larger than the element areas reported by Malkus for clouds reaching 50,000 ft in Hurricane Daisy of 1958. However, the penetrations above the tropopause were somewhat greater in the case of the tornado storm.

The rapid increase in echo volume above the tropopause and in the computed total energy required for penetration of the tropopause, along with a slight decrease in echo top height, suggests a progressive modification of the lower stratosphere in such a manner that the actual penetration energy does not increase so markedly with time, or perhaps even decreases. The mixing of the early storm tops with the lower stratosphere would moisten and cool the region so that subsequent penetrations would require less energy. The vertical exchange of momentum supported by the strong updrafts necessary for penetration of the tropopause by particles large enough to give a radar echo would tend to maintain and enhance the modified region above the main body of the storm. In effect, the tropopause would seem to be bulging upward in the vicinity of the storm, though the authors have no knowledge of direct measurements of temperature or humidity which would confirm or deny this picture.

Characteristics of New England hail—During
three summers, 317 hail reports were received with one or more useful bits of information about the hailstones or their timing. The most frequent maximum size reported (Fig. 16) is \( \frac{1}{4} \) inch. The median is 7 mm, and less than five per cent of the reported hail equals or exceeds one inch in diameter (see Fig. 17). Hail sizes in New England are considerably smaller than those found to the east of the Rockies by Beckwith [1959] and by Douglas and Hitscheid [1958], but are comparable to those observed in Illinois by Stout and others [1959].

Information on size distribution within a given hailfall was received from a limited number of observers who reported maximum and 'average,' or maximum and minimum sizes (Fig. 18). The representative size distribution is narrow. A typical max/min size ratio is 2, with nearly 90% of all distributions within a spread of 3. One should expect the size range to be relatively narrow at the ground because of the greater melting rate of small stones during their fall.

Many observers reported hail durations (Fig. 19), but a certain favoritism was evident for multiples of five minutes. A three-minute running mean smoothed out this periodicity somewhat. The median hailfall duration is three to four minutes. The five-minute periodicity was also a strong feature of the timing relationship between the beginning of hail and heavy rain (Fig. 20). In a very few cases hail began falling before the heavy rain, but in the median case
hail started falling four minutes after the heavy rain began.

Hailstone shapes were reported only infrequently, and are summarized in Table 5. Almost 75% of the reported shapes are about evenly divided between spheres and oblate ellipsoids.

Some of the observers reported the number of stones left on the ground per unit area, as well as size and duration of fall. Using Ludlam's [1958] formula for fall speed and his estimates of drag coefficient, the number concentration of hailstones was found. These are spotted on Figure 21 as a function of diameter. Apparently the two parameters are related only slightly. The median concentration decreases from 0.14 to 0.1 to 0.07/m² in the 5–10, 10–15, and 15–20 mm diameter size groups. A few mass concentrations are included. It is interesting to note that the maximum value, 6 g/m³, agrees with Ludlam's statement that the concentration of water in the form of hailstones has a maximum value of several g/m³.

General conclusions—New England hailstorms exhibit a wide variety of characteristics. Some of these features have been presented here, derived from radar measurements and the observations of a network of cooperating observers. The most surprising result, and the least reliable one, is the derivation, from the intensity of the maximum radar echo, of extremely high water concentrations in the cores of many hailstorms at medium altitudes. Even if a possible calibration error of a factor of five is taken into account, however, this maximum concentration reaches a value of 34 g/m³ at a height of 23,000 ft in a tornado-producing hailstorm.

Perhaps the most interesting result concerns the penetration of the tropopause by many storm echoes. The median rain-thunderstorm echo top falls short of the tropopause by 4000 ft and the median hailstorm with hail less than 3/4 inch in diameter has its echo top coincident with the tropopause. However, the median hailstorm having hail ¾ inch or larger penetrates 5000 ft into the stratosphere, and ⅓ of these storms penetrate 9000 ft or more. Extreme penetrations of 10,000 to 15,000 ft occurred on five days; four of these were tornado days, with the tornado storms heavily involved. The increase of echo

<table>
<thead>
<tr>
<th>Characteristic shape</th>
<th>Number of reports</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oblate</td>
<td>23</td>
</tr>
<tr>
<td>Spherical</td>
<td>22</td>
</tr>
<tr>
<td>Irregular</td>
<td>10</td>
</tr>
<tr>
<td>Conical</td>
<td>4</td>
</tr>
<tr>
<td>Cylindrical</td>
<td>1</td>
</tr>
<tr>
<td>Prolate</td>
<td>1</td>
</tr>
</tbody>
</table>

Ratio of maximum/minimum dimensions for oblate hailstones (six observations): 1.25, 2, 2.1, 3, 3, 8

Total 61

**Table 5—Shape of New England hail (from reports of Cooperative Observer Network during years 1956–1968)**

**Fig. 21**—The relationship of the number concentrations of hailstones to diameter

**Fig. 20**—Frequency of occurrence of time between start of hail and heavy rain
volume in the stratosphere with time during the course of one of the tornado storms suggests a progressive modification of the low stratosphere in the vicinity of the storm, allowing easier penetration by the storm tops during the latter part of the storm life.

Acknowledgments—The authors are deeply appreciative of the helpful criticism and many suggestions furnished by David Atlas. We are also appreciative of the constructive criticism of Edwin Kessler. We are pleased to record our gratitude to the volunteer thunderstorm observers of New England, whose many excellent reports have made this study possible.

References

Discussion
(Note: Discussion of this paper is combined with those of the preceding and following papers at the end of the following paper.)
Hail Studies in Illinois Relating to Cloud Physics

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Abstract—Three independent hail studies during the past year have provided considerable basic knowledge concerning Illinois hailstorms. Analysis of climatological records from 85 stations indicates marked spatial and seasonal differences in frequency of hailstorms within the State. Considerable variability also occurs from year to year and decade to decade.

One hundred eighty days with hail which caused damage to crops during 1953–1957 have been studied using insurance records or crop-loss, radar, and synoptic data. The crop-loss data were plotted to determine the time, location, and areal extent of the hailstorms. Variations in hail intensity (per cent of crop damage) were also examined.

Detailed case studies of the most significant hailstorms were made using radar and severe local storm volunteer observer data. It was found that distortions in squall lines corresponded closely with areas of greatest hail and wind occurrence. A case of hail formation in advance of a squall line was examined. Radar and rawinsonde data were used to formulate an hypothesis of the advanced hail formation process.

Introduction—Three studies of hail are currently in progress at the Illinois State Water Survey. The first study involves the use of the records of the U. S. Weather Bureau cooperative substation and first-order stations. These records provide data on the frequency of hailstorms throughout Illinois. The data are studied to determine if the climatological differences can be related to physical processes involving the formation and dissipation of hailstorms.

The second of these studies, supported by the Crop-Hail Insurance Actuarial Association of Chicago, utilizes insurance company records of paid hail losses. The purpose of this study is to determine whether meteorological parameters can be used to define variations in the hail hazard in Illinois, and, consequently, be employed in establishing hail insurance rates for various areas of the State.

The third study, supported by the Air Force Cambridge Research Center (AFCRC), is concerned with collecting and analyzing radar data and volunteer observer reports of hail occurrence to evaluate the utility of radar for identifying hailstorms.

This paper presents examples of the data used and the types of analysis being performed which are of importance in determining the physical processes of hail formation.

Climatological distribution of hailstorms—A study was made of the hail distribution in Illinois on an annual and seasonal basis. The purpose of this study was to provide a climatological description of hailstorms in Illinois in respect to their average and extreme occurrences. Emphasis was placed on the period from March through August, when most hail occurs in the State.

The analysis was based on satisfactory hail records between 1901 and 1950 from 85 stations in Illinois including 12 first-order stations [Huff and Changnon, 1959]. Fifty of these 85 stations had 20 or more years of reliable hail records.

Annual hail distribution—The average frequency of annual hail days is illustrated in Figure 1. In this figure, the average has been expressed in terms of number of days with hail in an average ten-year period. The annual hail maximum occurs in the region west of Springfield. Secondary maxima are indicated in extreme northwest and in southern Illinois. Areas of minimum occurrence are indicated in eastern Illinois and west of Peoria.

The areas in northwest and southern Illinois do not have elevations exceeding 1200 ft above mean sea level. The regions are quite rugged locally and the hills rise abruptly from the flatlands to the south in the path of the prevailing wind flow. Consequently, it is quite likely that the hail maxima in these areas are partially induced by the abrupt differences in the local relief. The area of maximum frequency west of Springfield has no pronounced changes in relief in relation to the topography of the land surrounding it. The U. S. Weather Bureau [1947] has pointed out that this maximum area, which

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is a part of a non-orographic hail belt extending from Oklahoma to Pennsylvania, coincides roughly with a region of strong frontal activity.

Summer hail distribution—The average frequency distribution of hail days during the three summer months of June, July, and August is similar to the annual hail frequency shown in Figure 1. However, the maximum in extreme southern Illinois is not present. Furthermore, the area of highest hail frequency is located in northwestern Illinois rather than in the area west of Springfield. The annual hail minimum west of Peoria appears pronounced in the summer.

Months of maximum hail occurrence—In general, the month of maximum hail activity is March, April, or May in the southern portion and April, May, or June in the northern half. Eight days with hail are average for May, six days for April, and between five and six days per month for March and June. The frequency of hail days declines through the summer and fall seasons, with an average of four hail days in July, three in August, and between one and two in September and October.

Reliability of distribution patterns—Since there was some question whether the hail maxima in northwestern, southwest central, and southern Illinois were persistent or had been induced by a few years of unusually heavy hail occurrence in these areas, the reliability of the hail distribution pattern was investigated. Hail maps for each of the five decades in the 1901-50 period were drawn. It was found the three maxima areas persisted throughout this 50-year period, although some deviations from the pattern of Figure 1 were apparent in the various decades. In general, the five ten-year periods showed patterns similar to the average pattern of Figure 1 with respect to outstanding features. However, some major pattern differences did appear from decade to decade.

To illustrate time variability further in the number and distribution of hailstorms in the State, hail occurrence maps for separate years were prepared. It was found, for example, that a relatively large number of hailstorms occurred over a large portion of the State during 1911. Individual station amounts ranged from zero to nine. In contrast, in 1935 hail was relatively sparse. The maximum number of storms recorded at any station was four, and the majority of the stations had none or only one storm. In 1936, again relatively few hailstorms occurred with a maximum number of four in northwestern Illinois and large areas of zero or one over the rest of the state.

Analysis of crop-hail loss data—Variations in the hail hazard in Illinois are being studied using records of paid hail losses compiled by the Crop-Hail Insurance Actuarial Association (CHIAA) [Roth. 1955]. The purpose of the study is to determine whether or not meteorological parameters can be used to define variations in the hail hazard, and, consequently, be employed in establishing hail insurance rates for various areas.

The records of paid hail losses include the year, month, day, and hour of hail occurrence; the county, township, and section (square mile) where hail occurred; the rate or percent of loss caused by the hail; and the type of crop damaged. In addition to the hail loss records for the period 1952-57, radar data of the Illinois
State Water Survey and surface and upper air data of the U.S. Weather Bureau are also being studied.

From these various types of data, information has been extracted on the location and extent of hailstorms, the movement of hailstorms, and the intensity of hailstorms in Illinois.

Location and extent of storms—Days having greater than 20 paid losses were defined as hailstorm days. The number of hailstorm days from late May to early October during the period studied was 201 and ranged from 21 in 1952 to 49 in 1956 [Stout, Blackmer, Changnon, and Huff, 1959]. Maps showing the location of hail damage to crops on each of these hailstorm days were plotted by marking the section in which hail damage occurred. A color code was used to designate the hour of hail occurrence when plotting the larger storms (greater than 150 paid losses). Examples of the plotted maps showing the time and location of damaging hail are shown in Figures 2 and 3.

Figure 2 shows the locations where hail damaged crops on August 7, 1953. On this date hail destroyed one per cent of the crops in Illinois insured by CHIAA companies. The storm was widespread enough to affect three per cent of the crops in the state and the average loss to the crops affected was 33%.

Areas of relatively continuous hail damage, shown on the map of August 7, 1953, will be referred to as 'swaths.' On this date there were six outstanding examples of these swaths, while other swaths which are smaller, or more scattered, are discernible.

The locations of crop damage during the hailstorm of August 9, 1954, are shown in Figure 3. This storm had the largest number of paid losses during the 1952-57 period and is of particular interest because of the extreme length of the longest swath. This swath is at least 160 mi long, perhaps longer, since it may extend into Iowa. This storm destroyed one-half of one per cent of the crops insured by CHIAA companies in

![Figure 2](image_url)
Illinois in 1954. It was comparable in extent to the storm of August 7, 1953, since it also affected three per cent of the crops.

All of the maps showing location of hail for the period 1952 to 1957, inclusive, were examined and information tabulated concerning the number, length, and width of swaths. In addition, the rectangular area necessary to enclose all reports of hail on a storm day was computed. This area is called the hail region. Table 1 presents statistics relating to the areal extent of the storms in various years.

The reason for the variations from year to year, as shown in Table 1, has not been investigated but may be related to long-period variations in such atmospheric parameters as zonal wind speed, temperature, and moisture distribution. Detailed information of this type is necessary to define the extent of hail more accurately. Information on the extent of hail provides knowledge of the volume of the atmosphere which is contributing to the formation of the hail. The need for more detailed data on hail occurrence for use in evaluating radar for identification of hailstorms has led to the establishment of hail observer networks [Donaldson, 1958; Douglas and Hirschfeld,
Fig. 4—Percent of storms moving from various directions (open bar) and percent of years damage caused by storms moving from various directions (solid bar)
1958: Wilk, 1959). In these networks, however, observers are often spaced so widely that exact definition of the areal extent of hail is impossible. Thus, crop-loss records are the best available data for defining the extent of hailstorms in heavily insured areas.

Hailstorm movement—The movement of each hailstorm was determined by examination of radar echoes at the time and location of hail occurrence. The direction of movement showed a seasonal change. In May and October, hailstorms moved from directions south of west. In July and August the preferred direction of movement shifted to the northwest quadrant.

The direction of movement of hailstorms varies greatly seasonally and annually. Figure 4 shows the percent of hailstorms moving from various directions each year, 1952 to 1957 inclusive, and also shows the percent of the total damage each year which occurred with storms moving from the indicated direction. The most damaging storm in a given year may cause a high percentage of the total damage. The percent of the year's damage caused by the major storm during each of the six years is indicated beside the damage bar of the direction from which the major storm moved. For example, in 1953 ten per cent of the storms moved from the west-northwest, causing 44% of the year's damage. Most of this damage was caused by the storm of August 7, 1953, which produced 43% of the year's damage.

Hailstorm intensity—Variations in hailstorm intensity may be studied using the percent of crop destroyed. This may be done by year, month, day, or hour for the heavily insured areas of the state. In this study, individual hailstorm days have been classified according to the areal extent of the storm and the intensity of the storm. The areal extent is given by the percent of the total insured crop in the state which was exposed to hail on a given day. The storm intensity is given by the percent of the exposed crop which was destroyed by hail. The distribution of storms in 1953 and 1957, classified in this manner, is shown in Figure 5. The figure shows that storms had a tendency to be larger and more damaging in 1953 than in 1957.

Local variations in intensity may be studied by plotting the percent of exposed crop in each section which was destroyed by hail. Maps showing intensity variations have been plotted for ten days with hail. Two of these maps are shown in Figures 6 and 7.

![Fig. 5—Comparison of 1953 and 1957 storm area and intensity](image-url)
Data from only two crops (corn and soybeans) have been plotted on these maps. By restricting the plotted data in this manner, apparent hail intensity variations resulting from differences in hail susceptibility among crops is minimized.

Figure 6 shows variations in hail intensity during the storm of August 7, 1953. Examination of the figure shows that the regions of heaviest damage are scattered along most swaths, except for the one in the center of the map. This swath had relatively heavy damage along a substantial portion of its length, and, according to a newspaper account, hail the size of baseballs and oranges was observed near the end of the swath. The large size of the observed hailstones may indicate that the thunderstorm causing the hail damage along this swath was more severe than the other storms on this date. However, detailed examination of the radar echoes revealed that at least three thunderstorms moved across parts of this hail damage swath during the reported hours of hail occurrences and that a single thunderstorm could not have produced all the hail. The fact that several different thunderstorms each produced hail in this area and that the areas overlapped in some parts of the swath probably explains the higher rates of destruction in this area. A more detailed study is being made of the radar echoes on this hailstorm day.

Figure 7 shows the distribution of crop losses during the storm of August 9, 1954. Most of the hail observed with this storm was the size of marbles according to newspaper accounts. The rates of damage in individual sections with this storm were generally less than with the storm of August 7, 1953.

Studies of hailstorm intensities are complicated by the fact that there are variations among crops in susceptibility to hail and that a single crop may vary in susceptibility to hail depending on its stage of development. Two methods may be used to overcome this margin of
error. First, losses to a single crop may be used in detailed studies; second, periods when all crops are equally susceptible to hail may be studied. Fortunately, one of the major crops in Illinois, soybeans, does not vary appreciably in susceptibility to hail during its life history. To show that two crops may be equally susceptible, the distribution of various rates of damage were compared for corn and soybeans for the storm of August 7, 1953. This comparison showed no significant difference in rates of damage to the two crops per unit area.

Currently, a study is being made to compare hail intensity with slope of terrain over which the storm is moving. This study is not sufficiently advanced to report at this time. One of the limitations in studying hail intensity using crop-loss data is the lack of information about the size and concentration of the hailstones. Identical crop damage could result from many combinations of hailstone size and concentration. It is at this point that the networks of volunteer observers become necessary.

Analysis of radar and network data—A network of volunteer observers was established in central Illinois in the summer of 1958, under AFCRC sponsorship (AF Contract 19(604)-4940, The Determination of Techniques for Radar Identification of Severe Thunderstorms). Approximately 1000 volunteer observers, located within a 30-county area and 100-mile range of the radar, furnished detailed reports concerning the time of occurrence, size, and concentration of hail, as well as reports of damaging wind, lightning, and heavy rain. Examination of these data in conjunction with PPI (CPS-9) and RHI (TPS-10) radar data yielded several enlightening facts concerning hailstorm formation.

Mid-tropospheric jet penetration—PPI radar display of intense pre-cold frontal squall line activity often exhibits major distortions, or bulges, indicative of zonal acceleration within
the line. A recent investigation [Nolen, 1959] suggests a close association of tornadic activity with this echo pattern. Nolen has appropriately named these line distortions 'line echo wave patterns,' referred to hereafter as LEWP.

Two storms in 1958 which exhibited LEWP were well documented by data from radar and network observers. Detailed analyses of the storms were made to determine: the association of hail and strong winds with the LEWP; and, whether the LEWP was a function of echo height.

Figure 8 illustrates the radar display on July 27, 1958, with the LEWP development shown at four-minute intervals on normal and reduced gain. The total time for development was approximately 30 min. The sharp increase in echo intensity along the southern extreme is a normal product of LEWP development. An RHI cross section indicated little or no echo height variation through the LEWP. This suggests that the line distortion was not a result of vertical growth into a steering level of higher wind speeds, but rather a consequence of a horizontal discontinuity in the wind field, in the form of a narrow jet penetration. Analysis of the general wind profile, with emphasis on the Peoria rawinsonde observation, provided support to this conclusion.

Figure 9 is a radar-synoptic schematic of the network, hail and wind reports, the LEWP, and the wind maxima that occurred on July 27, 1958. The wind value of 263°, 58 knots, was observed at 13,000 ft over Peoria at 18h 00m CST. A greater maximum of 245°, 64 knots, was observed near the tropopause at 29,500 ft. The northerly drift of the LEWP suggests that the higher maximum had limited directional control; however, the surface wind observations, as well as the LEWP, are aligned with the direction of the lower maximum.

The second example, August 7, 1958 (see Fig. 9) supports the LEWP correlation with the hail reports. Post-analysis of surface synoptic
observations confirmed the stationary front position in northern Illinois at 06h 00m CST and 12h 00m CST. The single wind maxima of 200°, 62 knots, occurred at 16,000 ft.

Pre-squall line hail shower occurrence—The PPI radar display for August 7, 1958, shown in Figure 10 indicates a less continuous line than that associated with the LEWP on July 27, 1958. It is apparent that further loss of line continuity results in the LEWP becoming ambiguous, as is highly probable in the case of isolated radar echoes associated with severe weather.

Most of the hail occurred simultaneously with the heavy rain associated with the principal thunderstorm cells. However, one detailed report was received of hail falling in advance of the disturbance. An examination of the RHI radar data disclosed a unique echo over this location. The echo at 349°, as shown in Figure 10, exhibited distinct streamers which suggested the existence of hail shafts. The only other echo in proximity was a suspended cell at 342°. Analysis of the Peoria rawinsonde data for 06h 00m CST and 12h 00m CST provided the general airmass temperature and moisture distributions south and north of the LEWP development. The profile in Figure 11 shows the streamers becoming continuous near the minus five-degree centigrade level, diffusing, and losing their identity near the wet-bulb freezing level. The echo depth decreased below the convective condensation level suggesting that limited evaporation was occurring. The surface observation indicated very light rain associated with one-fourth inch hail.

It was initially assumed that the hail was thrown out from the main thunderstorm. How-
ever, further examination of the RHI radar data suggests the formation of hail was aided by the auxiliary shower development at 342°.

The echo configuration and apparent hail formation process is illustrated in the schematic in Figure 12. At 09h 33m CST, echo A was distinguishable from 340 to 347° and sloped upward to 15,000 ft. Five minutes later, at 09h 38m CST, echo A was entirely suspended, with the base at approximately 7500 ft and the top entering the cirrus overhang at 26,000 ft. The strong vertical transport ahead of the LEWP had carried the super-cooled droplets along the trajectory (1) into co-existence with the ice crystals in the anvil. The graupel which was formed in this region (2) was released to the lower levels and would have terminated as virga, except for the fact that it was in juxtaposition with the origin of echo B. The growth stage of this new cell provided vertical support and droplets for continued particle growth (3) and hail formation.

**Summary and conclusions**—Studies of hail are being made using climatological, crop-hail loss, radar, and volunteer observer data. No single type of data provides sufficient information to conduct a comprehensive study of hail formation. For example, the crop-loss data can define the areal extent of damaging hail in regions where large areas of insured crops are located, but cannot provide information on hail size or concentration. The radar data can provide information on precipitation location, extent, movement, and intensity, but radar is not capable at present of definitely distinguishing between rain and hail. The volunteer observer network data are useful in determining the size and concentration of hailstones and the relationship of hail to the occurrence of rain, but do not permit the evaluation of the areal extent of hail.

Although the primary purpose of the investigations being carried out at the Illinois State Water Survey is the determination of hail distribution and radar detectability, it is hoped that the preliminary case studies presented in this paper will contribute toward a better understanding of the physical processes involved in hail formation.

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**References**


Douglas, R. H., and W. Hitschfeld, *Studies of...*

Discussion

(Relating to three preceding papers by Beckwith; Donaldson, Chmada, and Sheackford; and Stout, Blackmer, and Wilk.)

Dr. W. Hitschfeld—i think Messrs. Beckwith, Donaldson, and Blackmer have given us three excellent papers in which a great deal of factual material was presented. So, if I may lead off the discussion with Donaldson's paper, in Figure 10, he plotted the product of height and radar intensity measured in suitable radar units. I think this is the kind of approach that could be extremely valuable in the discrimination by radar techniques, of rain from hail, and possibly the recognition of tornadoes by radar.

Mr. Beckwith showed us some surprising things. The photographs of the dents that were produced by large size hailstones in aircraft at high levels are strong evidence that four inch hailstones occur at such enormous levels as 40,000 ft. This may suggest updrafts as high as 30 or even 60 m sec⁻¹, depending on hailstone shape and density.

Mr. Blackmer has shown us how one can use insurance records, which (Dr. Douglas tells me) are difficult to get at and analyze; and he has done a lot with them. I only want to sound a word of warning. When one comes to relating these insurance records with storm patterns, one may be in for a surprise. For instance, a 100-mile long streak of damage, may turn out to have been caused by three or four different storms.

Mr. Roy Blackmer—I have looked a little bit at the radar echoes of that particular storm. It is a little difficult to tell just how much echo development was going on, because the radar beam was a little wide and the place where the echo first appeared was out toward the maximum range of the radar. But there appears to be really good echo continuity along that whole path.

Dr. Hitschfeld—One of the nicest features, by the way, is the fact that you have suggested in showing from what square miles you had a right to expect the reports, and from what square miles you did not. The negative information is just as important as the positive. With the usual kind of ground observers, the negative information is either not available or very spotty but I think the way you did it gives a lot of credence to the continuity.

Mr. Blackmer—With the 1950 data, we can put two different types of observations together: the volunteer observer network gives us the distribution of hail size and accurate time of occurrence. The crop-loss data will fill in the gaps between those to give the exact areal extent.

Dr. L. J. Batten—I am curious about the observation of a thunderstorm that went 12,000 ft above the troposphere. If one sets down the approximate equation and considers an isothermal layer, one needs very extreme vertical velocities, in order for a parcel to get very far above the tropopause. As a rule of thumb, Youngent has suggested 20 m sec⁻¹ per kilometer of penetration. This means that the ascending air must have a vertical velocity of 70 m sec⁻¹ or more when it passes the tropopause. Taking into account all the possible errors, what would you say was the accuracy of the height measurement?

Mr. R. J. Donaldson, Jr.—Plus or minus 2000 ft average for all of these measurements.

Dr. Hitschfeld—How accurate is your measurement of the height of the tropopause?

Mr. Donaldson—An arbitrary tropopause was taken using the point above 300 mb at which the lapse rate first becomes less than two degrees C per thousand feet. Most of the time it was fairly clearcut. The sounding either became isothermal, or bent backwards, indicating a nega-
tive lapse rate. There are some cases where this
is not so simple, and I am sure there are errors
in trying to decide where the tropopause was.

Major Currie S. Downie—Tropopauses in gen-
eral are not like a wall or curtain that stay in
one position from one sounding to the next.
These things can vary considerably over a 12-
hour period; time and space variations may be
appreciable.

Mr. Donaldson—I am sure we can have a
possible six-hour error or a 150-mi error from
the time it is measured. But the time and space
effects probably average out over hundreds of
items of data.

Dr. R. Wexler—I enjoyed the paper very
much, but I want to put in this suggestion about
the use of statistics. I have noted that down in
the tropics, for example, it is very difficult to
tell where rain will develop on the basis of sta-
bility alone; even with a very unstable sounding
no rain may develop. Whenever anybody
determines the frequency of hailstorms asso-
ciated with large positive energy, I would like
to ask the frequency of clear weather also asso-
ciated with large positive energy.

Mr. Boynton Beckwith—Forecasters know
better than the dynamic meteorologists, that
thermodynamic curves are very poor forecast-
ing tools by themselves.

Dr. Hitschfeld—I agree with you. Positive
energy areas are a very poor indicator for hail
occurrences both in Denver and Alberta. I
nevertheless believe that the Fawbush and Mil-
er criteria are perfectly valid for the region
that they are worked out for. I think they have
a physical basis and have been verified empiri-
cally, but they do not seem to be exportable to
Canada or the Rockies.

Major Downie—Recently I made a study con-
cerning the frequency of occurrence of two-inch
hailstones, and find in this country the maxi-
mum frequency occurs in the Great Plains re-
region; they almost never occur along the sea
coast and only occasionally in the Middle West.
I would like to find out if there is an explana-
tion for this distribution. How can we explain
the fact that practically all the two-inch hail in
the country occurs over the land with an eleva-
tion of 3000 to 5000 ft, whereas in southern
United States practically no large hailstones are
reported.

Mr. Jerome Namias—Actually we are here
dealing with large-scale dynamic climatology,
and a very complex problem. However, there
are some points which are fairly straightforward.
In the first place very deep layers of unstable
air are required for hail and tornadoes whatever
way you look at it. These are manufactured
especially in spring and early summer in a
manner that favors the Middle West. Usually,
the deep moist layers are produced this way:
As the spring sun begins to warm southern
United States and the Bermuda High anticy-
cone begins to build, a shallow layer of warm
moist air flows northward (see warm moist
tongue in Fig. 13). At the same time mid-

![Fig. 13—Schematic differential flow pattern in spring leading to severe thunderstorms, hail and tornadoes, especially over shaded area](image-url)
tropospheric wind patterns over the west frequently are different during spring (broken lines of Fig. 13). This upper-level pattern, of course, depends on circulations in other parts of the hemisphere. The combination of surface and upper-level patterns results in differential advection over a broad area (especially shaded area). In the low layers we get warm moist air flowing into the Great Plains in the form of great warm moist tongues. At upper levels, 500 mb or so, the air is frequently very cold, having arrived from the eastern Pacific or even the Gulf of Alaska. The lower warm moist air is shielded by the Rockies, so the cold air overruns. This cold air has low specific humidity, and it is relatively dry; so that over a large area where this differential flow pattern obtains, there is increasing instability, both conditional and convective. The lapse rate increases because of the cold air overrunning the warm, and also convective instability increases because moist air from below is being overrun by relatively dry air. Also, one of the most favored storm tracks in spring when the polar front moves north is the 'Colorado Low,' so that synoptic-scale disturbances which have the vigor to release the energy of this instability are frequent. The synoptic activity is favored by geographical considerations. The presence of the Gulf, the cold air from Alaska and the Pacific, and the Continental Divide all fit into the dynamic climatology of hail and tornadoes. In the East we rarely have such meteorological conditions, since this type of differential advection rarely takes place. The rain situations are of a quite different character, and by the time the moist air and trough arrive, there has been a warming, and the conditional and convective instability have been released. There are many factors which make the East climatologically quite different from the standpoint of stability. Occasionally, tornadoes occur when the flow pattern at upper levels stalls so as to permit a moist current from the Gulf to enter below an eastern jet stream and into confluent pattern. After a period with very warm lower air and high troposphere, some storm may rip through and do the job of releasing the pent-up instability. But this is a very infrequent case and most of the time the pattern I earlier described is present.

Mr. Blackmer—Recently when looking at variations in hail occurrence across the state of Kansas, I plotted a cross section using ten-year mean June soundings for Columbia, Mo., Dodge City, Kan., and Grand Junction, Colo. On this cross section I marked the height of the convective condensation level and the height of the freezing level. At Grand Junction, where hail is quite frequent, the low humidity at low levels resulted in a convective condensation level at sub-freezing temperatures. At Columbia the convective condensation level was at a temperature considerably warmer than freezing. Thus the cross section showed considerable change in the thickness of the layer between the cloud base and the freezing level.

I believe that the temperature and thickness of the layer between the cloud base and the freezing level is very important in the formation of hail. In humid regions where there is a thick layer of cloud beneath the freezing level, the cloud water will be quite warm. To freeze any of this water it must be lifted considerably and the cloud drops could possibly grow to fallout size before being lifted to the freezing level. In a cloud with a base near the freezing level the cloud water will be cold and will not have to be cooled much more to freeze.

The cross section showed that the thickness of the cloud layer between the cloud base and the freezing level varied from zero in western Kansas to 3000 ft in eastern Kansas. Long-period insurance company records show an average crop loss of ten per cent in western Kansas while in the eastern part of the state the loss is about one per cent.

Obviously, much more work needs to be done before any definite conclusions can be reached. One very important step would be to obtain long-period average sounding using data only on days with hail. These soundings would provide data which could be compared with insurance company records of hail loss to learn more about just what conditions are most favorable for hail.

Dr. G. D. Kinzer—In bringing the session to a close, I would like to read to you from today's newspaper, "Eighteen inches of hail falls in Kansas." "A hail storm battered this small prairie town of Seldon in northwest Kansas for two hours last night covering the town with 18 inches of ice." I think that is a very remarkable event, and yet in the light of what we have discussed tonight it does not seem unusual. (A report on this spectacular hailfall can be found in the Monthly Weather Review, 87, 301-303, 1959, by A. D. Robb, entitled "Severe Hail, Seldon, Kansas, June 3, 1959."—Ed.)
Future Research in Weather Modification

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Abstract—The paper outlines the present status of research in cloud physics and weather modification, and attempts to emphasize the need for expanded research in certain areas of cloud physics. Several suggested outlines for research projects are given.

Introduction—Weather modification as treated in this discussion deals with competent research designed to increase precipitation, prevent hail, inhibit lightning, or dissipate fog. Present scientific efforts at weather modification follow quite closely experimental methods and techniques developed by Langmuir and others [1953]. These methods use nucleating agents dispersed into the cloud by ground generators or by aircraft flying over, in, or under the base of clouds to create artificial precipitation or to bring about other changes in state in the clouds. Solid carbon dioxide (dry ice) or silver iodide are the two most commonly used nucleating agents although others such as cupric sulphide, hydrogen chloride, and a number of clay minerals are effective as freezing nuclei.

Many uncertainties exist as to the exact manner in which the nuclei affect the clouds to bring about the change of state from vapor to water or solid form when they are seeded. This, then, brings us to the field of cloud physics.

Cloud physics and weather modification—Houghton [1959] recently has reviewed the current problems in cloud-physics research. One cannot read this excellent review without becoming singularly impressed with the urgent need for expanding and accelerating the research effort in cloud physics and the precipitation processes. Not to be overlooked are studies of the chemistry and electrical effects in the atmosphere.

Houghton’s review of homogeneous nucleation, condensation nuclei, freezing nuclei, precipitation processes, and natural precipitation mechanisms repeatedly emphasizes the broad gaps in our knowledge. He concludes that there are many challenging problems that only can be answered by both laboratory and field research. He particularly stresses the need for more instrumentation to measure the parameters of cloud physics.

A casual review of the papers presented at this meeting serves to impress the reader with the many unsolved problems of atmospheric processes. Almost every paper reports lack of observational or experimental data. Frequently, assumptions are made to justify later derivations or arbitrary criteria are set up to make it possible to justify the conclusions. It seems to me that many of the assumptions are the same ones used when I was a student 30 years ago.

Of course, much progress has been made but in comparison with other fields of science, medicine and atomic energy, for example, the progress in understanding atmospheric processes seems quite insignificant. Until more progress is made in our understanding of precipitation mechanisms, efforts at weather modification are limited to our present knowledge.

Need for basic research—The Advisory Committee on Weather Control [1957] recommended that encouragement be given for the widest possible competent research in meteorology and related fields. The report pointed out that a vigorous research program should be established with adequate provisions to maintain continuity and reasonable stability for long-term projects. It stressed the fact that emphasis should be placed on sponsoring talented men and their projects.

A National Institute of Atmospheric Research—The Committee on Meteorology [1958] of the National Academy of Sciences-National Research Council has proposed the establishment of a National Institute of Atmospheric Research. This Institute would be devoted exclusively to basic atmospheric research. It recognizes the formidable problems that the meteorologist faces in the field of cloud physics and would establish a research program commensurate with the global importance of weather. The program proposed by this Committee deserves the strongest possible support by all scientists having an interest in atmospheric research.

Cloud physics and planetary research program—Ackerman [1959] has suggested a two
pronged attack on the problem of atmospheric research. He suggests a cloud-physics program and a planetary program.

The cloud-physics program would be concerned largely with the problems discussed in this volume: dynamics of clouds, precipitation mechanisms, condensation nuclei, freezing nuclei, and other small-scale phenomena of the atmosphere. This program would be a long-range one with stable financial support with complete freedom of action and latitude to scientists in choosing areas of research. It would include experiments on any appropriate scale to insure orderly research. Such a program, Ackerman thinks, should be conducted by an agency of the government separately from those agencies now having meteorological services and heavily committed to day-to-day weather problems. An independent research program, he thinks, would be conducive to essential depth, balance, and originality.

The second program, the planetary one, suggested by Ackerman is a large-scale, long-range one which would be primarily concerned with global atmospheric circulation problems. It would be heavily weighted on the theoretical and observational basis and would eventually require international scientific cooperation. Guidance for such a global program might well be placed under the United Nations with a planning and review board made up of eminent meteorologists, mathematicians, physicists, and other representatives of other disciplines to insure widest possible cooperation. Such a program would provide an opportunity to show scientific leadership which might well play an important role in mitigating the problems in diplomacy. While such a program today might seem visionary, by tomorrow it might have revolutionary economic and military implications.

Fiscal problems—If we are ever to develop a multiphased atmospheric research program to get the answers to the thousands of questions now confronting us, we must present a bold, imaginative, and unified front that will command strong public support. At the present time the entire research effort can be seriously jeopardized by the whims of one or two public officials in prominent positions in government. In 1957 many of us saw research funds reduced to a trickle by the adverse decision of economy-minded officials and only the historic launching of Sputnik in October saved this program from almost complete obliteration.

Only two weeks ago two very important items were disallowed in the National Science Foundation fiscal 1960 budgetary requests, by the House Appropriations Committee. One item was for $500,000 for the initial phases of the establishment of the Institute of Atmospheric Research; the other a request for two million dollars to carry on the research and evaluation of weather modification as directed by Congress in Public Law 510 [American Meteorological Society, 1958]. Such actions on the part of unformed public officials account for the weak, uncoordinated, and halting atmospheric-research effort today.

One can cite many more examples where suddenly funds have been reduced or eliminated and have forced the release of good research teams. If and when funds are again granted, the chances of re-assembling a competent scientific team are very remote. Under these conditions, meteorology is not an attractive field for the young man interested in science.

The foregoing problems are given as examples of difficulties which prevent the establishment of a strong stable research program in cloud physics. Without the necessary answers gained from this research, weather modification efforts must necessarily be limited.

Research tools—Radar was found to be invaluable in studying the effects of cloud seeding in Arizona [Battan and Kassander, 1959]. As a research tool radar has become indispensable. Since World War II it has played a role of increasing importance for operational and research studies of atmospheric phenomena. Color radar combined with an electronic memory system is destined to become indispensable in future research projects in cloud physics.

Instrumented aircraft have been used [Cunningham and Atlas, 1954] to study the structure of hail at 32,000 ft as observed from the aircraft. Cunningham's long experience in observing frontal systems, thunderstorms, tornadoes, and hurricanes is a strong endorsement for instrumented aircraft as a research tool for future research. Simpson and others [1958] confirm the value of aircraft observations in studying the structure of hurricanes. Much of the experimental data obtained in the University of Chicago cloud-physics research program were obtained from specially instrumented aircraft.

Constant-level balloons, manned and unmanned, have not been mentioned in the present discussions. These large inexpansible polyethyl-
ene balloons are, in my opinion, ideal platforms for carrying on very important observations from altitudes of 50,000 to 100,000 ft. Capable of carrying a payload of 2000 lb and remaining aloft for more than 100 hr assures that the balloons can be used to track a front continuously across the United States. Radar, television, nuclei count, and air samples, as well as the usual cloud-physicists parameters, can be obtained throughout the life history of a frontal condition. Observations aloft coordinated with lower-altitude aircraft observations and a ground network of weather stations might give some very important clues into the atmospheric processes. Future research projects should include constant-level balloons for atmospheric research. The cost of these balloons would be much less than experimental aircraft and they offer a platform to obtain continuous observations at high altitudes for periods of a week or ten days. It would be feasible to use them for global oceanic observation platforms. Present electronic capabilities would permit the use of unmanned balloons in many areas but eventually all balloon observation stations should be manned with at least two observers.

Future research programs—Future research programs in weather modification should plan for and provide the following:

(1) An exhaustive cloud-physicists program on a continuous basis for a period of five to ten years. The program should be developed by competent scientists so that projects are well designed and carefully controlled to yield maximum amount of experimental data.

(2) Projects should be established in at least nine areas of the United States: Florida and Gulf Coast, New England, Iowa and South Dakota (flat lands), Utah and Colorado; California and Washington. The location is based primarily on orographic and climatic features to provide maximum opportunity for observing all types of weather situations.

(3) Instrumented aircraft and manned and unmanned constant-level balloons should provide observations on a coordinated basis to supplement ground observations.

(4) Additional mountain observations should be established to furnish most representative surface observations. High-speed cameras will find increasing use in observing change of state in the laboratory, and eventually their use in field experiments will become an essential part of any future research program.

(5) Radar of the latest type designed specifically for cloud-physicists observations should be available to each of the research projects. Color radar, magnetic-memory systems, and closed-circuit television should be provided when the science of the art can provide such ultra modern equipment.

Funds to implement the entire program outlined should not exceed 30 million dollars annually after the initial facilities have been established.

References


Battan, Louis, and R. Kassande, Artificial nucleation of orographic clouds, Institute of Atmospheric Physics, University of Arizona, June, 1959.


Discussion

Dr. Bernard Vonnegut—I agree with you that high-altitude balloons are a very useful tool for meteorological study. One application of balloons that should be of value in cloud physics is their use for obtaining time-lapse pictures of cloud development as viewed from above. These observations can be carried out with fairly small balloons as the cameras weigh only 10 or 20 lb. It is much more instructive to look down on clouds and see how they grow and move than to look up at an uninformative cloud base.

Dr. Helmut Weickmann—It is a very good
suggestion to look for areas which are specifically favored for the study of key problems of cloud physics or cloud modification. One is the Olympic Mountains near Seattle, Washington, for a study of orographic rain. These mountains have characteristics of a tropical rainforest on their windward side and desert-like stretches on the downwind side. Another is the downwind shore of the Great Lakes where studies on winter snow showers could be made. They are due to continental polar-air outbreaks across Canada at a time when the Lakes are still unfrozen. The unstable dry cold air eagerly takes up water vapor from the warm lakes which is readily being dumped on a 20-mile deep strip along the downwind shores. The Great Lakes region invites another study: large-scale seeding of supercooled cloud layers. The region produces very persistent stratus cloud decks during the winter, which can be dissipated through dry-ice seeding. A study appears feasible of the influence of large-scale seeding (areas of 1000 sq mi or more) on weather. This would have to be a cooperative project with the Air Force furnishing ‘flying boxers’ loaded with dry ice and the U.S. Weather Bureau and other institutions in the area making measurements of meteorological parameters in the air and on the ground, such as, for instance, the variation of net radiation flux, albedo, and of the local windfield before and after seeding.

_Dr. B. J. Mason_—It would be presumptuous for me to make comments on some aspects of Captain Orville’s paper, but nevertheless I would like to express what I feel very strongly about research in this field and the programs. There are three outstanding thoughts in this talk—continuity, stability, and building research around the promising individual. One of the greatest problems we have to face in this field of cloud physics is the shortage of competent people. There are many more projects than there are good people, and there are certainly many more problems than there are people competent to talk about them. We have spread the available talent far too thinly. This means we must do everything we can to make the maximum use of the people we have available and to recruit people from other physical sciences. This brings me to underline my view on how important continuity and stability are, so you can build up a team around a promising individual, and he can be sure he has funds for the work over a period of years. In my view one can do very much more research if one has ten million dollars over ten years, rather than ten million dollars in one year. There are no problems worth tackling in this field that can be solved by a crash program in one year. The trivial problems we are not interested in anyway. It is this continuity and stability that is so important, and I believe that one does not solve problems by building grandiose institutions in the first place, and finding staff afterwards. You have to build the institution around the staff. It is the people that matter and this means, of course, one must plan very carefully. In the long term, it is better to do five things properly than to make a half-hearted job of fifty things.

_Dr. Douglas K. Lilly_—It appears to me that for the most part we are trying to modify the weather after it was largely already quite settled and in order to modify a large convective system or something of this kind energies are required of any astronomical figure you want to take arbitrarily. I know, however, that in chemical reactions we have very often a situation in which it is necessary for a system to acquire a certain level of energy before it can do something. At this level it can possibly do several different things—the results of which may be quite different. But at exactly this peak, the activation peak, it takes really very little additional energy to push it in one direction or another, or in fact, back where it came from. I would like to suggest the possible place to look for a modification is way back at the beginning when an eclipse has not even started. There must be a point of indecision in the system at which a very small push in just the right direction could be perfectly enormous in its final effect, and this is something we looked at very little.
The Swiss Randomized Hail Suppression Project in the Tessin

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Abstract—The first two years of the Swiss randomized antihail project does not yet allow any statement having statistical weight about the effect of the AgI seedings to prevent hail, neither in a positive, negative, nor indifferent sense. However, the precipitation patterns display indications for a strong increase of the rain as a result of the AgI seeding by ground generators. According to the general design the project will be carried through for the time of five years in the expectation to arrive until then to some definite decisions as to the effects of the silver iodide on hail formation.

Since 1957 the Swiss Federal Commission for the Study of Hail Formation and Prevention [Sänger and others, 1957, 1958] has been carrying out its Third Hail Suppression Project in the Tessin. The test area lies in the southernmost part of the Alps and covers a region of about 3000 km² having a pronounced orographic character. The project is financed jointly by a Federal Grant from the Department of Public Economy (Division of Agriculture), by the Canton Tessin, and by two private firms which have cooperated in the project: Essagra S. A., Balerna, and the Oerlikon Machine-Tool Works, Bührle and Co., Zürich, the latter maintaining a large agricultural estate in the test area.

The seeding of the atmosphere is done from 20 silver iodide ground generators which surround the test area at distances ranging from less than one up to 30 km; seven of these are on Italian soil. The output of the generators is in the order of 10⁶/sec of active ice-forming nuclei, measured at a temperature of -10°C, which is in keeping with the figures usual in such experiments. The generators have been made available by the Oerlikon Machine Tool Works, Zürich, until now two different types have been used, the first a copy of the model developed by North American Consultants, Inc., Goleta, Calif., and the second a new type of construction in which the AgI solution is sucked in by the flow of the fuel gas. After some initial trouble the second type well proved its worth, so that from the beginning of the third annual experimental period this year it alone will be used. The experimental period runs from the middle of May to the beginning of October.

A strict process of randomization governs the occasions when seeding is to be carried out. Every day at 16h 30m a weather forecast is received from the meteorologists at the Osservatorio Ticinese in Locarno-Monti; if they consider that there is a danger of thunderstorms on the following day, then this is considered to be a test day. A randomization experiment then determines whether seeding is to be done or not. In this way two groups of test days are obtained, for which the observational results can be compared with one another on a statistically satisfactory basis.

Seeding always takes place between the hours of 07h 30m and 21h 30m, with the generators working to the rhythm of a ten-minute interval between five-minute operational periods. In accordance with the seeding time the observational or test period runs from 08h 00m to 22h 30m; only hailfalls occurring within this observational period are included in the later statistical evaluation. A special network of 29 observation posts, distributed over the entire area of southern Switzerland, has been set up in order to collect the necessary data in sufficiently local detail.

In the first two years of the project there were 55 test days, 42 seeded and 13 not. Hail was observed on 15 of the seeded days and on 13 of the unseeded days, but this statement takes no account either of the recorded intensity or of the geographical extent of the hailfalls. It only means that, during the test period, hail was reported at some place in the test area. It must further be noted that thunderstorms were actually observed only on 33 of the 42 seeded days and on 26 of the unseeded days. This allows us to see that the group of seeded days was more heavily burdened with storms than the unseeded group, so that the latter benefited from a higher rate of false forecasts. Generally speaking the amount of hail which has fallen during the first two years has been within modest limits,
if we restrict our record to the actual test period of the day, and does not permit any analysis in terms of intensity and location. The sparsity of the hailfalls compels us therefore to put aside any idea of an exact mathematical statistical treatment of the observational data, to wait at least for the results of the third test year, and for the time being to insist only that it is not now possible to make any statement having statistical weight, as to the effects, either positive, negative, or neutral, of the AgI seeding which has been done to prevent hail.

Two heavy hailstorms in the years 1957 and 1958 which lie outside the prescribed observation period deserve to be mentioned. These therefore will not figure in a later statistical evaluation of the test results. They are the hailstorm of August 13, 1957 and that of August 11, 1958. The first occurred on a day for which the forecast had been that there would be no storms (and which was therefore not a test day). In the case of the second hailfall, the storm did not begin until 22h 35m, five minutes after the end of the test period, and cannot consequently, if we are to be correct in our methods, be taken into account in the evaluation, despite the fact that this was a test day when there had been no seeding!

Particular interest attaches to the question whether the effect of AgI seeding may not perhaps be registered in the volume of precipitation. Apart from the records of the pluviograph

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**Fig. 1**—Daily precipitation in the test region in millimeters, 1957; upper values, seeded test days; lower values, unseeded days
in the Osservatorio Ticinese and of two other pluviographs which were put up during the course of the project, the only available figures for the daily amount of precipitation are those of the usual rainfall recording stations of the official Meteorological Service. If one considers, however, that seeding lasts for 14 hrs, that is, for more than half the day, then it is reasonable to expect that, if loading the atmosphere with silver iodide really has a strong influence on the amount of rain that falls, this must show up in the daily amount of precipitation.

The values obtained for mean daily rainfall in the years 1957 and 1958, are entered respectively on Figures 1 and 2. The upper figure

written beside the recording stations gives the mean quantity of precipitation for the seeded days, and the lower figure the mean for the unseeded days.

The astonishing result for 1957 (Fig. 1), namely that the increase in precipitation apparently caused by seeding is in the order of 100% (though there is some falling off towards the Alpine Divide), led to the enquiry being extended to find out how much rain had fallen in the nearby Italian districts and in the zone south of the Po. The necessary data was most obligingly put at our disposal by the Servizio Meteorologico della Aeronautica Nazionale through the good offices of Ezio Rosini. Since

![Fig. 2—Daily precipitation in the test region in millimeters, 1958; upper values, seeded test days; lower values, unseeded days](image-url)
Fig. 3—Daily precipitation in northern Italy in millimeters, 1957; upper values, seeded test days; lower values, unseeded days

Fig. 4—Daily precipitation in northern Italy in millimeters, 1958; upper values, seeded test days; lower values, unseeded days
these figures were entirely taken from pluviographs, it is possible to refer the amounts of rainfall exactly to the prescribed test period. In Figures 3 and 4, the results from this source have been assembled. It is extremely surprising to find that, for a majority of the recording stations south of the Po, the mean daily rainfall is again approximately doubled for the seeded days by comparison with the unseeded days during the test year 1957. This phenomenon was significantly not repeated in 1958. From Figures 2 and 4 it can be seen that in 1958 it is possible only to establish a doubling in the mean rainfall as due perhaps to seeding in the northern half of the Tessin. Yet at the same time even in the southern half the question of the mean rainfall between the seeded and the unseeded days is still considerably above unity. However, in the case of the Italian recording stations it does show a considerable fluctuation around unity.

That the recorded 100% increase in the amount of precipitation on seeded days even (in 1957) at the Italian recording stations south of the Po should have been brought about by the AgI seeding, is a physical impossibility; the observation can be explained in no other way than as the result of statistical scatter. And indeed these recording stations show a relatively small number of days on which there was any

Fig. 5—Kolmogorov-Smirnov test applied to the daily precipitation quantity of the test region, 1957 and 1958; percentage probability $P$ of an error of the first type
Fig. 6—Kolmogorov-Smirnov test of northern Italy, 1957 and 1958; percentage probability $P'$ of an error of the first type

precipitation, and as chance would have it the seeded days were unduly favored. Statistical treatment of the observational data, which in conclusion we shall now consider, clearly shows that this is, in fact, the correct explanation.

The experimental results with regard to rainfall were evaluated statistically according to the test procedure developed by Kolmogorov and Smirnov [Darling, 1955]; the evaluation was carried out by P. Schmid, a statistician at the Swiss Federal Institute of Technology. The results of these calculations are reproduced in Figures 5 and 6 for the recording stations on Swiss and Italian territory respectively. The figure written beside the station represents the probability $P$ that there would be at least the same result as regards an increase in precipitation if the experiment were repeated, but without any actual seeding. At the same time the differences of the values to 100% indicate in some way how much reliance we can place in the recorded increase in precipitation as a sign of the effect of seeding.

From Figure 6 it is immediately clear that all the recording stations south of the Po have probability values of more than 50%; thus the amounts of precipitation gathered by these stations have no significance from a statistical point of view in terms of the possible effects of seeding, a result in keeping with the supposition we have just made. The recording stations in the test area, however, present an entirely different picture; their probability values vary between 30 and 0%, and in no fewer than eight cases the probability is less than 5%. The recording stations involved here all lie in extremely mountainous terrain between the Magadino Plain and the central Alpine Massif.

From the point of view of physical meteorology the result of the statistical evaluation is obviously more consistent than the earlier figures given for the recorded rainfall. The possibility of effecting an apparently strong increase in precipitation as a result of seeding the atmosphere by means of AgI ground generators, makes it potentially likely that in similar orographic conditions hail formation can also be interfered with in some way. At all events it may be legiti-
mate to assume that, in the kind of conditions existing in the test area, dispersed silver iodide from ground generators is able to rise to heights where it can have an active nucleating effect. This means that when the experiment has been continued for a sufficient length of time it will be possible to arrive at some definite decision as to the effects, whether positive, negative, or neutral, of silver-iodide seeding on hail formation.

Dr. O. Essewanger—Did I understand correctly that you said there is no statistical evaluation?

Prof. R. Sänger—We have too little data to undertake an evaluation of the occurrence of hail.

Dr. Essewanger—That is what I wanted to say. The more samples you have, the more possibilities exist to get a significant result. One may get an insignificant result because of the shortness of samples?

Capt. H. T. Orville—Were your generator sites exactly the same in 1957 and 1958?

Prof. Sänger—We decided not to change anything during the whole period, mainly because of statistical reasons.

Mr. Jerome Namias—I do not mean to stick my head into a hornets nest here, but I have often wondered about the question of evaluating rain-making experiments from the rainfall records themselves. While I think the techniques used are legitimate from the standpoint of statistics, there are two possibilities that I would like to suggest for evaluations of this sort. One is that the synoptic characteristics of the periods be considered. Nature sometimes has a way of defeating our procedures of randomization. It might be that, because of the small number of seeding cases, during one long-period weather regime, the recurrent synoptic situation favors one period with respect to another. The second thought I have to offer is perhaps more controversial. One is often dissatisfied with the knowledge that weather forecasting has amassed up to the present. Nevertheless, there is positive skill on the part of the forecaster. I have wondered, therefore, why the forecaster can not be considered as some sort of a control in evaluating seeding effects. If one has sufficient cases over a long period, and the forecaster has positive skill, one can ask him to predict amounts over a long period of time and the discrepancy between his predictions and the observed amounts can be evaluated in terms of the effect of seeding operations.

Or it can be done on a ‘post-mortem’ basis. Give him various maps, let him make predictions or hindcasts of what might have happened, and use these as a control. So, I think there are methods other than statistical that might be put to use throughout the world.

Prof. Sänger—Every year we prepare a large report including the meteorological situations, but here I present only the statistical part. It is also our intention to discuss the observations made in a manner similar to Mr. Namias’ proposal. On the other hand we want to have an evaluation which is completely free of subjectivity.

Dr. Fred Decker—Will your future work on this include enough observation on hail so that you establish the area where the hail fell?

Prof. Sänger—We intend to evaluate the hail fall in three areas separately, and also with regard to intensity. Our observations have been made according to the international scale of intensity.

Dr. Helmut Weickmann—I am particularly happy about your comment, Mr. Namias, because it emphasizes the importance of meteorology in the problem of the evaluation of the seeding experiments. I like to recall the days of the great argument between Dr. Langmuir and the U.S. Weather Bureau on the subject of the generation of a seven-day rainfall period as a consequence of periodic seeding with one AgI generator in New Mexico. In those days the U.S.
Weather Bureau unfortunately did not employ synoptic meteorology analysis to the problem as you just suggested, but it tried to fight back with Langmuir's own weapons, with statistical analysis. Here, of course, Langmuir's genius could not be defeated and in the bombardment of his ever new and different significance tests and correlation factors which however contributed so little to the basic physical meteorological argument did the Weather Bureau lose rather than gain ground.

*Mr. Namias*—Could I just make one last remark? I meant to point out that in reality this whole problem indicates the extreme importance of numerical forecasting procedures. If we can get an objective rainfall forecasting method, we can find answers and in ways involving much less emotion.
A Project for a Formation of Cumulonimbus by Artificial Convection

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Abstract—The researches on the weather control should not be restricted to the nucleation of the clouds. The convection process plays a first importance part in the formation of the precipitation. After several experiments with great brush fires, an experimental installation capable of modifying the local system of convection sufficiently to produce rain has been planned. This installation called 'meteotron' will consist of a pump serving 100 burners with fuel oil, so that the total consumption will be of the order of one ton per minute. The results of the experiments with great brush fires indicate that the formation of artificial thunderstorms will probably be obtained, at least in equatorial regions, by working with the meteotron.

Historical introduction—At the conference in Tucson, in April 1956, Byers recalled the suggestions made by Espy [1841] that it would be possible to stimulate the formation of rain-giving Cumulus clouds by means of large fires. In 1938, 100 years later, Gorog and Ropo [1938], on the bases of some observational data, proposed that gasoline could be burned in large quantities with the same effect. In 1930, Dessolières [1930] showed how the men should be able to increase rainfall progressively in arid regions "en créant et multipliant au milieu des terres et des eaux des aires de surchauffe solaire de quelque cent hectares destinées à provoquer l'appel, la convergence d'énormes masses d'air et leur ascension vers le zénith ou, plus brièvement, en créant et multipliant les centres de coordination atmosphériques."

Preliminary experiments—The first experimental trial for the stimulation of convective ascending air currents consisted in clearing in the equatorial forest a surface of 0.2 km². In effect, in the surrounding forest the thermal capacity of the vegetation and the high evaporation tended to reduce the daily variation of temperature, in particular to reduce the daily maximum temperature. Over the cleared area, the solar radiation heated the soil to high temperature and the air temperature is also raised. In fact, it was observed that Cumulus clouds did tend to form over the cleared area and the study of these clouds has allowed a theory called that of 'coupled convection' [Dessens, 1956]. But it appears that to obtain Cumulus clouds sufficiently developed to give rain, there were also the following requirements: (1) the cleared area should be at least 1 km² in extent; (2) it is not sufficient really to clear or to burn the vegetation, but it would be necessary to make use of an uniform artificial covering for the area; and (3) the convective effect is only noticeable when low winds are very light.

To cover the terrain will be an expensive proposition and the results at least uncertain. For these reasons, instead of following the idea to use only the solar energy, we preferred to have recourse to another source of energy, very much cheaper, for short series of experiments: the energy liberated by the burning of the vegetation. We were immediately struck by the case with which, in the equatorial regions, large Cumulus clouds can be obtained by this method, on a condition solely that the fire should be started at a suitable time in the day. These fires should be made with careful regard to the meteorological conditions. When these are favorable, large Cumulus clouds can easily be stimulated. These observations confirm entirely the ideas expressed in the last century by Espy [1841].

Artificial cloud seeding or artificial convection—Since 1955 we have made experiments to attempt to increase precipitation on an area of the Congo Basin at Lukolela (1°S, 17°E). Two methods have been employed: (1) the use of rockets to send large sodium chloride particles up to cloud base; and (2) the use of silver iodide emitted by burners situated at ground level. Beginning in 1956, we have undertaken these simultaneously with experiments of artificial convection. The preliminary conclusion which we have reached after these two series of experi-
ments, is that the only falls of rain which we can attribute with certainty to our intervention are those which have been stimulated by artificial convection [H. Dessens and J. Dessens, 1956, 1957].

As far as the cloud seeding is concerned, the results are particularly disappointing, because the conditions under which we work seem exceptionally favorable: (1) the base of the Cumulus clouds was less than 1000 m above the ground, so that the rockets can easily reach the cloud base. (2) In this region there was a complete absence or at most very few natural ice-forming nuclei active at temperatures above -30°C. This was shown by measurements made with an ice-nuclei counter, at ground level, by Soulagé [1956], by the photogrammetric study of the clouds themselves by J. Dessens [1959], and also by observations of large subcooled Cumulus over the southern Congo and Rhodesia made by Schaefer [1958]. A third factor in favor of the seeding was the fact that the winds are generally light and of very steady direction, which facilitated the evaluation of the results of seeding. The last favorable condition is that the precipitation is principally due to local convection clouds.

In contrast with the apparent lack of any effect on the precipitation of the seeding, the evidence for the positive effect on the Cumulus of the artificial convection is very striking. Furthermore, these first experiments have permitted us to calculate approximately the amount of energy which has been liberated by combustion to trigger the formation of Cumulus or Cumulonimbus in definite local conditions in the dry and in the rainy season.

The ‘Meteotron’: a brief description—I use the word ‘meteotron’ to describe any experimental arrangement capable of modifying the local system of convection sufficiently to form frequently clouds which may produce rain. Our experience with brush fires suggests that the energy to be liberated must be comparable with that received from the Sun, that is to say of the order of 10^9 kw/km^2. This energy must be immediately available and ready for application to take advantage of the meteorological situation; that is, that it is necessary to use some liquid or gaseous fuel. We chose to use fuel oil. Our ‘meteotron’ therefore includes a station for the preparation of the fuel with a pump driven by a diesel motor. The motor drives also a small generator which produces alternating current necessary for automatic firing of the fuel at a distance.

We are, at present, thinking in terms of a distribution of 100 burners connected to the central post by tubes, these 100 burners situated on a circle of approximately 250 m in diameter. Each burner will burn 600 kg of fuel oil per hour so that the total consumption will be of the order of one ton of fuel oil per minute in the whole system.

Site of the experiments—The most suitable region for a first trial of this method, would, I think, be between latitudes 5°N and 5°S, in a region of plains during the rainy season. The severe droughts observed in 1958, south of the equator, justify such an experiment for the economic point of view. Regions such as the Mamybe, the Nari Valley or the Congo Basin will be very suitable. The relatively light winds, the absence of cyclonic perturbations, the permanent presence of a very humid air layer at low levels, the low cloud-base level are favorable factors. It is important to realize these differences between conditions in the tropics and those in temperate regions.

Method of use of the ‘meteotron’—The most delicate part of the operation is the choice of the moment to start. Our first objective is modest: just before the moment at which the general instability becomes sufficient to permit the natural formation of Cumulonimbus, we must light our burners, that is to say, anticipating the natural formation of Cumulus by a short period. Six minutes after lighting the burners, the first artificial Cumulus cloud appears. After 15 min the development of this cloud is greater than that of the other Cumulus in the region. So that, instead of leaving it to chance the location of the first formation of Cumulonimbus over the area, we have chosen and imposed this location. In brief, what we do, is to add a little artificial impulse to the natural instability of the air, which is due to the heating of the soil by the solar radiation of the whole area. This impulse starts half an hour before the time at which the normal Cumulonimbus will begin to form generally over the whole region in radius, say, of 50 to 100 km without intervention.

Despite the fact that the experience of previous days gives us some ideas of the moment at which Cumulonimbus will begin to form, it is necessary to be able to make, each day, a short-term forecast, in order to know exactly when the ‘meteotron’ should be put in the action. This
forecasting for periods less than an hour may be a delicate matter, but various methods are already envisaged to make it more objective. Such methods might include the observation of 'sferics,' the use of radiosondes at the critical moments and the study of the intensity of diffused light from the zenith.

When the operation started, the experiment will consist of observing low-level conditions, how long it would be possible to go on localizing the convection over the experimental area, and how long it would be possible to continue influencing the precipitation over the area.

Possible future development of artificial convection—From the foregoing discussion, it appears that our first objective would be to try to concentrate the rainfall over a cultivated area, possibly at the expense of a slight deficit of rainfall over surrounding regions. For example, if we succeeded in increasing the rainfall over 500 km² of cultivated area by 30%, it is possible that we will cause a decrease of 3% over 5000 km², which we assume will be relatively wild, that is to say forest or savanna. This decrease in the rainfall on the surrounding regions could have no economic disadvantage because it is not cultivated.

Before finishing this exposition of this project, I would like to discuss some of the future perspectives which the artificial convection opens to us and which have much wider application than that which we considered so far.

(1) In tropical regions, during the dry season, it is commonly found that the humid air at ground level, is overlain by dry air. It sometimes occurs that this humid layer is not sufficiently deep to permit the natural formation of Cumulus Nimbus. In this case the artificial impulse given by the burning of fuel, may provoke a local deepening of the humid air layer and thus the very local formation of cloud. In this case, it is possible to imagine, when during a dry season, the cloud formation is not general, that local cloud and precipitation may be produced. We have already succeeded in Africa, by brush fires, in producing some local light rainfall (1 mm) in situations when no natural rainfall occurred in the region.

(2) In temperate regions, in situations where instability occurs along a chain of mountains such as the Pyrenees, our intervention at critical moments might result in a concentration of stormy precipitation at higher areas useful to hydroelectric installations.

(3) From a study of the very heavy storm rainfall in southern France, sometimes accompanied by devastating hail storms, this phenomenon results from the coupling of the thermal energy of low air with the kinetic energy of the jet stream, through quite stable and localized ascending air currents. By analogy, some coupling effect might be obtained, in situation with strong jet-stream, with artificially stimulated convection. There would be advantage that, in this case, the location is controlled by the choice of place for the experiments.

The 'solar meteotron'—The first experiments have to be made with a 'fuel meteotron' because its cost is relatively low, less than $100,000, while the preparation of the surface for the differential heating by the Sun with the proper materials will require a very much higher expenditure. I hope, nevertheless, that some participants at this Second Conference at Woods Hole, will agree that, in the end, it will certainly be necessary to consider either the 'solar meteotron' or a 'mixed meteotron' in which both solar energy and fuel are used. The real domestication of solar energy will come only with the use of a 'meteotron' to provide a rain-giving cloud.

References


Discussion

Mr. W. A. Mordy—I lived for a number of years in Hawaii. In the process of harvesting sugar cane in the Islands they set fire to the fields to get rid of the leaves and waste material before they harvest the cane. They burn 15 to 20 acres of cane at a time. Frequently a cloud forms over these fires. At the appearance of these clouds such as we have seen in Dr. Dessens’ slides, we have occasionally made time-lapse movies. When I first went to the Islands, I heard a report that such a cloud had in fact yielded a half an inch of rain on one occasion. This, of course, was interesting to us, since these clouds were very frequent. They could be seen about fifty times a year. It therefore became a pastime of mine to follow up on these whenever possible. As I traveled around the Islands I did this. And I was very interested to see if I could find another instance when a cloud had, in fact, formed rain. I never found another instance, and I could not confirm the first report either. In ten years we found not a single case when one of these clouds produced rain.

Dr. H. Dessens—It is necessary to start the fire under favorable meteorological conditions; these are latent instability and no wind. Both conditions occur in our equatorial regions. Rain has been obtained by us in the first experiments but it was light rain, amounting to less than one millimeter.

Dr. Tor Bergeron—I have in fact another project for getting more precipitation in central and northern Africa: the main thing is that one should not let the waters of the Rivers Congo and Nile run out into the sea, but one should use any available source of energy for keeping them inland, and especially within the regions where there is already some vegetation. The second point is that the water should certainly not be used for irrigation in the northern part of the Sahara or lower Egypt, because it will be evaporated into dry air without clouds to a great extent. But further south, of course, the project of Dr. Dessens would probably come in very handy. I just wonder if these two projects could not go hand in hand and help each other, and I hope, in fact, we shall have an opportunity to do so, but then I have one question about the possible cost.

Dr. Dessens—From the economic point of view this project seems to me more useful than the project of sending rockets to the Moon.

Dr. C. E. Junge—I think it is a very good idea to start making use of the huge amounts of energy which are available in an unstable atmosphere just by initiating the vertical convection by means of bringing in some heat. I think this may really be the first step in a quite new direction of weather modification which should be much more exploited.

Major C. Dowin—We should also look into the chemistry of this process. Fuel oil during combustion produces about one and one-quarter (1¼) pounds of water per pound of fuel, depending on the particular hydrocarbon mixture involved. Thus an appreciable amount of moisture, in addition to the heat energy, is released into the atmosphere. That both of these factors are important is borne out by the Geophysics Research Directorate study of aircraft condensation trails, another example of artificially produced clouds.

Mr. Mordy—We did make estimates on the basis that one adds very much water. One would expect the cloud base in those cases would be different from other clouds in that area. But if you compare the effect of the added water and the effect of released heat, you find the energy added makes the difference. It is not the water. Actually, there are three possible variables: the nuclei, the water, and the energy.

Dr. Bernard Vonnegut—In his comments a few minutes ago, Dr. Bergeron hinted at some ideas he had on the control of climate, and he has consented to give us a brief, but more extensive outline of this.

Dr. Bergeron—As you know, there are great plans for building a new dam at Aswan, in Egypt. For immediate needs, this dam will be very useful, but not for more far-sighted planning of water resources. In fact, the irrigation water spent in the arid northeast trade region will only be utilized by vegetation to a very small percentage. Part of it will go through plants and be useful once, and then never more. But the rest of it will not even go through plants once: it will go directly up in the air and from there, anywhere. It will not come down again in North Africa, because there are no clouds there. True enough, we should not allow the
Fig. 1—Schematic picture of rain distribution and general circulation in lower troposphere over Africa in July

waters of the Nile, Niger, and Congo to reach the ocean. However, they should not only be kept inside Africa, but inside the southwest monsoon, between the two tropical fronts shown in Figure 1. We should use any sources of energy, in the future atomic energy, for distributing it rationally. If the water is spent where we already have vegetation and clouds, most of it will go through plants, and then that humidity will come down again and again.

Figure 1 shows the structure of the general circulation over the whole of Africa in July as observed from the ground. This analysis is based on ideas I already had in 1925; part of it was published in 1929, but nobody here has seen it. The southern tropical front moves down to Rhodesia in January. The northern one lies along 18-20°N in July, and there must be remains of it near the equator at the opposite time of the year. So, it seems as if the southwest monsoon were enclosed between two tropical fronts. The northern one we know very well by now; this was confirmed by the work of Brooks and Mirrless in 1934.

The precipitation distribution in Figure 1 is shown by shading. Evidently, the equatorial precipitation is not at all directly connected with the front. It is a pure air-mass precipitation, which falls within the region where this very humid and unstable air mass is deep enough to be able to produce Cumulonimbus. In fact, it forms a flow of air that one may really follow. So far as I know, there is no other place on Earth where you have an air mass that is so well defined. You have the even flow of moist air coming in from the ocean. It can not pass through the northern tropical front at lower levels, since this front surface is a well-established and marked one. It leaks out above, at the top of the Cumulonimbus clouds, but not so very much, because the temperatures are so low there, and thus amounts of humidity are very small. So, we can use the water of these rivers for important irrigation projects within the outskirts of
the monsoon region, in the savannas. There, it will go up and down many times before it leaks out east of Ethiopia and aloft, and perhaps also somewhat through the southern front. This picture holds good not only for July, but for all the time from April or May until October. In fact, the tropical front passes Khartoum (15°N.) in the middle of May on its way north, and passes the same place again moving south in the middle of October.

In the future, when we have economical atomic energy, we can even have factories evaporating sea water, thus enriching the southwest monsoon with humidity, and let the monsoon itself carry and distribute the humidity over the region, if huge irrigation plants prove too costly to construct. Moreover, the present cooling effects within the monsoon area will be intensified both indirectly by increasing cloudiness, shutting off the sunshine, and indirectly by the increased precipitation. Then the northern tropical front will move north a little. How much, we do not know, but it is quite likely to move north until it reaches a position where it cannot go further north for reasons of the general planetary dynamics. If you now have increased amounts of humidity, this project of Dr. Dessens is very useful because then it will pay even more to utilize the latent instability for releasing even more showers. So, I think these two projects might be able to cooperate nicely in the future when we have more sources of cheap energy.

Mr. Jerome Namias—It is refreshing to see a man of Dr. Bergeron's stature becoming interested in weather modification problems. Supporting his discussion, I want to bring out a thought which may have been overlooked in dealing with this problem. In the first place, of course, the problem will not achieve a desirable solution until the numerical part of it is solved, so that we know exactly the sequence of events after we have treated weather or cloud conditions artificially. Now, in this conference, where all sorts of ideas are set forth, we may consider possibilities which extend further into the future. Presently, a science is emerging from research in extended forecasting which deals with 'tele-connections,' the long-distance interactions of weather regimes where one or more effects in one area affect other areas in a very dramatic way. One of the aims of long-range forecasting is to find such areas which are in resonance with other areas. It appears possible that certain forms of the general circulation exist which are rather precariously balanced, and which might be influenced if man acquires the amounts of energy commensurate with what is believed to be necessary and, of course, knows how and where to apply it. Then those people who have worked many years on these problems, namely, those who have experience in extended long-range forecasting, should be consulted as to the weak spots and precariously balanced states in the atmosphere; that is, the circulations that are on the verge of doing one thing or another. Here is an example: The weather over the North American continent, and particularly the United States, is greatly influenced by a special type of intense cyclone which often forms in the Gulf of Alaska. If it is there, it tends to bring about one type of weather through its influence on the large-scale circulation over North America; if it is not there, another quite different, meteorologically opposed, type occurs. The development of this cyclone is very explosive. It takes place rapidly, and is sensitive to the general upper-air flow. If this flow comes from Alaska (with extreme cold air heated over the Gulf) the rapid heating and distortion by the coastal ranges induces strong cyclonic vorticity. Since this appears to be a rather sensitive mechanism, it may lend itself to artificial control. I therefore feel that synoptic studies of the general circulation and its weak spots should be carried on right now, so that if the day comes when commensurate supplies of energy are available, we will be in a better position to undertake projects of large scale weather modification.

Dr. Bergeron—May I just declare that I am more than in agreement with my friend on these points. I have always thought of these critical thresholds in the general circulation, how the pattern may just jump over from one preferred shape to another; and that there may be rather small quantities of energy needed. And, of course, I am also quite aware that dynamists, the long-range forecasters, and those working with numerical computers are the people to take care of certain important sides of that problem. The reason I chose that part of the world (North Africa) was, that down there the problem seems to be even simpler, because you do not need to change very much, just a little increase in the humidity, and that would be perhaps the best region in the world at present where one could utilize the humidity directly without too much loss.
Physics of Precipitation in Winter Storms at Santa Barbara, California

Clement J. Todd

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Abstract—The immediate purpose of the physical analysis of storms at Santa Barbara, is to increase the sensitivity of the statistical analysis, (a) by finding physical explanations of as much of the variability between target and control as possible, and (b) by finding physical methods of classifying which precipitation mechanisms are operating in order that the storm periods may be stratified according to a reasonable hypothesis of seedability for a statistical test.

Introduction—Meteorology Research, Inc., is engaged in a study using radar and other tools of physical analysis to see what light can be shed on the physics of natural and seeded precipitation in Southern California winter storms. The study is in conjunction with the Santa Barbara Cooperative Seeding Project. The Cooperative Seeding Project was designed as a statistical study to evaluate a commercial seeding operation. In the design of the statistical experiment the primary concern is on safe guards against all sources of bias.

The seeding operation is as follows; if the commercial operator forecasts the ensuing twelve hour period to be seedable he is instructed to seed or not to seed according to a random procedure controlled by the statistician. The physical observations are all passive.

Physics of rain distribution—Probably for a long time to come the best estimate of what would have been the unaffected precipitation over a target will come from the comparison of the actual precipitation in an unaffected control. There have been many attempts to make this comparison more sensitive by stratifying the storms according to synoptic scale types. Our approach is to go to the mesoscale and vertical structure and classify the storms into short periods that have physical similarity.

Operating a 3-cm radar at La Cumbre Lookout on a 4000-ft ridge six miles north of Santa Barbara for the last three years has convinced us that there are several different types of storm periods. The types are apt to persist for several hours at a time and they recur in different storms and in different years. The consistency of types is no doubt due to dominance of terrain effects. Only the surface has been scratched in understanding the significant physical causes of the different types, but we have a faith that understanding will come from inquiry.

Just scratching the surface has shown that a most pronounced rain anomaly is associated with the height of the stability layer. If it is low with respect to the ridge of the Santa Ynez Mountains, the surface air does not flow easily over the range. It piles up and is forced out to the west, as indicated by strong east surface winds at Santa Barbara. Figure 1 shows a cross section of the February 25, 1958 storm, an un-

Fig. 1—Time-height cross section over Santa Barbara for February 24-25, 1958, a typical low-level stable layer storm; broken lines represent air temperatures in °C; solid lines, the difference between air temperature and dew point.
complicated example of this structure. Figure 2 shows the hourly rainfall distribution of this storm, over (a) the island control area and four east-west strips of target, (b) the coastal plain, (c) the first ridge, (d) the valley, then (e) the next ridge to the north. The maximum occurs in the coastal plain and the first ridge. These same storm characteristics have been observed in Sweden by Bergeron [1949] and are there discussed in illuminating detail.

Figure 3a shows the scatter diagram of the island control stations against the mainland target stations for the stable storms of 1958. The rain pattern is consistent from one stable storm period to another.

Figure 3b on the other hand shows the scatter diagram for the unstable storm periods of the same year. Note that the slope of the regression
line is markedly different and that the correlation has degenerated considerably.

The storm of April 2, 1958, Figure 4, is an example of deep instability. Figure 5 shows hourly rain distribution. Here the lift starts close to the coastal ridge and there is little precipitation on the coastal plain, and the islands are missed completely during long periods of the storm. This might be interpreted to mean that at times cyclonic component of the storm produces no precipitation while the orographic boost sets off substantial rain.

There are many significant phenomena that repeat several times during a season, that we have not begun to tie down quantitatively yet. A distinct frequent and spectacular one consists of plumes of precipitation that point back to the high points on the ridges of the Channel Islands. There are silver iodide generators on the Channel Islands, and the first year we thought that we had a simple observable effect of seeding. The plumes, however, occur when the generators are off as well as when they are on. Another interesting phenomenon which has appeared several times is a plume of precipitation that points back to the channel between the two main islands, apparently identical to what Fujiwara [1958] has observed in Japan.

Precipitation mechanism—The possibility of discriminating between the condensation-coalescence and the ice-precipitation processes from ground observations seems good. As reported by

![Fig. 5—Hourly rainfall averaged over orographic regions for the typical convective storm of April 2-3, 1958](image-url)
MacCready and others [1958], both the drop-size distribution and the potential gradient have characteristics indicative of the precipitation origin. When the radar and vertical cross section indicate that the precipitation must be warm-cloud origin, the drop distribution follows the Blanchard type and the potential gradient shows fair-weather field. When the evidence points to ice-precipitation origin, the drop spectrum fits the Marshall-Palmer distribution and the potential gradient goes negative, or fluctuates, and may show discharges [MacCready and others, 1958]. In Figure 6, Drop Samples 1 to 9 are given as dots and show good agreement around the straight line representing idealized Blanchard distribution; while Samples 10 and 11, the broken lines, show marked disagreement. Figure 7 shows samples 10 and 11 agree well with Marshall-Palmer distribution line. Figure 8 shows the potential gradient record for the period. Note the change just before 20h 00m. Radar indicated that Samples 1 to 9 were warm cloud

![Graph](image_url)
FIG. 7—Raindrop size distribution expected from ice origin rain is shown by straight line; dashed line shows goodness of fit of Sample 10 and 11.

origin and that ice precipitation started just before 20h 00m and is represented in Sample 10 and 11.

Figure 4 is a time-height cross section of this storm. The warm cloud precipitation occurs when the top of the moisture layer is lower than −5°C. Ice precipitation starts when it goes above −5°C. Incidentally, this storm was seeded. Ice precipitation becomes general when the top of the moisture layer is above −15 or −20°C.

Sometimes the drop size distribution indicates that the two precipitation processes occur together.

We have not yet been able to go far enough to separate the artificially induced and natural ice precipitation on anything approaching the general case. We will have to intensify our physical measurements. Equipment is now on hand, such as vertical radar, electronic raindrop size counters, along with more potential gradient devices and nuclei counters. These did not receive full field application this year due to the history making dry period the latter half of the seeding season. In addition to the above it would be very helpful to make coordinated aircraft observations on top of the storm.

Comments—Mesoscale analysis combined with appropriate physical observations will certainly be the backbone of seeding analysis. In order not to use up all available degrees of freedom it is of utmost importance to make intensive use of mesoscale analysis of historical storms. This is especially true here in Southern California where a relatively good network of upper-air soundings provides an extraordinarily good chance of relating mesoscale structure and seeding effects, where seeding has gone on for ten years now and where two steel foundries are highly suspected of producing seeding anomalies [MacCready, 1957].

Conclusions—As we become skillful at using our physical tools, we have the possibility of developing greatly increased sensitivity of test, to a degree where we will be able to check hypothesis of when, where, how much, and under what conditions seeding is effective.

Acknowledgement—This research was supported by the National Science Foundation and the Department of Water Resources, State of California.

References


Discussion

Dr. Tor Bergeron—I like this paper very much. Especially, I appreciate the fact that we see here the influence of the orography. That is very much to be appreciated, and of course, you have also used every other means at your disposal for the physical analysis of these precipitation patterns.

I would like to repeat what I said in my open-
Fig. 8—Potential gradient trace during storm when Samples 1 through 11 were taken; note sharp swing to negative with Sample 10 and 11 when drop size distribution changes from typical warm cloud to ice originated size distribution.
ing lecture, that I should like to have more connection between the different scale ranges from micro over meso to macro scale. In my terminology this means from the cloud particles to the individual clouds and from there to the precipitating systems as seen in the synoptic scale. During most of the discussions on radar studies, which were quite admirable in themselves, I could not determine what the general synoptic situation looked like. I think it is important to have this in papers.

Dr. Walter Hilschfeld—I want to know what Mr. Todd meant by the plumes that were continually coming from the island toward the coast?

Mr. C. J. Todd—Well, they were streams of echo, like smoke plumes. But they did not start at the island; they would line up in straight lines of precipitation and the extrapolation of these lines would point right back to the ridges on the island.
Artificial Nucleation of Orographic Cumulus Clouds

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Abstract—A brief summary is given of the results of randomized cloud-seeding experiments carried out during two summer seasons. Orographic Cumulus were seeded with an airborne AgI generator. The results to date have suggested that seeding causes important changes in the physical processes involved. It is planned to continue the experiments.

Introduction—During the last two years, the University of Arizona has been carrying out a program of cloud-seeding research directed towards the study of the effects of silver-iodide nuclei on supercooled orographic Cumulus clouds. Clouds over the Santa Catalina Mountains in southeastern Arizona were studied.

Since the program is still in progress, this note is a brief summary of some aspects of the research. When the research is completed, the results will be published in the geophysical journals.

Design of experiment—The design of the seeding experiment was evolved with the assistance of K. A. Brownlee and W. Kruskal of the Department of Statistics of the University of Chicago. Briefly, the procedure involved an objective prediction, made prior to 09h 00m MST of each day, as to whether or not Cumulus congestus or Cumulonimbus clouds would form over the Santa Catalina Mountains. The main criterion for the prediction was whether or not the precipitable water at Tucson, Arizona, equalled or exceeded 1.10 inches. When this occurred, the day was considered to be suitable for seeding, and an envelope was opened which specified which of two suitable days would be seeded. If more than one unsuitable day separated two suitable days, the first day of the pair was rejected and a new pair was started. The scheme of randomized pairs was adopted in order to take into account day to day correlations and to assure that there would be an equal number of seeded and not-seeded days.

The actual seeding was carried out with an Australian-type airborne silver iodide generator suspended under the wing of a Super cub airplane. The generator has been made available through the cooperation of the University of Chicago. The flight plan involved repeated passes at about the -6°C-level along a track upwind from the mountain range. The pilot normally started the generator at about 12h 30m MST and continued his flight until all the seeding material was exhausted or the burner went out. Normally the seeding period was of the order of four hours. The generator consumed a 20% solution of silver iodide in acetone at a rate of 2 to 2½ gal/hr.

Observations—In order to permit studies of cloud and precipitation processes the following observations were taken: (1) visual cloud properties were recorded on a pair of carefully calibrated ground-located K-17 aerial cameras from which accurate estimates of cloud locations and dimensions could be measured; (2) the location and spread of precipitation echoes were observed with a vertically scanning 3-cm radar set; (3) rainfall was noted with a network of 29 recording rain gages; and (4) visual observations were made of the time and location of cloud-to-ground lightning strokes.

Results—The experiments conducted during the first two years suggest that AgI seeding caused some important changes in the natural cloud processes. During each summer 16 pairs of days were studied.

Rainfall—When data from both years were combined it was found that the mean rainfall per gage was 30% higher on the seeded days; however, the probability that the observed differences in the mean rainfall occurred by chance was quite high, about 0.14. This value was obtained from a sign-rank test which made use of a ranking of the differences of the mean rainfall of pairs of days. A comparison of the extreme rainfalls on seeded and non-seeded days showed greater differences, but the statistical confidence of a real difference was still not sufficiently high to be acceptable.

Heights of thunderstorms—An objective way to measure the relative frequencies of large
Fig. 1—Fraction of clouds observed to extend to the indicated temperature levels which contained precipitation echoes; data on which the curves are based are shown at the top of the diagram; ten clouds, five in each sample, had temperatures above +6°C and have not been plotted.

thunderstorms is to take radar observations every 30 min and note whether there is at least one cloud extending above any particular altitude. When this was done, it was found that during the seeded days there were about twice as many echoes extending above 30,000, 35,000 ft, and 40,000 ft. A sign-rank test in this instance showed that the probability that the differences in the number of clouds extending above 30,000 ft occurred by chance was 0.05.

Lightning—Lightning observations were not taken in 1957. However, in 1958, it was found that on the seeded days there were about nine times more lightning strokes than on the non-seeded days. A sign-rank test revealed that the probability of chance occurrence of the observed ranking of the differences of strokes on pairs of days was about 0.015. It was interesting to find, that notwithstanding the large difference in lightning frequency, there was little or no difference in the number of lightning-caused forest fires. One might offer the explanation that the higher lightning frequency was offset by more rain which reduced the likelihood of the formation and spread of fires.

Initiation of precipitation—By means of the cloud camera and radar data, it was possible to note the vertical extent of clouds (and thus cloud-top temperatures) and whether or not they contained precipitation. When a sufficient number of clouds have been examined it becomes possible to speak of the 'probability of precipitation' in clouds whose summit temperatures are between −12 and −15°C, or any other temperature interval. Figure 1 shows a summary of the observations made during 1957 and 1958. The smoothed solid and dashed curves were drawn in by eye. It is quite obvious that on the seeded days, the likelihood of precipitation was greater than on the non-seeded days. The fairly uniform shift of the curve towards the left lends support to the belief that the effect is real, and that in fact, the AgI seeding caused the formation of precipitation in clouds which would not have precipitated naturally.

If the nearly straight parts of the curves are extended to the abscissa (dotted lines) it is found that the 'not-seeded curve' intercepts the abscissa at about −17°C. It might be argued that this result is reasonable because observations of ice nuclei in the atmosphere show that in general, the concentrations at temperatures above −15°C are small. As the temperature is reduced the concentration increases. It appears reasonable to assume that the dotted-dashed curve represents clouds in which the ice-crystal mechanism was effective in causing precipitation.

An extension of the 'seeded curve' shows that it intercepts the abscissa at about −9°C, a temperature just below the value at which AgI crystals can be expected to become effective as ice-crystal nuclei.

If the interpretations of the significance of the dotted curves are correct, then one is led to the assertion that those precipitating clouds which fall to the left of the projected curves produced the precipitation by the condensation-coalescence process.

Summary—In view of the fact that this research is still in progress, the authors feel that they are still not ready to draw final conclusions. Although the results to date point towards the conclusion that seeding produced important effects, it is vital that more data be compiled in order to be sure that the results have not been
brought about by chance. This brief note has been written as a progress report for others working on similar problems. After more data are collected, it is hoped that some of the pressing questions in the important area of cloud seeding can be given definite and unequivocal answers.

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Discussion

Mr. J. Namias—The evidence looks very interesting here, and very convincing at first glance. I would just like to ask about the method of selection. You pointed out, if I understood correctly, that the criterion of whether you would have clouds was determined from the amount of precipitable water?

Dr. L. J. Battan—That is correct.

Mr. Namias—And then you chose a pair of days with the seeded day determined by the process of randomization. The question is whether you had a bias when you seeded on either the first or the second day. It is therefore very important that the data on which the evidence is based is evenly distributed between the first and the second day.

Dr. Battan—The data were very well distributed. This was one of the things the statisticians looked up right away.

Mr. Namias—That would be a rather important consideration, for if there were any bias, your criterion would imply a certain synoptic situation, the second day of which is not at all independent of the first, so you could have a certain situation that would either augment or inhibit. But if the evidence is evenly distributed between the two days, my point does not hold.

I was thinking also of the possible factor involving the preceding day’s precipitation and its effect on the surface. As I pointed out earlier, a relationship between thunderstorms occurring in southern Texas on one day and their lack in the same area on the subsequent day has been ascribed to wetting of the ground the first day.

Dr. Battan—This is exactly why we use randomization by pairs of days, because if you randomized by day, one might be led to seed or not seed on three or four consecutive days. With the present scheme the most you can get is two consecutive seeded or not-seeded days.

Dr. R. D. Elliott—I might add that the State of California Forestry Service has conducted cloud seeding experiments in the northern part of the State for a number of years to see if lightning could be reduced, and the answer so far seems to be that the lightning is actually increased on the seeded days. The experiments are randomized, but the outstanding factors are that the precipitation is increased, and this fits very well also with what Dr. Sänger said about the Swiss Hail Suppression Experiment.

Mr. C. E. Anderson—I want to raise a point about whether or not these experiments are truly randomized from the seeding-agent standpoint. This would apply to both the Swiss experiment and to the ones you conducted in New Mexico, where you released silver iodide on a particular day for seeding, and then did not seed the next day. I recall the seeding experiment that the Weather Bureau conducted on the West Coast as part of the Artificial Cloud Nucleation Project which the United States Government sponsored some five years ago. During this time we had made some trials in cooperation with Ferguson Hall who was in charge of that project in the State of Washington, with release of zinc sulphide tracer materials from the aircraft at 15,000 ft. We were quite surprised to find that the zinc sulphide turned up (and this was during precipitation at the release level) in the rain gauges, not that day, but the next. Therefore I am just wondering whether or not, when you release this in the atmosphere, you can depend on its being removed overnight, so that the next day is truly a fresh sample.

Dr. Battan—We do not have measurements, which can dispel this argument; but I find it hard to understand how it can happen.
Cloud Seeding in the American Tropics

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Abstract—Cloud-seeding projects that have been carried out in the American tropics are described, and evaluation of the results are summarized and discussed in the light of existing models of precipitation formation. A new model based on a field of competition in which convective clouds develop is proposed and discussed to explain the effectiveness of seeding under weather conditions where clouds develop relatively slowly toward the precipitating stage.

Introduction—We should hope that an examination of the results of cloud-seeding trials in the tropics will reveal essential agreement between the results observed and the expectations based on observations of the natural mechanisms of rain formation in the tropics and in laboratory and theoretical studies of natural and artificial mechanisms for the release of precipitation.

Most of the trials of cloud seeding in the American tropics have been of a practical nature, intended to increase the rainfall for a specific purpose or need, or, in a few instances, to diminish the force of locally destructive winds. The operations have been for the most part conducted by commercial firms and in a few cases by the agencies directly interested. The instances of cloud seeding performed by persons entirely uninvolved with the outcome have been with few exceptions incidental to other aspects of the investigation of cloud physics. In no instance known to the author have such seedings resulted in data suitable for evaluation to determine whether seeding applied for practical purposes could be effective.

Up to the present time, no evaluations have been performed by any competent independent agencies of the practical seedings carried on in the American tropics, nor has any specific rebuttal been made to the reports by the cloud seeders of practically successful results. It would be desirable, perhaps, to limit examination of seedings trials to those carried out under impartial auspices in conformity with certain standards as to seeding agent used, the control of its dispersal into the air or cloud, the adequacy of rainfall data, and the objectivity of the procedures of analysis. For the present, however, such a limitation would end our examination at once, and so we must examine what is at hand, and substitute suspicion for the luxury of impartiality.

Seeding experiences—A series of commercial projects conducted for the Francisco Sugar Co. is described next because it will serve to illustrate the conditions under which a majority of the projects were conducted and the manner of their operation. The target covers some 500 sq mi of nearly flat land for about 20 mi along the south coast of Camagüey Province in east central Cuba, and extending 15 to 20 mi inland. It is shown as the two right-hand areas in the map, Figure 1. The climate has been described in detail elsewhere [Howell, 1953]. Operations were begun in the summer and fall of 1951, using six ground-based silver iodide smoke generators of the acetone-propane type which dispersed about 50 g/hr of silver iodide. These were located at various points in and immediately upwind from the target at points where instructions for their operation could be given by telephone, and they were operated in groups of two to four at a time on 33 days during a period of three months at the times and places judged appropriate by the field meteorologist on the basis of forecasts and local pilot-balloon observations. Subsequently, in 1952, similar operations were conducted over the same target for 11 months, encompassing most of the dry season and all of the showery season, and during several of these months operations were conducted also for the adjoining properties of several neighbors.

In subsequent operations a different type of smoke generator was used, which burned string, impregnated with silver iodide, in a propane flame and dispersed silver iodide at a rate of about 10 g/hr, with about 15 generators situated in and upwind of the target. Operations were conducted in 1953 for five months which showed
a 15% increase in rainfall. In February and March of 1954, experiments were conducted on the spray seeding of warm clouds, but operations were suspended because of adequate field moisture. They were not resumed again until May of 1956, because of sugar marketing and milling restrictions, but have been conducted much of the time since then. Operations were somewhat upset by local political disturbances late in 1958, and some of the generators were put to uses other than those intended by the sugar mill, including perhaps some connection with an isolated dry-season shower that bogged down a column of tanks as it was about to close in on a large rebel encampment located within the target. Perhaps some future historian will be able to tell us if this was the first direct use of cloud seeding to discomfit an enemy.

There were several rain gages located in various sections of the Francisco property, and the adjacent properties also each had several rain gages. It was therefore possible to assemble a considerable network of gages, as shown on Figure 1, with at least ten years of record, 1941-1950, before the start of seeding. The first attempt at an evaluation was made after the 1951 season by comparing the percentage of normal rainfall received during the operating period on the target with the percentage normal received on the surrounding plantations, and it was estimated that an increase of 26% had been achieved. After the longer operations of 1952 over both Francisco and some adjacent plantations, the evaluation was improved by using a regression analysis using the seasonal total rainfalls for the operated portion of the year for each year of the historical record. This evaluation indicated an increase of 28% with significance for the group of operations well beyond the one per cent level. By this time we had thoroughly scoured the countryside for rain gages, and from then on the composition of the control data was not changed except to drop the Lagarenos gages—located under the cartouche of Figure 1—when it was found that they had no useful degree of correlation with the target. The evaluation of later years, however, was further refined by basing the regression on individual monthly rainfall totals instead of seasonal, and by transforming them to normality. The two evaluations made subsequently have shown respectively 15% and 25% increases. Figure 2 shows the regression with the results of the 1956 operations. The probabilities of chance occurrence are arrived at by a one-tailed t-test, allowing for the size of the sample and the departure of the control variable from its mean. Where several months of operation of a project have been tested against a single regression, the t-tests have been combined into one.

Seeding of warm clouds has been carried out,
<table>
<thead>
<tr>
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<th>Year</th>
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<td>Rio Mantaro</td>
<td></td>
<td></td>
<td>Category</td>
<td>Duration</td>
</tr>
<tr>
<td>Cuba, Francisco</td>
<td>1951</td>
<td>300</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Cespedes</td>
<td>1952</td>
<td>100</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Ermita</td>
<td>1952</td>
<td>50</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Francisco</td>
<td>1952</td>
<td>300</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Macareno</td>
<td>1952</td>
<td>150</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Najasa</td>
<td>1952</td>
<td>50</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Hawaii</td>
<td>1952-53</td>
<td>...</td>
<td>Uninvolved</td>
<td>Water</td>
</tr>
<tr>
<td>Cuba, Los Canos</td>
<td>1953</td>
<td>100</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Macareno</td>
<td>1953</td>
<td>150</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Cuba, Baltony</td>
<td>1953</td>
<td>30</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Preston, Boston</td>
<td>1953</td>
<td>300</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Francisco</td>
<td>1953</td>
<td>400</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Puerto Rico, Fajardo</td>
<td>1953</td>
<td>250</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Nearby waters</td>
<td>1953-54</td>
<td>...</td>
<td>Uninvolved</td>
<td>Water</td>
</tr>
<tr>
<td>Cuba, Macareno</td>
<td>1954</td>
<td>300</td>
<td>Commercial</td>
<td>AgI, wa-</td>
</tr>
<tr>
<td>Puerto Rico, Fajardo</td>
<td>1954</td>
<td>250</td>
<td>Commercial</td>
<td>water</td>
</tr>
<tr>
<td>Cuba, Baltony</td>
<td>1955</td>
<td>30</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Puerto Rico, Fajardo</td>
<td>1955</td>
<td>250</td>
<td>Commercial</td>
<td>AgI, wa-</td>
</tr>
<tr>
<td>Cuba, Francisco</td>
<td>1956</td>
<td>550</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Colombia, Santa Marta</td>
<td>1956-57</td>
<td>350</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Category</td>
<td>Duration</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Uninvolved</td>
<td></td>
</tr>
<tr>
<td>Cuba, Havanu-Mantanzas</td>
<td>1956</td>
<td>4000</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Manati</td>
<td>1956</td>
<td>400</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td></td>
<td>1957</td>
<td>400</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Puerto Rico, south coast</td>
<td>1957</td>
<td>400</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Cuba, Baltony</td>
<td>1957</td>
<td>30</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Hispaniola, Romana</td>
<td>1957</td>
<td>600</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Cuba, Esperanza</td>
<td>1957</td>
<td>120</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Los Canos</td>
<td>1957</td>
<td>100</td>
<td>Commercial</td>
<td>AgI</td>
</tr>
<tr>
<td>Florida, Boca Raton</td>
<td>1957</td>
<td>...</td>
<td>Uninvolved</td>
<td>AgI</td>
</tr>
<tr>
<td>Average of all AgI seedings</td>
<td>(24 cases)</td>
<td></td>
<td>Category</td>
<td>Duration</td>
</tr>
<tr>
<td>Average of all AgI and water seedings</td>
<td>(3 cases)</td>
<td></td>
<td>Uninvolved</td>
<td></td>
</tr>
</tbody>
</table>

---

a Excepting 1952.  
b On commercial basis prior to 1955.  
c Separate evaluations of runoff showed increase of 35%.  
d Water seeding with coarse spray in tops of trade Cumulus. Probability figure refers to likelihood that seeding caused precipitation in a cloud that would not otherwise have precipitation [Braham and others, 1957].
in the projects included in this study, by spraying either plain water in droplets with a mean diameter about 70 microns or by spraying sodium chloride solutions with a mean droplet diameter about 40 microns. In either case the spray was introduced at or near the bases of the clouds in the updrafts. In the discussion the process has been referred to simply as water seeding.

In Table 1, which shows the results of cloud-seeding trials in the American tropics, distinction has been made among three classes of project: (1) commercial, those conducted by a cloud seeder for a client where the cloud seeder presumably is motivated to show a positive result; (2) practical, those conducted by an agency which is motivated by pecuniary interest to obtain an accurate evaluation of the result, either positive or negative; and (3) uninvolved, those conducted by an agency having no involvement other than the search for knowledge. The table is not exhaustive, and I know that there are other trials about which my present information is too meager to make inclusion of them in this examination useful.

The individual percentages of increase cited in this table are extremely varied in reliability, some of them seem well founded, others are quite dubious. Hence it is unlikely that great confidence will be placed in the mean increases shown, but it is nevertheless interesting that the combined water and AgI seeding seems to show a clear advantage over AgI seeding alone.

Certain among these data are worthy of closer attention, however. They represent the cases where a target-control relationship has been established, usually subsequent to a first season of seeding; and then the same relationship, unaltered in any significant degree, has been used to evaluate subsequent seedings. There are six such cases, and they are tabulated separately in Table 2. These represent seeding trials where the game has been played according to pre-established rules. It will be noted that the mean increase of 19% among them compares closely with the 22% among all cases.

These values, especially the ones representing probability of chance occurrence, should still probably be regarded with reservations. It is conceivable that there are secular changes in the climates of target and control zones that would alter the target-control regression sufficiently between the historical period and the seeding period to affect the value of the test, though such changes remain to be demonstrated, and it is doubtless true that other possible selections of control region or historical period or both would have produced different results. However, it seems that it would take an extraordinary degree of prescience on the part of the evaluator to select a control that would at some future time be biased in favor of a false indication of seeding influence.

Leaving to the statisticians the question whether the figures tabulated afford reliable proof that the seeding increased the rainfall, one thing nevertheless is clear: a hypothesis that silver iodide seeding is incapable of increasing the rainfall in tropical clouds over land is open to serious doubt. The rest of this paper will be devoted to examining the implications of this statement.

Should AgI seeding affect tropical Cumulus?—The development of precipitation in shower clouds, both in the tropics and in middle latitudes, has been the subject of extensive study.

<table>
<thead>
<tr>
<th>Location</th>
<th>Year</th>
<th>Seeding agent</th>
<th>Rainfall increase</th>
<th>Probability by chance</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mexico, Necaxa</td>
<td>1950-58</td>
<td>AgI</td>
<td>+9</td>
<td>0.0001</td>
<td>Orographic, maritime air</td>
</tr>
<tr>
<td>Peru, Rio Chieana</td>
<td>1952-58</td>
<td>AgI</td>
<td>+25</td>
<td>0.0001</td>
<td>Orographic, continental air</td>
</tr>
<tr>
<td>Cuba, Baitony</td>
<td>1955</td>
<td>AgI</td>
<td>+12</td>
<td>0.03</td>
<td>Valley location</td>
</tr>
<tr>
<td>Puerto Rico, Fajardo</td>
<td>1955</td>
<td>AgI, water</td>
<td>+27</td>
<td>0.07</td>
<td>Windward shore, mostly flat</td>
</tr>
<tr>
<td>Cuba, Francisco</td>
<td>1956</td>
<td>AgI</td>
<td>+25</td>
<td>0.005</td>
<td>Flat, lee shore</td>
</tr>
<tr>
<td>Cuba, Manati</td>
<td>1957</td>
<td>AgI</td>
<td>+15</td>
<td>0.21</td>
<td>Flat, windward shore</td>
</tr>
<tr>
<td>Average</td>
<td></td>
<td></td>
<td>+19</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
In a census of tropical clouds Byers and Hall [1955] have shown that over the sea the likelihood that a cloud contains rain is related closely to the size of the cloud, and that all clouds were found to contain precipitation before they reached a size that would bring their tops above the freezing level. In clouds over the land, both in Puerto Rico and in the central United States [Battan, 1953], weaker connection was found between the size of clouds and the probability of precipitation being present in them, but it was concluded from the timing and position of appearance of radar echoes in developing shower clouds that the onset of precipitation was due to a coalescence process. A fairly well-developed model for the formation of precipitation by a coalescence process has emerged from the suggestion by Houghton [1938] that the few particles at the large end of the drop-size spectrum play an important part and the work of Bowen [1952], Mason [1952], Ludlam [1951, 1952, 1956], Keith and Arons [1954], East [1957] and many others in developing quantitative expressions for the growth of droplets by coalescence. These have been used by MacCready and others [1957] to develop graphs which he employed successfully in Project Shower to predict the time of onset of precipitation and to indicate that this precipitation originated by coalescence rather than by the Bergeron-Findeisen process. It has been widely concluded as a result of these observations and analyses that the presence of ice crystals plays no part in the onset of precipitation from most convective clouds and only a secondary role in subsequent developments, and that therefore, although a basis is established for supposing that water seeding may sometimes stimulate Cumulus clouds to produce showers, the supposition that AgI or dry-ice seeding could have such an effect is discouraged. If the apparent effectiveness of AgI in stimulating precipitation is to be given a reasonable explanation, it must be through the development and verification of a somewhat modified model of shower formation in Cumulus clouds.

Considerations for a new model—Our first survey of cloud conditions in Cuba, undertaken in the summer of 1951, showed the regular occurrence of convective clouds over the land which reached the freezing level a considerable time, frequently hours, before the first onset of precipitation, and the phenomenon of 'cirrus pumping' was also observed which appeared to indicate that some clouds reached even to the level of −40°C temperature where homogeneous nucleation occurred and caused the cloud tops to glaciate without increase of particle size, detaching themselves and floating off as separate cirrus umbrellas (see Fig. 3). The marked difference in characteristics between the clouds over the land and those over the water was noted, as has later been well documented by Malkus [1953, 1955], with the generally higher level of convective activity in the land clouds. These observations appeared to indicate that considerable volumes of cloud frequently endured for some time in a supercooled state without raining, or before the onset of rain, and led us to believe that AgI seeding had at least a reasonable chance of being effective in releasing rain.

These and subsequent observations led us to distinguish between two typical sequences of cloud development leading to showers. Sequence I is characterized by the nearly continuous and rapid growth of the cloud from humble beginnings, through the congested stage and on to the formation of Cumulonimbus, the entire development taking place in perhaps half an hour to an hour. Such a sequence occurs typically when the conditional instability of a deep moist layer is released by a definite impulse such as the arrival of a sea-breeze front. As the cloud develops, rudimentary precipitation particles are formed at a much faster rate than they are lost by evaporation at the top and edges of the clouds, and precipitation begins promptly as soon as the first-formed particles are sufficiently aged. The time interval during which artificial influences could operate to speed up the formation of precipitation is extremely short, with consequent small likelihood of any considerable effect being produced by seeding. Rainfall is likely to be widespread in the general region where cloud development occurs.

Sequence II is characterized by a much more gradual increase in the state of development of the clouds, even though convective activity remains high all the while. Typically many clouds of nearly the same size may be seen that display great activity of growth in their lower parts and dissipation in their upper parts, the average size of the clouds gradually increasing and their number decreasing. This state of affairs may continue for many hours until it is ended by the afternoon decrease of diurnal heating, or it may be brought to an end by the development of precipitation in one or another of the clouds. When
under circumstances such as these precipitation does become well established in one cloud, it is hard to avoid the impression that a marked increase in the rate of growth of the cloud is often connected with the onset of precipitation, accompanied by the degeneration of other clouds in the vicinity. New towers rise from the top of the precipitating cloud to much greater heights than formerly and take on the aspect of Cumulonimbus; the cloud base darkens; and the impression of roiling, tumbling growth-and-dissipation activity is replaced by one of swift organization of a large-scale convective cell. The ensuing rainfall, while sometimes heavy, is likely to be much more spotty than that accompanying Sequence I.

Both sequences, together with gradations between them, were observed during our seeding activities in Cuba. After the first season’s work we drew daily isohyetal maps [Howell Associates, 1952] and examined them for connections between the pattern of rainfall and the pattern of seeding. It was immediately noticed that about a third of the maps showed good correspondence
between the areas of rainfall and the positions of the smoke plumes at from one to two hours of wind travel from the generators. Another third of the maps showed some weak connection, and in the remaining third no connection was found. Figure 4 illustrates an exceptionally good connection. Furthermore, it was noticed that most of the days when there were isolated heavy showers showed a good connection, while most of the days when rain was more or less general showed weak connection or none at all. The analyses suggested that the seeding was most effective under Sequence II conditions, and it is these conditions that form the basis for a new model of shower development. Although it was not found possible to treat these results rigorously, they together with the pilot balloon runs and local cloud observations gave the field meteorologist a feeling of considerable confidence in directing the seeding effects onto the target.

The field-of-competition model—We have taken for our model not a single cloud but a portion of the atmosphere, overlying a uniform ground surface, extensive enough to contain several convective cells. We suppose that heat and moisture are added slowly at the bottom of this atmosphere, creating a "moist layer" of conditional instability which gradually deepens through the upward transport of heat and moisture by convective clouds. Horizontal transport is presumed sufficient to maintain more or less horizontal uniformity throughout the region. Above the moist layer the air is assumed to be relatively dry and with slight positive stability.

This part of the atmosphere constitutes a field within which a number of convective cells compete for the potential energy that they can convert to air motions. At each stage in the heating and deepening of the surface layer there is an optimum size of convection cell that represents a balance between the advantages of larger cell size for drawing energy from more air and the disadvantages of lengthening horizontal transports, the optimum size becoming larger as the surface layer deepens; and all the competing cells will tend to approach this optimum size. Within the field of competition there will therefore be a number of clouds that in their mature stage are of nearly equal size and in which the probability of precipitation is nearly equal.

Following the model developed by Ludlam [1956] and others, we consider the clouds as formed by a series of bubbles or ring vortices that entrain air from their environment, producing energy in the lower part of the cloud, while at the top and sides of the cloud exchange of air with the environment causes evaporation, cooling, and subsiding motion. The bubbles, while within the cloud, lose water by mass exchange but not by evaporation, but once they reach the cloud top they begin evaporating rapidly, cooling and falling back as they dissipate but leaving a moister environment for the next bubble. Cloud droplets form in a variety of sizes depending on
the updraft at the condensation level and on
the size on the condensation nuclei, and we may
regard some fraction of the largest cloud droplets
as rudimentary precipitation particles that,
if they survive long enough within the cloud,
will become raindrops.

The rate of formation of rudimentary precipita-
tion particles in a given cloud depends on the
concentration of giant hygroscopic nuclei, oc-
currence of collisions, etc. Survival of them in
the evaporating part of the cloud, or if they are
thrown out of the cloud, depends upon their
being large enough to maintain their existence
until they are re-entrained or fall back into the
cloud. Ludlam [1956] has estimated that a di-
diameter of about 150 microns is critical for dro-
plets carried by a bubble out the top of a cloud;
those smaller will be lost; those larger will fall
back into the cloud and continue growing. The
establishment of precipitation, we assume, re-
quires the survival of some critical number of
rudimentary precipitation particles in the region
of the cloud where the liquid-water content is
high. Or, to express it another way, the cloud
must accumulate a certain quantity of particle
seniority in its rain factory. If the operating
force is insufficient or the turnover is too high,
this necessary quantity of seniority will not be
accumulated, and although some few raindrops
may fall, precipitation will not be established.

But when this critical seniority is reached and
precipitation is established, a mass of water that
approximates, according to Braithwaite [1952], 20% of
of the water entering the convective circulation
(and probably with respect to the water in
the active portion of the cloud a somewhat
higher percentage) is removed by the rain and
falls from the cloud. This water represents not
only a loss of mass from the cloud but also
recipient of the heat that would otherwise have
gone to re-evaporate this water in the upper
branch of the circulation. Both these effects
operate in the direction of increasing the convec-
tion in the cloud in which rain forms as com-
pared with similar clouds nearby.

Now, during the time that there are a num-
ber of clouds having approximately the same
size, the ascending air currents will tend to oc-

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hence the instability it experiences will be much
greater; its convective circulation will increase
rapidly in magnitude and depth. The energy-
producing part of the circulation is then able to
draw on the whole depth of the moist layer as
a source of moisture and energy, whereas for-
merly only the lower portion of it was available
so. These several effects working in concert cause
the precipitating cloud to grow rapidly in size
and in the intensity of its circulation, suppressing
the unsuccessful competitors. Figure 5, which
is a photograph of a Cumulus cloud 13 min after
being seeded by dry ice published by Kraus and
Squires [1947] and reproduced here through the
courtesy of P. Squires, appears to show this ef-
effect occurring under conditions similar to those
of our model.

At any time during the gradual growth of the
convective clouds, we may describe their collec-
tive approach to the rain stage as a frequency
distribution of the property that we have called
quantity of seniority in each cloud's rain factory,
which can be represented in the manner of Fig-
ure 6 as an ogive of cumulative probability of
the seniority having surpassed the value critical
for production of rain in some one of the clouds
within the model. If there are many clouds in
the group, by the time the percentage of clouds
surpassing the critical limit rises to a few per-
cent (at a 'seniority' of about 12 on Figure 6), it
becomes very likely that rain will have begun
somewhere, in one of the clouds within the field
of competition. We then presume that this oc-
currence will be followed by rapid growth of the
successful cloud and depress the level of convec-
tive activity elsewhere, causing the average sen-
iority of the remaining clouds to diminish again.

Let us consider the effect on this model of seed-
ing one of the clouds with water droplets or
hygroscopic particles during the time that the

clouds is approaching the stage critical
for the formation of rain somewhere in the group.
The seeding in effect employs in the seeded cloud
a large number of pre-aged rudimentary pre-
cipitation particles and thereby increases the
quantity of seniority in the operating force, and
places the seeded cloud somewhat higher on the
seniority scale than its neighbors, perhaps only
by a few percent, or one unit higher on the arbi-
trary scale of Figure 6. But it will be noted that
before the group as a whole reaches the point
where the formation of rain in it becomes likely,
the seeded cloud will have a very good chance
of becoming a rain-producer.
Fig. 5—Photograph of a Cumulus cloud 13 min after seeding with dry ice (Courtesy of P. Squires)

Fig. 6—Ogive illustrating an assumed typical relationship between the mean aging of a group of clouds in the field-of-competition model and the probability of precipitation onset somewhere in the group.

And now let us consider the effect on our model of seeding one of the clouds with AgI. No effect is to be expected until the top of the cloud reaches a temperature of about $-5^\circ$C. But when this temperature is reached in the cloud, some ice crystals will appear and some of the rudimentary precipitation particles will freeze, either because of infection with AgI or through collision with an ice crystal. The volume of air within which these frozen particles will continue to grow encompasses all that within which the air is saturated with respect to ice and therefore is considerably larger than that within which the unfrozen particles can grow; and even in dry air the frozen particles evaporate more slowly. Furthermore, as Douglas [1960] has shown, the frozen particles may enjoy as much as a two-times advantage in growth rate by accretion over their unfrozen neighbors. The result is that the rate at which the cloud loses rudimentary precipitation particles is diminished and the rate at which these particles grow or ‘acquire seniority’ is in-
creased. In exactly the same way as with the water-seeded warm cloud, then, the seeded cloud is given an advantage that expresses itself in the form of a much improved chance that the seeded cloud will be the one in the group that will first develop precipitation. It will be noted that this effect of AgI seeding operates through the coalescence mechanism and that it is unnecessary to postulate the independent growth of new rudimentary precipitation particles in the form of ice crystals entirely by sublimation in order to account for an effect of the AgI seeding.

When the seeding releases precipitation within the field of competition, the ensuing chain of events draws to the seeded cloud the energy from a much larger share of the atmosphere than it could otherwise have reached. In effect, the "signal" energy released directly by the seeding is amplified many times over, and the magnitude of the output is determined not so much by the strength of the signal as by the energy resources of the system. The outcome will be sometimes the initiation of rain on a day when rain would not otherwise have fallen, in which event the precipitation efficiency of the model as a whole is increased; and sometimes the outcome will be to direct onto the target area the center of shower development that might otherwise have been elsewhere, and to a certain extent to increase the precipitation efficiency of the shower mechanism.

Discussion—Of course, natural occurrences present all gradations between the Sequence I and Sequence II types of development described above. On the other hand, certain locations are more suitable for the occurrence of one sequence or the other. For example, we may contrast the situation of the Francisco target, already discussed, where the usual trajectory brings air for some considerable distance over a rather uniform level terrain, with the situation of the Necaxa Watershed in Mexico and with the Boca Raton, Florida, site of the Project Seabreeze studies. In the latter two, strong localized impulses are available to initiate convection, the orography in the Necaxa location and the sea breeze at Boca Raton, so that one might expect more frequently to find Sequence I followed at these sites. Indeed, examination of the plots of cloud top height against time for the Project Seabreeze observations [MacCreedy and others, 1957] shows that in most cases the clouds more than doubled in height and reached altitudes greater than 25,000 ft within less than an hour. The effect of AgI seeding might therefore be expected to be less at these locations than on the Francisco target. Referring back to Table 2 we note that the rainfall stimulation at Necaxa is indicated to be less than that at Francisco, in accordance with this expectation.

It has often been noted that horizontal shear in the wind sometimes decapitates growing Cumulus clouds, carrying away the portion of the cloud containing the best-developed rudimentary precipitation particles and thereby preventing or delaying the onset of precipitation. However, shear also contributes to the energy available for convection when the cells reach an appropriate size; and if the field of competition is characterized by vertical shear, a sudden jump in the growth rate of the clouds will appear when they approach the size at which they can utilize the energy of the shear for their growth. However, this jump will probably appear throughout the field at about the same time, and indeed it may have much to do with the occasional development of Sequence I clouds causing more or less general rainfall. However, our experience in the tropics suggests that the jump usually does not occur until a cloud has begun precipitating, and that the shear energy then contributes to the selection of the favored cloud for still more vigorous development.

The operation of the field-of-competition model suggests that the effects of cloud seeding will be extremely variable, often ineffective but sometimes extremely effective, with an average effectiveness that depends to some extent upon the habitual sequence of cloud developments over the seeded region. These characteristics are indeed suggested by the data on actual seeding so far accumulated. The model further suggests that, if it is practical to isolate for experimentation those situations characterized by Sequence II developments, effects of seeding will occur that may be large enough to make their reality immediately apparent.

Another implication of the model is that experiments performed in clouds situated over localized sources of convective impulses, such as isolated mountain peaks or small tropical islands, while capable of making trial of the physical changes that may be produced within a single cloud by seeding, will tend not to demonstrate to full advantage the significance of the com-
petitive advantage that seeding may give to one cloud in relation to others.

It is planned for the present to develop further the model that has been proposed here and to attempt to bring it to the point where experimental verification of it can be sought.

References
BRAHAM, R. R., Jr., The water and energy budgets of the thunderstorm and their relation to thunderstorm development, J. Met., 9, 227-242, 1952.
DOUGLAS, R. H., Growth by accretion in the ice phase, this volume, pp. 264-270, 1960.

Discussion

Dr. Joanne S. MALKUS—I think this is one of the best descriptions of tropical clouds. I have read Dr. Howell’s paper in some detail, and have listened to his lecture, and I like this approach on cloud modification very much. I think this is because Dr. Howell is not just dumping in stuff and making statistical analyses, but he is also asking questions concerning the processes at work.

Dr. Bernard VONNEGUT—Our observations in New Mexico show that at about the same time that the cloud suddenly grows and gives precipitation there is also an extremely rapid buildup of electricity. The electric field frequently doubles every minute or two so that in ten minutes it can increase by two or even three orders of magnitude. I think it would be exceedingly interesting in observations of clouds to observe not only the onset of precipitation but also the development of electrification.

Dr. Walter HITSCHFIELD—I would like to ask a question about the system of organization that you mentioned. You started out with a population of small clouds; one of the clouds for some reason is able to develop ahead of the others, and then somehow is able to organize the development of the remainder of the clouds. What is the scale? What is the range over which such an organization could be active; and further, have you any idea of how it might work?

Dr. W. E. HOWELL—First as to scale and range. Dealing with impressions rather than measurements, I would say that clouds approaching the critical stage have a depth, typically, of 15,000 to 20,000 ft, and are roughly the same in diameter, three miles or so. After passing the critical stage they may grow perhaps five times their horizontal size and to a depth of perhaps 50,000 ft, still guessing. This impression was gained from
repeated flights at 20,000 ft altitude: at that altitude, you do not seem to be even halfway to the tops of the mature clouds.

As to the mechanism, of course the latent heat released by freezing was suggested very early, but I think this may be only a small part of the story and that the more important effect is the removal from the precipitating cloud of the condensed water that would otherwise cool the air off again by re-evaporating in drier air entrained into the cloud. You might say that precipitation removes latent cold from a cloud and leaves the heat locked in. I think it is an interesting and fruitful field for investigation to study other possible feedback mechanisms. For instance, a cloud, once slightly favored on account of latent heat, will then also have a competitive advantage in utilizing the energy of shear. Also, the downdraft chimneys set up by precipitation shafts may be important in increasing the efficiency of convective overturning.

The behavior of the clouds strongly suggests that feedback of some sort is working here, and I do not believe that Dr. Weickmann’s plumbing model of precipitation stimulation can be considered complete without a valve controlling the supply to the spigot that can be operated by the water coming out the holes.
Artificial Precipitation Potential during Dry Periods in Illinois

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Abstract—The macroscale meteorological conditions of the atmosphere were studied during 31 dry periods which occurred in 1953–1955. A dry period is defined as at least five consecutive days with less than ten per cent of the normal precipitation over an area in east central Illinois.

The parameters investigated, measured at Rantoul, Illinois, were: precipitable water, low cloudiness, and the Showalter stability index. The upper-air flow, surface temperature, and general synoptic conditions were considered in individual case studies.

The results indicate that although there is near normal water vapor in the atmosphere during the majority of the dry periods, there was a deficit of low clouds. It is concluded from the study that in addition to present cloud-seeding techniques, much research is needed to determine means of initiating clouds, since, during dry periods in Illinois, large quantities of clouds desired for seeding are not available.

Introduction—Bergeron [1935] postulated nearly 30 years ago that precipitation was the result of microphysical processes in clouds containing a mixture of ice particles and subcooled water. However, little progress was made toward a better understanding of the precipitation process during the ensuing 20 years. Langmuir [1948] and Schaefcr [1948] began to make further progress in the study of precipitation physics with their laboratory and field experiments in cloud seeding. These experiments definitely illustrated the possibility that man could artificially affect the microphysical processes taking place within clouds.

Following these pioneer experiments, considerable research has been directed toward modifying clouds by attempting to increase the efficiency of the physical processes which initiate or enhance precipitation. While it is of the utmost importance to determine the physical structure of the clouds, it is equally important to examine the macroscale conditions attending these cloud formations.

The President's Advisory Committee on Weather Control [Orville and others, 1957] recently disclosed that the most effective cloud-seeding experiments have been performed in areas where pronounced orographic effects are present. Little evidence has been found for establishing the success of seeding clouds to increase rainfall in flat-land areas, such as the Midwest.

This paper summarizes an attempt to examine some of the more obvious macroscale parameters of the environmental atmosphere, such as precipitable water, low cloudiness, and stability during periods when artificial stimulation of precipitation is most needed; that is, during periods when the natural precipitation process is either not functioning or is very inefficient.

Analysis and results—The dry periods chosen for study in this investigation consisted of at least five consecutive days during 1953–1955 in which no measurable precipitation occurred at Rantoul, Illinois. To assure that no portion of overlapping wet periods was included in the dry-period analysis, the beginning and ending days of each dry period were deleted from the dry period. Thus, a defined dry period of five days was, in reality, a period of seven days without precipitation at Rantoul.

Inasmuch as a point observation of rainfall sometimes can be unrepresentative, the dry periods, as determined from the rainfall records at Rantoul, were investigated further on an areal basis. An average areal value was calculated from the precipitation values for seven stations within a 50-mi radius of Rantoul. The dryness of each period was then evaluated by comparing the daily average areal precipitation of the period to long-term daily averages. If less than ten per cent of the average amount was observed, the period was accepted as a dry period. This selection yielded 31 cases for investigation as dry periods.

To conduct this study, the Illinois State Water Survey obtained IBM punch cards from the
U. S. Weather Bureau containing upper-air soundings at 12-hr intervals during 1953-1957 at Rantoul, Illinois. The precipitable water content for the layer from the surface to 400 mb was computed from the punch-card data by a digital computer at the University of Illinois.

The Showalter stability index was computed in the standard manner for the twice daily observations at Rantoul during the five-year period. A negative index value indicates unstable conditions and the numerical value is related to the degree of instability. Monthly median values were determined from these computations and are used as normals throughout the remainder of this paper.

Cloud data for this study were obtained from a previous study of cloud distributions in Illinois made by Changnon and Hoef [1957] of the State Water Survey. The hourly observations of clouds were summed to provide a total number of tenths per day of individual cloud types. In discussing cloudiness during dry periods, only low-cloud data were used since the observation of higher clouds is very dependent on the amount of low cloudiness. Furthermore, cloud seeding techniques to initiate or increase precipitation are more readily performed on the low clouds.

The clouds used in this investigation are: Stratus, Stratocumulus, Cumulus, Cumulonimbus, and Nimbostratus.

Daily normal values of precipitable water, stability index, and low clouds were calculated and used as a basis for determining the normality of observations obtained during the dry periods. The data obtained for this investigation were studied on an annual, seasonal, and individual case basis. Table 1 illustrates the median values of precipitable water, cloudiness, and stability for the 31 dry periods during the years 1953-1955.

At the same time nearly 80% of the observations were more stable than the five-year median values. Thus, very stable atmospheric conditions, which are unfavorable for the formation of precipitating clouds, are predominant during dry periods.

An examination of the 31 cases involved in this study disclosed that in 90% of the dry periods the cloudiness experienced was less than the normal amount. However, in only 71% of the periods was the amount of atmospheric moisture below normal. Eighty per cent of the periods investigated were more stable than the five-year seasonal median values of the stability index.

In an attempt to determine facts more pertinent to the problem of weather modification, these data were studied on a seasonal basis. Table 2 illustrates the results of the seasonal investigation.

It is noted in the Table that during dry periods in winter there is considerable more cloud cover than in any of the other seasons, even though the winter dry periods are 31% below normal. The percentage of precipitable water in winter is relatively high, indicating a supply of moisture available for the precipitation process if it is initiated. However, the stability index is more stable than would be expected for the winter months which have a five-year median of +12.7. Admittedly the meaning of the Showalter stability index in the colder months of the year is somewhat dubious, but it does serve to

<table>
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<th>Duration</th>
<th>Precip. water</th>
<th>Clouds</th>
<th>Stability Index</th>
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<td></td>
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<tr>
<td>8.0</td>
<td>90.3</td>
<td>58.0</td>
<td>+8.0</td>
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* Per cent of normal determined from daily averages.

<table>
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<th>Summer</th>
<th>Fall</th>
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<td>8.0</td>
<td>8.5</td>
<td>10.0</td>
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<td>Clouds, per cent of normal</td>
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<td>19.2</td>
<td>49.5</td>
<td>49.3</td>
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<td>76.8</td>
<td>95.4</td>
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<tr>
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</tbody>
</table>
illustrate a comparison between the dry periods and the median values obtained from upper-air observations during 1953–1957.

The spring months are clearly below normal in cloudiness while the precipitable water is still relatively high with 77% of the normal value. The remaining two seasons have over 90% of the normal precipitable water, but have less than 50% of the normal cloudiness. Since cloud modification practices to initiate or increase precipitation are dependent on the existence and amount of low cloudiness, the data presented indicate that investigations on methods of initiating low clouds are equally as important as the seeding of existing clouds. van Straten [1958] recently attempted seeding cloudless air with carbon black to initiate clouds. The results of these experiments are encouraging, and certainly justify more research on the physics of cloud initiation. In all the dry periods investigated, there appears to have been an ample supply of moisture in the gaseous state, but for some reason the condensation process was not operating efficiently. It was suspected that the lack of cloudiness was due to the absence of vertical motions which are necessary to the condensation process. However, an inspection of the individual dry periods does not indicate an absence of the conditions which would support convective activity in a number of cases. As an example, a ten-day dry period in July–August of 1953 was above normal in precipitable water content, and relatively unstable with a median stability index of 4-0.5. However, only 33% of the normal low cloudiness was observed. The surface maximum temperatures during the period were above 90°F, and a visual inspection of the thermodynamic structure of the atmosphere did not reveal any unusual, stable layers. Therefore, many of the conditions conducive to convective cloud formation were observed, but the clouds did not form.

An examination of the upper-air flow during this dry period revealed predominantly northwesterly flow over Illinois. In view of the fact that the macroscale conditions leading to cloud initiation and precipitation were present, it seems likely that the flow from the northwest was deficient in condensation nuclei. The Illinois State Water Survey, under a grant from the National Science Foundation, is now engaged in an airborne particulate sampling program to determine the distribution of nuclei during a wide variety of synoptic conditions. The results of the particulate sampling research will provide data on the distribution of condensation nuclei during dry periods similar to that described.

It is evident from the data presented that, in the majority of the dry periods investigated, the clouds and instability which are necessary for initiation of precipitation were not present and therefore, cloud-seeding practices would not have helped to alleviate the problem.

Summary—This study has brought to light some of the interesting features of the conditions of the large-scale atmosphere during periods of little or no precipitation in the Midwest. The study is presently in its initial stages and will be developed more fully, but the preliminary results present some interesting aspects of the problem of weather control. If we are to alleviate drought in the agricultural areas of the Midwest, it appears that in addition to seeding existing clouds to increase precipitation, methods for initiating clouds must be developed.

If Illinois is considered representative of the Midwest, a large quantity of clouds desired for seeding are just not available during dry periods.

Acknowledgments—The writer wishes to thank William C. Ackermann, Chief, and Stanley A. Changnon, Climatologist, Illinois State Water Survey for their helpful comments, discussions, and careful review of the manuscript. This work was accomplished under the immediate supervision of Glenn E. Stout, Head of the Meteorology Section of the Illinois State Water Survey.

References


Discussion

Dr. W. E. Howell—May I speak to this question from the point of view of rain-making experience. Of course anyone in my position makes 90% of his new contacts during the 10% of the driest weather. One is always asked the question, “Can you make it rain in the middle of a drought?” I can honestly say that our reply is “No.” However, that is not the end of the matter. These people do not come to us in the first five dry days; they are not particularly appalled by any five-day period without rain. It is when they have gone 30 or 60 days with only a fraction of the normal precipitation that they begin to worry about drought. By then, the question is not, “How suitable is a typical drought day for cloud seeding?” but “How frequently during an extended drought do seeding opportunities present themselves?” We have made, therefore, a rather detailed study for Louisiana defining a drought situation as a period beginning with the sixth day after a general rain, and continuing as long as the cumulative amount of precipitation was less than a tenth of an inch a day. In the 39-month period, January 1949–March 1952, we found 400 drought days, covering a trifle over a third of the time, in 37 individual drought periods, mostly of a few days duration; but there were several more extended periods of drought: for instance, August 26–December 20, 1950; April 28 to June 10, 1951; and November 8 to December 17, 1951. Now, in the longest of these, August 26 through December 20, 1950, there were 31 out of 116 days, about ¼, during which small amounts of rain fell somewhere on or near the target. It was similar for the other drought periods.

It is our feeling that these occasions that punctuate the typical drought are quite good for cloud seeding when the precipitation mechanism almost reaches the stage of rain but does not quite get there; our experience working in this kind of condition is on the whole quite favorable. We feel that we can not actually break a drought, but we also feel that we can considerably diminish its intensity during certain critical parts of the drought.

Mr. Jerome Namias—I would like to say a little about this problem of drought, and also mention a few things about criteria for the formation of rain. The criteria Mr. Semonin used were the Showalter stability index and the total precipitable water. Those two indices are certainly not the best criteria for the occurrence of rain. It is well known that the vertical component of air motion is the central factor in the formation of rain, and Mr. Semonin’s criteria would often not be very wise choices. Also, variations in precipitable water of five per cent have to be considered against the knowledge of the normal variability; and not considered small. We have to know the frequency distribution of it, particularly in the Midwest, where this distribution is extremely important. Now, with regard to the situation Mr. Semonin described; I have tried to look it up, although the precise dates were not mentioned in his abstract. I could get just a broad scale picture, but I have studied that drought period as well as many others.

Figure 1 depicts the 700-mb conditions attending the great Dust Bowl drought of 1936. It is somewhat familiar in nature to the period that arose during the seven-year drought from 1951 to 1956 over Texas and adjacent areas of the southern Plains. The main thing to note in this figure is the great anti-cyclone in the mid-troposphere into which dry air from the westerlies is recurrently injected. A good deal of subsidence takes place, resulting in lack of clouds, and the excess insolation during the daytime raises temperatures as much as 10° above normal; this means that temperatures between 100 and 110°F are frequent over the drought area. The two conditions, very hot and very dry, go hand in hand.

An important point is that this great drought-producing cell depends in large part on the existence of two neighboring cells, one in the Atlantic and one in the Pacific; both have to be anomalously strong, so that they are, so to speak, in resonance. If one or both of the oceanic cells should vanish, the U.S. cell would die. But once this mechanism is set up, there are also some life-sustaining properties for the continental cell which remain to be explained. It might be, for example, that Squires’ suggestion of the different nuclei counts and kinds in continental and maritime regions might be important here. There are also some other factors which might be raised, such as changes in the characteristics of the underlying surface itself, and presence or lack of water on this surface. In other words, solar heat may be used for evaporation or for
building the upper level anti-cyclone, depending on the character of the surface.

Figure 2 shows an isentropic analysis for this particular month (August 1936). This is a sloping surface of constant potential temperature (315° A) whose heights are given by the broken lines. Shown also are the circulating moist and dry tongues which are largely responsible for the outbreak or inhibition of showers. You will note in this great drought-producing cell that dry air is frequently flung from the westerlies over Canada, around the cell, finally entering into the central portion of this great anticyclonic eddy. At the same time, this dry air moves from higher to lower elevations as it spins into the cell. The dry subsiding air inhibits the formation of clouds. As a matter of fact, if Cumuli do form, when they penetrate this dry layer a sort of entrainment takes place, so that they are dissipated quickly, by the mixing with this dry air. Now, in the southern part of Illinois it is extremely dry. There are deficits on the order of two to four inches in the Plains area of the United States. In some areas there is no rain at all, and the rain is confined to the left-hand portion of the moist tongue, where the air is to some extent forced up slope. A similar case was observed during August 1955, and Figure 3 shows
the mean 700-mb pattern and the precipitation anomaly. This is the August which I presume is one of the periods treated in Mr. Semonin's talk. You will note the anticyclonic cells are also pronounced, as in the 1936 case.

What I am trying to emphasize is that drought is generally a very large-scale phenomenon, and its modification may require us to deal with events quite remote from the immediate drought area. One should be very careful not to draw conclusions about his efforts to modify drought on a large scale by rather localized seeding, particularly in areas where there are few clouds, and these not of the proper type. As I indicated in my earlier talk, all of us as meteorologists must be aware of these large-scale problems.

Mr. R. G. Semonin—The dry period was in August 1953, and was very similar to the mean pattern you have shown. You had asked how the amount of precipitable water is related to precipitation. I have been trying to find out for two or three years now, and I have found no definite relationship. I just wanted to show what the conditions were during these periods rather than imply any relationship between them, but the water vapor is present, at least in these dry summer periods, and in many cases the latent instability is also there. However, there is no mechanism to release this. We can not burn cornfields, of course; they are quite expensive; so we have no way of initiating convection.

Dr. Tor Bergeron—I want to point out the High over central Europe and southeastern Sweden that co-existed with the American drought in 1955. In fact it was one of the driest summers on record in southwestern Sweden; and that was one of the first summers that we had our project Pluvius working. It almost ruined our project.

Mr. Namias—Incidentally, I had an article published in the Monthly Weather Review (Some meteorological aspects of drought, with special reference to the summers of 1952-4 over the United States, September, 1955) describing the 1952 to 1954 drought in relation to the general drought problem.

Maj. C. Downie—Experiments carried out by the Geophysics Research Directorate have yielded little encouragement that a drought
Fig. 3—Mean 700-mb contours (above) and precipitation pattern (below) for August 1955
situation can be remedied artificially and, to the best of my knowledge, the artificial production of precipitation over a large area has never been conclusively demonstrated. However, the potentiality of particular techniques for producing localized precipitation has been proven.

Dr. Howell—In none of the experiments I referred to, did we attempt to produce large-scale changes, or look for large-scale changes; and I believe that the points I was making were concerned entirely with small interruption of the drought in small-scale areas.
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