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## PAPERS TO BE PRESENTED

SESSION ONE

INVITED REVIEW PAPERS ON INSTRUMENTATION

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# THE MEASUREMENT OF AIR MOTION IN CLOUDS by James Telford, Desert Research Institute, Nevada, U.S.A.

The measurement of the trajectories and motion of air parcels in clouds is of importance in the study of most aspects of cloud, rain and hail development. The rise of the cloudy air above cloud base is closely related to the liquid water content at each level. In cumulus clouds the observed dilution of cloudy material, so that it is less in liquid water content than expected for the adiabatic rise from cloud base, can only be attributed to mixing of dry air from the -surrounding environment into the cloud. Qualitatively the rise of the cloud and its attribution to the buoyancy of the less dense warmer cloudy air is well understood but the understanding of the quantitative balance between upward velocity, buoyancy and entrainment rate is still relatively obscure. The fluid flow models of cumulus convection using eddy diffusivity avoid the real question of how turbulent motion functions. The infinite gradient between cloudy and clear air properties at the upwind edge of a cloud, for example, implies a zero eddy diffusivity in a place where the turbulence is probably greatest and entrainment is a maximum.

The problems of the measurement of air motion in clouds thus depends on building a measuring system capable of gathering information on these questions in realistic detail. The entrainment of air through the sides and top of a cloud implies that precise horizontal and vertical velocities need to be measured in the surrounding clear air, without unknown zero drifts. The need for detailed understanding of the nature of turbulent transfer and consequently the transfer of dry air and momentum into and throughout the cloud means turbulent motions must be studied. Thus reliable short time constant measurements of velocity as well as moisture and temperature are needed. Since turbulence is highly variable in nature and clouds are highly variable from one to another the recording of data and the processing techniques must allow a large amount of data to be reviewed quickly. Digital recording and processing techniques are needed.

These requirements, when considered with the knowledge that vertical velocities are less than a few metres per second mean that only a modern digital computer inertial platform can meet the overall requirements.

Hardware schemes involving lesser equipment all involve compromises; but if carefully used may give novel and useful results. When a free gyroscope is used as a vertical reference precise vertical velocities are only possible over periods of less than a few minutes if errors are to be kept less than say 1 ft sec<sup>-1</sup>, and then there is still considerable doubt about the absolute velocity since the accuracy to which pressure altitude can be measured is barely sufficient to provide absolute reference velocities in less than a minute. The gyroscopic performance is crucial for all systems and no system accuracy estimate is acceptable until this performance is established.

Real time calculations and displays of virtual potential temperature position and motion velocities may not be essential but are certainly very desirable, and are quite feasible.

Radar Measurements of Air Motion and Hydrometeors in Clouds

by

### David Atlas and R. C. Srivastava Laboratory for Atmospheric Probing Department of the Geophysical Sciences The University of Chicago

Vertical incidence Doppler measurements in rain provide an excellent method of deducing drop-size spectra, but small updraft errors can lead to large errors in the concentration of large drops. In snow, size spectra estimates are subject to a variety of errors. Errors are also likely to be large in the case of hail; however, Battan has deduced limiting hail size spectra using reasonable assumptions.

Updrafts in convective storms have been deduced from vertical Doppler measurements from the "lower bound" method, in which the slowest falling particles are assumed to fall at 0.5 to 1 m sec<sup>-1</sup> in the absence of drafts. In rain, reflectivity provides an <u>estimate</u> of the expected mean Doppler velocity so that the difference between the actual and expected  $\overline{v}$ 's is an estimate of the draft speed to about  $\pm 1$  m sec<sup>-1</sup>. Possible improvements based on attenuation and polarization measurements are noted. Basic features of updraft and reflectivity fields are treated.

The most dramatic advance in convective storm observations comes from simultaneous use of two Doppler radars by Lhermitte. These provide ambiguity-free measurements of the 2-D velocity field; drafts come from divergence. Though single Doppler radar observations are subject to ambiguities, reasonable 2-D flow patterns may be deduced. Examples are presented. Reports of cyclonic vortices in association with the hook echo and echo weak vault, and a funnel cloud below the vortex in one case, are noteworthy. Real-time tornado detection will be greatly aided by the rapid advances in signal processing such as the plan-shear indicator of Donaldson and Armstrong.

The Velocity-Azimuth Display (VAD), which provides winds, divergence, and deformation, has been applied with particularly great success by Browning, Harrold, and colleagues to map 3-D flow patterns in extratropical storms. The combination of VAD and RVI (Range-Velocity Indicator) data by the latter has also provided new understanding of the circulation across sharp cold fronts. With supporting radiosonde data, Browning et al. have also found a limiting Richardson number (Ri) of ~ 0.3 in frontal zones; turbulent breakdown prevents further intensification of shear. Borreson's VAD data also show well-defined turbulent layers, but only in the up- and down-shear directions; 2-D overturning is implied. Lhermitte and Wilson have also devised interesting schemes of measuring turbulence from VAD data.

Doppler measurements of turbulence in the Soviet Union give the turbulence spectrum and the eddy dissipation rate; these are relevant to entrainment and diffusion calculations. Techniques have been devised by Mel'nichuk et al. and Atlas and Srivastava to adapt incoherent radars for turbulence measurements.

## MICROPHYSICAL MEASUREMENTS IN CLOUDS by R.E. Ruskin Naval Research Laboratory, Washington, USA

The choice of instrumentation for investigating detailed drop-size spectra lies between the established replication methods and one of the newer expensive optical detection methods with digital data-sorting. A reliable method (given satisfactory calibration) is to use the CSIRO soot-covered round rod sampler. With this method, however, it is difficult to obtain a sample which is large enough to provide good statistical significance in a cloud which has much variation. The continuous film Formvar sampler provides a much larger sample, but has uncertain collection efficiency for small droplets and suffers from break-up of an unknown fraction of larger droplets. This type also provides a more or less satisfactory method for determining the ice crystal fraction, although it is quite questionable for distinguishing ice in some situations near freezing.

Optical counting and sorting instruments which avoid the tedious data reduction of the replication methods have been developed. These include methods that are dependent upon the scattering from a laser beam, and upon the extinction or the shadow from each droplet to produce the signal pulses. The configuration and limitations of these instruments are discussed. Electrostatic charging methods are also under development for measuring both cloud droplet spectra and ice crystals. The limitations of optical ice-discriminating instruments are also discussed.

Simpler instrumentation is available for those measurements limited to the concentration, n, and the average droplet radius, r, of the spectra. The optical transmission over a 20m path (a function of  $nr^2$ ) is combined with a measurement of liquid water content (LWC) (a function of  $nr^3$ ). This method is strictly rigorous only for monodisperse drops but may be quite representative of young clouds. Methods of applying such techniques to spectra more relevant to older clouds are described, together with an assessment of various ways of measuring liquid water content.

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The various methods of measuring raindrop spectra are also critically reviewed. These include the charged screen probe, the microphone type and the more reliable foil samplers and the large optical scattering disdrometer.

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SESSION TWO

INSTRUMENTATION

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Charles E. Abbott

James E. Dye

J. Doyne Sartor

National Center for Atmospheric Research

Boulder, Colorado

The electrostatic disdrometer, a droplet sizing device, based on the original model by Keily and Millen (1960) for the Air Force Cambridge Research Laboratories was tested, modified, and calibrated in our laboratory for use as an automatic, continuous cloud droplet probe on aircraft. Laboratory investigations have shown that soon after entry an incoming droplet is broken into many fragments, the number of which is dependent on drop size. These fragments impact and splash on an electrode raised to 510 volt potential producing a voltage pulse which is dependent upon size of the original droplet. These pulses are amplified and fed into a ten-channel pulse height analyzer that accumulates and reads out the data each 0.5 sec. The range of the instrument is from 4  $\mu$  to about 50  $\mu$  radius. In practice, however, the upper limit is fixed primarily by the low number density of larger droplets found in clouds.

Airborne tests of the instrument on "The Explorer" sailplane and NCAR Queen Air have shown the present instrument to operate reliably with minimal maintenance. Comparisons in warm clouds of the electrostatic disdrometer with the Johnson-Williams hot wire liquid water content meter and with impaction slide replicas have shown that the disdrometer measurements generally agree quite well with the other methods. Comparisons of droplet size distributions measured by the disdrometer and impaction slide are shown in Figs. 1a and b. A comparison of liquid water content as measured by the three techniques is shown in Fig. 2.

Because of its small size, simplicity of operation, low power consumption, and operational reliability, it is proving to be of much value

\*The National Center for Atmospheric Research is sponsored by the National Science Foundation.

in the study of the microphysics of droplet spectra in airborne investi-

gations of clouds.

## Reference:

Keily, D. P., and S. G. Millen, 1960: "An Airborne Cloud-Drop-Size Distribution Meter," J. Meteor., Vol. 17, 349-356.



Figure 1. Comparison of measured droplet distributions between the
electrostatic disdrometer and slide
impactor (Hatched Area)
a) Slide taken at 1450:12.5 E.S.T. at 3200 ft., 17 C.

b) Taken at 1446:53.5 E.S.T. at 5200 ft., 13 C.





# NEW AUTOMATIC EQUIPMENT FOR INVESTIGATICS OF DROP AND CRYSTAL MICROSTRUCTURE IN CLOUDS.

## A.G.Laktionov, N.K.Nikiforova, V.V.Smirnov, G.I.Shelchkov O.A.Volkovitsky

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Recently photoelectric sensors allowing automation of the measurement process have been used extensively for investigation of the cloud and fog microstructure (Laktionov, 1957; Nevzorov, 1963; Akulshina et al., 1967; Shelchkov, 1970; Knollenberg, 1970). Television counters of cloud particles are coming into use (Smirnov, 1969). The present report gives description of the equipment used for investigation of the microstructure of the drop and crystal cloud medium and developed on the basis of both methods. Also, some measurement results are presented. 1. The complex of photoelectric devices is intended to measure

1. The complex of photoelectric devices is intended to measure the cloud drop concentration and size and to investigate time and space fluctuations of microstructure parameters. Drop radii are measured from 0.4 to 100 mm and concentrations up to 2.10<sup>3</sup> cm<sup>-3</sup>. The microstructure measurement and recording simultaneously carried out in several points are fully automatized. The technique for the processing of results with the help of the electronic computer is developed.

2. A modification of the photoelectric device with the optical formation of the counting volume serves to measure the microstructure of supercooled, mixed and ice-crystal fog. It is based upon the difference in polarization characteristics of lightscattering by crystals and drops. Concentrations of crystals of 10 to 200 mm in size are measured. The maximum particle concentration measured by the device with an error of no more than 20% constitutes 500 cm<sup>-3</sup>.

3. The television counter of cloud drops serves to visualize suspended drops and to measure the concentration and size distribution of drops in flows having velocities from 4 to 10 m sec<sup>-1</sup>. The transmitting tube (superorthicon) of the television sensor operates under the short light pulse illumination. Using the counting volume up to 5 mm<sup>2</sup>, the drop diameters are recorded from 4 to 40 mm and concentrations up to 10<sup>4</sup>cm<sup>-3</sup>. Video information is analyzed with the help of the television computing system and multichannel analyzer. The rate of analysis is 50 counting volumes a second. Data are recorded by the perforator and decimal typer. 4. The television counter of crystal particles is intended

4. The television counter of crystal particles is intended to visualize and to count crystals of 4 to 400 mm in size which are in suspension or in the flow having velocities up to 30 m/sec. Data are recorded by the video tape-recorder and counter. Photographs of TV monitor screen with crystal particle images on it are given.

5. Automatic photoelectric and television counters have made it possible to obtain the data on the character of fog microstructure nonuniformity. Regularities in the drop concentration fluctuations during the cloud drop spectrum formation as well as drop space distribution are investigated. TV counters allow visualization of the cloud drop and crystal formation processes when some seeding agent is introduced into supercooled fog.

### An Airborne Optical Ice Crystal Counter

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The measurement of the concentration of ice particles in clouds remains a significant problem in cloud and precipitation physics. Previous devices designed for this task have generally provided either data of extremely limited spatial resolution or data which required lengthy post-flight analysis (e.g. continuous particle samplers). During the last few years we have developed an automatic counter for use on an aircraft which detects ice particles in the air but does not count even large water drops. This device provides a continuous recording in real-time of the concentrations of ice particles in clouds.

The present instrument (Figure 1) is an improvement on an earlier design of the authors (see Hobbs and Ryan, 1969) and consists of two polarizers set for maximum extinction through which a narrow beam of light is directed. An ice crystal passing between the polarizers causes a slight rotation of the incident plane polarized beam either by the optical activity of the crystal or by scattering from its anisotropic shape and results in a light pulse being detected by a photomultiplier. Cloud droplets, being spherical and not optically active, do not produce a rotation of the plane polarized beam. Extensive laboratory and field tests have established that ice crystals with maximum dimensions in excess of 100 µm are detected with this device. The sample area is adequately de-iced and the flow through it is isokinetic.

The signal from the photomultiplier is processed through two signal conditioners, a pulse height discriminator and a pulse rate integrator. Both of these signal paths are routed to logarithmic amplifiers and analog tape recorded. The integrated signal is displayed to give direct read-out of the concentration of ice crystals in the air. The results of recent measurements in natural clouds will be presented along with other studies designed to determine the minimum detectible size for different types of ice crystals.

#### Acknowledgements

Thanks are due to Prof. P. V. Hobbs for his help during the course of this work. This work was supported by grants from the National Science Science Foundation (NSF GA-27637) and the U. S. Dept. of Interior (Contract 14-06-D-5970).

#### References

Hobbs, P.V. and B. F. Ryan, 1969, <u>Contributions from the Cloud Physics Laboratory</u>, Research Report III, University of Washington, Seattle, Wn.



Fig. 1. Schematic representation of the optical ice crystal counter (not to scale)

## A HOLOGRAPHIC SYSTEM FOR DROPLET SIZING IN CLOUDS AND FOGS

## R.J. Adams, Meteorological Office, Bracknell

A holographic instrument has been developed and used in naturally occurring fogs. Droplets of diameter  $d \approx 10 \ \mu m$  and greater have been detected and measured.

The holographic technique overcomes many of the limitations of other sampling methods, offering less disturbance to the airflow than impactor devices, and providing more information with less ambiguity than alternative optical systems. The technique can be regarded as providing a means for bringing into the laboratory a sample of around half a litre for subsequent detailed analysis, and thus information such as number, size distribution, shape and position of individual particles, and total water content may be extracted.

The system for recording, using a Q-switched ruby laser, generally followed the simple in - line layout, but, for comparison, some holograms were made using an imaging lens. Reconstruction, incorporating a zero order stop in the Fourier transform plane of a projection lens, also includes a rotating diffuser which has been found to reduce speckle effects.

Holograms have been produced showing the size and spatial distribution of the droplets from a number of different laboratory sources, including the spinning top, aerosol spray and ultrasonic atomiser types. The instrument has also been operated in fog, when the variations in droplet size spectra obtained were in good agreement with near simultaneous measurements from an impactor device. Additionally, the holographic instrument gives spatial distribution information, showing both small and larger scale variations in drop concentration which the impactor does not record. Further, the impactor records a volume in the form of a very long thin cylinder averaged over several tens of seconds, whereas the hologram records an instantaneous sample, with length and breadth approximately equal. It must be remembered, however, that the impactor has a lower limit of detection considerably better than the holographic system. One of the advantages of holography, namely that the information is presented, as in reality, in three dimensions leads however to a major problem when rapid analysis is required. As a first step towards the development of an automatic analyser, a system employing spatial filtering in the reconstruction stage has shown promise as a means of rejecting images appearing in planes other than those on which the detector is focussed. Nevertheless, the design of an analyser involving no operator intervention still appears to be a formidable problem, although the production of a system capable of recognising the existence of isolated events of interest, say the presence of a few large droplets among large numbers of smaller ones, presents a much lesser problem, and one to which the application of holography seems particularly well suited.

Because of its many advantages, particularly those arising from relatively low sample disturbance and short exposure time, the system has the potential of being developed into a most powerful aircraft instrument. The advantages and technical difficulties of operating such a system on an aircraft are described.

TETHERED BALLOON CLOUD INSTRUMENTS By H. T. Bull\* J. J. Colls D. L. Dolman Cavendish Laboratory, Cambridge

A set of instruments has been developed for investigating St and Sc clouds by means of the tethered balloons at R.A.E. Cardington. The scientific objectives relate to the droplet-size spectrum and radiative properties of these clouds. The instrument system consists of:-

- 1. Continuous droplet-size sampler
- 2. Angular albedo telescope
- 3. A net radiometer, and up and down looking solarimeters on the instrument platform and also on the ground
- 4. Wet and dry bulb psychrometer
- 5. Cup anemometer for mean wind
- 6. Hot wire anemometers for the turbulence intensity

These instruments and the mounting platform are shown in the attached photograph. The maximum height attainable is 7500 ft and the instrument platform can remain aloft for as long as required providing the wind speed is less than 25 knots.

The cloud-droplet sampler is an impactor device, following the designs of May. It is considerably extended however for use on the tethered balloon to allow for a large sample volume. The instrument details are as follows:

- 1. The input tunnel establishes isokinetic flow of 2.7 m/sec over an outside wind speed of 3 m/sec to 15 m/sec.
- 2. Useful drop diameter range is  $3 \mu$  to  $60 \mu$
- 3. A space resolution of one meter, allowing for an adequate count of cloud droplets.

The hot wire anemometers are mounted on a damped pendulum and have an overall frequency range of .01 cps to 1 cps.

The instruments are used in order to investigate the role of mixing at the cloud boundaries by relating the turbulence intensity to the degree of broadening of the droplet spectra which are observed even in clouds which have been in existence long enough for a near mono-disperse spectrum to develop.

The radiation instruments are used to obtain the angular distribution of back scattered visible radiation; and also, to measure the absorption of sunlight in clouds when both the thickness and drop size spectra are simultaneously measured.

Current result are shown on the accompanying slides.



THE DESIGN AND USE OF AIRCRAFT DROPSONDES IN THE STUDY OF CLOUD

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Aircraft dropsondes provide a means of determining the physical properties of weather systems on a grid scale of tens of kilometres over areas whose typical dimensions are hundreds of kilometres. In particular sondes which are designed to respond to air motions along their trajectories and which are capable of being tracked in three dimensions provide a means of investigating the dynamics of frontal systems. The present Meteorological Office dropsonde depends upon radar positioning and consists of an octahedral target. Wind tunnel tests have shown that the horizontal air motion relative to the target is less than  $0.5 \text{ m sec}^{-1}$  for the largest shears likely to be encountered in the low troposphere, and trajectories of one sonde determined by two independent precision radars suggest that one minute winds may be obtained to  $\pm 0.3$  m sec<sup>-1</sup> at ranges of 40 to 50 Km. During experiments to determine air motions within warm frontal situations these dropsondes have provided wind data at 25 to 30 Km intervals and have enabled the specification of vertical velocities to  $\pm 2 \text{ cm sec}^{-1}$  at altitudes up to 4 Km over areas measuring 150 Km x 100 Km. Although the target has been tracked automatically out to ranges of 100 Km in the presence of moderate rainfall, the use of radar for windfinding is clearly range limited and has a severe practical limitation of tying the observations to a closely defined location. There are also problems of wind determination at low elevation angles.

An alternative approach has been taken for the next generation of Meteorological Office dropsondes by using the tracking facility provided by the LF navigation aids such as Loran C, Omega or Decca. This NAVAID sonde is described and preliminary trial results are presented. Position finding is achieved by phase comparison of radio signals received from at least three fixed coherent transmitters. As applied to windfinding
dropsondes the LF signals received at the sonde are retransmitted on a VHF link to the dropping aircraft where the phase comparison is carried out. The altitude of the sonde must also be determined and the suitability of various methods are discussed.

The accuracy with which winds may be defined again depends on the ability of the sonde to follow the air motions, on the stability of the LF propagation to the sonde, and on its position relative to the existing transmitter network. The results of trials designed to test the relatively short term stability, necessary for windfinding, of the Loran C network are presented. It is found that areas, typically 700 Km across, exist to the south of transmitting stations on the Faroe Islands, Iceland and Greenland within which 1 minute winds are expected to be determined to better than 0.4 m/sec during the daytime. Therefore, because it is now only necessary to maintain a satisfactory RF link between the aircraft and sonde, experiments extending over much larger areas may be undertaken in the North Atlantic from an aircraft such as the Meteorological Research Flight C 130.

Finally, possible uses of the NAVAID sonde within GATE to determine divergence profiles and hence vertical motions within cloud clusters are discussed, and, as an extention of past work on warm frontal systems, future projects which sample an atmospheric volume an order of magnitude greater than hitherto possible are described.

# SESSION THREE

# NUCLEATION AND CLOUD DROPLET GROWTH

#### PROBLEMS OF CLOUD - AND CONDENSATION - NUCLEI RESEARCH

Prof. H. -W. Georgii, Frankfurt/Main

The review of recent research on condensation and freezing processes related to cloud formation will be concentrated on the following problems:

## 1) Distribution of background aerosols

Recent measurements of aerosols in continental areas far from human activity and over the oceans reveal interesting results with respect to the number concentration and size distribution of the background aerosols. These measurements have recently been supported by investigations of the size distribution of aerosols in higher layers of the atmosphere. The results of these investigations are compared with microphysical properties of clouds and related to numbers and sizes of cloud elements. There is growing concern about the long-term increase of particles on a world-wide scale due to anthropogeneous processes. This global increase in particulate matter may change the microphysical structure of clouds. New evidence will be reported on.

## 2) <u>Cloud Nuclei</u>

It is still an open question which fraction of aerosol particles is preferably consumed as cloud nuclei during the process of cloud formation. The second part of the review will therefore deal with new results of investigations of cloud nuclei. We shall briefly discuss the methods of observation and try to relate the cloud-nuclei spectrum to the aerosol spectrum which is in our opinion an important step to coordinate the efforts of cloud-physicists and aerosol-physicists. Little is known of the chemical composition of nuclei active in cloud formation. The knowledge of the composition of cloud nuclei is of interest in order to indicate more precisely their sources and their production rates. Direct analysis seems difficult owing to their small size. There are indications pointing to ammonium sulfate as being a predominant substance of cloud nuclei. New results on the vertical and horizontal distribution of sulfate-containing particles in different size ranges will be reported. They may allow some conclusions on the importance of these particles in the cloud formation process. The results show also that cloud-droplets pick up gaseous traces from the atmosphere, incorporate them and may act as secondary sources of nuclei after evaporation. Measurements carried out over the Atlantic Ocean make it seem doubtful that the majority of nuclei active in formation of maritime clouds are sodium chloride particles. There is growing evidence also that over the oceans sulfate particles may play an active role as nucleants. Finally the relation between cloud nuclei and cloud elements will be discussed.

# S.C. Mossop

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New developments in this field in the past three to four years are surveyed.

The theoretical treatment by Fletcher of the relationship between particle size and nucleating ability has been improved by him to take account of the fact that nucleation takes place at sites scattered over the particle surface, and that these sites vary widely in activity one from another.

The phenomenon of pre-activation of nuclei, by which they are able to nucleate ice at much higher temperatures after certain pretreatment, has been satisfactorily explained by Evans and his coworkers. Pre-activation is ascribed to the ordering in an ice-like structure of an  $H_2^0$  monolayer adsorbed on the nucleus surface, which takes place below a certain critical temperature.

Considerable interest has been shown in contact nucleation. This does not seem to involve any new physical process but may be more effective than bulk freezing for nuclei which are in some way inactivated by immersion in water.

In the case of AgI there is a conflict of evidence as to whether it is inactivated by immersion in water or not. Where the AgI is mixed with a soluble iodide as in some cloud-seeding aerosols, it seems likely that there will be some loss of activity due to dissolution of the AgI, particularly when released below cloud base at temperatures warmer than 0°C.

The relative importance of the deposition, condensationfreezing and contact-freezing mechanisms is being explored for both natural and artificial nuclei between -10°C and -20°C. 20

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Intercomparison workshops have brought about a recognition of the serious errors to which some ice nucleus counters are subject. When counting natural ice nuclei several instruments give reproducible results, and of these the membrane filter technique is the most suitable for field use. For artificial nuclei, particularly AgI, the position is less clear, but it seems likely that the real cloud situation can best be simulated in chambers where long-lasting clouds can be made.

Measurements of ice particle concentration in natural clouds show satisfactory agreement with the measured ice nuclei in certain stratiform clouds and in cumulus clouds of continental nature. In cumuli where the drop spectrum is more maritime, the ice particle concentration may exceed that of the nuclei by a factor of 10<sup>4</sup> at temperatures near -10°C. The exact nature of the enhancement mechanism is unknown but presumably it involves the larger water and ice particles which are present in the type of cumuli in which it occurs.

We discuss the conditions under which one would expect rainfall to be related to ice nucleus concentration. This has hardly been investigated at all in the past, which is surprising, considering that the existence of such a relationship is a prerequisite for success in the artificial stimulation of rain by the Bergeron-Findeisen process.

## A Case Study of Ice Crystal Multiplication by Mechanical Fracturing

Larry Vardiman and Lewis O. Grant Colorado State University Fort Collins, Colorado

It has been hypothesized by Mason (1955) and Koenig (1968) that the disparity between concentrations of ice nuclei and ice crystals at temperatures cold enough for dendritic growth might be explained reasonably well by the fracturing of fragile dendritic crystals. However, documentation of the conditions under which dendritic crystals fracture is limited. This paper will describe a case study of an orographic cloud whose top was in the dendritic region. During portions of its existence this cloud produced ice crystal concentrations within a factor of 10 of the ice nuclei concentrations, while at other times the two concentrations differed by a factor of 1000.

A continuous formvar replicator was used to obtain the shapes and concentrations of snowfall from orographic clouds at the High Altitude Observatory (HAO) near Climax, Colorado. Cloud top temperatures were obtained by measuring the height of the cloud with a 3-cm radar and picking off temperatures corresponding to these heights from soundings adjusted for space and time. Ice nuclei concentrations were observed with a Bigg-Warner type expansion chamber.

Figure 1 shows the concentrations of fragments, irregulars, and total ice crystals at HAO as a function of time on February 14, 1969. Cell 1 showed the highest concentration of fragments followed by lower concentrations with each succeeding cell. The concentration of irregular crystals between cells increased throughout the period, however. The low wind conditions and perfect operating characteristics of the replicator during this period indicated that the crystal fragments observed in the cells actually came from within the cloud and were not due to extraneous sources. Figures 2, 3, and 4 show typical 22

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examples of crystal fragments observed. Figure 2 is particularly interesting as it shows a crystal which had one branch broken off and the two nearest branches growing at an increased rate relative to the others. Figure 4 shows a crystal with regrowth after one branch was broken. Throughout the entire period the ice nuclei concentration remained between .1 and .5 nuclei per liter effective at -20°C.

Although the dendritic crystals observed in this study were not of the most fragile class of dendritic crystal, fragmentation obviously occurred. It would seem apparent then that clouds which contain dendritic type crystals have a dramatically favorable region for ice crystal multiplication. Further work is in progress to define the exact mechanism which produces the fracturing and the prevalence of this mechanism in various classes of clouds.





Fig. 2



Fig. 3



Fig. 4

# MODEL INVESTIGATION OF THE INITIAL STAGE OF CONDENSATION

# IN THE CLOUD

# N.V.Klepikova, Y.S.Sedunov, O.A.Volkovitsky

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One of the most important stages of dynamic development of clouds is the so-called initial stage at which a definite number of condensation nuclei pass into drops and the initial drop spectrum is formed. Analysis shows that numerical calculations of the set of equations describing this stage (Howell, 1949; Mordy, 1959; Neiburger and Chien, 1960) and semianalytical solutions of the problem (Twomey, 1959; Buikov, 1966; Sedunov, 1967; Volkovitsky and Sedunov, 1970) did not allow one to establish finally the influence of many factors upon the cloud spectrum formation process. Experimental investigations of the initial condensation stage in real clouds are mainly of a qualitative nature (Twomey and Warner, 1968; Marwitz and Auer, 1968; Warner, 1969). Model investigations are performed under comparatively little change of ver ical velocity and condensation nuclei (Volkovitsky and Lak ionov, 1969).

In the present work the equations most completely describing the initial condensation stage in the cloud are numerically solved. The calculated drop concentrations, maximum supersaturation, change of drop spectrum form and other parameters are compared with those determined in model experiments which have been carried out in the cloud chamber 3200 m<sup>2</sup> in volume. In calculations as the initial condensation nucleus size distributions the distributions are taken which have been constructed from the condensation nucleus supersaturation distributions measured during two runs of experiments. Calculations are performed for vertical velocities of 5 to 100 cm/sec which have been used in the experiments.

Results of calculations and experiments are in quite satisfactory agreement. Numerical solution of the set of equations with the account taken of only regular condensation has made it possible to obtain a broad drop spectrum for rather a short period of time.

The errors are determined which arise when using the formulas for drop concentration calculations which have been suggested earlier. The role of individual effects in the process of cloud drop spectrum formation at the initial stage is discussed.

#### NONSTEADY-STATE KINETICS OF DROPLETS IN CLOUD PHYSICS

## N. Nix and N. Fukuta University of Denver, Colorado, U. S. A.

Droplet growth and evaporation under rapidly changing environmental conditions were studied theoretically. The solutions for the vapor and temperature fields were found in closed analytic expressions, assuming Maxwellian conditions at the droplet surface. They describe the mass and heat transport simultaneously. Changes in vapor density and temperature in the environment (i.e., far away from the droplet) with respect to time were simulated by exponential functions. The processes in question are described by one dimensionless complex parameter. By using these solutions, criteria are shown for deviations from the quasisteady-state by 1%, 5%, and 10%.

For the droplet growth rate, the new theory shows an overshooting effect before the condition finally reaches the quasisteadystate. This overshooting effect is a function of the rate of environmental change, and is proportional to the droplet diameter so that it becomes dominant for larger droplets.

Using a similar approach, the transient supersaturation wave with respect to water is described when a supercooled droplet freezes. In the theory, the temperature of the freezing droplet surface is simulated by an exponential function of time. As freezing proceeds, a large single wave of nominal supersaturation is created and moves away from the droplet with steadily decreasing amplitude. The larger the droplet supercooling, the higher the head of the supersaturation wave, i.e., the supercooling determines almost entirely the maximum amount of supersaturation. For larger 25

droplets, the supersaturation wave penetrates farther into the space and lasts longer. The formation and movement of the supersaturation head after freezing explains the generation of secondary fog droplets observed around a rapidly freezing supercooled droplet. As an example of the computation, a droplet of 50  $\mu$ m radius at -15°C creates a supersaturation head of 16% at a distance of 50  $\mu$ m, 0.1 sec after freezing. The relationship of variables will be graphically presented.

2. Call

#### ON THE POSSIBLE PRODUCTION OF SUBMICRON ICE FRACMENTS DURING RIMING

OR THE FREEZING OF DROPLETS IN FREE FALL.

J. L. Brownscombe, P. Goldsmith Meteorological Office, Bracknell.

One explanation of the observed discrepancy between the number of freezing nuclei in the atmosphere and the number of ice crystals in relatively warm supercooled clouds is that a secondary ice crystal production mechanism during droplet freezing exists.

Recent observations of droplets 50-250 µ diameter freezing during free fall in air have shown that insufficient visible ice fragments are produced to account for this discrepancy. However, these observations provide conflicting evidence on the possible production of larger numbers of ice particles of a size below the limits of direct optical detection. It has also been shown that very few visible splinters are produced during rime formation in air.

The experiments reported here are designed to detect submicron ice particles produced during the freezing of droplets.

# (1) Droplets freezing in free fall.

Droplets from 20 to 70  $\mu$  diameter were produced from deionised water, 10<sup>-4</sup>M NaCl, 10<sup>-3</sup>M NaCl and 10<sup>-4</sup>M Ag I + Na NO<sub>3</sub> solutions. The droplets were frozen in free fall at ambient gas temperatures around -30 to -40<sup>o</sup>C (and also at -10<sup>o</sup>C to -15<sup>o</sup>C for droplets containing Ag I) in air, helium and carbon dioxide atmospheres. The warm droplets were allowed to fall through a steep temperature gradient into the cold environment to enhance the probability of shattering in air. Small ice crystals or water droplets ejected from the freezing droplets were detected by counting the number of submicron-sized evaporation residues. Splinters or droplets as small as  $\sim$  0.2  $\mu$  diameter were detected using this technique.

When water droplets were frozen in a carbon dioxide atmosphere both at -30 to  $-40^{\circ}$ C and at -10 to  $-15^{\circ}$ C large numbers of small particles were thrown off. In helium and air, however, there was no evidence for a significant production of sub-micron particles.

## (2) Droplet freezing during riming.

The experiment was performed in a 300 litre cold box approximately 1 m deep. A cloud was produced by condensation. Loe covered insulated rods  $\sim$  3 mm diameter were rotated in this cloud so that the riming surface was moving at a velocity of up to 80 cm s<sup>-1</sup>. Individual frozen droplets in the resultant rime had a mean volume diameter  $\sim$  15  $\mu$ . Droplets of mean volume diameter  $\sim$  40  $\mu$  could also be injected using an ultrasonic atomiser.

Any small ice crystals produced during riming would grow in the supercooled cloud and be detected when they fell out onto a tray of supercooled sugar solution which covered the base of the chamber.

In carbon dioxide at temperatures between -4 and  $-10^{\circ}$ C large numbers of ice crystals were always detected when riming took place. However, in terms of ice crystals produced per drop freezing on the rime surface, the maximum observed was only one in three hundred.

In air under similar conditions the maximum ice crystal production rate was 1 crystal for every 5 X  $10^4$  droplets frozen.

Microscopic examination showed that the rime had a very open chain-like structure which should produce heat transfer conditions favourable for the ejection of splinters as droplets freeze.

# (3) Conclusions.

These experiments do not support the view that large numbers of submicron ice fragments are produced when droplets freeze during riming or in free fall.

# SESSION FOUR

# PHYSICS OF RAIN

#### THE PHYSICS OF RAIN

By P. Squires Desert Research Institute, Nevada, U.S.A.

The formation of rain in warm maritime clouds by droplet collection has been observed to occur in times which are far too short to be explained by computations which ignore the stochastic Although improved computational procedures aspect of the process. have been developed for studying stochastic collection, the implications of this process have not yet been adequately described. In very general terms, however, theoretical results so far available appear to come fairly close to explaining the time required for the formation of rain in a maritime cumulus. While it remains a possibility that the effects of microturbulence or of electrical forces may not be negligible, or that mixing of some parcels of cloud air with the dry environment may locally modify the droplet spectrum in such a way as to accelerate the formation of rain, there seems to be little reason to suspect that these phenomena will be found to play a crucial role.

In continental clouds, on the other hand, the calculations indicate that the formation of rain by droplet collection requires much longer periods of time - probably exceeding the lifetime of a parcel of cloudy air - so that warm rain would not easily form in such clouds, in apparent agreement with observations.

Improved calculations of droplet collision efficiencies have somewhat modified Hocking's earlier results, indicating that in the case of a collector droplet of radius less than 18 microns the collision efficiency is not zero, although it is small. These changes have not greatly altered the overall physical picture, the main outlines of which are determined by the strong dependence on radius of the volume swept by a cloud droplet in unit time.

The effects of man-made pollution on the distribution of precipitation have recently been studied in several locations. Even though such effects seem physically plausible, convincing statistical demonstration is still lacking. Such statistical investigations are apparently complicated by the occurrence of climatic drifts due to natural causes and the lack of adequate long-term precipitation records. Some progress has been made toward understanding the economy of cloud nuclei. It seems clear

that on a global scale, the anthropogenic production of such particles constitutes only a minor addition to the natural Twomey's investigations on the volatility of production. cloud nuclei have reached an important stage with his recent conclusions that most cloud nuclei in both maritime and continental air consist of some rather volatile material such Taken together with evidence such as the as ammonium sulphate. geographical distribution of cloud nuclei, and their decreasing concentrations aloft, this points to the probability that cloud nuclei do not in the main originate at the surface, but rather are formed in the lower troposphere from gaseous trace components (such as ammonia) emitted from land surfaces. The vapor pressure of ammonium sulphate even at atmospheric temperatures is such that cloud nuclei-sized particles would evaporate in a few hours unless vapor saturation exists. These results would imply that the relationship between cloud physics and some aspects of the chemistry of the atmosphere may be closer than was formerly thought.

Much effort has been expended in the modelling of convective clouds, often in connection with weather modification experiments. Our poor understanding of many aspects of the physics, such as turbulent mixing or the initiation of ice particles and their subsequent growth, still poses a barrier to fundamental advances; but there is reason to believe that the critical application of such insights as we now have will improve our ability to evaluate the results of such experiments. Moreover, this ability will improve as studies of sub-problems forming part of the overall study develop and produce usable parameterizations for use in the larger context, such as that Berry has provided for stochastic drop collection.

# SLIP FLOW CORRECTIONS TO THE COLLISION EFFICIENCIES OF SMALL CLOUD DROPLETS

## M. H. Davis

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One consequence of Stokesian hydrodynamics is that the surfaces of two spheres sedimenting in a viscous fluid cannot be brought into actual contact by a constant force such as gravity. In order to compute "collision efficiencies" for cloud droplets, it has been necessary to make some assumption which will allow for new physical phenomena when the gap between droplet surfaces is very small. Hocking and Jonas (1970) discussed the problem and computed two sets of collision efficiencies based upon the assumption that a "collision" occurred when the gap between the droplets was  $10^{-3}$  or  $10^{-4}$  of the radius of the larger. For drop radii larger than 30 microns their results showed little dependence on the minimum gap chosen, while for smaller drops a marked dependence was found.

The present analysis was undertaken in an effort to eliminate the need to make an arbitrary choice of minimum gap for collision efficiency calculations in the Stokesian regime. The creeping flow equations for two approaching spheres were solved using slip flow boundary conditions. The resulting force coefficients are functions both of geometrical quantities and of the "slip parameter," which can be related through experimental data to the mean free path of air molecules in the neighborhood of the droplets.

The principal result of the slip flow analysis is the removal of the "collision barrier" noted previously. Newly calculated collision efficiencies for small droplets are presented in Fig. 1 along with comparable results from Hocking and Jonas (1970). While the new results

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<sup>\*</sup>The National Center for Atmospheric Research is sponsored by the National Science Foundation.

exceed somewhat those obtained previously for drops smaller than 30 microns in radius, it appears unlikely that the differences are sufficient to affect the development of large drops by the coalescence mechanism.



Hocking, L. M., and P. R. Jonas, 1970: The collision efficiency of small drops. <u>Quart. J. Roy. Met. Soc.</u>, <u>96</u>, 722-729.



Figure 1. Comparison of present results (solid lines) for several larger drop radii with those reported by Hocking and Jonas (1970) with  $\varepsilon = 10^{-3}$ .

# THE COALESCENCE OF DROPLETS IN A TURBULENT CLOUD

# P.R. Jonas and P. Goldsmith - Meteorological Office, Bracknell

It has been demonstrated that the observed spectrum of cloud droplet radii and the production of rain from warm continental clouds in their observed lifetime cannot be explained on the basis of current theories of droplet growth by condensation and coalescence. This conclusion is not modified when the effects of the electric fields in clouds are considered. Calculations (Bartlett & Jonas, 1972) have also shown that turbulence in clouds cannot appreciably affect the growth of cloud droplets by condensation. This paper describes investigations of the effects of turbulence on the coalescence process in clouds.

It has been shown by Woods, Drake, Goldsmith (1972) that turbulence in clouds acting over distances comparable with the distance fallen by pairs of interacting droplets can be approximated by a linear shear in the air flow. The magnitude of this shear in moderate clouds being festimated as about 5-10 s<sup>-1</sup>. The collection efficiencies of cloud droplets falling in a sheared air flow have been measured for a wide range of droplet radii. The results, some of which are shown in Fig. 1, indicate that the collection efficiencies of droplets of radii less than about 20 µm are appreciably increased by the presence of the shear. This increase is shown to be proportional to the shear above some critical value. The experiments also indicate a dependence of the collection efficiency on the direction of the shear.

Calculations of the collection efficiencies of cloud droplets falling in a sheared air flow are also described. These predict a much smaller increase in the collection efficiencies than the experiments suggest and also that the increase is larger for larger droplets, in contradiction to the experimental results. Some possible reasons for these discrepancies are discussed together with the results of further experiments in which the collection efficiencies of cloud droplets in unsteady air flows were measured. The results of these experiments were similar to those obtained in the linear shear and suggest that the relaxation of the droplets into the changing air flow may be important. This was ignored in the calculations.

It is shown that if these collection efficiencies are typical of those in a moderately developed cloud then the droplet size spectrum will be sufficiently broadened to produce rain in times more consistent with the observations than has been suggested previously.

Fig. 1



The Collection Efficiencies of Several Droplet Sizes for Smaller Droplets. The **curves** indicate the theoretical results for collisions in still air and the points, the measured values in a shear of 14 s<sup>-1</sup> in a horizontal air flow.

<u> </u>					
	40	μm	radius	large	drop
			·Δ		
	30	μm	radius	large	drop
<b></b>					
	20	۲m	radius	large	drop
~~~~~					
	10	μm	radius	large	drop

Bartlett J.T., Jonas P.R. 1972

On the dispersion of the sizes of droplets growing by condensation in turbulent clouds. Quart. J. R. Met. Soc. <u>98</u> pp. 150-164

Woods J.D., Drake J.C., Goldsmith P. 1972 Coalescence in a Turbulent Cloud Quart. J. R. Met. Soc. <u>98</u>, pp 135 - 149

THE EFFECT OF COALESCENCE EFFICIENCY AND SATELLITE DROP PRODUCTION UPON RAINFALL RATES AND DISTRIBUTIONS

P R Brazier-Smith, S G Jennings and J Latham

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Recent experimental and theoretical work by Brazier-Smith, Jennings and Latham (1972) has shown that the permanence of the coalescence of a pair of colliding water drops of size comparable with that of small raindrops depends upon their incident relative velocity, U, their radii R and r (R>r) and the perpendicular distance X between the centre of one drop and the undeflected trajectory of the other. A critical value of X,  $X_c$ , exists above which there is sufficient rotational energy in the system to enforce separation and below which the coalescence is permanent. The separation efficiency S, is defined as  $(1-\epsilon)$  where  $\epsilon$ , the coalescence efficiency  $[=(X_c/R+r)^2]$  is shown to be given accurately by the equation

$$r = \frac{173f(R/r)}{r U^2}$$
(1)

$$f(R/r) = \frac{(1 + \alpha^2 - (1 + \alpha^3)^{\frac{2}{3}})(1 + \alpha^3)^{\frac{11}{3}}}{\alpha^6 (1 + \alpha)^2}$$
(2)

and  $\alpha = R/r$ .

The present study is concerned with the effect that these experimentally verified values of coalescence efficiency have upon the development of rainfall in a cloud. In addition, since the experiments showed that the separation of a drop-pair after collision was generally accompanied by the formation of satellite drops, calculations were made of the role that these might play in the transformation of cloud water to rainwater. Although

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the numbers and radii of the satellites produced were variable, typical values were 3 and 80µm respectively.

The development of a raindrop spectrum is treated in the following manner. It is assumed that at time t=0 there exists within the cloud a distribution of drops of maximum radius 100µm which are uniformly distributed in size per unit logarithmic interval. Cloud water is being released by condensation at a rate, dependent upon the updraught velocity, which remains constant throughout the duration of one set of calculations. The initial drop spectrum sweeps out cloud water with a collection efficiency of unity and also interacts with itself to produce coalesced drops and satellites which themselves collect cloud water and interact with the other particles. These interactions are treated theoretically by means of a variation of the method of Bartlett (1966) for integrating the stochastic growth equation. In this way the drop-spectrum, the total precipitation intensity, the concentrations of cloud water and rainwater and the radar return signal are calculated as a function of time for several different values of release rate by condensation. In addition, in order to establish the sensitivity of the rainfall development to the coalescence efficiency e and to the formation of satellite drops, series of computational runs were made for  $\epsilon = 0, \epsilon = 1$  and  $\epsilon$  given by equation (1) with satellites either accompanying separation, or not being produced.

The results of the calculations show that the rate of increase of precipitation intensity is insensitive to the value of  $\epsilon$  or the presence of satellites, being governed predominantly by the rate of release of cloud water by condensation. However, the raindrop size spectrum and the

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rate of production of drops approaching maximum size is very sensitive to these factors. In particular the radar return signal, illustrative of precipitation growth, can vary by a factor of about 5 according to the selected value of  $\epsilon$ . It is therefore important to use accurate values of  $\epsilon$ , given by equation (1), in calculations of rainfall development. They are currently being applied to calculations of the changing size distribution of raindrops as they fall from cloud to ground.

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A<u>326</u>, 393-408

# Droplet Collection Rates for

# Double Initial Gaussian Distributions

Edwin X Berry

Richard L. Reinhardt

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The growth patterns of double distributions show the importance of initial Telford-type stochastic spreading of the larger drops as a part of overall stochastic growth.

We define three extensive variables of the droplet number density function f(x) over droplet mass x by the following integrals:  $N=\int f(x)dx$ (cm<sup>-3</sup>),  $L=\int xf(x)dx$  (gcm<sup>-3</sup>),  $Z_f=\int x^2f(x)dx$  (g<sup>2</sup>cm<sup>-3</sup>), where N, L,  $Z_f$  are the number density, the liquid water content, and a "normalized" radar reflectivity, respectively. We also define two intensive variables by the ratios  $x_f=LN^{-1}$  (g),  $x_g=Z_fL^{-1}$  (g), and use  $r_f$  and  $r_g$  to be the radii (cm) corresponding to these two masses.

The initial droplet spectrum of Fig 1 is a linear combination of two Gaussian functions,  $f(x)=Nx_f^{-1}e^{-X/x}f$  for 0.8 gm<sup>-3</sup> of  $r_f=10\mu m$  and 0.2 gm<sup>-3</sup> of  $r_f=20\mu m$ . The pattern and the progress of  $r_g$  are similar to those for single functions (described in the Reinhardt-Berry paper in these proceedings) but does not fall into the same quantitative, parameterized form.

Fig 2 shows each part of the double distribution of Fig 1 considered separately. Spectrum 1 decreases regularly and stays stationary. Spectrum 2 increases regularly as it moves to the right. Fig 3 shows the  $r_g(t)$  for Spectrum 2 as if by itself for three conditions: L=1.0 gm<sup>-3</sup>, L=0.2 gm<sup>-3</sup>, and for L increased at the rate indicated in Fig 2. This last curve lags behind the curve for the actual growth rate for the Spectrum 2 when interacting with Spectrum 1 (shown) but takes up the same slope as  $r_a \approx 30 \mu m$ .

This difference indicates that Spectrum 1 does more than simply add water mass to Spectrum 2. It also causes the spreading of Spectrum 2 by what must be described as Telford-type spreading due to the stochastic nature of the collections of Spectrum 1 by Spectrum 2. (Compare the paper by Scott in these proceedings.) The shape of Spectrum 2 (Fig 2) at 5 min shows the shape of stochastic spreading.



Fig. 1. Time evolution of a linear combination of two Gaussian functions having 0.8gm<sup>-3</sup> for  $r_{f0}$ =10µm and 0.2gm<sup>-3</sup> for  $r_{f0}$ =20µm.



Fig.2. The data of Fig.l as two separate parts, spectrum l on the left, spectrum 2 on the right, time above each curve.



Fig.3. The change in time of  $r_g$  for spectrum 2.

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# NUMERICAL MODELLING OF THE PROCESSES OF TRANSFER

AND COAGULATION OF CLOUD PARTICLES BY THE MONTE-

CARLO METHOD IN THE THEORY OF PRECIPITATION

# FORMATION

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Stochastic character of processes of transfer and coagulation of cloud particles and precipitation formation gives the possibility of using the Monte-Carlo method for numerical modelling of these random processes and numerical solution of the kinetic equations of coagulation of cloud particles.

There are two possible variants of using the Monte-Carlo method.

The first approach- the trajectory of motion, collision and coalescence of the trial cloud particles are modelled by an electronic computer.

The function of distribution of field cloud particles with which the trial particles collide is set by previous iteration. By means of the trial particles the function of distribution is searched at the following iteration. The trial particle the trajectory of motion and collision of which are traced is chosen at random.

Consideration of the random character of collision and coalescence of the trial cloud particles with the field particles is attained by means of the random choosing of the length of free path of the observed trial particle and the random choosing of the volume of the field cloud particle which collides with the trial particles.

Unknown function of the distribution of cloud particles is proportional to the number of revealing trial particles in the given cell of phase space.

During the solution of the linearized kinetic equation of coagulation the necessity of the iterational scheme no longer arises.

The second approach includes the methods in which the Monte-Carlo method proper is used for the computations of collisional integrals. The computation of integrals of free path probability and function of appearance coupled with the integral iterational scheme of the kinetic equation allow us to express unknown function of the distribution of the cloud particles and its moments as a series of consecutive approximations directly through the function of distribution of the zero approximation and doesn't require memorization of the functions of the cloud particles distribution at previous iteration.

The zero iteration is set in analytical form.

Testing of numerical methods of solution of the coagulation kinetic equations has been carried out by means of confrontation of obtained numerical solutions with existing exact solutions in case of space-homogeneous task, allowing to get analytical solutions in the easier cases. By means of the Monte-Carlo method numerical solutions for the linearized kinetic equation of coagulation for the spaceinhomogeneous task describing the transfer and coagulation growth of the system of large cloud particles moving amongst small droplets with a certain function of the distribution are obtained.

The numerical solutions of non-linear kinetic equation of coagulation of the cloud particles in case of space-inhomogeneous task with set boundary condition at the cloud base have been obtained by means of the Monte-Carlo method.

The results of numerical experiments have been under the discussion.

# ON THE CRITICAL STATE OF THE WARM CLOUD FOR THE FORMATION OF RAINDROPS

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When cloud droplets have grown somewhat large by condensation in an ascending air, they will be put into random motions by eddies of air they are falling through, begin to coalesce by colliding with each other and go on to grow large enough to capture lots of small droplets as they fall through the cloud [1].

Before droplets become so large that their difference in fall velocity never be neglected for their random motions in turbulent air, collisions may be simply regarded as those between uniform droplets of an average size. Fig. 1 shows probability p that a particular droplet will collide per second with other ones of the same size whose number is unity per cm<sup>3</sup>. Eddy size has been taken such as makes p maximum but never less than lcm—a possible minimum size of an eddy [1], replacing eddy velocity with eddy size by Weizäcker's law, (eddy velocity)<sup>3</sup>/(eddy size)=const. The constant has been taken 8--the most favourable value for getting p=0.10 and 0.17 for the drops of radius r=0.20 and 0.25mm respectively. But it will be larger in convection or storm clouds.

Now, it seems that cloud droplets begin to grow fast into raindrops when they have got a certain size. This would be a matter of probability. From this point of view, such a critical size may be estimated about  $25^{0}30\mu$  in rsdius corresponding to the point of maximum curvature of p--curve in Fig. 1.

The collision process seems to be like that in which the transfer of liquid water is taking place from smaller droplets to larger ones at the rate of  $mn^2p/2$  or  $w^2p/2m$  per second, where m, n, and w are drop mass, number per cm<sup>3</sup>, and liquid water per cm<sup>3</sup> respectively.

Therefore,  $w^2 p/m$  may be available for the criterion of the expectation that cloud droplets will be able to grow up into raindrops before they are carried out of the cloud. Here, m, n, and w must be mean mass, number per  $cm^3$ , and liquid water per  $cm^3$  respectively of the droplets which have grown larger than the above-mentioned critical size.

 $w^2 p/m$  is estimated  $1.1 \times 10^{-10} \text{g/cm}^3 \text{sec}$  for the drops of r=0.20 mm at the level of about 1,500m within the drizzling cloud on one occasion [1]. Suppose that this quantity is in this case invariant with respect to r, then w=0.23 and  $0.19 \text{g/m}^3$  for the droplets of r=25 and  $30 \mu$  respectively.

Thus, it may be stated that raindrops would hardly be expected to form within clouds in compararively weak turbulence before some number of cloud droplets grow larger than about  $25 \sim 30 \mu$  in radius and in addition partial liquid water content referred to such large droplets becomes about  $0.2 \sim 0.25 g/m^3$ .

This paper is intended to reply to Dr. H. Dessens's remark on precipitation physics in 1965 "The process of condensation, Bergeron-Findeisen, coalescence are efficient only after another process has been accomplished; that being the concentration of water content in the cloud" [2].

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Fig. 1.

THE ROLE OF TURBULENCE IN THE DEVELOPMENT OF PRECIPITATION

By J.T. Bartlett, (M.R.E., Porton, England)

There are at least three distinct ways in which turbulence in clouds may possibly accelerate the development of precipitation:

- By modifying the numbers of cloud nuclei activated in the initial stages of condensation.
- By increasing the distribution of the sizes of droplets growing by condensation.
- By increasing the collection efficiency of small droplets.

Each of these effects has been considered in recent papers by different authors, not all of whom have reached the same conclusions.

This paper will attempt to critically review progress on this subject to date and to highlight the aspects which require further consideration. Particular attention will be given to the effects of mixing between a cloud and its surroundings and to the possible recirculation of droplets.

# SESSION FIVE

PHYSICS OF SNOW

# THE FORMATION OF SNOW CRYSTALS

Ву

# Choji Magono

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The review will consist of three parts:-

- (a) Ice crystals which are nucleated from nuclei other than ice forming nuclei, namely microscopic ice particles in air, including splinters.
- (b) Growth rate of snow crystals.
- (c) Crystal habit.

## THE SHALLOW PRECIPITATION SYSTEMS

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Snow crystals can be regarded as minute analogue computers which reveal in great detail certain atmospheric conditions in precipitation systems. The replacement of these analogue computers through digital numerical simulation is only possible for the most elementary forms such as prisms, plates, simple stars, which occur seldom in precipitation. So far, we cannot as yet unravel all detail of atmospheric conditions which realistic snow crystal forms reflect but their study and interconnecting analysis permits to draw conclusions to the existance of certain precipitation mechanisms. Cirrostratus and Altostratus, the "releaser" clouds in Bergeron's nomenclature, typically discharge prisms, often hollow, prism bundles and plate crystals. Bergeron's "spender" cloud which forms the supply of most of the precipitable water converts the Altostratus to Nimbostratus through the addition of a more or less convective Stratus or Stratocumulus layer. The prisms then acquire side plates and endplates and deform through riming whereby a rimed droplet often forms the center of a new branch which at the existing temperature and humidity is dendritic. The spatial forms promote further riming and growth by diffusion resulting in very complex endforms which no longer resemble the snowstars which have captured the fantasy of men throughout the centuries.

It is interesting that these stars, so prominent as the symbol of snow precipitation, form in entirely different and as yet little explored cloud system. They are the typical product of the shallow cloud systems which form in the planetary boundary layer and which are most typical of these regions on the globe where continental air flows across warm bodies of water such as the Japan Sea west of the peninsula Hokkaido, the lakes of the Great Lakes Easin in the USA, in the Mediterranean regions and probably in many other European regions downwind of the warm Gulf Stream waters and in the regions around the Black and Caspian Seas. While in

these areas a warm water surface provides moisture for the layer whose stability is nearly neutral, other cases of such systems occur in continental areas where moisture is either supplied by reaching the condensation level after a long upslope trajectory as occurs in the High Plains area east of the Rocky Mountains in the USA, in Bavaria, Germany, or simply by the low level convergence of a cyclonic disturbance.

The precipitation mechanism in these cloud systems is determined by convection and in-cloud updrafts which exceed the fall velocity of the snowcrystals. Upon reaching the cloud top the updraft spreads out under the boundary layer inversion and forms a thin Stratocumulus cumulogenitus layer. In this layer the ice crystals are collected and seed the clouds beneath whose updraft have ceased.

The ice crystal forms are then determined by the top temperature of the cloud system which is usually between -15 and  $-20^{\circ}C$ . At these temperatures stars are the prominent crystal habits produced in a water cloud. Examples of such types of precipitation are shown.

# The Size Distribution and Concentration of Cloud

Particles in Arctic Clouds

#### Ъy

# K.O.L.F. Jayaweera <sup>1</sup>

Clouds over land in the polar regions near the sea are typically stratus and occur in layers about 3000 feet thick and separated by about 1000 feet. Any other cloud formation is only exceptional and is rarely seen. The composition of these clouds were investigated using an MRI cloud particle sampler fitted to a Cessna 180 in the vicinity of Barrow, Alaska, during the month of September 1971. During this month the clouds extended from 800 feet to 12,000 feet and 35 passes were made at different temperature levels ranging from +2C to -11C. At each pass the volume of air sampled were about 500 liters. The purpose of these investigations were to establish the conditions under which ice particle multiplication does or does not take place and to determine the importance of ice nuclei on the ice crystal development in clouds.

The ice crystals observed in these clouds were all columnar and unrimed. The concentration of ice crystals at -11C and -7C were about 40 per m<sup>3</sup> and 10 per m<sup>3</sup> respectively and at -4C it was less than 2 per m<sup>3</sup>. These values agree well with the average ice nuclei concentrations at the corresponding temperatures, measured using Millipore samples collected at Barrow during this period and developed using the technique of Stevenson (1970). The ice crystals at -7C were sheaths and were larger than the solid columns observed at -11C. The sheaths had diameters near 30 µm and lengths about 100 to 200 µm, while solid columns were about 20 µm diameter and 50 µm length. The water droplet concentrations and spectrums were analysed for each pass. The size spectrums did not vary significantly at different levels, hence in Fig. 1 the overall size spectrum is shown. These values were corrected for the collection efficiency of the sampler which has a zero value for droplets 1. Geophysical Institute, University of Alaska, College, Alaska 48

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less than 3  $\mu$ m and almost 100% for drops greater than 35  $\mu$ m. The concentrations on most occasions were about 90 per cm<sup>3</sup> while the highest were 160 per cm<sup>3</sup>, giving a liquid water content of 0.2 to 0.3 gm per m<sup>3</sup>. Therefore, our analysis shows that the characteristics of the arctic stratus does not vary from day to day or at different levels, has a narrow spectrum without any big drops, and lie between the stratus clouds observed in Germany and Hawaii (Mason, 1971).

In conclusion, the results of the study at Barrow suggest that in clouds with narrow droplet spectrum and without any big drops the concentration of ice crystals can be determined from the laboratory measurements of ice nuclei and indicate that ice particle multiplication is found only in clouds that has big drops or rimed ice particles or both. These observations agree with the contention of Mossop (1971) on the factors necessary for ice multiplication in clouds.

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Figure 1. Average size distribution of cloud drops +2 to -110 September 1971, Barrow, Alaska
Initial Growth Process of Snow Crystals from

Frozen Water Droplets

by Akira Yamashita and Chuji Takahashi

(Geophysical Institute, Tokyo University, Tokyo)

The cloud chamber of about 15 meter high (Yamashita, 1971\*) made us possible to observe the growth of frozen water droplets in free fall. It was humidified by the ceaselessly falling supercooled water droplets of about  $1 \sim 2/cc$ . In this humidified condition, once a part of supercooled droplets were frozen they grew in free fall.

Almost spherical frozen water droplets were estimated to grow through the polyhedral shape having two basal

 $\{0001\}$ , six prismatic  $\{1\overline{1}00\}$  and twelve pyramidal  $\{1\overline{1}01\}$  faces to usual hexagonal double plates at about  $-11 \sim -18$  °C or to columnar ice crystals at about  $-4 \sim -8$  °C. This growth process is explained in Fig.1 and an example of the polyhedral crystal having 20 circular faces is shown in Fig.2.

\* Yamashita,A., 1971: Skeleton ice crystals of non-hexagonal shape grown in free fall. J. Meteor. Soc. Japan, 49, 215-231. A little change of growing conditions produced ice crystals of various interesting shapes, and those ice crystals estimated to be grown from shattered or deformed frozen water droplets were also obtained.

It was also found that air bubbles were frequently enclosed at the basal plane, at the pyramidal plane and at between every two pyramidal planes next to each other (see Fig.3).



Fig.1 The estimated growth stages and growth processes. The figures  $A \sim H$  are drawn as c-axes look up. Observed size range : about 30 ~150 µ in diameter.



Fig.2 Growth stage C (83  $\mu$  in diameter)



c-axis

10 µ

Fig.3

Hollows formed on pyramidal planes. These hollows are estimated to become air bubbles if this crystal grows a little more. Growth stage D.

### THE ICE PROPERTIES OF CONTINENTAL CUMULUS CLOUDS

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The details of airborne measurements of the ice crystal properties of the local continental cumulus clouds are presented and discussed in view of similar measurements in maritime clouds elsewhere. The data presented, based upon measurements in 59 clouds in the cloud-top temperature range of  $-8^{\circ}$ C to  $-25^{\circ}$ C, suggests that:

- Ice crystal concentrations at about 1000 ft. below the tops of the clouds correspond to the concentrations of ice nuclei active at the temperatures of these tops.
- No evidence has been found that suggests the eximtence of time dependent enhancing mechanisms.
- 3) Partitioning of the clouds according to their horizontal dimensions and the observed spatial distribution of ice crystals in the various cloud groups do not show any preference of certain clouds to multiplying mechanisms.

An analysis is presented also which discusses with some detail the formation of graupel elements in the clouds. Finally it is concluded that ice crystals may play a major role in the initiation of precipitation in the local continental clouds.

### STUDIES OF THE MICROSTRUCTURE OF CLOUDS AND THE DEVELOPMENT OF PRECIPI-TATION IN CYCLONIC STORMS OVER THE CASCADE MOUNTAINS OF WASHINGTON STATE

by

### Peter V. Hobbs

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During the past four years the Cloud Physics Group at the University of Washington has been studying the microstructure of clouds and the development of precipitation in winter cyclonic storms over the Cascade Mountains of Washington State. A review of this work will be given in this paper. The studies include extensive airborne and ground observations, the development of a theoretical model to describe the growth and fallout of precipitation, and investigations into the feasibility of modifying the distribution of snowfall across the Cascade Divide by artificial seeding.

A description will be given of the airborne and ground facilities which are being used in this program, with particular emphasis on a number of new techniques for the collection and automatic counting of ice particles in clouds.

Following a brief account of the synoptic character of Pacific Northwest storms detailed information will be presented on the microstructure of the clouds, the nature and concentrations of the ice particles in the air, and the mechanisms by which precipitation particles grow. The effects of the Cascade Mountains on the airflow, the structure of the clouds, and the distribution of precipitation will be discussed.

Finally, field and theoretical studies to investigate the feasibility of modifying the distribution of snowfall across the Cascade Mountains by artificial seeding will be presented.

### SNOW CRYSTAL GROWTH HABITS

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Modern theories of snow crystal habit tend to pass over the problem of the extremeness of habit variation at fairly high supersaturations. c/a ratios of from 500/1 to 1/500 have been found, and it seems difficult to get anywhere near this amount of variation from any of the theories proposed. If habit development is considered to be due to competition between basal and prism faces at edges of the crystal, then Marshall and Langleben's idea of increasing supersaturation causing increasing habit exaggeration does not apply, because very near an edge, the conditions at basal and prism faces will never be very different. A growth mechanism is needed that can give differences in growth rate of a factor of 500 or so at the same temperature and supersaturation, between prism and basal faces.

If we apply two-dimensional nucleation theory to snow crystal growth, and examine the effect of surface energy differences between prism and basal faces, we find that even small surface energy differences can give these extreme growth anisotropies, at reasonable supersaturations. To make these calculations give reasonable growth rates, however, it is necessary to use an ice surface energy much smaller than that generally assumed. This is not unreasonable, because, as pointed out by Hillig, the surface energy to be used in two-dimensional nucleation can be much less than the "bulk" surface energy when the thickness of the surface layer is greater than one monolayer, as is expected in ice. The four, familiar habit regimes may now be obtained by invoking surface energytemperature curves for the prism and basal faces like those already

suggested by J. D. Bernal.

Another growth mechanism that seems not to have been suggested before applies only to growth at temperatures fairly near the melting point and at conditions of supersaturation with respect to liquid water. In these conditions ice growth may well be fastest by first nucleating supercooled water on the ice surface and then growing ice from the liquid. This mechanism may account for the one or two highest temperature habit regimes. If it can account for the two highest, then a surface-structural transition need not be postulated to give the habit changes.

The relevant supersaturation is that at the crystal edges, not that of the environment. In the growth range where two-dimensional nucleation is fast, a large change in environmental supersaturation may give only a small change in edge supersaturation, because diffusion is rate-controlling. There is, however, some evidence that at extremely high environmental supersaturations c/a does approach unity, which is to be expected.

At low supersaturations the crystal growth rate by two-dimensional nucleation is vanishingly small. Therefore the crystals that do grow must have the appropriate imperfections, and the habits will not be as consistent or uniform as those at high supersaturation.

Tests of these ideas must be measurements of the relative surface energies of the prism and basal faces of ice as a function of temperature. Possible approaches might be through contact angle measurements or equilibrium form determinations, though both of these will give information on bulk surface energy and not necessarily that appropriate to the theory.

MOLITEATER ADJORFTION AND STEP GROWTH ON THE

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The molecular basis for the development of ice crystals in their multivaried forms has not yet been found. It appears that the vertical growth of a face is the result of a lateral growth of steps. In turn, the step growth is determined by the opposing mechanisms of surface diffusion and desorption of water molecules with the step velocity given by:

$$v = 2\delta F \chi_s / h\rho_{ice} \tag{1}$$

where  $\delta F$  is the net impingement flux,  $\chi_s$  is the mean migration distance of a molecule on the surface, and h is the step height.

Unfortunately, it does not appear that the simple Einsteinian relationship,

$$\chi_{s} = \sqrt{D\tau}$$

(D is the surface diffusion coefficient and  $\tau$  is the mean lifetime on the surface) can explain the temperature dependence, even in a qualitative way (see



Fig. 1).

In 1965 Hobbs and Scott attempted to explain this temperature dependence in terms of an adsorbed monolayer and a "blanketing" effect due to nearly total coverage of the surface. At first it appeared that the revised expression could explain the experimental results in a qualitative way. But, looking at the relative sizes of the terms more carefully, it can be shown that, for the low supersaturation used in the experiments, the revised expression degenerates to Eq. 2 (see Lakoma, 1969).

In further pursuit of this problem, we recently considered the case of multilayer adsorption, and the experimental

results do seem to be explained both qualitatively and semiquantitatively. The treatment assumes that the surface is not a table top, but is diffuse, perhaps with the molecular density profile shown in Fig. 2. It is interesting that Edwards and Evans similarly considered an adsorbed phase in relation to nucleation processes.

Consider the classic theory of multilayer adsorption of Brunauer, Emmett and Teller (1938). The adsorption isotherm treats molecules bound directly to the surface as being most strongly attached or having the largest bonding energy,  $E_1$ , to the surface. Any number of layers can then be attached to this lowermost layer and are attached with a bonding energy equivalent to the heat of condensation of the liquid, E1. When the number of layers becomes quite



large, liquid appears on the surface and the vapor pressure above the surface becomes equal to the equilibrium partial pressure over the liquid, po. Written in terms of the total number of adsorbed molecules n and the pressure ratio  $X = p_v/p_o$ , the basic equation is:

$$n/n_{\star} = CX/(1 - X)(1 + (C - 1)X)$$
 (3)

where  $p_v$  is the vapor pressure over the surface,  $n_t$  is the number of adsorption sites per cm<sup>3</sup>, and the sorption parameter C is defined by the equation

$$C = C_1 EXP[E_1 - E_1/RT]$$
(4)

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(2)

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where  $C_1$  is a constant generally considered to be of the order of unity.

Assuming only small excess pressures are used, the net impingement flux of molecules is

$$\delta F' = K_{p} \delta n \tag{5}$$

where

$$K_{B} = \frac{k_{T}(p_{0} - p_{e})^{2}(p_{0} + (C - 1)p_{e})^{2}}{Cn_{t}p_{0}(p_{0}^{2} + (C - 1)p_{e}^{2})}, k_{T} = (2\pi m kT)^{-\frac{1}{2}} \text{ and } p_{e} \text{ is the}$$

equilibrium vapor pressure of ice. The expression (5) relates the net impingement flux  $\delta F'$  to the excess of adsorbed molecules. The constant K<sub>B</sub> is the B.E.T. counterpart of the desorption constant K =  $1/\tau$  used by Hobbs and Scott. It defines an alternate form for the mean distance  $\chi_S$  (Eq. 2) and a different temperature dependence for v (Eq. 1).

Assume for the present purposes that the diffusion coefficient D has the temperature dependence used previously and that  $C \approx 1$ . The new desorption constant K<sub>B</sub> has the temperature dependence shown in Fig. 3 and, adjusting the parameters in the equations, the velocity of step growth has the form shown in Fig. 4. Indeed, the data of Hallett and Kobayashi are well described by the revised theory.

This result appears to be caused by the relative saturation of the surface by adsorbed molecules. As 0°C is approached the surface becomes water-like and can accumulate an infinite number of adsorbed molecules.



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## SESSION SIX

PHYSICS OF HAIL

### PROGRESS IN HAIL RESEARCH

### Roland List University of Toronto, Toronto, Canada

The progress in our understanding of the physics and dynamics of hail formation has been considerable over the past four years. The major achievements are considered to be as follows:

The hypothesis of growth of embryos from frozen drops, which has been more convenient from the computing point of view, could not be established conclusively because it is normally contrary to the structure of hailkernels, - except for the few cases where air bubble arrangements clearly point to frozen raindrops. Therefore, the dendrite - graupel - small hail cycle is essentially consolidated as the major growth mode.

Investigations on the free fall behaviour and the heat and mass transfer of the different embryo shapes have provided the basis to improve some newly developed graupel growth models and to insure that numerical simulation can start from more appropriate assumptions.

The study of hailstones was enriched by further micro analysis of internal features and by the use of x-ray diffraction for the determination of densities. The big step forward, however, came from the recognition that hailstone shells do have distinct and different distributions and concentrations of air bubbles. It was, therefore, not surprising that preliminary icing experiments in a wind tunnel showed that the bubble characteristics may represent a powerful and simple vehicle to supplement or maybe even replace the improved isotope techniques used to interpret the life history of hydrometeors.

The long suspected insufficiency of hailstone aerodynamics based on drag alone, and thereby, neglecting lift and moment effects, was finally tackled; it can be shown now how and when particles may rotate, spin, precess and nutate. For cases where these non-translatory motions are important, effective drag coefficients have to be established as well as new corresponding heat and mass transfer rates. Another point was considered too: the icing of model hailstones under naturally occurring particle gyrations.

The complexity of more or less complete simulation of hailgrowth has led to a wide variety of model types, each considering special and important features of natural hailclouds, but it will take quite some time until their conclusions can be checked and expanded in three-dimensional and time dependent models where the dynamics is based upon proper thermodynamics and where adequate treatment of particle distributions and their motion and growth is provided.

Modelling has to be closely linked to field observations, and again progress can be reported on single storm studies and hail climatology. It is disturbing, however, that only scanty data are available to extrapolate from hail on the ground to hailstone concentrations in the atmosphere. At present there are indications that concentrations of  $\sim 10^{-2}$  m<sup>-3</sup> are possible and maybe even normal. If this would be true, it means that a 'competition concept' in hail prevention might require increases in embryo concentrations by a factor of  $\sim 10^3$  over the natural one.

In the recent past the projects in hail prevention have not produced new ideas or concepts beyond the Russian operation. However, a new practical dimension was added by extensive studies of the economic aspects of hail prevention.

## MASS TRANSFER MEASUREMENTS ON MODELS OF SNOW CRYSTALS, CONICAL GRAUPELS AND SMALL HAIL

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In order to further our understanding of the initial stages of hail formation, a study of the mass transfer by diffusion and convection to models of snow crystals, conical graupels and small hail was undertaken. Measurements were made over a Reynolds number range applicable to such particles falling in the atmosphere, namely 25  $\lesssim$  Re  $\lesssim$  200 for the snow crystals and 400  $\lesssim$  Re  $\lesssim$  2000 for the other models. Measurements at lower Reynolds numbers could not be made due to the limitations imposed by free convection.

The measurements of the simulated mass transfer were made in a liquid tunnel by an electrochemical method, in which the current flowing in the electrolytic cell depends only on the diffusion and convection of ions to the model. This current can be measured with a high degree of accuracy and local transfer measurements can readily be carried out. Due to the similarity of heat and mass transfer, conclusions can also be drawn about the heat exchange.

Table 1 gives the ratio K of the transfer per unit area to the snow crystal models to the transfer to a sphere at an equal Reynolds number. The characteristic length chosen was always taken as the major diameter. The K values listed range from 0.95 for the hexagonal plate to 1.92 for the stellar crystal. Models with deep insections and with little shielding of the downstream face have high specific (per unit area) mass fluxes. The local transfer study showed that the specific mass flux to the side of a disc may be, for certain Reynolds numbers, up to four times that of the upstream face. A corresponding increase of approximately 2.5 times

was found for the stellar crystal with plates. The mass flux to the downstream face of a disc is less than that to the upstream face; however, for the stellar crystal with plates the situation is reversed.

Table 2 gives the K values for the conical graupel and small hail models. The transfer study showed that the effect of changing the particle orientation or adding roughness elements depends on the particle shape. Either main particle orientation, apex facing flow (Aff) or base facing flow (Bff), may result in the highest mass flux. Adding roughness elements to the  $90^{\circ}$  teardrop shaped model increases the specific mass flux by 50% in the Aff orientation; for the  $90^{\circ}$  cone-hemisphere the increase is negligible.

The transfer measurements can easily be combined with drag coefficient measurements. This leads to the first real understanding of the growth conditions experienced by small hail, originating from snow crystals and growing through a graupel stage. The corresponding general figures will be shown.

MODEL SHAPE	K	MODEL SHAPE	K
Hex. plate	0.95	90 <sup>0</sup> T Bff	0.76
Cylindrical	.96	90° T Aff	.74
Disc	.99	90 <sup>0</sup> C-H Bff	.88
Hex. cylinder	1.01	90 <sup>0</sup> C-H Aff	1.05
Broad branched	1.04	90° C-SS Bff	1.04
Stellar with pl.	1.49	90° C-SS Aff	.96
Dendritic	1.62	70 <sup>0</sup> C-SS Bff	.93
Stellar	1.92	70° C-SS Aff	1.12

Table 1. Transfer per unit area normalized to that of a sphere for snow crystals. Table 2. K values of teardrop (T), cone-hemisphere (C-H) and cone-spherical sector (C-SS) particles, with given apex angles.

## A WIND TUNNEL INVESTIGATION OF THE CRYSTALLINE STRUCTURE OF SMALL, FROZEN DROPS FALLING AT TERMINAL VELOCITY IN AIR

Ву

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The present investigation was undertaken to study the conditions under which water drops freeze single- or poly-crystalline, and in this way to clarify the role of frozen drops in the glaciation of atmospheric clouds. A frozen drop, whether single or polycrystalline, which continues to grow by collision with supercooled cloud drops develops into a polycrystalline graupel or hailstone. A frozen drop which continues to grow by vapor diffusion develops into a crystal cluster if it is polycrystalline and into a hexagonal, single snow crystal if it is singlecrystalline<sup>1-3</sup>. Most data available in literature on the polycrystallinity of water drops were derived from experiments with drops frozen by contact with  $ice^{2,4-8}$ . No measurements are available for the case of drops nucleated while falling freely at terminal velocity in air. Our present investigation consisted of three series of experiments: In a first series, drops of highly purified water and of sizes between 90 and 420µ radius were freely suspended in the air stream of the UCLA cloud tunnel. In order to stably suspend supercooled drops and ice particles at temperatures below O<sup>O</sup>C a mobile inner tunnel was devised which consisted of a contraction and observation section as the main tunnel but with smaller dimensions so that an air gap was left between the inner tunnel and the external tunnel. All air suspended drops could be supercooled to temperatures below -25°C before freezing. The supercooled drops were nucleated by contact with a dilute aerosol of dry clay particles. The frozen drops were removed from the tunnel air by a specially constructed, refrigerated "slurp-qun" which aspirated the ice particles into a small glass dish where they were caught at the interface of two organic liquids. The dish was then removed from the slurp gun and taken into a walk-in cold chamber where the frozen drops were investigated for polycrystallinity under a polarization microscope with a gypsum plate (Red Order 1) between polarizer and analyzer. In a second series of experiments 9µto 1 mm radius drops of various aqueous suspensions of Ag l particles were suspended at the interface of two These were contained in glass dishes and placed in the organic liquids. walk-in cold chamber. The air in the chamber was cooled in steps of 1°C. After each degree of cooling the drops frozen were removed from the dish

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and investigated under the polarization microscope. In a third series of experiments drops of highly purified water and between 3µ and 1 mm radius were supercooled at the surface of very visc ous paraffin oil contained in glass dishes. These were placed in the walk-in cold chamber in which the air was set at various temperatures to +0.2<sup>0</sup>C. The supercooled drops were brought into contact with a single-crystal ice plate prepared by a special technique. The drops frozen on the plate were then analyzed under the polarization microscope. The results of the three experimental series, documented by colored photographs, are summarized by curves 1 to 3 in To the left of and below each curve the drops freeze single Fig.l. crystalline, while to the right and above they froze polycrystalline or with a crystallographic orientation different from that of the ice seed crystal. Figure 1 suggests the following conclusions: (1) The results for water drops froze on ice qualitatively agree with those of previous authors. (2) A plot of the present results on double logarithmic scale shows that the condition for drops to freeze polycrystalline follows a law of the form  $r_{c} = \left[\beta \left(\Delta T\right)\right]^{3}$  while  $\beta$  follows a law of the form  $\beta = a/k^{b}$ , where a=1/8, b=23.0,( $\Delta T$ )<sub>c</sub> is the median supercooling for a population of drops of size  $r_{c}$  that freeze polycrystalline or with an orientation different than the ice seed, and where k is the heat conductivity of the medium outside the drop. (3) The obtained relationship shows that the formation of new orientations on ice growing through supercooled water is controlled by the heat flux away from the growing ice crystal. (4) At a given supercooling, a cloud drop frozen by an ice forming nucleus can be considerably larger and still be an ice single-crystal compared to a drop frozen by contact with an ice particle.

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Figure 1. □ drops frozen in air; ● drops frozen in oil; △ drops frozen on ice: present results, x Hallett, o Magono and Aburakawa. + Aufdermauer and Mayes.

# RESULTS OF HAIL PROCESS INVESTIGATIONS I.I.Gaivoronskii, L.A.Dinevitch, B.I.Zimin Central Aerological Observatory, U.S.S.R.

The present paper deals with the results of the physicomcteorological and radar investigations of hail processes for many years, including simultaneous observations which have been carried out using three radars set in the meridional direction in the southwestern part of the USSR and covering a distance of about 150 km. Meteorological conditions, which are favorable to the evolution of hail processes are defined. In these cases the situation is characterized by cold advection in the middle troposphere, convergence of flows that the direction is near a meridional one, while divergence of flows is observed for the more upper levels. An analysis of conditions under which the development of hailstorms doesn't occur has also been made. Probable relationships between storms and hail phenomena are established. Some particularities are noted in the structure of the lower atmospheric layer thick of five kilometers, when the hail processes are developing. The conditions of temperature and hygrometry during the days with hail are more variable as compared with days without hail.

Radar observations of hail processes were made using a 3,2cm radar. Synchronized observations by several radar stations permitted to determine a number of important features of the frontal zone and to follow the evolution of hail clouds both under their natural development and artificial modification. A great deal of attention was given to the regeneration of the zones of radar reflectivity from hail clouds. On the basis of the observation data, obtained with a specially created hail and rain measurement network, as well as the radar observations, a relationship was established between the paths of displacement of regions containing large liquid drops and the orientation of hail swathes. The configuration and the extent of the latter were defined as well as the duration of hail fall and the stone size distribution. For the period of 1965-71 the data of radar observations were obtained from more than 400 hail clouds and about 1000 shower clouds. Criteria determining a probable hail danger of clouds were found for a given region taking account of the aerosynoptical conditions. These criteria were successfully used in hail suppression.

By a method of combined areas the regions of hail growth and fall and the mesocharacteristics of a hail cell were determined. An attempt has been made to interpret the causes of the non-uniform distribution of hail along a hail swath. Data of observations for many years and isolated experiments, obtained with a dense precipitation network and related to the rainfall regime over protected and control areas, are presented. These data show that in the regions of hail suppression the artificial modification doesn't lead to decrease of the amount of precipitation.

### HAIL AND LIGHTNING FROM TRANSVAAL THUNDERSTORMS

### A.E. Carte and G. Held

Simultaneous observations of radar reflectivity structure, detailed surface patterns of hail and the location of lightning were made on thunderstorms in the vicinity of Pretoria and Johannesburg, South Africa during 1971/ 72. A 10 cm radar with a 1.1° conical beam was used. Some 4000 persons provided details of hail occurrences within an area of 40 x 70 km. Lightning at night was photographed by a network of five all-sky cameras spaced between 10 and 60 km apart. Visible ground flashes could be located by triangulation to within 1 km when photographed by two or more cameras.

The complexity of some hail fall-out patterns is illustrated in Figure 1. A high density of reports and accurate times are necessary to reveal such detail. Such studies have shown that intermittency of point hailfalls can arise from movement of irregularly-shaped hail areas, contraction and expansion of a hail cell and more than one cell traversing the point.

Storms studied so far have shown a wide variety of characteristics but common features are emerging, such as a tendency for hail and lightning to occur near the leading edge of radar echoes and for new cells to form on the left flank.

Figs. 2 and 3 show two examples of storms tracked by radar. Fig. 2 depicts an isolated storm that remained small for  $1\frac{1}{2}$  hrs then developed two more cells. All cells produced lightning but no hail. Heavy precipitation can obscure flashes, as evidently was the case between 1905 and 1920 hrs for a camera at the radar station (lowest histogram, Fig. 2), when other cameras recorded flashes from cell 3.

A 240 km line of storms travelled northwards on 3 November 1971. Some merged to produce the complex shown in Fig. 2. Others ahead of the line moved south-eastwards. Hailstones up to 2,5 cm diameter were observed, the largest on the left edge of the hail area.



FIG. 1 Details of hail areas at one-minute intervals within a 10 x 10 km square, densely covered by hail observers in Johannesburg. This area falls within a hail path more than 50 km in length that occurred on 29 September 1970. The density of points from which the hail areas for 1546 hrs were derived is also shown (lower left).



FIG. 2 Radar PPI sections (near ground level) at various times for an isolated storm that was tracked for 3 hrs on 30 October 1971. It gave rise to lightning (see histograms on right) but no hail. Echo top heights against time are also shown. The vertical cross-section is through Cell 2, in the direction of motion.



FIG 3 Radar PPI sections (near ground level) are shown for 1650 and 1810 hrs on 3 November 1971. Only the envelopes enclosing the most intense echo are indicated. The hatched areas are where hail fell on this day and the solid black area shows where it was hailing at 1810 hrs. The locations of lightning flashes to ground are indicated for the period 1810 to 1815 hrs. The oval envelope encloses all points at which lightning was located between 1800 and 1840 hrs. The earliest flashes occurred at the southern end of the oval and the latest ones near the northern extremity. Echo top heights and lightning frequency are also shown (right hand side). The cross-section (top, right) shows the most intense echo and contours for weaker echo.

## SUPERCELL AND MULTICELL ALBERTA HAILSTORMS

by

A.J. Chisholm<sup>1</sup> and J.H. Renick<sup>2</sup>

The Alberta Hail Studies Project has observed a broad spectrum of hailstorms over the past five years with a 1.15° beam 10 cm radar. This paper reviews these observations with respect to the influence of the ambient wind and storm energy on the development and persistence of the two characteristic storm types; namely the organized multicell and supercell storms, for which descriptive models of the reflectivity structure, airflow and hailfall are presented.

Supercells are characterized by a vault or Bounded Weak Echo Region with a diameter of 5-10 km and a lifetime of 15-45 min. The BWER is coincident with a broad strong updraft that begins in the sub-cloud air on the RH storm flank, proceeds upward through the storm core while moving toward the LH flank and is subsequently sheared downwind (see Fig. 1). Hail forming on the updraft periphery results in a high  $Z_e$  gradient around the BWER. Due to organization by the wind pattern, parallel but separate updraft and fallout zones develop which result in a stable, persistent storm airflow. As a consequence of the strict conditions which must be satisfied, less than a half dozen supercells occur within the 50,000 km<sup>2</sup> project area each summer.

By contrast, multicell storms may or may not be organized depending upon the wind pattern and intensity. Multicell storms develop in vertical stance so that the precipitation loads the updraft causing it to decay rapidly. The organized multicell storm propagates by means of a recurring

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pattern of cell development on the RH storm flank (see Fig. 2), with one or more cells exhibiting a WER at any one time. As the WER and updraft are time dependent and short-lived, the instantaneous WER may only be used as an air <u>streamline</u> indicator. An air parcel <u>trajectory</u>, on the other hand, is delineated by tracing the history of a single cell from cumulus stage to dissipation. Such an air parcel, originating in the cumulus stage nears the top of the storm in the mature stage and is the volume responsible for the initial echo. After releasing its precipitation load, this air exits forming the cloud anvil as shown in Fig. 2.





Fig. 1. (Top) Supercell Storm. Contours of  $Z_e$  labelled in  $dB_z$ .

Fig. 2. (Bottom) Multicell Storm. Heavy arrow depicts air parcel trajectory.

### AIRCRAFT MEASUREMENTS OF AIR MASS AND WATER VAPOR CONVERGENCE IN A NORTHEAST COLORADO HAILSTORM J. C. Fankhauser National Center for Atmospheric Research, Boulder, Colo., U.S.A.

Sequential (~15 min) position and configuration of PPI radar echoes from a hailstorm in NE Colorado on July 14, 1971 are shown in Fig. 1. Average height of the terrain is 1.5 km, and observed cloud base was near 3.2 km (MSL). Radar echo height reached 14.8 km at 1817 MDT, some 2.5 km above the tropopause (200 mb) as determined from nearby rawinsondes. Environment winds were easterly in the subcloud region and backed through north to west-northwesterly in the cloud-bearing layers. Movement of the most persistent element ( $280^{\circ}/10 \text{ m sec}^{-1}$ ) was 5° to the left of and 6-8 m sec<sup>-1</sup> less than the mean ambient air flow from cloud base to the tropopause.

Observation of this system by aircraft began at 1700 MDT, shortly after the first precipitation echoes appeared on the Grover radar, and continued throughout the life of the storm, at altitudes ranging from 2.2 to 10.5 km MSL. A deHavilland Buffalo equipped with an inertial navigational platform, operated in the subcloud region (850-700 mb) and completed 9 constant-pressure circumcloud orbits between 1700 and 1920 MDT. A North American Sabreliner flying at 8.0, 9.25, and 10.5 km and relying on Doppler navigational equipment for horizontal wind measurement, accomplished four similar patterns before 1900 MDT.

Flight patterns are designed to enclose the cumulonimbus cloud so that flow normal to the aircraft track, relative to the storm, may be integrated to provide estimates of horizontal divergence (Table I, Column 4) at intervals of 10-20 min. Dew point measurements from the Buffalo are used with wind measurements to assess water vapor convergence (Column 5), which, when compared to radar and rain gauge data, provides an evolving estimate of the storm's precipitation efficiency (Column 7). The confidence level is highest for the bracketed value and this is close to the mean over the storm as a whole. Efficiency in excess of 100% late in the storm's life reflects diminishing water vapor convergence and heavy rainout during decay.

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(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
TIME	P	А	D	M	R	E	Н
INTERVAL	(mb)	(km <sup>2</sup> )	(10 <sup>-4</sup> sec	<sup>1</sup> )(ktn sec	-1)(mm hr <sup>-1</sup> )	(%)	(km)
1704-1711	767	93	-12.4	2.00	77.5	9	9.2
1708-1719	767	174	-9.6	3.13	64.7	20	10.5
1720-1732	788	263	-12.4	5.18	71.0	24	12.0
1728-1741	789	303	-8.8	4.35	51.4	39	13.0
1729-1747	380	1321	0.8	Rúth	-	-	13.2
1741 <b>-</b> 1753	788	255	-12.2	5.00	70.6	32	13.5
1747-1808	380	2086	2.8	· -	-		14.1
1753-1806	788	292	-12.8	5.25	64.7	[42]	14.2
1806-1821	789	398	-10.6	7.21	65.2	48	14.8
1821-1840	790	691	-5.1	5.62	29.3	>100	13.8
1836-1854	267	2862	2.3		-		12.6
1907-1921	786	331	-1.9	0.74	8.0	>100	11.0

TABLE I. Measurements and calculations from aircraft and radar observations. High altitude aircraft data in italics.

1) Time (MDT) interval for individual closed loops.

2)  $\overline{P}$ ; the mean static pressure over the interval.

3) A; the area enclosed.

4)  $\overline{D}$ ; horizontal divergence.

5)  $\overline{M}$ ; water vapor convergence into the subcloud layer.

6)  $\overline{R}$ ; precipitation rate, assuming  $\overline{M}$  condenses and falls within A.

7) E; precipitation efficiency; ratio of measured rainfall rate to  $\overline{R}$ .

8) H; height of radar echo top; from Greelev radar.



# THE ELECTRICAL BEHAVIOUR OF SEVERE HAILSTORMS IN ALBERTA, CANADA

by

J. Maybank and J.E. Pakiam Saskatchewan Research Council, Canada

During the summers of 1968 to 1970, the potential gradient was measured beneath severe storms in Alberta, Canada, using "field mills". The programme was carried out in conjunction with measurements made by the Alberta Hail Studies Organization, these including PPl data from a narrow beam 10.4 cm radar located at Penhold and hailfall and rainfall data from a volunteer farmer reporting network.

Two storms have been analysed in some detail. From an analysis of the radar film, the storm of July 11, 1970 was determined to be multi-cellular, with new cells growing on the south side of the storm complex and maturing as they moved through the complex. The potential gradient measurements indicated a very high lightning flash rate (up to 40/min) with a complex potential gradient pattern. Radar storm tops reach 43 kft ground level and maximum hailstone size associated with the storm was walnut (mean diameter ≈ 2.4 cm). On the other hand, radar evidence showed that the storm of July 9, 1970 was a steady-state Browning "supercell" with propagation occurring in a continuous manner. The potential gradient records showed very low lightning flash rates (a steady 1-2/min) and a simple pattern to the potential gradient variation. This storm resulted in a hailfall of stones larger than golfball size (mean diameter of golfball size  $\approx$  6.0 cm) while radar storm heights were 🛪 50 kft.

The evidence shows that the electrical behaviour of a storm is a function of the complexity of a storm, a storm with many cells contributing to the electrical activity having more lightning than

one with only one cell in it. In addition, the analysis of the potential gradient records of other storms suggests that the amount of lightning from a storm complex is proportional to the height of the storm, except in the case of the Browning "supercell" (Fig. 1). Here, either the monocellular nature of the Browning storm or some weak electrical charging mechanism appears to keep the lightning flash rate at a very low level.





## SESSION SEVEN

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# STRUCTURE AND DYNAMICS OF CLOUDS

### THE STRUCTURE AND DYNAMICS OF CLOUDS

by:- H. D. Orville, South Dakota School of Mines & Technology, Rapid City, South Dakota, U.S.A.

This paper attempts to summarize the advances over the past four years in describing the initiation, growth and dissipation of cumulus type clouds. As a point of departure for the review, the third equation of motion is written so that the various effects on the cloud's vertical velocity are illustrated. These include a pressure gradient term, thermal and water vapor bouyancy terms, a drag force term due to cloud, rain and hail particles, an electrical force term, and a turbulent mixing term. In addition, the vertical motions within a cloud are influenced by large scale convergence divergence patterns which at this time are simulated in numerical models through initial and boundary conditions. Research concerning these effects and the interactions among the various terms are discussed.

Several studies concern the production of cloud, rain and hail particles, which ties together the microphysical and dynamic aspects of the cloud. Water conservation equations stressing the mixing ratio concept and utilizing assumed particle size - distributions, or actual calculations of the particle size - distributions as they evolve and their implied water conservation are the two most common methods for describing this interaction.

Turbulent mixing within the cloud and between the cloud and its environment are crucial aspects of the problem with no definitive answers as yet. Some two-dimensional numerical models use constant eddy coefficients and others nonlinear eddy coefficients.

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One-dimensional models use mixing that is inversely proportional to the radius of the cloud element. These last models are proving extremely useful in operational weather modification projects. A comparison of the mixing effects in the two types of models can be made.

Rudimentary electrical effects on cloud dynamics are being modeled. These effects, certainly negligible in fair weather cumuli, may become of great importance in certain portions of mature storms.

Numerical models of one and two space dimensions have been most common but with the advent of more powerful computers the emphasis is now switching to 3-dimensional time-dependent models. Some conceptual models and hydrodynamic analogies to storms are also receiving attention. However, these models lack the important coupling of the cloud microphysics and dynamics.

Observational tools are being improved. Weather radar including Doppler, is the most useful for studying large storms, but various research programs use aircraft also, particularly for nonprecipitating clouds. Nevertheless the state of the art of aircraft instrumentation for measurements within clouds is still far behind what it should be. In addition, the volume sampled within a cloud via aircraft is a small fraction of the space-time volume of a cloud.

The development of numerical and conceptual cloud models, the refinements of weather and Doppler radar, and the organization of cloud physics programs utilizing radar and instrumented aircraft signifies a major effort at understanding the structure and dynamics of clouds.

THE STRUCTURE OF RAINBANDS WITHIN A MID-LATITUDE DEPRESSION

K. A. Browning, C.W. Pardoe - M.R.U. Malvern M. E. Hardman, T. W. Harrold - Meteorological Office Bracknell

A detailed case study has been made of a winter depression over the British Isles in which extensive banded structure was observed in the precipitation above the warm front. Measurements of the mesoscale airflow and precipitation structure of the rainbands were made using a variety of radar techniques together with multiple aircraft and radiosonde observations. The measurements were made over the sea to avoid confusing effects of topography. The dominant rainbands were found to be oriented parallel to the surface cold front, were typically 100 km wide, and they moved with a velocity faster than the underlying warm front. For the most part the bands were characterized by clusters of weak small-scale convective cells owing to the release of potential instability produced where tongues of relatively dry,  $low-\theta_w$ , air in the middle troposphere overran low-level moist air which was beginning its slantwise ascent above the warm frontal zone. Although there was the usual large-scale, and thermally-direct circulation associated with the active warm front, the air which was pumped up by the small-scale convection within the rainbands entered a region of weak positive baroclinity (as opposed to the negative baroclinity of the warm frontal zone) whereupon it participated in a thermally direct circulation of its own. This led to each rainband having a rearward-sloping anvil cloud canopy characterized by ascending air with colder drier air descending beneath. Precipitation falling from the canopy evaporated within the underlying drier air thereby intensifying the descending branch of the circulation. In one of the rainbands the mesoscale transverse circulation was associated with very large ageostrophic winds, and frictional forces are suspected to have played a major role.

The important ingredient responsible for the formation of the rainbands appears to have been the incursion of the relatively dry, cold, air in the middle troposphere above the moist warm-sector air in a region where the resulting potential instability could be realised by large-scale ascent. Although the potential instability was very weak in the present case, the origin of the rainbands appears to have been essentially similar to that of pre-frontal squall-lines. The intensity of the convection within rainbands depends on the stability but the <u>existence</u> of rainbands depends on other dynamical factors leading to widespread ascent.

The fact that the rainbands were observed over the sea shows that they were a characteristic feature of the system. Overland, however, convection may be triggered preferentially over (sometimes continuing downwind of) hills, thereby leading to a partial reorganisation of the rainbands. Further studies are needed to investigate the susceptibility of convective rainbands to oreigenic effects as a function of such factors as the magnitude and height of unstable layers in relation to frontal zones.

### Observations and Study of Convective Storms by Doppler Radar

### R. M. Lhermitte University of Miami

The paper presents Doppler radar observations of precipitation particles velocity fields inside convective storms occuring in south Florida and their interpretation in terms of the storm's threedimensional circulation.

A brief discussion of the dual-Doppler radar method, the accuracy of velocity measurements and the relevance of precipitation particles for tracing air motion is included in the paper.

A new real-time method for the digital processing and display of the mean Doppler velocity at a speed comparable with the rapid evolution of convective storms is presented. The method is capable of processing mean radial velocity fields at the rate of several thousand mean Doppler samples per second thereby allowing excellent space continuity in the presentation of velocity data in a two-dimensional plane associated to radar beam scanning.

The complete Doppler spectra are also sampled and processed but at a much slower rate (100 spectra per second) by a digital real-time unit evaluating the Fourier transform of time signals at selected ranges. The information is presented and stored in the form of Doppler spectrum/ range functions obtained for selected radar beam positions.

The analysis of the data acquired inside convective storms reveals that indeed very wide Doppler spectra (10 m sec<sup>-1</sup> or more) are often observed, especially in the case of a strong shear in the low levels of the storm environment. Such spectrum width exceeds the expected contribution of the environmental wind shear across the radar beam and seems to imply concentration of the shear due to the storm velocity processes and also the possible contribution of velocity fluctuations at smaller scale. However, the mapping of mean motion obtained from mean Doppler data exhibits a very consistent and well organized structure of the velocity fields showing time persistent features, such as strong, local, vorticity and convergence.

# SOME NEW DATA ON THE EXTERNAL STRUCTURE AND DYNAMICS OF CUMULUS DEVELOPMENT OVER THE CAUCASIAN MOUNTAINS

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Since 1967 to 1971 the systematic theodolite and terrestrial photogrammetric observations were undertaken in the area of the South Georgian upland (the altitude- from 2 to 3 km above the Sea level) in Caucasus for the purpose of establishment of the dynamic characteristics and external structure of mountain cumuli.

As a result of the observational study of Cu con the mean v values of the following parameters were determined: vertical velocities of cloud summits and their distribution by the height, the period of cloud-summit oscillation in the vicinity of stably stratified layers, the frequency of thermal generation in clouds and their life-time in an unsaturated environment.

Along with the establishment of horizontal and vertical dimensions of clouds and their towers the tower expansion rates were determined and the available energy of their growth was estimated.

The observational data on the outer structure of Cu con allowed us to reveal 3 characteristic scales of the horizontal extent of various cumulus elements-the composite towers, single thermals on the cloud surface and the inhomogeneities of the thermal surface, due to the turbulent interaction between the cloud and its environment.

Connection between the average size of the third type elements with the processes of cloud development was determined. The minimum sizes were found to occur on the surface of rapidly growing cumulus towers, while the size of the elements grew by the time on every fixed level in the passive state of the cloud.

The comparison of these results with the data obtained by others in Sweden, USA, Australia (Saunders, Anderson, Brown) and the European part of the USSR(Vulfson) showed that in spite of comparatively small vertical extent of clouds observed by us, they have much common in the dynamics of growth. The aircraft observations on the structure and the distribution of vertical velocities in cumuli (Ackerman, Kornienko) showed the close agreement between the results of the determination of cloud element scale according to inner and outer structure of clouds.

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## CUMULI STRUCTURE AT VARIOUS EVOLUTION STAGES

### N.I.Vulfson, A.G.Laktionov, V.I.Skatckij

## Institute of Applied Geophysics, Moscow, U.S.S.R.

1. Heterogeneous cumuli structure is mainly stipulated by convective columns in clouds. Numerous measurements showed that those convective columns differ from environmental cloud air not only in temperature and motion velocity but in microstructure parameters as well. The present paper is an endeavour to investigate connection of some parameters of macro- and microstructure of clouds at various evolution stages: on the basis of the measurement results obtained in 57 cumuli during one day convective columns in clouds various size drops concentration and cloud water content are considered.

2. Convective columns in clouds were identified by a temperature field; convective columns were determined as zones with a slightly higher temperature compared with that of environmental cloud air. Cloud microstructure was determined by means of two inflow photoelectric measuring devices for cloud drops with diameters more than 3, 5, 10, 15, 27, 32, 38, 52, 57, 70 > m and water content measuring device. The devices listed had time constant of about 0,03, 0,1 and 2 sec correspondingly. The reading of all instruments was recorded on an oscillograph tape.

The above said apparatus was mounted at an aircraft, type IL-14. The flights were conducted in the region of Siktivkar under conditions of air mass weather on sunny day of July 10,1970. The measurements were mostly conducted in the upper part (one third) of the cloud. The clouds investigated were divided into four groups:

Group		2	3	26 10
'Measurement time	9-2-920	9 <sup>22</sup> -10 <sup>27</sup>	11 <sup>04</sup> -12 <sup>24</sup>	16 <sup>20</sup> –18 <sup>10</sup>
, clouds	5	17	15	20
Cloud thickness,	100-150	500-1000	1000-2100	300-500
			1 	1

3. The consideration of the drops concentration on the oscillogrammes showed that drops zones, particularly when sizes exceeded 10 µm did not fill clouds completely. In this case drops concentration in some zones (regardless of their sizes) can significantly differ from each other even in the same clouds.

4. The presence of drops zones is connected with ascending convective columns in clouds. Though drops zone sizes depend on fixed drop sizes (decreasing with a drop size increase), the distances from the edge of a cloud to a point of a drop zone where maximum drop concentration is observed coincide as a rule with that to a point in a corresponding convective column where there is maximum temperature excess.

The coincidence of above said distances takes place in all zones with all fixed sizes drops.

5. It is possible to distinguish mean size drops zones (where drops exceed some size) at all cloud development stages besides original one (where droplets with diameters exceeding 10  $\mu$ m do not form separate zones) which coincide with mean sizes of convective columns. For the clouds of 2-4 groups the mean sizes of convective columns appeared to be equal to 140, 160 and 150 m correspondingly. Approximately the same sizes have zones with drops diameters more than 30, 40 and 30  $\mu$ m.

Thus the drops of above said sizes are mainly concentrated inside the convective columns.

6. During the intensive cloud development (2 and 3 group) there is rather pronounced relation between maximum temperature of convective columns and maximum concentration of droplets of all measurable sizes. As it was said above both are observed in the same cloud points.

7. Drops with diameter more than 70 µm were observed at all cloud development stages (though not in all clouds at the initial stage). The mean concentration in the considered cloud groups was correspondingly 0.1; 7; 24 and 0.8 particles/litre.

At the initial stage of cloud development the drops of said above sizes did not form separate zones. At the further development stages mean area of the zones with drops which diameter exceed 70 µm was approximately equal to 0.30; 0.35 and 0.02 of cloud cross-section at flight levels. A precipitation growth process for stratiform warm cloud in presence of a vertical wind shear suggested by 8.6 mm radar observations

### Henri Sauvageot

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Thin altocumulus or stratocumulus frequently show billows formed by the combined action of a small vertical wind shear and of an unstable stratification produced by radiative heat fluxes (Scorer 1951 - Ludlam 1967). Some observations above the Plateau de Lannemezan of such billows in a warm cloud layer were obtained with the use of a vertically pointed 8,6 mm narrow beam radar, and were correlated with upper-air data ; recordings show that, in spite of its weak thickness (800 to 1000 m), the layer produces precipitation trails, with one virga associated to one billow. Moreover the reflectivity pattern is asymmetrical. Two types of asymmetry associated with two inverted vertical wind shear profiles are observed.

These observations suggest the following process of droplet growth : in the billow the droplet trajectory is determined by the air movement and by the weight of the droplet ; small droplets with negligible fall velocity follow air movement and have a nearly circular trajectory ; droplets with largest fall velocity have a more complicated trajectory : the rotation center moves horizontaly towards the ascending part of the billow, and the radius decreases (see Fig.1). Moreover, the vertical component of the air velocity at a given radius in the billow is maximum in the median level. For this reason, above this level a very favourable region for the drop growth exists, similar to the accumulation zone in cumulonimbus pointed out by Sulakvelidze. According to this process the ascending part of the billow present the strongest reflectivity and the asymmetry is governed by the shear.

This mechanism gives a possible explanation of some contradictory observations (Facy 1959) about precipitation under thin stratiform warm clouds. It also implicates correlations between the cell reflectivity, the vertical wind shear and the precipitation fall velocity ; the study of these observed correlations is in agreement with the proposed mechanism.



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Author : P. C. Sinclair, Colorado, U.S.A.

Aircraft Measurements Near and Within Severe Storms

Specially instrumented jet and propellor type aircraft have been used to measure wind and temperature fields near and within severe storms in northeastern Colorado during the summer hail season. Both aircraft are equipped with boom mounted, moving vane, relative wind sensing systems. A gyro reference and center of gravity accelerometer system provides data on the aircraft motions. The measurement systems are advanced models of those described by Sinclair<sup>1</sup> (1969). They will be further elaborated on here primarily in terms of system errors, calibrations, and flight test results.

Because of the need to more fully understand the structure and intensity of the updraft region of a severe storm, the aircraft measurements were concentrated in this area both at cloud base and at upper levels between 28,000 ft and 35,000 ft MSL. The measurement plan was designed to provide direct quantitative information for realistic cloud model initial conditions as well as independent upper level boundary conditions. While the sub-cloud aircraft obtained wind and temperature measurements at or near cloud base, the upper-level jet aircraft made direct penetrations through the maximum reflectivity region as indicated by ground-base radar. The upper-level storm penetrations were initially started at a "high" <sup>1</sup>Sinclair, P. C., 1969: Proceedings of the International Conference on Cloud Physics, Toronto, August 26-30, 1968.

altitude with each successive penetration flown at a lower altitude until damaging hail was encountered and/or aircraft g limitations were approached. All aircraft tracks were flown essentially at constant heading and constant attitude both in the sub-cloud and upper-cloud regions.

The measurements further substantiate previous results that indicated a relatively smooth cloud-base updraft region as compared to the more turbulent upper-level updraft measurements. The upperlevel penetrations encountered maximum vertical air velocities of  $38 \text{ m sec}^{-1}$  and velocity variances of  $34 \text{ m}^2 \text{ sec}^{-2}$ . The maximum velocity variances occurred near the maximum updraft region which is in agreement with the results of doppler radar measurements. However, maximum horizontal shears of the vertical velocity exceeded those derived elsewhere by doppler radar techniques. In general, the upper-level measurements indicate a relatively low turbulence, sometimes subsiding, region outside of the visible cloud, a higher turbulence, updraft-downdraft region between the visible cloud boundary and the main updraft region, and a high turbulence, strong updraft, warm core region of the storm. The sub-cloud updraft size and structure features will be compared with those found at the upper levels.

# OBSERVATIONS FROM AIRCRAFT PENETRATIONS OF HAILSTORMS

By

D. J. Musil, R. A. Schleusener, W. R. Sand and J. H. Hirsch South Dakota School of Mines and Technology, U.S.A.

Observations have been made during penetrations of active hailstorms by an armored T-28 aircraft in the region between 24,000 ft MSL and the melting level. These observations were made in conjunction with the National Hail Research Experiment being conducted in northeastern Colorado.

The observational platform is an extensively modified North American T-28 aircraft which has been armor plated to withstand hailstones up to 3 inches in diameter. The principal objectives of the measurements are to determine the composition of high reflectivity zones, to measure updraft velocities in the storm, and ultimately to determine the ice-water budgets.

The size of hailstones is currently being estimated by pilot's comments and the noise of hail striking the windshield, both of which are recorded on a dual-track tape recorder. Instrumentation is under development for a quantitative measurement of hailstone size and concentration based on the concepts of momentum sensing and light interruption. Prototype sensors will be flown during the summer of 1972.

Other instrumentation includes equipment for measurement of altitude, airspeed, rate of climb, vertical acceleration, temperature, liquid water content, aircraft position and time. Meteorological data are recorded on magnetic tape and computer analysis procedures for data reduction have been developed. Analysis of a typical data sample is presented, which includes such parameters as updrafts, temperature, altitude, and hail encountered.

# SESSION EIGHT

NUMERICAL MODELLING OF CLOUDS

## NUMERICAL EXPERIMENTS IN

## SIMULATION AND PARAMETERIZATION OF SMALL-SCALE

#### THERMAL CONVECTION

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In a series of numerical experiments, small-scale non-hydrostatic convection is first studied by integration of the nonlinear system of equations for conservation of mass, momentum, and thermodynamic energy. The integrations are carried out in two and three space dimensions and time, and the equations include terms representing the effects of turbulence on the sub-grid scale and of phase changes of water. The results are compared with observations and with previous numerical studies of moist convection. In simplified cases simulating dry laminar convection, the numerical results are compared in detail with the heat transports and cell structures observed in laboratory experiments. Quite good quantitative agreement is found, and the advantage of including three space dimensions is demonstrated. The results are illustrated by a computergenerated film of the evolution in time of the velocity and temperature fields. Some implications of these studies for the parameterization of small-scale convection in models of the large-scale atmosphere are discussed.

## OF CONVECTIVE CLOUDS

#### R.S. Pastushkov

## Central Aerological Observatory U.S.S.R.

It is known from observations that the ambient wind and especially its vertical shear  $\mathcal{U}'_o$  play an important role in the dynamics of convective clouds. However hitherto there is no three-dimensional model of convective cloud in the atmosphere with a wind shear. In the present study this model have been constructed and the effects of  $\mathcal{U}'_{i}$  on the evolution of an isolated three-dimensional convective cloud are studied with numerical integration of hydrodynamic and thermodynamic equations. The model included: three equations of motion, continuity equation, thermo-dynamic equation, water vapour and liquid water equations, equa-tions concerned with the eddy diffusion coefficients and terminal fall velocity of the water drops. A special feature of the model was the dependence of all eddy diffusion coefficients upon the field of motion and the average scale of turbulence. Closed lateral and no slip vertical boundary conditions will be adopted. The dimensions of the domain under consideration were a height of 12 km, a width of 14 km and a length of 22 km. Mesh sizes were 1 km in all directions. The time interval was 20 sec. Computational scheme was two-order accuracy. Calculations are terminated at 40-60 min after initialization of convection.

Four series of calculations have been performed, which in-

clude sixteen runs differing from one another both in the value of  $\mathcal{U}'_{\bullet}$  and the initial value of static energy of the atmospheric instability per unit of the height of unstable layer  $\mathcal{F}_{\bullet}/_{\Delta}H$ . In every series of calculations the value of  $E_{o/A}H$ was constant 24' varied from 0 to 3 or 6 m/sec.km. and the value of

It is found that convective clouds may be grouped into two

It is found that convective clouds may be grouped into two different types ("weak" or "strong") in accordance with the ef-fects of  $2\ell_o'$  and  $\frac{\epsilon_o}{_{\Delta}H}$ . It has been shown that if  $\frac{\epsilon_o}{_{\Delta}H} < 0.65 \cdot 10^2$ , then convection is "strong". For "weak" convective cloud there is the critical value of shear  $2\ell_{occ}'$ , which depends on the value of  $\frac{\epsilon_o}{_{\Delta}H}$ . The value of  $2\ell_{occ}'$  becomes smaller when the value of  $\frac{\epsilon_o}{_{\Delta}H}$ is decreased. At  $\frac{2\ell_o'}{_{occ}'}$  in agreement with previous numerical studies of other authors there is inhibiting effect of  $2\ell_o'$  on the development of weak convective cloud and at

on the development of weak convective cloud and at  $|2l_o'| \ge 2l_o'$  its evolution is completely suppressed. 210

The essential new result of the present study is that the development of "strong" convective clouds under the effect of  $2'_o$ are intensified. It is manifested first of all in formation of new convective cells on the down-shear side of the cloud and in subsequent increase of kinetic energy of vertical motions and rates of various energy transformations. For "strong" convective clouds there is a resonance value of shear  $2/2_{***}$ , which also depends on  $\frac{\epsilon_{*}}{4}$  and at which the degree of the intensification of convection has its maximum (30-40 %). It was found that for values of 2/0cz reasonable values of E./AH and Uores. are close to 2-3 m/sec.km.

So it has found that ambient wind shear not only inhibits

the development of convective clouds and prolongs their life cycle, as had been shown by previous authors, but may intensify the convective clouds. All these features are in good qualitative agreement with observations.

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# NUMERICAL EXPERIMENTS ON THE RELATION BETWEEN MICROPHYSICS AND DYNAMICS IN CUMULUS CONVECTION

by Francis W. Murray and L. Randall Koenig

A two-dimensional hydrodynamical cloud model with cylindrical symmetry, incorporating Kessler's (1969) and Berry's (1968) parameterizations of cloud microphysics, is used to study several aspects of the interaction between dynamics and microphysics. The basic model is that of Murray (1970). The parameterization of the microphysics includes condensation and evaporation of small droplets, conversion of small droplets to large drops, evaporation of large drops (at a slower rate), and fallout of large drops. Three runs were made for different drop concentrations and dispersions, and three additional runs were made to determine the consequences of (a) suppressing formation of large drops, (b) evaporating large drops as fast as small ones, and (c) suppressing conversion of latent heat during evaporation. These runs with artificial conditions help to isolate the effects of evaporation, water loading, etc. on the dynamics of the cloud. In all of the runs a sounding was used that produced a mediumsized tropical cumulus cloud. In all but the last-mentioned run (which showed explosive growth) the life cycle of the cloud was simulated from inception to complete dissipation.

The three ordinary runs produced very realistic cloud simulations. The differences among them were what one would expect; e.g., a maritime cloud grows more vigorously than a continental cloud, produces a higher proportion of large drops, and is more efficient in terms of amount of rainfall on the ground compared with total condensation.

It was discovered, however, that evaporation plays a much larger part than water loading or drag due to entrainment in controlling the ultimate size of the cloud and in initiating and sustaining the sub-cloud downdraft. Rapid evaporation about the summit and upper periphery of the

cloud, which is usually associated with strong entrainment, produces a large temperature deficit, and the resulting negative buoyancy inhibits further growth of the cloud. Similarly, evaporation about the lower periphery and below cloud base cuts off the inflow of new moisture to sustain cloud growth. However, the upper and lower parts of the cloud are, to a large extent, decoupled dynamically, so the development of a strong sub-cloud downdraft has limited effect on the ultimate extent of cloud growth. When the outer parts of the cloud (where the air is not saturated) contain a larger proportion of large, slowly evaporating drops than of small, rapidly evaporating droplets, these effects are diminished, and the cloud can grow more vigorously. Interestingly, the total amount of water condensed in the vigorously growing clouds was not substantially greater than that in the weaker clouds, further emphasizing the importance of the rate of evaporation.

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#### The Dynamics of Convection in Shear

by J.S.A. Green and M.W. Moncrieff Department of Meteorology Imperial College, London

Considerations of the energy transformations in steady, two-dimensional, finite-amplitude convection in shear show that such systems have a propagation speed which depends on the shear, density scale-height and available potential energy of the large-scale flow. Comparison of the predicted speeds with the observed in a variety of convective systems including squall lines, severe cumulonimbus and cold fronts, show agreement even in systems that appear to be distinctly threedimensional. This model is therefore useful for predicting the displacement of such systems.

If the Richardson number is larger than a critical value, steady propagation is impossible and only transient systems are allowed. This has some bearing on the transition between cumulus and cumulonimbus convection.

The eddies have distinctive transfer properties which are also functions of the large scale flow field. Energy is transferred upwards but the law of transfer is essentially distinct from a diffusive process. Momentum is transferred in a sense that enhances the original shear and increases the kinetic energy of the mean flow - a feature likely to be important for the parameterisation of convection in synopticscale models.

Time dependent, three-dimensional models of convection, which include detailed parameterisation of cloud micro-physics

show similar properties tending rather rapidly towards the steady-state configuration.

The analysis suggests that observational studies of convection in the atmosphere do not necessarily demand detailed measurement on the sub-cloud-scale in order to define some important properties, at least if the Richardson number is less than the critical value for steady motion.





## THE GROWTH OF PRECIPITATION AT A COLD FRONT

# P. R. Jonas, Meteorological Office, Bracknell.

Observations have shown that heavy precipitation, often accompanied by small hail, can fall from cloud systems only a few kilometers deep. Extensive radar measurements of the air motion at a sharp cold front with cloud 3 km deep are available (Browning & Harrold, 1970) and these have been used as a dynamical basis on which to calculate the growth of the precipitation particles.

Small embryo precipitation particles were assumed to fall into the air flowing into the frontal zone from the extensive cloud ahead of the front. The motion and growth of these particles was then followed as they moved through the front and fell to the ground as precipitation. Accretion and sublimation were considered as the major growth mechanisms and the freezing and melting processes were included in the model.

The calculations showed that recycling of the precipitation particles in this shallow system was a very important factor in their growth. The observed small hail pellets could only have been formed if the particles underwent at least two ascents in the strong updraught at the frontal region. It is suggested that the recycling process may be sometimes as important for the development of heavy precipitation from shallow cloud systems as it is for the development of large hail in severe storms.

The model also demonstrated the production of a short burst of heavy rain immediately following the passage of the surface cold front. A major factor in determining the intensity and duration of this burst of precipitation being the concentrating effect of the air motion on the particle trajectories. The model also gave good agreement between the observed and calculated values of such parameters as the total precipitation, precipitation particle size and density and the maximum terminal velocity of the precipitation particle.

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# Author: B. Morton. Monash University, Australia. Turbulent entrainment into cumulus clouds

Turbulent mixing processes are of central importance in cumulus modelling but are at present poorly understood, and have been inferred from laboratory scale studies of turbulent plumes and thermals. In a paper at the International Conference on Weather Modification (Canberra, September 6-11, 1971; of which a copy is enclosed) we summarised the deficiencies of turbulent plume and thermal models as bases for cumulus modelling and indicated a new approach to entrainment into a growing cumulus cloud. This work on entrainment has been continued and has led me to look at some relevant aspects of high Rayleigh number buoyant turbulence.

I should first give a brief description of fresh experiments on plumes and thermals that demonstrate some of the limitations on the earlier ideas of self-similarity, and suggest a process of scale separation at large Rayleigh numbers in which the energy-containing scales of motion do not increase in size proportionately with the convective entity (plume or thermal). Supporting evidence is available from fire studies. There are implications for high Rayleigh number turbulent mixing.

I should then the very brief account in our earlier paper on vortex entrainment, and seek to bring out the essential difference between entrainment into a plume which is a steadyin-the-mean process and entrainment into a thermal which is an essentially unsteady process caused by time-development of the structure of the thermal and associated with changes in its pattern of mean vorticity. My aim would be to concentrate attention on the physical ideas, introducing no more equations than strictly necessary, as I regard the concept of unsteady entrainment as an important point for discussion amongst those working in all aspects of cloud physics. In particular, this approach to entrainment directs attention away from the knobbly exterior of the thermal (or cloud) towards the interior diffusion processes; the surface mixing is shown to be only one part of a more general process of turbulent diffusion, and the region exerting strongest control over the motion and growth of the thermal lies in that part of the interior where the gradients of mean vorticity are greatest. Diffusion in this region of the thermal is due to turbulence with typical length scales appreciably smaller than the element itself, and an improved understanding of high Rayleigh number turbulence is now important. We are thus led to a relatively more complica-ted entrainment mechanism, but one that allows us to explore the role of entrainment of air displaced from above the growing cloud as well as that entrained laterally; and I should finish with a brief comment on the relative importance of lateral and overhead entrainment in cloud models.

My approach is intended to emphasise the fluid dynamics of cumulus modelling rather than the construction of formidable (often numerical) models, as I regard our current knowledge of cumulus dynamics as quite inadequate to support either the weight of the microphysics or the speed and ease of large computer models.

# Hail Growth by Stochastic Collection in a Cumulus Model

by

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A one-dimensional, time dependent numerical model of a cumulus cloud is presented that generates hail and radar reflectivities at realistic rates. The distribution of hydrometeors evolves with time and varies with height at a given time as a result of condensation, sublimation, stochastic collection, sedimentation, drop freezing, and drop breakup. A total of 40 mass categories, each twice the mass of the former, corresponding to radii from 2.5 µm to 2 cm, are used to determine the ice and hail distribution. The first 31 categories up to 2.5 mm radius are used for the water drop distribution. Radar reflectivities are computed from Mie scattering theory for water and ice spheres in each category, then summed to give the reflectivities that can be compared to those observed by radar. Only the updraft radius at the earth's surface, the mixing coefficients and the initial droplet distribution at cloud base are arbitrarily specified.

Results for a June and August storm simulation are presented in a time-lapse colored movie. Included in the movie are the heighttime evolution of the hydrometeors' mass mixing ratio and radar reflectivity factor in each category and in total, and the number densities of water drops and ice spheres at the changing level of maximum radar reflectivity.

#### A PRECIPITATION MANAGEMENT SYSTEM BASED ON COMPUTER ASSISTED APPLICATION OF CLOUD PHYSICS

Clement J. Todd Division of Atmospheric Water Resources Management Bureau of Reclamation U.S. Department of the Interior

As part of an overall program of managing the hydrologic cycle which includes large-scale water storage, diversion, and delivery systems particularly in water deficient regions of the United States, the Bureau of Reclamation is involved in developing an operational precipitation management program. A systematic approach has been applied in developing a cloud seeding hypothesis which relies heavily on a computer-assisted assessment of cloud microphysical and dynamical processes. This approach consists of (1) developing simple cloud dynamic/cloud physics models, (2) verifying the validity of the models in real cloud experiments involving both treated and untreated clouds, (3) using the models on historical data to develop climatological frequencies and amplitudes of expected effects, (4) designing a basic framework to define the general domains under which the various modes of cloud seeding are more effective than the natural cloud precipitation mechanism and to delineate the relative effectiveness between the different seeding techniques, and (5) applying the models in real-time decisionmaking using remote time-share computer terminals at the field project sites. The models predict when the desired effects are best achieved by treating to stimulate cloud growth, or when the treatment should be to enhance

precipitation development through ice-phase or hygroscopic seeding.

The data base used as an input in the real-time models is provided by the National Meteorological Center, Suitland, Maryland, via a computer-to-computer communications link. It consists of radiosonde data and 12- and 24-hour forecast gridded fields depicting the vertical structure of the atmosphere. Also, input to this system is a surface heating function. The output is the time-height structure of the atmosphere for a specified target area throughout the day. A convective cloud model is run for different cloud widths upon this structure to determine (1) when to expect significant convection; (2) the cloud-base height; (3) the updraft profile, liquid water content profile, and cloud-top height for a natural cloud and one with enhanced buoyancy from ice-phase seeding; and (4) the precipitation amount and efficiency for the natural and stimulated cloud. A hygroscopic precipitation mode is chained on the predicted cloud parameters (updraft profile, liquid water content profile, etc.) to determine whether this form of seeding can be used effectively in increasing the precipitation efficiency. The convection model is one-dimensional and timedependent, with compensating subsidence which suppresses convection when the atmosphere has been stabilized. This model is designed to predict the amount and efficiency of natural precipitation, as well as the effect of a number of management options.

Using the model in decisionmaking assures a quantitative application of accumulated experience. The model is designed to easily accommodate any changes considered to be better approximations to real atmospheric processes. Thus, the incorporation of new knowledge into decisionmaking can progress at a rapid rate.

# SESSION NINE

# ELECTRIFICATION OF HYDROMETEORS

# THE ELECTRIFICATION OF HYDROMETEORS

By B.J. Mason Meteorological Office, Bracknell

This lecture presents a critical review of recent experimental and theoretical studies of several different mechanisms by which electric charge may be generated and separated within clouds. Laboratory experiments show that the freezing of supercooled water drops, the melting of snowflakes and hailstones, the disintegration of large raindrops, and collisions between ice crystals, water droplets and hail pellets are all likely to be accompanied by significant charging, and an attempt is made to assess their relative importance in the electfification of the thunderstorm.

At a more fundamental level, these phenomena involve a number of very different mechanisms of charge transfer, e.g. by ion segregation at the ice-water interface during freezing, by proton migration in ice down a temperature gradient, by conduction of charge between ice crystals and water drops colliding with and rebounding from hail pellets, and by the shearing of the electrical double-layer at the surfaces of air bubbles bursting in water and of water drops bursting in air.

The polarity and magnitude of charges and electric fields in thunderstorms are accounted for by a mechanism in which a small fraction of the cloud droplets impinging on the undersurfaces of polarized hail pellets rebound and carry away some of the induced positive charge and leave the hail pellet with a net negative charge.

# CALCULATIONS OF THE ELECTRICAL DEVELOPMENT OF

## CONVECTIVE CLOUDS

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In collaboration with the British Meteorological Office airborne studies are currently being conducted of the coincident dynamical electrical and microphysical properties of convective clouds. It is hoped that analysis of the records of electric field strength, hydrometeor charge, temperature, cloud particle size, type and concentration and other parameters obtained using instruments located on an airplane and on drop-sondes, together with radar information, will expose the mechanism or mechanisms responsible for the early electrical development of the clouds. In an effort to ease this formidable task of interpretation numerical models have peen constructed from can be determined which the charge and field distribution produced within the cloud as a function of time by various alternative charging mechanisms as a function of its meteorological characteristics such as updraught velocity, precipitation intensity, cloud water content (liquid and solid) and the numbers, types and sizes of the cloud particles. In addition to providing a possible technique for eliciting, by comparison with the measured records, the dominant mechanisms involved inconvective cloud electrification the model provides a much more realistic description of the electrical development than former computations of field growth, which have taken no account of the dynamical structure of the clouds and have been essentially one-

dimensional, dealing only with average values of parameters which often are extremely sensitive to altitude, such as ionic conductivity, particle concentrations, and existing field strength, which must be divergent in the presence of charged particles. The model is being developed steadily through successive stages of increasing complexity and realism and at present can predict the time variation of the vertical distributions of charge and field strength resulting from selected precipitative charging mechanisms as a function of the instantaneous magnitudes of the following parameters; updraught velocity, existing field strength, cloud particle characteristics and ionic conductivity (which are altitudedependent); and precipitation intensity, increasing with time according to established theory. The feedback process between field strength and charging rate can be taken account of and it is hoped that current laboratory experiments will yield values of separation probabilities as a function of particle sizes and field-strengths, which can be utilised in these computations. This work will also permit the sensitivity of the electrical development to the values of particular parameters to be assessed; this may assist the design of future experiments. Ultimately it is planned to incorporate inclined updraughts and threedimensional distributions into the model. A typical example of the calculations will be described.

Let us envisage an inductive charging mechanism in which a fraction, B, of the naturally occurring collisions between cloud particles, radius r, and precipitation particles, radius R, results in a transfer of charge  $Ar^2$ , where A and B are independent of field. The precipitation particles are produced at a given height and grow by

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collecting cloud water as they fall, in the absence of an updraught, with radius-dependent velocity V. If  $R_m$  is the size of the largest precipitation particles at a time t then it can be shown that the charge-density on precipitation at a level Z where their radius is R is approximately proportional to  $R^2$ , whereas the cloud-charge density is proportional to  $R^2 ln(R_m/R)$ . These differing functions lead to a net charge density and integrations over Z gives an analytic expression for the vertical field distribution field. The introduction of an updraught leads to increased maximum value of field because the larger more highly charged precipitation particles are contained within a smaller range of height. Extending the model using experimentally determined values of the field-dependent parameters A and B requires simple numerical integration.

# CHARGE TRANSFER ASSOCIATED WITH DROP-DROP AND DROP-DROPLET

INTERACTIONS IN POLARIZING ELECTRIC FIELDS

S G Jennings

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Studies have been made of the charge transfer accompanying the collision and separation of water drops falling in an electric field. The charge transfer q was measured for values of field strength E, impact velocity V, drop radius r, radius ratio R/r and angle  $\theta$  ranging from 5 to 800 V.cm<sup>-1</sup>, 50 to 300 cm.sec<sup>-1</sup>, 200 to  $600\mu$ m, 1.0 to 3.0 and 0 to 90 degrees respectively, where  $\theta$  is the angle between the field and the line of centres of the drops at the moment of separation.

The uniformly sized water drops were ejected from hypodermic needles by modulating the flow of water through them, and then collided in the volume bounded by two plane electrodes, across which an electric potential existed. The drops coalesced temporarily, swung around each other and separated, each resulting stream being collected in a vessel connected to an electrometer in order to measure the transferred charge. The measured values of q were generally in good agreement with theoretical values derived from the equation  $q=\gamma \Xi r^2 Cose$ , where  $\gamma$  is a function of R/r, as computed by Latham and Mason (1962) for a pair of uncharged conducting spheres of radius R and r (R>r) which come into contact and separate in a polarizing electric field of strength E.

Though calculations indicate that collisions between drops of precipitation dimension falling within a cloud can separate appreciable quantities of electric charge, the process envisaged, in which a larger raindrop overtakes a 106

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smaller one, coalesces with it temporarily, swings around it, and separates from beneath it, will act to dissipate existing electric fields. It is possible that the absence of strong fields in all-water clouds may be a consequence of this dissipative process. The interaction of polarised raindrops below cloud base may also be responsible for the bipolar charge distributions measured at the ground by Smith (1955).

The more usual form of the general inductive process of cloud electrification, namely the transfer of charge resulting from the interaction of drops of precipitation size with cloud droplets, is currently being investigated. Measurements have been made of the charge transfer (and hence the separation probability) between water drops ranging in radii from 400 to  $800\mu$ m falling through droplet clouds, produced by a spinning disk device, in the presence of polarizing electric fields. The tentative conclusion to be drawn from the preliminary experiments conducted to date is that the separation probability possesses values between 0.1 and 0.01% for polarizing field strengths up to about 400 V.cm<sup>-1</sup> and becomes significantly reduced as the field increases above this value.

#### References

Latham J and Mason B J 1962 Proc.Roy.Soc.A<u>266</u>, 387-401 Smith L G 1955 Q.J.Roy.Met.Soc., <u>31</u>, 23-47

# The Hydrodynamic Instability of Raindrops and the Initiation

of Lightning -- a Review

by: G. A. Dawson, University of Arizona, U.S.A.

It has long been considered that the interaction of falling raindrops with regions of high electric field in the base of thunderclouds is responsible for the initiation of lightning strokes. As a result there have been many experimental and theoretical studies of such interactions. All the theoretical treatments have been forced to make simplifying approximations, some of them gross. All the experimental studies of water drop/electric field interactions have been performed under unrealistic conditions, e.g. pressures of one atmosphere, low humidity, room temperature, and usually also at velocities less than terminal (although the drop shape is inextricably involved in instability, and the aerodynamics of fall greatly affect the drop shape.) Until recently, discussions have also been restricted to single drops, whereas under realistic conditions, the interaction between the field and a whole population of drops is the process of importance. The differences between the behaviour of a single drop and a population can be fundamental. The single drop responds to electric stress via a dissipative process to reduce the applied field. A population of drops, can have a regenerative effect, concentrating the electric field to initiate the lightning stroke.

Nevertheless there are still serious discrepancies between the values of electric field needed to produce drop instability, (a minimum of 9 kV cm<sup>-1</sup>) and those that have been measured in thunderclouds, (a maximum of about 4 kV cm<sup>-1</sup>).

All previous experiments on drop instability have used either vertical or horizontal fields. However, a consideration of the factors contributing to instability, i.e., hydrostatic pressure, aerodynamic pressures, surface and electric stress, indicate that instability onset should occur much more easily at field angles other than the vertical or horizontal. Experiments have now been performed in a vertical wind tunnel with variable field angles of about 45°, using simulated large raindrops of about 1 mm radius falling at their terminal velocity. The results have substantially agreed with the analysis as regards instability onset, though have shown some important new subsequent behaviour. At oblique angles, the instability onset is appreciably lower than for vertical fields though still far greater than the measured cloud field values. Experiments are planned for a low pressure high humidity environment to more completely model the cloud conditions.

A résumé will be given of the conclusions of the various experiments and experimenters, and their application to the thundercloud base.

# QUESTION ON THUNDERSTORM ELECTRICITY GENERATION

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1. Attraction of the electrokinetic phenomenon discovered by Ribeiro, Workman and Reynolds seems to prove its value in explanations of the thunderstorm electricity nature.

Detailed experimental researches of the phenomenon are made and the theory is developed. A strong dependence of the phenomenon intensity on concentration of some admixtures in water samples is noted.

2. The considered process of thunderstorm electricity generation is simulated within a flow of supercooled aerosol. Results of the experiments show a possibility of effective controlling of the electrization intensity.

3. A plausible hypothesis of electric structure of a thunderstorm cloud must correspond its thermodynamical structure. A jet model of cloud convection is used for calculation of vertical profiles of thermodynamical characteristics of a cloud, vertical velocity field, and trajectories of cloud and precipitation particles. S.R. Shewchuk and J.V. Iribarne University of Toronto, Toronto, Canada

Droplets of pure water or solutions in the range of 35 to 150  $\mu$  diameter were produced at the top of a small tunnel by a vibrating needle and accelerated in an air stream to varying velocities, at temperatures ranging from -6 to  $-14^{\circ}$ C (Fig. 1). A certain amount of coalescence occurred before leaving the tunnel; e.g. an initial stream of 95  $\mu$  droplets had an average diameter of 105-110  $\mu$  at the exit. The original charge on the droplets was reduced by adjusting the potential of an inductive ring close to the tip of the needle; the residual charge and the charges obtained during riming were measured alternately by collecting the droplets in a cup or letting them impact on an ice covered target, both connected to a vibrating reed electrometer. The charge separation was obtained by difference. The experimental conditions corresponded to accretion in clouds with liquid water contents between 0.5 and 5 g/m<sup>3</sup>.

Fig. 2 shows typical results for the charge separation per droplet Q as a function of impact velocity, in the absence of an electric field. Appreciable charging (>  $3 \times 10^{-7}$  e.s.u./droplet) occurred over a velocity threshold of 10-12 m/s for  $\sim$  100  $\mu$  diameter between -6 and  $-10^{\circ}$ C. This threshold increased to about 20 m/s or more for 50  $\mu$  droplets. Magnesium oxide slides located close to the target showed evidence of splashing which increased with impact velocity; no traces of splinters could be detected. The influence of an external field E is indicated in Fig. 3. For 95  $\mu$  droplets,  $-10^{\circ}$ C and 8 m/s, dQ/dE  $\cong$  2x10<sup>-6</sup> in e.s.u. was obtained both for pure water and solutions. The slope was reduced about 5 times for  $\sim$  68  $\mu$  droplets and no detectable charge was

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obtained for  $\sim$  35  $\mu$ . Impact velocity increased the slope: at 18 m/s it was  $\sim$  5 times larger than at 6 m/s (pure water).

Regarding thunderstorm electrification, ice particles falling in a cloud at less than 10 m/s in the absence of appreciable fields will not produce any significant contribution. Above that threshold the effect will depend critically in sign and magnitude on the liquid composition and on the presence of enough droplets in the 100  $\mu$  range. For typical salt contents and velocities  $\geq$  12 m/s negative charges in the order required to explain thunderstorm fields can be expected. As a field develops, an additional charge separation by induction becomes significant, whose sign will tend to further increase the field. The effect of particle charge saturation has not been studied yet.

Fig. 1 - Tunnel. B: tight box; Th: thermoregulator; HW: heating wires; C: conductivity cell; D: loudspeaker driver; V: vibrating needle; A: air inlet; S: screen; W: steel wool; T: brass drift tube; H: honeycomb; AT: acceleration tube, inside diameter 1 cm; E: 6° conical exit; L: metallic liner; Ta: target; Cu: cup; El: to electrometer.

Fig. 2 - Influence of chemical composition and impact velocity on charge separation. Temperature: -10°C. Average drop diameter: 100-105  $\mu$ . x: 10<sup>-4</sup> N (NH<sub>4</sub>)<sub>2</sub>SO<sub>4</sub>; 0: pure water (containing  $\sim$  0.7x10<sup>-5</sup> M CO<sub>2</sub>);  $\nabla$ : 2x10<sup>-5</sup> N NaCl;  $\Delta$ : 10<sup>-4</sup> N NaCl;  $\Box$ : 10<sup>-3</sup> N NaCl.





Fig. 3 - Influence of electric field on charge separation. Temperature:  $-10^{\circ}$ C. Impact velocity: 8 m/s. •: average drop diameter 105  $\mu$ , pure water and solutions; o: average drop diameter 68  $\mu,~10^{-4}$  N NaCl.

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## A MODEL FOR THUNDERSTORM ELECTRIFICATION BY POLARIZATION CHARGING

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Recent experiments by Scott and Levin (1970) show that charge transfer in collisions between polarized ice particles can, under some conditions, be very efficient. In the following model we assume that polarization charge transfer occurs only in collisions between ice crystals and graupel or hail particles, and that there is some (as yet unspecified) temperature range where polarization charge transfer is efficient. Thus, the region in the cloud where charge transfer occurs has upper and lower boundaries which are determined by temperature and/or the coexistence of ice crystals and graupel or hail particles. For polarization charging the existence of upper and lower boundaries implies that only over some limited range of updraft velocities the generation of electric fields can reach maximum efficiency. The latter occurs when updraft is such as to maintain the separated positive and negative charge centers above and below the region where charge transfer can take place. Given the observed spatial and temporal variations of updraft velocities in thunderstorms, complex electric field structures are to be expected.

Our present mathematical treatment is restricted to regions where updraft velocities allow electric field buildup to take place at maximum efficiency. We compute the growth of electric fields due to charge transfer in successive ice particle interactions, taking into account the effects of electrical forces on gravitational charge separation. Our model applies to cloud conditions where the number of ice crystals is large compared to the number of graupel particles. At present, only two ice particle sizes are considered and both particles are assumed to

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be spherical. Expressions are derived for charges carried by ice particles experiencing successive collisions in a spatially limited exponentially rising electric field. Neglecting charge inhomogeneities near the boundaries of the region where the electric field buildup takes place, the growth rate of the field is approximated by the parallel plate equation.

The growth rate equation is solved numerically for a number of different cloud conditions. The computed field rises exponentially until the electrical forces become strong enough to slow down the process of gravitational charge separation. When gravitational charge separation just counterbalances the charge loss due to various leakage currents, a steady maximum field is maintained. The computations give reasonable field growth rates for graupel particle concentrations above  $2 \text{ gm/m}^3$ . When graupel particles are 2 mm in radius or larger maximum electric fields can be reached that are higher than fields so far observed in thunderstorms. For nonspherical ice particles we expect the field growth rates to be shorter and maximum fields to be lower than the values obtained in our computations.

As the gravitational charge separation slows down ice particles carrying charges of both signs begin to accumulate in the field. If through a lightning stroke the electric field strength is suddenly decreased, all the accumulated ice particles are released, and a gush of rain or hail may result.

#### Reference:

Scott, W. D., and Zev Levin, 1970: "The effect of potential gradient on the charge separation during interactions of snow crystals with an ice sphere," J. Atmos. Sci., Vol. 27, May 463-473.

# SESSION TEN

# CLOUD PHYSICS PROBLEMS OF GARP

## AIRCRAFT INFRARED OBSERVATIONS OF THE EFFECTS OF CLOUDS ON RADIATIVE TRANSFER

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For many years the radiative properties of clouds have posed a problem when attempts have been made to include the effects of these properties on radiative exchange through clouds. Perhaps one of the most successful methods of attempting to analyze cloud properties has been that of direct (in situ) observations on the clouds themselves. Even prior to 1940 aircraft observations were made on at least the solar spectral region of lower cloud reflectivity. It is obvious that radiative properties of higher clouds, to perhaps 45,000 feet, were out of the question in the late 30's and early 1940's. However by 1946 measurements by the UK Meteorological Office to 40,000 feet provided excellent data on the Infrared emissivity of cirrus clouds. These direct observations certainly equal in importance to the early success in 1940 in the U.S. in determining cloud solar reflectivities of the tops of stratus clouds.

Rapid advances in cooled and other infrared detectors that followed the advent of the meteorological satellites permitted observations that, for the first time, could provide data of sufficient accuracy to aid radiative transfer calculations. Here one is saying that atmospheric infrared observations may well be within the same order of magnitude in accuracy as previous laboratory transmissivity measurements. In any event we now had the cloud penetrating platforms, such as NASA Ames Research Center's superb Convair 990 jet flying laboratory and the detectors. Detectors in radiometric systems that are flyable can readily reach minimum detectable radiances of  $1.0x10^{-7}$  w cm<sup>-2</sup> sr<sup>-1</sup> over narrow spectral intervals. In this connection a word should be said on the continuing rapid progress in the interference filter area. Taken in all, by 1970 airborne infrared

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and solar radiometric and spectrometric capabilities had reached a level of accuracy that clearly interested theoretical researchers in radiative transfer approximations.

Recent success in the U.K. and the U.S. in the combination of radiometric and lidar techniques based both on the ground and airborne is a very important advance in reaching or approaching a solution to the problem of radiative transfer through clouds. Not only cloud transmissions versus optical and geometrical cloud thickness have been analyzed but direct relations between transmissions and particle size and distributions have been studied. It is obvious that the combined technique provides information that cannot be obtained by either technique alone. Several additional U.S. airborne lidarradiometer surveys are planned.

As a result of the observations it is now possible to parameterize cloud radiation transfer with sufficient accuracy to reduce the computed versus observed discrepancy to + 15 percent. More advanced multiple scattering and transmission techniques underway in the UK and US for satellite and other applications will reproduce radiometric transfer through clouds to an accuracy equal to or better than + 10 percent. Carefully executed step climb with airborne radiometers is providing much usable data for radiative transfer in clouds. The UK lidar radiometric technique shows that cirrus emissivity varies more directly with cirrus particle number density than with cirrus thickness assuming that the particle size distribution did not vary significantly. Recent results in the US combining airborne lidar and radiometer observations through cirrus clouds also indicates a close correlation between lidar backscatter and the infrared cloud transmissivity. These results suggest a method for simplified modelling of the influence of clouds on infrared radiative transfer.
## RADIATIVE PROPERTIES OF TERRESTRIAL WATER AND ICE CLOUDS

AT NEAR INFRARED THERMAL WINDOW WAVELENGTHS

by G. E. Hunt Meteorological Office, Bracknell, England

There is great interest at the present time in the deviation on a global scale of the vertical structure of atmospheric temperature and composition from measurements of the radiation emitted from a planetary atmosphere in the infrared absorption bands of the gaseous constituents. The most serious source of inaccuracy in the interpretation of such measurements arises from the effects of clouds. This is a direct consequence of our present limited information on the microphysical properties of clouds and of our understanding of the manner in which a cloud interacts with a radiation field. Consequently, it is of fundamental importance that cloud physics observations are made simultaneously with radiometric observations of cloud layers.

In the context of a satellite experiment the most suitable regions of the infrared portion of the spectrum to study the radiative properties of clouds are atmospheric window wavelengths, where there is negligible absorption from atmospheric gases. The emitted radiation at these wavelengths will then depend upon the properties of the cloud layer and the surface of the planet.

We present the results of an extensive study of the radiative properties of terrestrial water droplet and ice clouds at the thermal infrared wavelengths of 2.3, 3.5 and 3.8µ and selected wavelengths in the range of 8.5 - 13µ. The theoretical study is centered around atmospheric models in which all the physical processes of radiative transfer are accurately incorporated. Computational techniques of great precision are used to study the multiple scattering of radiation by the cloud particles. The microphysical properties of water droplet clouds have been represented by characteristic distributions for cumulus clouds and the broader distribution for cumulonimbus. Since there is <u>no</u> available information on the microphysical properties of ice clouds we have selected a range of physically realistic distributions for this study. Our extensive studies show that water droplet and ice clouds do not behave like blackbody radiators at thermal window wavelengths. The emissivities of ice clouds are less than the corresponding value for water droplet clouds as a consequence of their lower water content. Indeed, The maximum emissivity of a thick ice cloud of liquid water content of  $5 \times 10^{-3}$  g.m<sup>-3</sup> is ~ 70% while the corresponding value for a water droplet cloud of water content  $6.3 \times 10^{-2}$  g.m<sup>-3</sup> is ~ 99% in the 8 - 13µ interval. In this region, the emergent radiation is dominated by that emitted from the cloud layer. There is, however, a small contribution to the emergent fields for radiation reflected by the cloud top. The radiative properties of the cloud vary slowly throughout with wavelength monitoring the small variation of the refractive indices of water and ice over this spectral interval.

At the near infrared wavelengths the emissivity of a cloud layer is considerably less than its value in the 8 - 13 $\mu$  region, since there is a large contribution to the emergent field of radiation reflected at the cloud top. Indeed at 2.3 $\mu$ , a thick water droplet cloud layer may reflect ~ 49% and emit ~ 50% of the emergent energy. Ice clouds behave similarly, although their reflectivity is smaller than water droplet clouds since they contain larger particles.

This investigation forms the first study of the radiative properties of terrestrial clouds at the near infrared thermal wavelengths and the most detailed comparison yet of water droplet and ice clouds in the thermal infrared. Detailed discussions of the theoretical results and comparisons with available observational data will be given.

#### A NUMERICAL SIMULATION OF WARM CUMULUS CLOUDS

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A time-dependent, Eulerian model of the life cycle of warm cumulus clouds is presented that combines the vertical equation of motion, the equation of mass continuity, the first law of thermodynamics, and the equations of continuity of water vapor, cloud droplets, and raindrops. The dynamic interaction between the cloud and its environment is modeled by two entrainment terms: turbulent entrainment representing lateral mixing at the side boundary of the cloud and dynamic entrainment representing the systematic inflow or outflow of air required to satisfy two-dimensional mass continuity. The model can, therefore, be regarded as having "one and a half" space dimensions. The unique feature of this model that makes it conceptually more realistic than previous cumulus models is the manner in which the microphysical processes in the cloud are treated. The formation and growth of droplets by condensation and stochastic coalescence are modeled in detail for 68 logarithmically-spaced Eulerian size classes covering a range of particle sizes from 2 to 4000 microns in radius.

With the specification of an atmospheric sounding, convection is initiated by invoking a temperature and/or humidity pulse in the surface boundary layer. The model predicts the height of the condensation level and generates the droplet spectrum at cloud base on a representative cloud nuclei spectrum. The dynamical and microphysical processes of the model

determine the further development of the cloud, predicting its ultimate height and the distribution of cloud droplets, raindrops, vertical velocity, temperature, and relative humidity as functions of time and space. The spectrum of droplets at each level in the cloud and sub-cloud layer changes with time under the combined action of convection, sedimentation, vertical diffusion, turbulent entrainment, dynamic entrainment, condensation-evaporation, and stochastic coalescence.

Model calculations are being made to investigate the effect that the interaction between the dynamical and detailed microphysical processes of a warm cumulus cloud has on its development and precipitation history. To accomplish this objective the fundamental parameters of the model such as cloud nuclei spectrum, stochastic collection kernel, cloud radius, vertical mixing rate, and the entrainment rates are being systematically varied over an observationally and/or theoretically reasonable range of values. Calculations with the model to date have resulted in the prediction of cloud base heights, cloud top growth rates, cloud depths, and profiles of temperature, supersaturation, liquid water content, and vertical velocity that are consistent with the observed values of these parameters for tropical cumuli as reported in the literature. The initiation and development of warm rain also appears to be in good agreement with the reported observations. It is shown that neglect of the dynamic entrainment term results in unreasonable values of the characteristic properties of warm cumulus clouds. Details of the results of extensive computations with the model will be presented.

# A CLOUD FIELD MODEL FOR SIMULATING THE DEVELOPMENT OF THE CLOUD LAYER IN THE TROPICAL ATMOSPHERE

by

#### Joseph Levine

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'Bubble' and 'jet' models in which the form and scale are parameterized are considered a valid alternative to less explicit spectral models in which transfer up or down the scale is parameterized. These scale models have already been used extensively to predict the gross characteristics of individual clouds in a given observed static clear environment regarded as infinite. The modelling of a cloud field by a distribution of 'bubble' and 'jet' elements has been proposed to predict the development of the temperature and moisture structure of the cloud layer in response to a given heat and water vapor flux through the sub-cloud layer and a quasi uniform large-scale subsidence in the troposphere. The horizontal momentum balance of the cloud layer may be similarly treated in the largescale horizontal equation of motion by introducing the exchange of horizontal momentum into the 'bubble' and 'jet' models. However, only the thermodynamic development of the cloud layer is treated here.

Individual clouds can be forced by inhomogeneities of the underlying heated surface or by the most energetic scale of turbulent fluctuations in the subcloud layer. This scale is considered to be roughly comparable to the subcloud layer depth. These irregular cells are regarded as the sources of randomly distributed momentum pulses at cloud base which act as sources of tubbles' roughly 100 to 200m in radius. These rise only a short distance above cloud base into the conditionally unstable air above and come to rest 121

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after losing their buoyancy. Since the cloud layer is stable for dry convection, the remains of these 'bubbles' which have come to rest persist as passive saturated diffusing cloud puffs.

As the heat flux increases the frequency of bubbles per unit 'area increases, and therefore the probability of puff superposition increases. This leads to the development of the dry cloud layer environment. The early stage of cloud layer development is designated the cumulus humulis stage.

As the cumulus humulis puffs increase in size the probability of weak larger-scale subcloud motions being superimposed at their bases increases. The proper coincidence of a weak perturbation with a large cloud puff base is regarded as the event necessary to cause the rapid growth of a quasi-steady state jet. This is designated as the cumulus congestus stage.

A numerical randomized model experiment for the development of the cumulus humulis stage is proposed. The 'static energy' conservation equation is written. Then a Reynolds transport equation is derived, which is valid for any type of cloud layer experiment.

### Numerical Experiments on the Initiation and Development of Cumulus Cloud Populations

### by G. E. Hill

### Pennsylvania State University, U.S.A.

In this study fields of cumulus convection are examined by integration of appropriate equations numerically. There is a three-fold purpose, namely, to prescribe realistic but simplified initial conditions, to determine the nature of cumulus growth, and to find out what is the natural size and spacing of cumuli at various stages of growth.

A two dimensional set of equations with compressibility retained are integrated using a modified Euler-backward scheme. To prevent the plumestem problem of Malkus and Witt, a modification of the finite difference scheme is introduced. Also, non-linear viscosity dependent upon both the deformation field and the local buoyancy is employed. Various test integrations are performed to ascertain the basic character of the model when a dry warm bubble is artificially introduced. The development of a classical convective circulation is shown. Prolonged integration results in the occurrence of an oscillation in the circulation direction at the Brunt-Väisälä frequency, and the propagation of a gravity wave-train outward from the original disturbance.

A complete moisture cycle is introduced so that convective disturbances in a conditionally unstable atmosphere may occur. The effect on cumulus growth of mixing both water vapor and cloud water with unsaturated air is compared with the case of no explicit mixing. The transfer relations between cloud water and hydrometeors are parameterized with at least two precipitation categories allowed. The resulting rainfall with two sizes is compared with the Marshall-Palmer distribution. The effect of evaporation from rainfall on the subsequent cloud population is also presented.

To simulate both initial and surface boundary conditions, a surface layer is established, wherein parameterized eddies transfer heat and moisture from the lower to the upper part of the surface layer. The surface heating is effectively that which would be found from solar radiation excess over surface emission, or from a heat reservoir such as the ocean underlying cooler air. Variations in surface heating, both in position and in time are allowed, so that plausible variations in the meteorological variables occur. The time span over which such randomized surface forcing occurs is tested to find out if the cumulus fields alone generate statistical time-space variations, such that the cumulus population evolves as is observed, for example, in time-lapse photography.

Several experiments are carried out with various representative initial distributions of temperature (inversion height, stability), relative humidity (depth of moist layer, absolute humidity) and surface heating (net radiation, oceanic heating).

Finally, a field of large-scale low-level convergence and upper-level divergence is introduced so that the cumulus convection portion of the CISK mechanism is simulated. Also, the intensity of the large-scale divergence is reduced to that appropriate to tidal motions to test whether cumulus activity is significantly raised beyond that without the effect.

## PAPERS TO BE READ BY TITLE

В

These papers have been placed approximately in groups appropriate to the various sessions. For instance, those papers which may be of interest mainly in the discussion during SESSION THREE are numbered B3.n . . . etc, and those for SESSION SIX B6.n . . . etc. Those not falling easily under the various session headings (e.g. papers on Weather Modification, Fog and Miscellaneous subjects) are listed at the end, and given paper reference numbers B.a, B.b . . . etc.

### Recent Cloud Microstructure Measurements with an Optical Device

Robert T. Ryan, Arthur D. Little, Inc.

Henry H. Blau, Jr., Environmental Research and Technology

### Cambridge, U.S.A.

We have recently described the operation of an optical instrument for counting and sizing cloud droplets (Ryan, Blau et. al. 1972). The instrument utilizes light scattering and operates as follows: the intersection of a laser light beam and photomultiplier field of view define a sampling volume in space which is swept along by the movement of an aircraft. Cloud particles passing through the volume scatter light from the laser beam into the detector. For cloud droplets the scattered light intensity is a monotonic function of size. The light pulses are counted and sorted in 12 geometrically spaced intervals in a pulse height analyzer. By knowing the cross sectional area of the sample volume and aircraft speed a cloud droplet size distribution can be found. Typically ~70 cm<sup>3</sup> of cloud is sampled per second and the diameter range of cloud droplets measured and sized is  $^{-2}$  3 -  $^{-70}$  µm. Because the scattering properties of ice crystals are not known, ice crystals are sized in terms of equivalent water droplet diameters.

During a field trip in January 1972, we measured the microstructure of a variety of cloud types including Pacific coastal stratus, cirrus, fresh contrails, tropical cumulus rainshowers and tropical stratiform clouds.

The environmental conditions of the Pacific coastal stratus were very similar to conditions observed during previous microstructure measurements of this cloud type (Ryan, Blau et. al. 1972). No dramatic changes in the cloud structure were seen during level flight over distances of the order of tens of kilometers. The vertical microstructure of the cloud did exhibit 125

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small changes with a shift in the distribution peak towards larger droplet sizes as a profile was made from cloud base to cloud top. Slight increases were seen in average liquid water content from cloud base to top.

A fresh contrail (~2 minutes old) at high altitude (~10 km) was measured and the apparent size distribution was observed to be dramatically different from any naturally occuring ice crystal cloud we have yet studied. The particle concentration was of the order of  $100/\text{cm}^3$  whereas in natural cirrus, crystal concentrations are seldom observed greater than  $10/\text{cm}^3$ . The apparent size distribution appears to be exponential and is narrow. The range of apparent size is ~3 - ~10 µm diameter. In natural cirrus the apparent size distribution is very broad and flat with particles larger than 50 µm quite common.

Measurements in tropical rain showers showed a typically very broad size distribution with frequent (~ $1/50 \text{ cm}^3$ ) particles larger than ~70m. In our pulse height analyzer all particles larger than ~ $70 \text{ }\mu\text{m}$  are counted in the highest channel. During these cloud penetrations we obtained a cloud droplet spectrum approximately every 100m.

Several altostratus (altitude  $\sim 5$  km) were penetrated over the Caribbean. These clouds were less than 100m. thick, the particle size range narrow and the concentration very low ( $\sim 10/cm^3$ ).

A detailed analysis of these preliminary results is in progress and will be presented in the near future.

#### References

Ryan, R.T., H.H. Blau, Jr., P.C. von Thuna, M. L. Cohen and G.D. Roberts, 1972: Cloud microstructure as determined by an optical cloud particle spectrometer. J. Appl. Meteor. 11, 149-156.

## PROBLEMS OF AIRPLANE DIVERGENCE MEASUREMENTS

## Heinz H. Grote

Atmospheric Physics and Chemistry Laboratory Environmental Research Laboratories National Oceanic and Atmospheric Administration Boulder, Colorado 80302

Two basic problem areas arise in the measurement of divergence from an airplane: a) Definition of the measurement as such, and b) accuracy of the wind determination.

In dealing with planar divergence one starts with the idealized mathematical model of a sharply bounded area of horizontal flow. Any measured value of the inor outflow in any closed horizontal loop divided by the area of the vertical flow provides then the constant value of divergence at that level. In practice this area is hard to determine. An additional problem arises from the finite time it takes to fly the loop and the motion of the storm incurred during this time. Generalization of the measurement at one level to other adjacent levels poses another problem. In measuring the wind from an airplane generally perfect conformity of the airplane motion with that of the surrounding air is assumed. Errors in measurement arise from the deviation of the airplane axis from the actual direction of motion and from errors made in the determination of the air- and ground speed and the drift angle and the heading.

The airspeed measurement is the largest contributor to errors in the wind determination, but errors in groundspeed and direction contribute considerable amounts. Doppler radars suffer from the difficulty of determining the center frequency in a band of return frequencies, especially at the edge of rain-systems; for inertial navigational systems the characteristic Schuler frequency of the erection mechanism is close to a quarter of the inverse loop time.

The influence of Gaussian-distributed and of bias errors are discussed.

Some examples of divergence flights around thunderstorms are given and investigated for their errors.

An important necessity in divergence determination is a near real-time computation. An on-board minicomputer system with immediate display of flight path, winds and radar returns provides the divergence at the instant the loop is closed. Such a system has been built and tested by the author and will be discussed briefly.

#### A CLOUD DETECTOR FOR USE WITH CONVENTIONAL RADIO SONDES

P. J. Cowley & P. Goldsmith Meteorological Office - Bracknell

A sensing system has been developed for detecting the presence of cloud drops of up to 100 microns radius when it is rising through the atmosphere at 1200 ft.min<sup>-1</sup>. Tests indicate that the system responds to both warm and supercooled droplets.

The detecting sensor comprises two parallel-wound coils of 0.002 in. diameter platinum wire on a  $\frac{5}{16}$  in. diameter cylindrical perspex former which has a two-start thread of 80 t.p.i. The cylinder is mounted in a square perspex frame in a housing made from 0.018 in. aluminium alloy as shown in Fig.1. The flap critically defines the maximum size of drop that may approach the sensor, and is so designed that rain drops are not detected. Absorbent paper is used to reduce splashing on the housing walls and to allow surplus collected water to be drained from the housing.

Calculations show that 2-200  $\mu$ g.sec<sup>-1</sup> of water may be collected, depending on the type of cloud. Any water collected on the sensor cylinder results in a change of resistance between the coils. The electronic circuit used to monitor this change in resistance is shown in Fig.2. Electrical heating of the sensor current from the  $67\frac{1}{2}$  v. battery evaporates water off the detector. If the detector resistance falls to less than 3 K ohm the zener diode D 1 maintains a high voltage across the coils irrespective of the current. This ensures a very rapid recovery when the detector leaves cloud. The maximum power dissipation is set by the capability of the  $67\frac{1}{2}$  v battery. The voltage V 1 varies with resistance between the coils. It is compared with voltages V<sub>2</sub> and V<sub>3</sub>. As it increases RL1 and RL2 are actuated in that order.

The system has been tested in use with Meteorological Office Mk 2B radiosondes. Contact closure of RS1 and RS2 connects capacitors into the sonde audio oscillator which lead to changes in the oscillator periodicity. The result of a typical ascent is shown in Fig.3. Level 2 indicates cloud of higher water content than level 1 which is set to detect clouds of very low liquid water content.

Results indicate that such a detection system with a single switch may be used on routine radiosonde ascents to indicate the heights of layers of cloud.



Fig 2: 2-level Circuit



Fig 3 Part of Tephigram of a Modified Mk 2B Sonde

THE DYNAMIC DIFFUSION CHAMBER FOR LABORATORY STUDIES OF CLOUD PROCESSES K. Gamara and J. Hallett, Laboratory of Atmospheric Physics, Desert Research Institute, Reno, Nevada

The extent to which laboratory studies of cloud particles are carried over to the atmosphere depends on the effectiveness with which laboratory design simulates the atmospheric process. Many experiments reproduce a minimum of atmospheric variables and whilst having the advantage of investigating some physical process have the disadvantage of not simulating fully the cloud process. Laboratory study of crystal growth rate from the vapour phase illustrates these limitations. Ice growth rate depends on the local supersaturation influenced by updraft and neighboring particles, by the air flow around a particular shape oriented in a particular direction to the flow field, by the temperature through kinetic effects at the crystal surface, and pressure through the diffusion coefficient. Several classes of experiments may be distinguished:

1. The fall of crystals through a supercooled water cloud with temperature as independent variable gives growth at water saturation and an ice supersaturation dependent only on temperature. The particle is ventilated at its terminal velocity.

 Growth on a substrate where temperature, air pressure, and supersaturation are independently controlled; the substrate however dominates the heat flow; ventilation is small and uncontrolled.
 The static vapour diffusion chamber with temperature, supersaturation and pressure as independent variables; ventilation by free convection.

The dynamic diffusion chamber simulates temperature, pressure, supersaturation and velocity; each quantity is varied independently. Conditioned air flows between two horizontal ice coated surfaces, with temperature and vapour pressure differences, see (A). There are two design criteria: (1) laminar flow (Re < 2000); (2) approach to thermal and diffusive equilibrium. For ice crystal growth studies minimum

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practical separation of plates is about 2cm. With air, this gives Re  $\approx$  10u (u - velocity cm sec<sup>-1</sup>), for a chamber length 1, separation d, the thermal equilibrium is approached in time  $\frac{1}{u} = \frac{d^2}{\pi^2 k} \approx 2$  sec. Plate temperatures of -1 to -25°C give working temperatures within the range -3 to -20°C and supersaturation 2% over ice to 30% over water, in the absence of nuclei. A fibre placed at the far end serves as a site for growth as in the static diffusion chamber. Dendrites grow at -15°C with an air speed of 8cm sec<sup>-1</sup> (B) growing into the wind direction; growth velocity varies with u<sup>1/2</sup>. Supersaturation can be monitored for short periods by observing interference of He-Ne laser from front and back of a suspended growing drop, to give moving interference fringes.

This system has the potential for investigating particulate behavior under conditions which closely simulate the atmosphere, and has been used qualitatively for: (a) growth by diffusion and accretion by cloud drops  $10\mu$  diameter; (b) scavenging of aerosol particles; (c) ice nucleation. Squires has used a dynamic chamber for monitoring the critical supersaturation for cloud nuclei at temperatures above  $0^{\circ}$ C; this system has the potential for quantitative studies at low temperature.

This work was supported by National Science Foundation Grant No. GA 27943.

B2.4

## CONTINUOUS CLOUD PHYSICS DATA OBTAINED IN UPDRAFT SHAFTS OF CONTINENTAL CUMULUS CLOUDS

J. Doyne Sartor Wim Toutenhoofd National Center for Atmospheric Research Boulder, Colorado

Unique cloud physics data were recently obtained with "The Explorer," an instrumented sailplane belonging to the National Oceanic and Atmospheric Administration and operated by the National Center for Atmospheric Research. The sailplane was centered in the updrafts of several cumulus congestus clouds while circling with a turn radius of  $\sim 250$  m.

The sailplane instrumentation includes an electrostatic disdrometer which measures every 0.5 sec a drop size distribution for  $\sim 1 \text{ cm}^3$  of sampled air, a Cannon cloud particle camera, an impactor slide gun and instruments for obtaining the vertical speed of the air and the ambient temperature. The instrumentation is discussed in more detail by Toutenhoofd, Dye and Sartor (1972).

A one-minute sample of disdrometer and vertical airspeed data obtained in a flight on 12 August 1971 is presented in Fig. 1. The sailplane was flown in the updraft through the base of a continental cumulus congestus to an altitude 1900 m above cloud base.

The high resolution of these data makes it possible to derive statistical parameters important for testing certain cloud models.

A summary of the data obtained in the cloud mentioned above is presented in Fig. 2, where  $\sigma_r$  indicates the drop radius standard deviation obtained from an 0.5 sec disdrometer sample. All data are averaged over 100 m intervals of pressure altitude and plotted in solid lines; standard deviations are plotted in dotted lines.

B2.5

<sup>\*</sup>The National Center for Atmospheric Research is sponsored by the National Science Foundation.

Auto and cross correlations and other statistical parameters have been derived from the data and will be discussed.

### References:

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Toutenhoofd, W., J. E. Dye, and J. D. Sartor, 1972: "An instrumented sailplane-A platform for cloud physics research with continuous rapid response measurements," submitted to Int. Conf. on Cloud Physics, London.

> VERT SPEED sec.

- Fig. 1 (right) Example of high resolution cumulus congestus data obtained with "The Explorer" sailplane.
- Fig. 2 (below) Summary of the data obtained in the updraft shaft of a continental cumulus congestus.

(km)

PRESSURE ALTITUDE



EXPLORER SAILPLANE

12AUG7

## AN INSTRUMENTED SAILPLANE - A PLATFORM FOR CLOUD PHYSICS RESEARCH WITH CONTINUOUS RAPID RESPONSE MEASUREMENTS

## Wim Toutenhoofd James E. Dye J. Doyne Sartor National Center for Atmospheric Research<sup>\*</sup> Boulder, Colorado

A Schweizer 2-32 sailplane, "The Explorer," has been instrumented as a platform for cloud physics research. The sailplane pilot can center his craft in the core of the updraft of cumulus clouds and in mountain wave clouds he can maneuver to a desired position in the wave. This gives the sailplane pilot the unique capability to investigate the microphysics, kinematics, and dynamics of clouds simultaneously in a steady state or quasi-Lagrangian sense. The relatively slow true air speed and choice of fast response instrumentation allow measurements to be made with a spatial resolution of 15 to 20 meters.

The scientific instrumentation mounted on "The Explorer" is summarized in Table I. The electrostatic disdrometer (Abbott, Dye and Sartor, 1972) measures and reads out the cloud droplet size distribution each half second during transit through the cloud. Ice crystal concentrations and droplets in the same volume are measured in situ every 1/2 sec using the Cannon particle camera (Cannon, 1970). The vertical speed of the sailplane is measured continuously and can be converted to vertical airspeed from a simple equation of motion for the aircraft. A diode thermometer mounted at the stagnation point on the nose of the sailplane gives continuous temperature measurements. The measured parameters and voice communications are transmitted to a mobile ground station where the information is received with a nine-channel telemetry receiver and is recorded on magnetic tape. The raw data can be viewed in real time on strip chart recorders or oscilloscope. The telemetered data is digitized

<sup>\*</sup>The National Center for Atmospheric Research is sponsored by the National Science Foundation.

from a playback of the magnetic tape and fed into the NCAR computer so that most of the analysis and reduction of the data can be accomplished with a minimum of effort a short time after local flights. Samples of data collected in summer cumulus and wave clouds will be presented to illustrate the unique capability for atmospheric research and the versatility of an instrumented sailplane.

Results obtained with the sailplane in the updrafts of continental cumuli are discussed elsewhere by Sartor and Toutenhoofd (1972).

Instrument	Range*	Accuracy*	Sampling Volume and/or Time Resolution
	For		2
Camera-in situ parti-	concentrations >3	3µ	5.0 cm <sub>3</sub> for $10\mu$ droplet,
cles, liquid & solid	For sizing <u>&gt;</u> 8µ	<u>+</u> 20%	130 cm for ice; 1/2 sec
Electrostatic cloud droplet disdrometer	4 to $22\mu^{1}$	<u>+</u> 10%	1.0 $\mathrm{cm}^3$ per 1/2 sec
Cloud droplet impac- tor slides	<u>&gt;</u> 2µ	<u>+</u> 15%	50 cm <sup>3</sup>
Variometer, vertical speed of sailplane	-40 to +40 m/sec	<u>+</u> 0.4 m/sec	<0.5 sec
Pressure altitude	1010 to 120 mb	<u>+</u> 10 mb	<l (0.5="" mb="" resolution<br="" sec="">obtained by integrating ver- tical speed from variometer)</l>
Temperature	-75 to +30°C	<u>+</u> 1.5°C	
Indicated airspeed	0-67 m/sec	<u>+</u> 4 m/sec	<0.5 sec
Vertical accelerometer-10 to +10 g $\pm 0.3$ g <0.5 sec			
To be installed by June 1, 1972:			
Lyman alpha humidiometer			
Polonium electric field meters			
* Particle sizes in radius			
<pre>1) All droplets with radii &gt;22µ are counted in one channel <u>References:</u> Cannon, T. W., 1970: "A camera for photographing airborne atmospheric particles," Image Technology, April/May.</pre>			
Sartor, J. D., and W. Toutenhoofd, 1972: "Continuous cloud physics data obtained in updraft shafts of continental cumulus clouds," submitted to Int. Conf. on Cloud Physics, London.			
Abbott, C. E., J. E. Dye and J. D. Sartor, 1972: "The electrostatic disdro- meter-An operational cloud droplet probe," submitted to Int. Conf. on			
Cloud Physics, Lo	ndon.		<b>R</b> 2 - 6

TABLE I

### RADIOMETRIC FREE AIR and IN-CLOUD

### TEMPERATURE MEASUREMENTS ABOARD NOAA AIRCRAFT

Bу

James D. McFadden NOAA Research Flight Facility Miami, Florida

and

Peter M. Kuhn NOAA Atmospheric Physics & Chemistry Labs Boulder, Colorado

In an attempt to evaluate the measurements from aircraft of free air and in-cloud temperatures radiometrically, a specially modified Barnes PRT-5 radiometer was installed on a NOAA Research Flight Facility DC-6 in November 1971. The sharp cut-on, cut-off filter used on this modified PRT-5 has a band pass of 14.62 to 16.15 microns at the half power points and has an average transmittance of 68 percent.

To date evaluation flights have been made in conjunction with the Atmospheric Physics & Chemistry Laboratory's weather modification program in the Great Lakes region around Buffalo, New York under varying mid-winter conditions, and comparisons of measurements obtained on these missions with other on-board temperature sensing devices, such as the Rosemount and vortex thermometers, have been made. This paper will present a discussion of the system and its installation in the aircraft, a comparison of the results from the above mentioned flights and the results of convective cloud temperature measurements to be made over South Florida in April and May 1972. An instrument for continuous measurement of the concentration of liquid water in fog.

J.A. Wisse, Royal Netherlands Meteorological Institute

An instrument (Fig. 1) is described for continuous recording, with a response time of about 10 sec., of the concentration of liquid water in fog. An impactor with an airflow of 0,1 m<sup>2</sup>/min samples fog droplets on a sensor. A constant power of 2,5 Watt gives the sensor an excess temperature of about 100°C if no fog is present. Evaporation of fog droplets cools the sensor. A cooling of 1°C corresponds to about 5.6 mgr water/m<sup>2</sup>. Dust and residues of evaporated fog droplets accumulate on the sensor. Therefore, the heat transfer from the sensor to the air might in some cases increase up to 10% after 12 to 18 hours of sampling. Assuming the increase of the heat transfer to be linear in time, the accuracy of the instrument is about 10% with a minimum of 10  $mgr/m^3$ . In order to minimize inaccuracies and to control proper operation of the device, a second identical impactor is placed in series with the first one. The difference between the excess temperature of the first and the second sensor is the measuring signal. For continuous measurements at several heights from meteorological towers isokinetic sampling is not feasible. Therefore the intake of the impactor is placed in the stagnation point of a sphere, with a diameter of 310 mm. In this way, the impactor samples from a decelerated air stream, in which the liquid water concentration is not significantly increased, because of the low catch efficiency of the sphere. In order to check this sampling method, two instruments with an air flow of 0.1 and 0.03 m<sup>3</sup>/min were placed 1.5 meters apart. The geometry of the instruments was the same, except for the diameter of the entrance. The entrance velocity was 100 m/sec. in both instruments. Measurements during 9 fogs showed that 204 five minutes averages, greater than 80  $mgr/m^3$  had a mean difference of  $6 + 14 \text{ mgr/m}^3$ . No relation between this difference and wind speed could be found. As an example Fig. 2a and b show the concentration of liquid water content measured at a height of 1.5 meters in an advection fog and in some low banks of radiation fog. In Fig. 2a the extinction coefficient measured by a nearby transmissometer with a baselength of 16 m is given. In Fig. 2a wind speed varied from 0.5 to 1.5 m/sec. In Fig. 2b wind speed was less than 0.5 m/sec., the extinction coefficient varied from 4 to 40 per km.





Fig. 2 concentration of liquid water  $\beta$  mgr/m<sup>3</sup> measured with a flow rate of 0.1 m<sup>-</sup>/min (+) and 0.03 m<sup>-</sup>/min (.); extinction coefficient  $\sigma$  km<sup>-1</sup>; d =  $\beta \beta / \sigma$  (predominant diameter of the liquid water content distribution).

**B2.**8

## F. PASQUALUCCI

## NATIONAL INSTITUTE FOR TELECOMMUNICATIONS RESEARCH of the COUNCIL FOR SCIENTIFIC AND INDUSTRIAL RESEARCH Johannesburg, Republic of South Africa.

A meteorological Doppler radar system of advanced design is described.

The radar operates in the continuous wave (CW) mode and uses separate antennas for the transmitter and the receiver (bistatic design).

To determine the target range and to resolve targets in range a digital pulse compression technique is used. A pseudo-random binary code modulates in phase the transmitted wave and the receiver computes the correlation function. When the reference code has a delay corresponding to the range of the target, the output of the correlation receiver is the peak of the autocorrelation function of the code.

The main advantages of such a radar system over a conventional pulse Doppler radar are:

- (1) Targets can be detected at very short ranges ( 15 metres).
- (2) The range resolution obtainable (15 metres) is many times better than that of a conventional radar.
- (3) The bistatic design allows the use of one transmitter and numerous receivers for multi-dimensional mapping of air circulation in clouds.

The block diagram of the radar system is given in Fig.1. The output of the radar is recorded and fed to a digital computer for real time processing. Using suitable interface circuitry a frequency resolution of 10 Hz in the Doppler spectrum was achieved. The radar is, at present, being used to study the variation of drop-size distribution in rainfall.



## 35 GHZ DOPPLER RADAR BLOCK DIAGRAM

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## THE PRODUCTION OF ICE SPLINTERS BY THE FREEZING OF ACCRETED WATER DROPS

### <u>D. A. Johnson - Meteorological Office, Bracknell</u>

#### 1. Introduction

Freezing water drops are able to fracture under certain conditions in the laboratory, and the fragments could be an important source of ice crystals if enough drops behaved in this way when accreted on an ice substrate in a cloud. The present work was designed to explore the conditions for drop fracture on accretion, since previous laboratory simulations of the process have shown conflicting results, some workers observing copious splinter production and electric charging and others none at all. A drop freezing on an ice substrate would normally freeze by conduction of most of latent heat into the substrate, only a small fraction of the heat being transferred to the air during freezing. Fracture is therefore unlikely unless the heat balance is redistributed in favour of the free surface of the drop.

### 2. Experiments

A stream of water drops  $20-100 \ \mu$ m diameter fell through a precooling unit into a coldroom where they froze on an ice substrate at an ambient temperature between -5 and  $-15^{\circ}$ C. The substrate could be ventilated at up to 10 m s<sup>-1</sup> and was connected to an electrometer and recorder. Many rime pellets were grown under a variety of conditions but no significant charge separation was observed above the background level equivalent to 1 event per 6 x  $10^4$  drops accreted. No ice orystals were observed when a dish of supercooled sugar solution was placed downwind of ventilated rime, when any particles exceeding 10 um diameter would have been detected. However, splinters were observed when the apparatus was filled with CO<sub>2</sub> gas, both visually as they were ejected and growing in sugar solution downwind, and were associated with changes in electric charge of both signs exceeding the noise level by a factor of 15.

The most favourable conditions for drop fracture are expected when drops freeze as spheres on top of a previously frozen drop or chain of drops to form an open, low density structure, instead of spreading on impact as they do at higher impact velocity and smaller supercooling. Even under the former conditions cine photography revealed that the expansion bulge on the frozen drop was opposite to the point of contact, showing that the heat was transferred mainly through the surface of the drop in contact with the ice. Calculations based on a simple model supported this observation, but showed that the proportion of heat transferred through the free surface of the drop could be doubled in an atmosphere of helium gas. Charging events and ejected splinters were observed at a maximum rate of 1 event per 2000 accreted drops when the apparatus was filled with helium.

## 3. Conclusion

Although ice splinters are ejected with significant electric charges when water drops freeze on an ice surface in carbon dioxide or helium, no evidence has been found for similar behaviour in air, and it is concluded that fewer than 1 drop in 6 x  $10^4$  will eject an ice splinter larger than 10  $\mu$ m diameter. The controlling factor is the heat transfer distribution around the drop, which is always heavily biased in favour of asymmetrical freezing from the ice substrate when drops freeze in air. The fracture of freezing of water drops is unlikely to contribute significantly to the ice crystal concentration in a cloud. J.C.H. van der Hage, Department of Meteorology, State University of Utrecht, Utrecht, The Netherlands

The classical nucleation theories are based upon the idea that the generation of a stable droplet is suppressed as long as the free energy of the vapor is not enough to supply the energy required for the liquid-vapor interface of the droplet. The influence of the released heat of condensation is neglected.

It will be argued here that this is not correct.

Condensation is a local convergence of vapor molecules to an extent that the molecules find themselves in each others attraction-spheres. By definition a vapor-liquid transition has then occurred. It is shown that the change in potential energy which accompanies this process is amply sufficient to provide the surface energy, no matter how small the droplet may be. In fact, the surface energy is that fraction of the total potential energy which is not released, while the heat of condensation represents the surplus which has to be dissipated.

The rate at which the heat can dissipate will decide whether nucleation will occur at a certain supersaturation.

Thomson's law on the critical drop radius stays valid because it describes a drop in equilibrium with the vapor during which there is no heat exchange. Therefore this law still determines the mass and temperature of a stable drop at the prevailing saturation rate, and therewith the amount of heat which must have been dissipated when such a drop is generated.

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Evidently heat sinks, e.g. particles in the vapor and even more so, macroscopic walls will promote condensation. It is demonstrated that for large condensation nuclei the temperature rise of the nucleating droplet becomes small, it may certainly be neglected for the case of nucleation on walls. This simplification allows the calculation of a minimum time of formation for a stable droplet on a wall at a given saturation ratio.

Finally a series of experiments on condensation on a macroscopic wall in a thermo diffusion chamber is described. The experimental results do not agree with the classical theory, but can be explained by the new theory. The predicted time of formation for a stable drop at various saturation rates corresponds roughly with the observed values.

B3.2

## GROWTH THEORY OF A POPULATION OF DROPLETS AND THE SUPERSATURATION IN CLOUDS

Norihiko Fukuta University of Denver, Colorado, U. S. A.

Growth of a population of droplets in which each droplet is surrounded by an impenetrable cell boundary has been studied for the case in which the temperature decreases linearly with time (as by adiabatic expansion). Explicit expressions for the temperature and vapor fields are obtained assuming Maxwellian conditions at the droplet surface, i.e., the temperature and vapor density at the surface are in equilibrium with the surface. The rate of supersaturation generation by cooling at each space point is assumed equal to the rate of supersaturation reduction by the growing droplet, so that a quasisteady-steady condition exists.

At the beginning of the growth process, the growing droplet modifies the temperature and vapor fields chiefly near the droplet surface. The influence of the droplet spreads as growth continues under increasing supersaturation, and eventually reaches the cell boundary. At this point the unlimited growth mode quickly and efficiently switches to a cell boundary-limited growth mode. The maximum supersaturation occurs at the point of mode switching.

The relationships among the supersaturation (S-1), droplet radius, cooling rate (-dT/dt), droplet number concentration n, time, and thermodynamic variables and constants are clarified. The maximum supersaturation,  $(S-1)_{max}$ , varies as

$$(S-1)_{\max} \propto \left(\frac{1}{n}\right)^{\frac{1}{2}} \left(-\frac{dT}{dt}\right)^{\frac{1}{4}}_{D} \left(-\frac{dT}{dt}\right)^{\frac{1}{2}}_{M},$$

where the subscript D stands for dry adiabatic cooling and M for

moist adiabatic cooling. When  $(-dT/dt)_{D} = 10^{-3} \text{ oC sec}^{-1}$ ,  $(-dT/dt)_{M} = 5 \times 10^{-4} \text{ oC sec}^{-1}$ , (corresponding to 10 cm sec<sup>-1</sup> updraft velocity), n = 100 cm<sup>-3</sup>, and the cloud base temperature is  $10^{\circ}$ C,  $(S-1)_{max} = 0.14\%$ . The time to reach the maximum is 25 sec, and the maximum occurs 2.5 m above cloud base.

n corresponds to the total number of cloud condensation nuclei active below  $(S-1)_{max}$ . A simple graphical method is developed to estimate the number concentration of droplets when (-dT/dt) and the activity spectra of the cloud condensation nuclei are given.

A cell boundary-controlled growth equation for a droplet whose surface does not satisfy the Maxwellian condition (the temperature and vapor fields show discrete steps due to the condensation and the thermal accommodation coefficients) has also been derived. The effect of these coefficients is to raise  $(S-1)_{max}$ . A method is suggested to determine the effect of the coefficients by comparing the theory and observed data for number concentration of droplets, (-dT/dt) and cloud condensation nucleus spectrum.

#### ON THE SIMULATION OF NATURAL CLOUD SUPERSATURATIONS IN COLD CHAMBERS

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Extensive intercomparisons of different ice nucleus counters undertaken during the Second International Workshop on Condensation and Ice Nuclei have revealed the necessity of critically assessing their design principles in order to model cloud conditions. In this perspective, an analysis of the operation of a cold chamber is made with regard to the saturation existing at the wall and in the sample, considering nonsteady-state vapor and temperature fields. An analytical expression for the nominal saturation ratio is obtained. It is found that the maximum supersaturation in the chamber reaches a value far beyond those known to exist in clouds, if the sample and the wall are both saturated with respect to water. Fig. 1 displays the supersaturation profiles in a chamber which has the wall saturated at -15°C and the sample saturated at 15°C is introduced into it. An explicit expression for the threshold value of the sample relative humidity, which avoids any excessive supersaturation in the chamber, is presented for the case when the sample and the wall are saturated with respect to water. If the wall is saturated with respect to ice, or undersaturated, the numerical computation of the threshold relative humidity of the sample can also be made with the help of the expressions obtained. The results of computations dictate the sample relative humidity for given operating conditions so as to avoid undesirable supersaturations in the chambers and thus help optimize their operation for simulating natural cloud supersaturations.



Fig. 1. Plot of supersaturation (S-1) vs distance (x) from the wall of the cold chamber at different time intervals (t) after the sample introduction.

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In the past two years a number of workers have developed detailed models for the growth of ice crystals in a cloud. Keenig and Jayaweera, for instance, have considered a microphysical model, including additional ventilation terms and, in a parametric way, additional information on the crystal shape (see Fig. 1). Cotton has fitted the ice growth model with 21 different categories of crystals into a dynamical framework including warm cloud processes.

Lamb measured the linear growth rates of the basal and prism faces in a <u>pure vapor environment</u>. These measurements allow a description of the crystalline habits and their variation with atmospheric conditions in an integral way. This includes both surface effects (resistance) and volume effects due to diffusion of vapor to the surface and heat away from the surface.

For the basal faces the linear growth rate  ${\tt G}_{\tt B}$  is determined by the net impingement flux  $_{\delta}\!{\tt F}$  :

$$G_{B} = \frac{\alpha_{B}m_{W}}{\rho_{ice}} \delta F$$
(1)

A similar equation applies to the prism face, so that the net rate of mass accumulation is given by:

$$\tilde{M}_{s} = \rho_{ice} [G_{B}A_{B} + G_{p}A_{p}]$$
(2)

And, if the impingement fluxes over the two faces are equal,

$$\dot{M}_{s} = k_{T} (\alpha_{B} A_{B} + \alpha_{p} A_{p})^{\delta} p = K_{s}^{\delta} p \qquad (3)$$

where  $k_T$  is the kinetic theory factor, the  $\alpha$ 's are deposition coefficients, the A's are areas,  $\delta p$  is the excess vapor pressure just above the surface, and  $K_s$  is an effective mass transfer coefficient for surface processes.

The problem is to calculate  $\delta p = p_i - p_e(T_s)$ , the vapor pressure just above the surface and the equilibrium vapor pressure at the temperature of

the surface. This requires a consideration of the volume effects. Using the classical approach,

$$M_v = K_m(p_{\infty} - p_i) = K_h(p_e(T_s) - p_e)$$
 (4)

where  $p_{\infty}$  is the environmental vapor pressure,  $K_{m} = 4\pi CDMf_{1}/RT_{\infty}$ ,  $K_{h} = 4\pi C\kappa RT_{\infty}^{2}f_{2}/L^{2}Mp_{e}$  and the symbols have their usual meanings. All these terms add in terms of series resistances so that the rate of mass accumulation is given by:

$$\mathring{M} = \frac{P_{\infty} - P_{e}}{1/K_{s} + 1/K_{m} + 1/K_{h}}$$
(5)

Eq. 5 has been programmed on the CDC 6600 computer to follow the integrated crystal growth; results are presented in Fig. 2. In this case the three extremes are considered, revealing the contribution of the three terms in the denominator of (5). We see that, at 50 sec., the K and K terms are large and have only a small contribution to (5). However, the surface term (see Eq. 3) determines the <u>relative</u> growth rates





of the two faces and the capacitance C which enters the  $K_m$  and  $K_h$  terms. The K term varies in proportion to the area (see Eq. 3), whereas the other terms vary as the crystal size (C). The result is that the  $K_s$  term is most important in the early stages of growth (see Fig. 3).

A comparison of the present data with the data of various workers is shown in Fig. 4 (50 sec. curve, dotted line). The general lack of pronounced peaks in the calculated values suggests that the calculations have omitted consideration of the positive feedback effect expected due to the crystal faces "reaching out" into the environment. In this regard, a full solution has been completed including separate capacities for each face with a form

 $\tilde{M}_{1} = C'_{11}\Delta\rho_{1} + C'_{12}\Delta\rho_{2}$   $\tilde{M}_{2} = C'_{21}\Delta\rho_{1} + C'_{22}\Delta\rho_{2}$ (6)

where the subscripts 1 and 2 refer to the basal and prism faces, the C's are modified capacitances, calculated from the full solution for a and the Ao's are excess vapor densities

cylinder (see Smythe, 1962), and the  $\Delta \rho$ 's are excess vapor densities.

This type of solution should care for the effect in the quasi-steady state. The results, however, were not significantly different from the earlier results. Apparently the enhanced growth expected due to the "reaching out" effect is approximately compensated by the negative cross terms,  $C'_{12}$  and  $C'_{21}$ . Also, none of the solutions predict values of the ratios of the axes greater than about 5. This result would indicate that, in natural growth, truly unsteady state, moving boundary effects, spurious values of excess vapor density, and perhaps enhanced ventilation are producing added deposition at the projecting edges (effectively increasing  $C'_{11}$  or  $C'_{22}$  or reducing  $C'_{12}$  or  $C'_{21}$ ) and increasing the growth (see dotted line in Fig. 4).



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PHOTOLYTIC ACTIVATION OF THE ICE-NUCLEATING PROPERTIES OF AGI HYDROSOLS Dennis M. Garvey, University of Wyoming, Laramie, Wyoming, USA

Droplet-freezing experiments have shown that irradiation can dramatically enhance the nucleating activity of AgI in colloidal suspensions. Spectra derived from experiments at constant cooling rates for two AgI suspensions under different conditions of illumination are shown in Figs. 1 and 2. The details of the experimental procedure and the method of calculating the freezing-nucleus spectra have been reported by Vali (1971). It can be seen that UV irradiation induced an increase of more than an order of magnitude in the number of nuclei in the first suspension and that its effect was even greater in the second. The source of illumination was a GE H85A3 mercury-arc lamp at a distance of approximately 25 cm. Experiments at constant supercoolings were performed after bringing the droplets to the desired temperature with no illumination. If the light was then left off, very few nucleation events took place. Fig. 3 shows what happened when instead the UV source was directed on the drops, in one case continuously and in the other intermittently. It is apparent that nuclei were being created with time and that this effect started within a minute after irradiation had begun. It can be noted, too, that the activity decreased equally quickly when the light source was turned off and that the activating effect was apparently completely reversible. In Fig. 4 we have converted the data from a similar experiment under steady illumination to the fraction of the unfrozen drops freezing per minute, a number related to the rate of creation of nuclei. It can be seen that this rate was nearly constant for as long as there were sufficent drops to give good statistics.

No attempt is made here to explain the physical mechanism responsible for the increased activity, though it is felt that the particlewater interfaces were being constantly altered by the radiation and that at any time there was a fair chance that the surface structure 152

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was such as to induce nucleation. Bryant and Mason (1960) have hypothesized that such an effect might be due to the removal of impurities by a photolytic process, though in the present case the effect may have its root in the reversibility of the photo-decay of AgI.

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Fig. 1. Freezing nucleus spectrum of powdered silver iodide in a colloidal suspension.



Fig. 3. Percentages of droplets frozen as a function of time with and without ultraviolet irradation.



Fig. 2. Freezing nucleus spectrum of colloidal silver iodide precipitated from a complex potassium iodide solution.



Fig. 4. Fraction of unfrozen droplets freezing per one minute interval.

### An In Situ Method for the Chemical Identification of Cloud Nuclei

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The method of flame scintillation spectral analysis involves the study of intensity and wavelength of light emitted when small particles are subjected to a high temperature. Calculations of the minimum particle size to give a significant signal-to-noise ratio for detection in the flame are described and based on the solid decomposition rate together with intensity relationships and spectral characteristics of the photodetector. The main conclusion predicts that under typical flame conditions the intensity of emitted light is proportional to the particle surface area or particle mass, depending on the initial particle size and the degree of cyrstal imperfection. The experimental part of the paper describes design and construction of a flame scintillation spectrophotometer and its application to the measurements of particles containing Na, K, and S at the sea shore of Hawaii and in the laboratory. The quality of this instrument is characterized by the following parameters: The smallest detectable particle size is 0.02 um, the smallest detectable particle size difference is  $\Delta \log r =$ 0.2, and the maximum particle concentration the instrument can handle without significant losses is  $N_{max} = 300 \text{ cm}^{-3}$ . The time required to detect a statistically significant number on a real time basis is of the order of minutes. A direct read-out of the slope of a Paretian particle size distribution is possible. Simultaneous measurements of K and Na in sea salt particles reveals a fractionation between these cations that apparently takes place in the bursting bubble formation of condensation nuclei. The sea salt distribution

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at the shore follows an asymptotic power law with a slope of -3.8 and a total population of 30 cm<sup>-3</sup>, or about 5 percent of the total Aitken population. The sodium aerosol that was able to penetrate the marine trade inversion is characterized by a slope of -4.5 and a total number concentration of only 0.2 cm<sup>-3</sup>. This suggests a removal efficiency of 99 percent with the larger particles being removed more efficiently. The sum of sodium containing particles, by which the sea salt component of the marine time aerosol is characterized, and of sulfur containing aerosols, hypothesized to be  $(NH_4)_2 SO_4$ , does not add up to the total Aitken nuclei population of 300 cm<sup>-3</sup>. This suggests the presence in the maritime aerosol of a third component which is probably of organic nature.

## THE KINETICS OF THE CONDENSATION GROWTH OF DROPLETS IN THE ADIABATIC PROCESS

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The condensation growth of a droplet population in the homogeneous air mass, rising up at a constant rate, is considered. Any time when the maximum water vapour saturation is observed, can be taken for the initial time reading t. Once the peak is passed the supersaturation  $\Delta(t)$  becomes a monotonously decreasing time function.

At the beginning droplets without any admixtures are considered. In such a case the droplets with the radius  $\mathcal Z$  , which at the  $Z_{K} = \frac{Z^{+}}{A(0)} (Z^{+} \simeq 10 \text{ cm})$ initial moment is smaller than the critical  $Z_{x}(t) = \frac{z}{t}$ will evaporate. As the critical radius monotonously increases, the point  $\mathcal{T}_{\kappa}(t)$ , moving to the ring along the size axis, can reach some near-by points, which move in the same direction. Then the droplets, corresponding to these points, will begin to evaporate. Assymptotically for large times the rate of transit of growing droplets into evaporating ones tends to zero. The radii of all growing droplets tend to one value 7. . Thus it is shown that in this case some droplets evaporate and the continuously growing droplet size spectrum gradually becomes  $\phi'$ shaped. The sixe axial coordinates of the point narrowing of the size spectrum are found. The droplet number which can reach  $\mathcal{Z}_c$ is defined. The characteristic time  $t_c$  of the drop size spectrum narrowing is defined. Then the case when all the droplet population have the same amount of soluble admixtures is considered. Thereby the initial drop size spectrum is being gradually divided into two parts. Both spectrum parts tend to narrow and move in different directions along the size axis. With 2 - ~ the droplet size spectra expressed by the monotonously decreasing continuous functions are to some extent specific. In this case there are always large droplets which change very little during periods considered. All the droplets are not able to reach the points of narrowing  $\mathcal{Z}_c$  . In general the size range from  $\mathcal{Z}_c$ will be packed with drops. The analysis of such a situato တ tion leads to the following conclusion. The point of narrowing  $\mathcal{Z}_c$ still exists as before. When the time is large (  $t \gg t_c$  ) in the

point  $\mathcal{T}_c(t)$  the size spectrum maximum is being formed. Its amplitude will get enlarged with time. The droplets considerably contributing into the fluid phase will concentrate in the near vicinity of the point  $\mathcal{T}_c \leq \mathbb{Z} \leq \mathbb{T}_c + \mathcal{E}(t)$  (with  $t \to \infty \mathcal{E}(t) \to 0$ ). These droplets influence the supersaturation course. For droplets with the radius  $\mathbb{Z} > \mathbb{Z}_c + \mathcal{E}(t)$  the size spectrum can be found analytically. With  $\mathbb{Z} \to \infty$  this drop size spectrum coincides with the initial one.

# THEORETICAL AND EXPERIMENTAL INVESTIGATIONS OF SOME REGULARITIES OF AEROION DIFFUSION ONTO CLOUD AND CONDENSATION NUCLEI

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For the interpretation of model experiments the basic assumptions of the diffusion theory of aeroion deposition on aerosol particles are analized. First of all assumptions about the independence of aeroion deposition on an individual droplet and about the use of the stationary diffusive equation for describing this process are considered. It is shown that for the time greater than the characteristic time of ion concentration relaxation these assumptions are true, provided the interactions in the system of drops and ions are taken into account. The experiments are conducted in the 3200 m volume cloud chamber in the conditions of natural ionization. The adiabatic cooling of the chamber leads to the adsorption of moisture by condensation nuclei and cloud formation. In the process of fog formation and further dissipation measurements of polar conductivities, spectral volume charge density and cloud microstructure are carried out. The radii R of the dispersed media particles during the fog life-time vary from  $R \ll \ell$  (  $\ell$  - free path length of the aeroions in the air) up to R»l therefore dependence of diffusion streams from Rand  $\ell$ is taken into account. Simultaneously the characteristic time of charge variations and cloud particle charging are estimated. The data from the first series of experiments where the initial humidity is equal to some 100% are compared with the theoretical calculations. It is found that the concentration variations of positive and negative ions due to the absorption by the droplets do not contradict the interpretation of the process in terms of ion diffusion.

In the second series of the experiments the air humidity varies from 40% to 80%. The measurement of the volume charge spectral density shows that the humidity increase in the chamber leads to an essential shift of the condensation nuclei mean charge to the negative sign. The shift value cannot be explained by different aeroion mobilities.

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The hypothesis on the absorption properties effect of the "Watered" condensation nuclei surface on the process of charging is being discussed. The reason for the rise of the infinite energy of the inductive attraction between ion and droplet with their approach is analyzed. This infinity does not allow to consider the selective properties of the droplet surface. It results from the fact that the inductive aeroion charges are assumed to be concentrated in an infinitesimally thin surface layer. The account of the aeroion interaction with free charges within a droplet shows that the thickness of the surface layer, in which charges are distributed, approximately equals the Debye radius screening. Simultaneous consideration of the real size aeroions and dielectric water permeability variations in the electric field gives the inductive energy values for pure water droplets and for droplets with soluble salts,  $10^{-3}$  ev for the first and  $10^{-1}$  ev for the second. This allows to consider the selective properties of a water layer adsorbed at the condensation nuclei surface, when aeroion diffusive streams are calculated.

## EPITAXIAL RELATIONSHIPS AND CHEMICAL KINETICS OF AgI-BEARING COMPLEXES IN HETEROGENEOUS NUCLEATION

By

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Extensive laboratory investigations during the past few years have revealed several negative aspects of the use of AgI seeding agents containing hygroscopic complexing agents. In the AgI-NaI-H<sub>2</sub>O system we have studied the crystallography and phase relations of three polymorphs having the composition AgI·NaI·nH<sub>2</sub>O (with n varying between three and four) and have revealed that all are monoclinic or of lower symmetry and appear to have no good epitaxial relationship with ice. Our studies show that generally two of the three polymorphs will breakdown as the saturation point is reached in the cloud but it is possible to have the third compound persist at temperatures lower than  $-5^{\circ}$ C and thus reduce the nucleating activity. Further deactivation might be anticipated because of the nature of the AgI precipitate coming from the breakdown of the complexes.

Although in the  $AgI-NaI-H_2O$  system the breakdown of the complexes yield free AgI surrounded by a solution envelope, in the corresponding  $AgI-KI-H_2O$  system in kinetics of the breakdown process are much lower and allow persistence of the phases into the region of normal activation of AgI nuclei.

Use of ammonium iodide  $(NH_{l_1}I)$  with AgI in acetone precludes the development of these complexes but still involves some contamination with slight hygroscopic properties. All evidence from both the laboratory and field observations made in our investigations points to the superiority as ice nuclei of the AgI particles obtained from AgI-NH<sub>h</sub>I-acetone solutions. 0ur work is in agreement with that of other laboratories showing the high activity ( $10^{12}$  nuclei/gm at  $-3^{\circ}$ C) of these nuclei in comparison with little or no activity at these temperatures from the other systems. We conclude therefore that in the order of decreasing efficiency or utility as cloud seeding agents we recommend (1) the  $AgI-NH_{L}I$  system, (2) the AgI-NaI system, and (3) the AgI-KI or other hygroscopic-bearing systems. In fact we strongly recommend that the KI bearing system not be used because of the observed low kinetics of breakdown of the complexes formed and symmetry analogous to that found for the NaI system which precludes support of any good epitaxial relationship.

Water Saturation in a Membrane-Filter Ice Nucleus Chamber - Fact or Fiction

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No entirely satisfactory method of measuring the concentration of ice nuclei in ambient air has yet been devised. As noted at the Second International Workshop on Condensation and Ice Nuclei (Fort Collins, Colorado, 1971), considerable disparity in ice nucleus concentration measurements occurs with varying apparatus. The millipore filter (MF) technique, as first developed by Bigg et al. (1963) and then substantially improved by Stevenson (1968), is being fairly extensively used. Despite the so-called volume effect problem, wherein ice nucleus concentrations are observed by some investigators to decrease with increasing sampled volume, presumed advantages of the MF method include its potential for large volume sampling, long activation time for nuclei, and precise control of relative humidity in the filter conditioning chamber.

Water saturation at the filter surface is desired in order to subject the ice nuclei to reasonably representative cloud conditions. In order for any diffusion chamber to reach and maintain prescribed humidities, it is essential that the supply flux of water vapor considerably exceed depletion by wall surfaces and the activated nuclei (ice and condensation). It was not apparent from the literature that vapor source versus sink calculations had been performed for "ice chambers", as had been done previously for cloud condensation-nuclei chambers (Twomey, 1959). Rather, one can only assume from stated temperature differentials and corresponding humidities that vapor losses by nuclei were ignored or considered negligible. The latter did not appear realistic, recognizing the high concentrations of

 $0.1\mu$  or larger particles typically captured by a filter (circa  $10^7 - 10^8$  cm<sup>-2</sup>). Consequently, a numerical time-dependent model of chamber conditions was developed to compute filter-surface relative humidity for various representative concentrations of ice and condensation nuclei. Subsequent experiments were run to test the calculations.

The following conclusions resulted from the finite-difference, iterative calculations:

1. Water saturation with representative concentrations of nuclei and volumes equal to or greater than 100 l is unlikely. Even with the chamber temperature differential set to give relative humidities in excess of 100% (with no sinks), the actual humidity is not likely to reach water saturation. After the humidity reaches its peak value, it then steadily declines as the ice crystals grow larger and become the dominant vapor sinks.

2. The maximum humidity decreases with an increasing number of sinks (nuclei of either type) on the filter. Thus as the sample volume is increased and chamber humidity reduced, one might expect fewer ice nuclei to be activated. This seems to offer another possible explanation for the well-known volume effect.

3. Peak humidities are achieved in approximately 3 to 7 minutes depending on the cooling time constant in reaching the desired base temperature. Small time constants allow the humidity to reach higher values.

4. Assuming the model simulates actual conditions in the chambers in use, one must conclude that the nuclei are generally counted under slightly subsaturated (water) conditions and are either deposition (sublimation) nuclei, contact nuclei, or "mixed" condensation-freezing nuclei. Though by no means unequivocal, the latter type seems the most probable; initial experiments are being conducted in an attempt to clarify the situation.

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Realistic Application of Telford's Stochastic Model for Chud Droplet Growth W. T. Scott and C.Y.Chen, Univ. of Nevada and Desert Research Institute, Reno.

A rapid, accurate saddle-point method has been developed for computing Telford's 1955 stochastic theory of cloud droplet coalescence growth. The accuracy has been checked against two exact methods, and the inclusion of Davis-Sartor collection efficiency<sup>1</sup> and Beard-Pruppacher velocity formulas has been arranged. The results enable the calculation of the probability that any fraction f of an initial large-size component of a cloud droplet spectrum of droplet volume kv<sub>o</sub> will grow by collection on droplets of volume v<sub>o</sub> to size  $lv_o$  or larger in dimensionless time T (k and 1 are integers). Results will be presented showing the effects of varying the efficiency E, the velocity u, and initial size on the growth of 15µ,20µ, and 25µ particles to 50µ or more.

The theory uses the probability  $dP_j = a_j dT$  that a droplet of size  $jv_o$ (j integral) becomes one of size  $(j+1)v_o$  in dT, where  $a_j = Ej^{2/3}(\frac{u}{u}-1)$ , where  $u_o$  is the velocity of the initial size droplet. By changing  $dP_j$  to dj, the time for continuous growth  $T=\int_{j2}^{j1} dj/a_j$  can be found. This integral can be converted into an integral over Reynolds number and thus be easily found from the Beard and Pruppacher formulas with E = 1. Continuous and stochastic growth rates are found to be closely equal for 50µ and larger drops. Consequently the time can be found for a fraction f of the particles of size  $kv_o$  to grow stochastically to say 5 times their diameter and then continuously to raindrop size,500µ or 1000µ radius. This time will be exhibited as a function of f from  $10^{-2}$  to  $10^{-6}$  for a variety of realistic conditions.

Figure 1 shows the time of combined stochastic and continuous growth for particles of radius 20µ falling through particles of radius 10µ for  $f = 10^{-2}$ ,  $10^{-4}$ , and  $10^{-6}$ . Also shown is the time for pure continuous growth from 20µ. Real time t in minutes is shown, given by  $t = (4a_0\rho/3wu_0)T$  where  $a_0$  is the initial small-droplet radius,  $\rho$  is the density of water, and w is the water content of small droplets in the cloud. The fairly small reduction of growth time from the stochastic effect is evident. Comparison will be made with corresponding results of Berry and the contribution of this method toward resolving the question of a "condensation-collection gap" will be discussed.

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## AUTOMATIC, CONTINUOUS MEASUREMENTS OF DROPLET SIZE DISTRIBUTION IN QUASI-MARITIME CUMULI

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The variations in cloud droplet distribution as a function of both altitude and horizontal position have been of interest for some time in the study of condensation and coalescent growth of droplets and entrainment of dry air into cumulus clouds. But, the lack of an automatic, continuous instrument or technique for making the measurements has restricted measurements to techniques based upon collection and/or replication of the droplets either with a continuous replicator or slide impaction device. Measurements such as those of Diem (1942-1966) and as recently as those of Warner (1969) have relied on these techniques. It is now possible to make the measurements automatically with real time readout with the electrostatic disdrometer, an instrument originally developed by Keily and Millen (1960) for the Air Force Cambridge Research Laboratories and modified, tested, and made operational for airborne use at NCAR (Abbott, Dye, Sartor, This paper reports the results of such automatic, continuous cloud 1972). droplet measurements made with a spatial resolution of about 35 meters in small cumulus clouds over the Florida Everglades.

A series of twelve flights were made on an NCAR Queen Air to test and compare the disdrometer with the Johnson-Williams liquid water content meter and impaction slides and to investigate the structure of small maritime cumuli. The clouds investigated were small cumuli with bases at about 3000 ft. and cloud tops around 7000 to 8000 ft., with cloud top temperatures above freezing. An example of some of the data collected is shown in Fig. 1, where the concentrations of drops in separate 3  $\mu$ 

<sup>\*</sup>The National Center for Atmospheric Research is sponsored by the National Science Foundation.

intervals are plotted versus time on sequential traverses of the cloud at different elevations. From this and similar additional data a number of conclusions are drawn. It is readily apparent that the cloud boundary frequently is quite sharp with little difference in the spectrum at the center of the cloud and only 35 meters from the boundary. There are regions or pockets in the clouds where the size distribution is wider than the surrounding cloud. The well known increase of median droplet radius with altitude is apparent, but there is not an obvious increase or decrease in droplet concentration with altitude.



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#### Figure 1:

Measured droplet concentration as a function of altitude and horizontal position in a small cumulus cloud over the Everglades, Florida. The number shown with each line is the channel number which corresponds to the radius intervals: 1 - 4 to  $7 \mu$ , 2 - 7 to  $10 \mu$ , 3 - 10 to  $13 \mu$ , 4 - 13 to  $16 \mu$ , and 5 - 16 to  $19 \mu$ .

#### THE WATER-DROP REBOUND PROBLEM

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Past experimental work indicates that during the collision of two water drops the existence of an air film trapped between the deforming water surfaces presents an initial barrier to the coalescence process by preventing the drops from making actual contact. The air film drains rather slowly until a critical drop separation is reached (on the order of 0.1 microns), at which time the air gap is suddenly bridged and coalescence proceeds rapidly. If the drop-rebound time is shorter than the film-drainage time, it is clear that the drops will bounce apart before coalescence can be initiated. The present research is concerned with giving a quantitative theoretical description of the bouncing problem, based on the above physical model. Estimates of rebound times and film-drainage times are presented here for the case of millimeter-size drops colliding at various speeds along their line of centers. The results compare favorably with experiment.

The drop dynamics are simulated numerically using a modification of the Marker-and-Cell method, a scheme developed to study the motion of a viscous liquid possessing a free surface. The full Navier-Stokes equations are utilized, and the geometry is here constrained to be axisymmetric. Capillary effects are included by sensing the curvature of the surface as determined by a Lagrangian coordinate system. The accuracy of the computations has been established by comparing the simulation of a drop oscillation with known analytic solutions, and by comparing the motion of colliding drops determined photographically with similar model predictions.

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The collision of two drops of equal size is simulated by allowing a single drop to impact on a rigid free-slip boundary. The computations have shown that for small fluid viscosities, the characteristics of such a collision, for example the deformation and rebound time, are functions only of the non-dimensional Weber number.

The dynamics of the draining air film are studied using an extension of a theory originally developed by Stefan and by Reynolds. The present theory takes account of the radial motion of the adjacent drop surfaces as the drops deform during collision. The adjacent surfaces are assumed to be flat and parallel (it can be shown that the dimpling actually observed in the film does not affect the drainage time). The theory involves several important time-dependent parameters that are evaluated from the drop simulation study, and would be extremely difficult to obtain experimentally. Good agreement is found between the predictions of this theory and actual observations [notably the work of Park (1970)], indicating that the bounce-off phenomenon can be explained by a film-drainage theory. The experimental dependence of the bouncing phenomenon on the impact speed,  $V_o$ , is found to be a result of the strong  $V_o^-$  dependence of the radial flow in the drops at the adjacent surfaces during deformation.

The role played in this problem by the finite mean free path of air molecules has been examined by using an effective reduced viscosity evaluated from experiments with rarefied flows. The effect of this reduced viscosity on predicted film drainage times is minor.

It is found that the effect of electrical forces on the bouncing phenomenon cannot be explained in terms of a uniform air film, and probably results from the acceleration of small surface perturbations which have lead to locally enhanced electric fields.

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**B4**.2

The Shape and Terminal Velocity of Low Surface Tension Water Drops

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During the last few years, several investigators have suggested that surfactants (surface tension reducing chemicals) injected into stratiform or convective clouds might alter the microstructure or behavior of these clouds. There are chemicals available which can reduce water surface tension by about a factor of 4 in concentrations of only 10 ppm and with chemicals effective at such low concentrations it is conceivable that one might reduce rainwater surface tension for a short time interval.

Before any field experiments are warranted, laboratory experiments are necessary. The only previous observations of low surface tension drops were those of Blanchard (1949). We have constructed a small vertical wind tunnel to stably support 1 -~ 8 mm diameter water drops falling at their terminal velocity. The tunnel is open ended and similar in design to some aeronautical and cloud physics wind tunnels (Pruppacher and Neiburger, 1968). The air flows into the tunnel through a smooth entrance section, moves into a 30 cm deep honeycomb section of 3 mm cells, through the contraction section (contraction ratio 20) into an observation region where the flow is diverging and finally through the air pump to the exhaust. The temperature and humidity of the entire room must be controlled in order to control the environment in the tunnel observation area. A divergent cross section in the observation region provides a position of vertical stability (Maybank and Briosi, 1961; Pruppacher and Neiburger, 1968). For horizontal stability a small velocity dip is formed by crossing three 50 µm wires in the tunnel section between the contraction section and observation section. At the point the drops are supported, the velocity dip is about 8 mm wide and the turbulence level in the dip varies

om 0.5% at 4 m/sec velocity to 0.6% at 10 m/sec velocity. Drops stably supported in the tunnel are photographed to determine shape and size and the tunnel velocity U is measured to determine terminal velocity  $V_{\infty}$ , since at the point the drops are supported  $V_{\infty} = U$ . Measurements are made with a pitot tubemicromanometer device and a hot film annemometer.

Dramatic changes are made in the drop shape and terminal velocity by reducing surface tension. Preliminary results indicate that by reducing the surface tension of 1-2 mm radius drops a factor of 3, the deformation (ratio of the drop minor axis dimension to the major axis dimension) is increased by about 20% and the terminal velocity decreased by about 30%. The high frequency drop oscillations previously observed (Blanchard, 1949; Maybank and Briosi, 1967; Brook and Latham, 1968) decrease as the surface tension decreases and deformation increases. These preliminary results indicate that by reducing surface tension only a slight increase in the sweeping action of individual drops can be expected. The terminal velocity is decreased by a factor almost equal to the increase in drop cross-sectional area. At present the experiments are still in progress and detailed quantitative results will be presented in the near future. This work is supported by the Army Research Office - Durham under contract number DAHC 04-70-C-0061.

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#### ACCELERATION OF DROPLETS DUE TO THE WAKE EFFECT

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### I. Introduction

Experimental measurements of collision efficiencies for equal size droplets performed by Telford et. al., 1955 and Woods and Mason, 1965 indicate that this collection is not zero. Recently, Cataneo et. al., 1970 have shown that the wake behind one droplet can significantly effect the trajectory of an equal size droplet falling over one hundred diameters behind it. In this paper, a procedure is outlined to determine the acceleration experienced by the rear drop due to the wake effect and this acceleration is reported for a 700um diameter droplet.

### II. Experimental Techniques

A standard technique is now readily available to generate single droplets of controlled sizes and production rates (Lindblad and Schneider, 1965). This technique has been modified so that pairs of equal and unequal size droplets can be produced at will (Adam et. al., 1971). The droplets in the pair can be produced with little or no charge, over a wide velocity range, and with variable spacing.

Consider a pair of droplets which are produced with a separation, S, and fall a distance, D, as in Figure 1. The equation of motion of the rear droplet is given by

$$\dot{y}_2 = g - \frac{F(Re)}{m} + \frac{F(S)}{m}$$

where g is the acceleration of gravity, F(Re) is the drag force on the droplet proportional to the Reynolds number and F(S) is the wake force proportional to droplet spacing. Since it has been experimentally determined that the front droplet is not influenced by the presence of the rear droplet for S greater than five diameters, the acceleration of the front droplet is

$$\dot{y}_{1} = g - \frac{F(Re)}{m}$$

An estimate of the wake acceleration can then be obtained by producing a pair of droplets with an initial velocity, V, and measuring the change in spacing,  $\Delta S$ , after the pair falls a distance, D. The average wake acceleration on the back droplet is then

$$\frac{F(S)}{m} = \frac{\Delta S}{S} \frac{H}{y_1}$$

where  $\ddot{\mathbf{y}}$  is the average acceleration of the front droplet.

## III. Results

A typical set of data are shown in Figures 2 and 3. The droplets used were 700um diameter, and fell with an initial velocity of 4 m sec<sup>-1</sup>. At a distance of 50 cm, the velocity was 3.4 m sec. This corresponded to an average drag acceleration

of  $-l^{1}$  m sec<sup>-2</sup>. The final versus initial spacing expressed in diameters is shown in Figure 2. Using these values, the average wake accelerations are for one initial velocity as a function of the spacing. This should be an accurate estimate of the acceleration experienced by one pair where the velocity of the rear droplet is a function of the spacing. This is true because the Reynolds number changes less than 30% from velocities of 4 m sec<sup>-1</sup> to terminal.

While the acceleration produced by the wake is less than 2% of the drag acceleration, it can add significantly to the relative velocity between droplets in the pair. This relative velocity, however, need not insure coalescence after collision. In fact, for most wake collisions examined, the rear droplet pushes the front drop to the side and passes it.

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## OSCILLATION FREQUENCIES OF FREELY SUSPENDED WATER DROPS

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In this paper, we report on an experimental study undertaken to investigate the natural oscillations of freely suspended water drops. Particular attention was centered on measurement of the main frequency of the drop oscillations over a substantially wider range of sizes (1 to 7 mm in diameter) than previously measured by other investigators. The results confirmed a theoretical relationship between drop oscillation frequency and diameter first derived by Lord Rayleigh in 1879.

The data analyzed consisted of film strips taken with a high speed 16-mm camera with close-up attachments for high magnification and a framing rate of up to 4,000 frames per second. The drops were photographed in a large vertical wind tunnel capable of suspending hundreds of drops simultaneously. Due to the large number of drops available for close, detailed study, it was possible to observe, in a qualitative manner, many forms of oscillation, rotations, and other more complicated

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patterns of distortion. However, quantitative measurements were attempted only for the pure cases of simple oscillation.

Our observations agreed well with Lord Rayleigh's prediction

$$f = \sqrt{n(n-1)(n+2)} \sqrt{\frac{2T}{\pi^2 \ell d^3}}$$

where f is the frequency, T is the surface tension, C is the density, d is the diameter, and n gives the mode of vibration. Our experimental data agree with both the power dependence and the coefficient of proportionality. This agreement over a frequency range of an order of magnitude is interesting, since the 3/2 power law is the first and simplest approximation for an oscillating drop. In particular, it represents the very small deformation of an inviscid, incompressible, spherical drop neglecting the effects of gravity and air flow, while our drops were suspended at their terminal velocity as in a rain cloud.

## WAKE EFFECTS AND DROP IMPACTIONS

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Introduction and Experiments: The development of precipitation and the determination of drop size distribution has led to speculation on the significance of drop impactions and drop wake effect interactions. Recent investigations on freely suspended large hydrometers (diameters  $\geq$  4mm) in a vertical wind tunnel (updraft diameter 1.8m) have revealed the relative importance of drop impactions and wake effects as microphysical mechanisms in drop growth. Hundreds of drop interactions have been recorded with high speed photography and then analyzed to determine the important parameters. A thousand or more drops were simultaneously suspended in the updraft and allowed to interact with each other through wake effects. For the drop impaction studies, the large drops were first suspended in the airflow and then impacted by a spectrum of smaller drops introduced from below.

<u>Wake Effect</u>: Generally, a 5mm equivalent diameter drop has a 100% chance of colliding when separated by a vertical distance of 5cm (10 drop diameters) or less. This is true when the drops nave a small relative horizontal velocity and when the center of the upper drop is not much more than a radius away from the wake axis. Of the 204 wake interactions studied, coalescence occurred in 33% of the wake collisions. For 20% of the collisions the drops bounced off each other. 46% resulted in break-up. The drops that broke up can be subdivided into 19% bag break-up and 25% fragmentation which resulted in several millimeter size droplets. For 54% of the

wake break-ups the two colliding drops while maintaining most of their original mass pulled out a filament which produced several small droplets. These experiments indicate that wake effects of large drops can effectively modify the drop size distribution only if certain rather specific conditions are satisfied. A number concentration of at least 200 drops  $\geq$  4.5mm in diameter per cubic meter, initially, must exist in a quasi-stable situation for times approximating ten minutes.

Impactions: The point and the angle of impaction are important in determining the number and size of drops produced during a non-permanent coalescing collision. However, the relative kinetic energy (r.k.e.) of the smaller impacting drop is the most important parameter. The r.k.e. was defined as 1/2 the mass of the smaller impacting drop times the relative difference in drop velocities(2). A critical r.k.e. of about 15 ergs demarcates permanent and non-permanent coalescing collisions. An impaction exceeding 15 ergs r.k.e. produces an average of 4 to 5 millimeter and submillimeter size drops. For all drops > 2mm in diameter there exists a unique spectrum of smaller drops whose r.k.e. of impaction exceeds 15 ergs, thus perventing permanent coalescence. However, the impactions seldom completely destroy the larger drops, although they do remove some mass. Drop impactions produce a rapid increase in the number of precipitation size drops while limiting the growth of larger drops. Impactions reveal the self-regulating mechanism in nature that enables collisions to influence both the initial growth and the determination of final size for large drops.

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## THEORETICAL AND EXPERIMENTAL STUDIES OF THE SHAPE AND

## TERMINAL VELOCITIES OF FALLING RAINDROPS

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Recently Pruppacher and Pitter (1971) showed that the shapes of falling raindrops could be deduced theoretically by using experimentally determined values of aerodynamic pressure about a falling sphere of the same Reynold number. However, to complete their calculations, they required the experimentally determined values of terminal velocities of raindrops. Using an iterative procedure, described by Brazier-Smith (1971) for theoretically determining the shape of a drop subject to external forces, we have been able to extend the work of these authors so that the terminal velocities need no longer be predetermined, but are derived along with the drop shapes. Once the equilibrium shape for a drop of given radius has been determined, its terminal velocity is then calculated from the stagnation pressure at the drop base.

The shapes of three drops of radii a)0.00cm, i.e. spherical; b)0.20cm, and c)0.25cm, as deduced theoretically are shown in Fig.1. This illustrates the characteristic flattening (case B) or depression (case C) of the lower pole of the falling drop. The predicted values of terminal velocities agree with experimental values to within 6% for radii less than 0.20cm (e.g. the predicted theoretical fall velocity of a 0.20cm drop is 9.43 m s<sup>-1</sup> compared with an experimental value of 8.83 m s<sup>-1</sup>). For a radius of 0.25cm the predicted fall velocity is 13% in excess of the experimental value of 9.09 m s<sup>-1</sup>.

The shape of free-falling drops has been studied experimentally. Using the technique described by Blanchard (1971),

large drops of a milk/water mixture of known surface tension were produced. After the drops had fallen 4m the initial oscillations had died away and the drops were photographed in reflected stroboscopic light. The acceleration, a, and the velocity of a given drop, radius R, at a given point can be deduced from the image positions. By dimensional similarity of the bond number,  $\rho(g-a)R^2/T$ , the shape is representative of a drop of radius R' at its terminal velocity given by R' = R[(g-a)/g]<sup>1/2</sup> for constant surface tension. Fig.2 shows a 0.4cm radius drop (followed by a smaller one in its wake) displaying a consistent, hence equilibrium shape from image to image and falling with a velocity of 7.33 m s<sup>-1</sup>. By dimensional similarity this corresponds to a water drop of 0.25cm radius falling at its terminal velocity.

Unlike previous experimental studies, for example Magono (1954) it was unnecessary for the drops to be at terminal velocity. Thus many experimental studies which were previously impractical now become feasible. The prospect of extending this study to include electrified drops is an important and promising one: work in this direction is currently in progress. It is known that these drops are related to the maximum attainable electric fields inside thunder clouds.

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References:



# Figure 1



Figure 2

## Droplet Collection Rates as Determined

by an Initial Gaussian Distribution

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Droplet growth rates due to stochastic collection evaluated for various initial conditions show a remarkably regular pattern of development. Each initial spectrum is a Gaussian number density function f(x) over droplet mass x, which is specified by a mean mass  $x_f$  and a relative variance var  $x=(1+v)^{-1}$ .

Fig 1 shows the time evolution of the mass density function g(lnr)=xf(Lnr) for the initial spectrum with  $r_{fo}=14\mu m$  and var x=1. Note that the radius,  $r_g$ , corresponding to the mass  $x_g=\int x^2 f(x) dx / \int f(x) dx$ , follows the progress of the spectrum mass quite closely. The time evolution of a narrower initial spectrum evolves slower and in a different pattern. Fig 2 shows  $r_g$  versus time for a variety of initial spectra. Fig 3 shows plots of the parameter  $B=(var x)^{-\frac{1}{2}}x_{fo}^{-1}$ versus T, the time required for  $r_g$  to reach  $50\mu m$ . Each circle on the graph corresponds to a different initial spectrum. The equation TL=3+11.4×10<sup>-8</sup>B, which describes the data shows the linear effect of the liquid water content L on the growth rate of a given spectrum.

The instantaneous time derivative of  $x_g$  is evaluated by observing the changes in the parameter  $b(x_g)=(x_gL)^{-1}dx_g/dt$ . Fig 4 shows a plot of  $b(x_g)$  versus  $r_g$  for four initial distributions.



Fig. 1. Time evolution of the Gaussian function distribution having  $r_{f0}=14\mu m$  and var x=1.0.



Fig. 2. Time change of r for several initial spectra: (1)  $r_{f0}=10\mu m$ , var x=1, (2)  $r_{f0}=12\mu m$ , var x=1, (3)  $r_{f0}=14\mu m$ , var x=.25, (4)  $r_{f0}=14\mu m$ , var x=1, (5)  $r_{f0}=18$ , var x=.25, (6)  $r_{f0}=18$ , var x=1.

Fig. 3. The time T required for  $r_g$  to reach 50um as a function of the parameter  $B=x_{f0}^{-1}$  (var x)<sup>-2</sup> and the liquid water content L.





Fig. 4. The parameter  $b(x_g)$  of the instantaneous rate of change of x as a function of  $r_g(x_g)$  for initial conditions given by  $(r_g, \gamma)$  where var  $x = (1 + \gamma)^{-1}$ .

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#### COMPARISON OF EXPERIMENTAL COLLECTION EFFICIENCIES OF CLOUD DROPS WITH THEORETICALLY COMPUTED COLLISION EFFICIENCIES

#### Morris Neiburger

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A quantitative theory still awaits definitive evaluation of the collection efficiency of falling cloud drops, among other factors. The collection efficiency  $E = E_s \cdot E_k$ , where  $E_s$  is the collision efficiency and  $E_k$  is the coalescence efficiency. Theoretical evaluations of  $E_s$  have been made, and some experimental evaluations of E carried out. The earlier experiments involved conditions which deviate markedly from those occurring in natural clouds. We have recently evaluated E over a range of drop sizes using the UCLA Cloud Tunnel, a vertical wind tunnel which provides conditions approximating those in natural clouds.

Both in the Stokes regime, assumed valid for radii up to 30 µm, and for larger cloud drops there have been new theoretical computations. In the Stokes regime Davis and Sartor found that the earlier computations by Hocking were faulty, and Hocking and Jonas have corroberated their findings. For larger cloud drop sizes Shafrir and Neiburger, revised by Neiburger (1967) carried out what may be the most valid computations using the flow superposition technique. Recently Klett, Davis and Neiburger (unpublished) attempted a solution using modified Oseen equations and computed collision efficiencies for collector radii from 30 to 70 µm. Comparisons of these theoretical results with the experimental results of Picknett, Woods and Mason, Telford et al, and Beard are shown in Figures 1 and 2. It is seen that for the 30 µm collector drop the Davis-Sartor theoretical curve fits the available experimental data better than the Klett-Davis-Neiburger one does, and that for the larger sizes the Shafrir-Neiburger curves fit the experimental data better than the K-D-N curves.

Table 1 shows the results so far available of the experiments using the UCLA Cloud Tunnel, compared to the theoretical results, for various values of the radius A of the collector drops and the ratio p = a/A of the radii of the two drops. The differences between the experimental values and the theoretical values are large, and their variation with A and p are also different than the variation of the theoretical values. A possible explanation is that under the conditions in the tunnel, corresponding more nearly to natural cloud, the coalescence efficiency  $E_{l}$  is quite small and decreases with increasing A and p in this range of p. Experiments using drops with charges large enough to influence the coalescence efficiency but not sufficient to affect the collision efficiency are under way to test this hypothesis.





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COMPARISON OF ODSERVED COLLECTION EFFICIENCIES WITH THEORETICAL COLLISION EFFICIENCIES

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Table 1

### Use of Equivalent Cylinders and Discs To Calculate

the Fall Velocities of Ice Crystals

K.O.L.F. Jayaweera<sup>1</sup>

It has been generally assumed that calculations of fall velocities of ice crystals, both columnar and plate-like, from the drag coefficient data of cylinders and discs can be made only as an order of magnitude calculations (List and Schemanauer, 1971). But recent laboratory experiments, performed with both single ice crystals falling in air and models of ice crystals falling in viscous liquids suggest that this approximation gives good indication of the true fall velocity provided suitable equivalent cylinders and discs are utilized for this purpose.

For columnar ice crystals, an equivalent cylinder having the same mass and external dimensions as the ice crystal has the same fall velocity as the ice crystal. Confirmation of this equivalence comes from the measured and calculated fall velocities of needles, sheaths and solid columns performed by Jayaweera and Ryan (1972). These experiments however were confined to small ice crystals (< 100 µm) but there is no reason why such equivalence cannot be extended for larger single ice crystals.

The inability to obtain dendritic type ice crystals with long branches in the laboratory has defied experiments to check for similar equivalence for plate-like ice crystals. But in a tank experiment, Jayaweera (1972) found that models of ice crystals and discs made of the same material having the same mass and surface area have very similar fall velocities. For a given shape a linear relationship exists between the two fall velocities with the star shaped model deviating most at 25% from its equivalent disc. These results are valid for various combinations of materials of discs and liquids and covers a Reynolds number range of 0.5 to 200, therefore applicable to ice crystals falling in air.

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Use of this technique to determine fall velocities is useful because

- (1) the drag coefficient data for discs are readily available and its relationship with Reynolds number does not depend very much on the thickness-diameter ratio, and
- (2) recent observations of Auer (1971) indicate that the surface area of basal face for plate-like crystals can be expressed in a simple expression involving only the crystal diameter.

Indeed in some dendrites observed by Auer the surface area is nearly independent of crystal diameter suggesting that if such crystals have the same thickness they will all have the same fall velocity as found by Nakaya and Terada (1935).

In conclusion, provided proper equivalence is adhered to, the fall velocities of both columnar and plate-like crystals can be calculated from those of cylinders and discs.

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Nakaya, U. and T. Terada Jr., 'Simultaneous observations of mass, falling velocity and form of individual snow crystals', J. Fac. Sci., Hokkaido, Imp. Univ., 1, Ser. 2, pp 191-201, 1935. Trigonal, Trapezoidal and V-shaped Ice Crystals Grown in Free Fall

### by Akira Yamashita

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Trigonal ice crystals were found to be formed by the seeding method to use adiabatic expansion technique as well as the method to use a very cold body. Especially the former method is interesting as it did not produce various complicated ice crystals but produced trigonal (see Fig.1), scalene hexagonal (see Fig.2) and hexagonal ice crystals at about  $-1 \sim -4^{\circ}$ C, at about  $-6 \sim -11^{\circ}$ C and at about  $-17 \sim -30^{\circ}$ C. Trigonal ice crystals are thought semi-stable because they were apt to change to scalene hexagonal ice crystals during their growth. A few trigonal dendrites (see Fig.3) and considerable number of hybrid crystals consisting partly of trigonal dendrites were also found at lower than  $-18^{\circ}$ C.

Their growth directions  $\langle 1\overline{1}00 \rangle$ , the second most closely backed directions in the basal plane, are thought to be the second most favourable directions in the basal plane for the growth of ice crystals.
Trapezoidal ice crystals and V-shaped ice crystals are shaped those found among various non-hexagonal ice crystals (Yamashita, 1971) grown by the seeding method to use a very cold body. V-shaped ice crystals and trapezoidal sheathes (see Fig.4) were estimated to be grown from the trapezoidal ice crystals as shown in Fig.5.

V-shaped ice crystals were formed only at about  $-3 \sim -4.5^{\circ}$ C, at about  $-6 \sim 8^{\circ}$ C and at lower than about  $-18^{\circ}$ C. Those grown at lower than about  $-18^{\circ}$ C (see Fig.6) were found to be a kind of twin crystals and the value  $\theta$  (see the figure) of those crystals decreased as the temperature became low.



Fig.1 -3.3<sup>o</sup>C (the triangular face ; (0001) )



Fig.2 -25.7°C (the scalene hexagonal face; (0001)



Fig.3 -26.8°c

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Fig.6 -26.8°C

Computer Simulation of Shape of Ice Crystal Growing or Evaporating in Air

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A differential equation governing the shape of the ice crystal growing or evaporating in air has been iteratively integrated with time. The crystallographic anisotropy has been taken into consideration. The temperature of the whole system was assumed to be uniform. The sublimation coefficient of water molecules was not introduced.

Two dimensional treatment in polar coordinates was made. The field of the water vapor around an ice crystal was expressed by the use of 5000 grid points. The radial distance from the origin of the coordinates to any given point on the crystal surface was expressed as a continuous variable. The angular spacing was every 5 degrees so that 27 was divided into 72 sectors for each of which the radial distance was computed. The length of the time step was 0.3 second.

Anisotropic property of the ice crystal was taken into consideration by expressing the dependence of the surface tension upon the direction of the normal of the local crystal surface as a function. Forms of the function were chosen so as to fit best the  $\eth$ -plot of ice with respect to the basal and the prismatic planes respectively. The local saturation vapor density at each portion of the crystal surface was determined from the local curvature, the local surface tension and the second derivative of the surface tension.

By altering the amplitude of the function expressing the anisotropy of the surface tension, the shape of hypothetical ice crystals of varying degree of anisotropies were computed from a nearly isotropic case to an extremely anisotropic case beyond the actual ice crystal.

The initial shape of the ice crystal was a circular or elliptic rod. The initial sizes for growth were about 5 microns in radius. Those for evaporation were about 12 microns in radius. The outer boundary of the air was a circular inner wall of a uniform vapor density. The wall was positioned at 525 microns from the center of the ice crystal. In computation of growing stages the vapor density at the wall was altered as follows; saturation over flat water at -15°C, 5%, 1%, 0.1%, 0.01% supersaturation over flat ice and finally just the saturation over flat ice at - 15°C. In computation of evaporating stages the same amplitudes were substracted from the saturation over ice.

Results of a series of the computations show that the shape of Wullf's equilibrium crystal and so called Curie's law of velocity of growth of the crystallographic facets took place prior to dendritic stages.

The Polarity of Electrical Charges on the Shapes of Snow Crystals

### Katsuhiro Kikuchi

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It is generally known that the polarity of the electrical charges on rain drops is positive and that the polarity of the electrical charges on snow crystals is negative. And it is also known that at times there is an inverse relation between the polarities of the electrical charges on precipitation elements and the polarities of the atmospheric electric field strength. This phenomenon has been explained by the ion capture theory. However, no explanation as to the shapes of snow crystal carrying negative charges has been given so far. On the other hand, it is well known that the habit of the shapes of snow crystals is decided by both temperature and supersaturation with respect to the ice surface.

The author carried out some observations in the field of cloud physics and atmospheric electricity when the author was a member of the wintering party of the 9th Japanese Antarctic Research Expedition at Syowa Station (69°00'S, 39°35'E), Antarctica throughout a whole year from February 1968 to January 1969. The observations of electrical charges on the natural falling snow crystals and blowing snow particles, the shapes of snow crystals and electric field strength were selected from a view point of importance and interest of the charge generation mechanism in clouds. The observations of the shapes of snow crystals were done by microscopic photography and the replica solution method and studies of the charges on snow crystals were made by vacuum tube electrometers for each crystal. At the same time, a field mill type of electrometer was used for the observation of the atmospheric electric field.

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As a result, a very interesting phenomenon was found in which the electrical polarity of snow crystals of dendrite, sector and plate was negative and the polarity of snow crystals of column, combination of bullets and side plane was positive. Taking into account that the temperature range of the growth of dendrite, sector and plate is between  $-10^{\circ}$  and  $-20^{\circ}$ C and that the other three shapes of snow crystals are between  $-20^{\circ}$  and  $-35^{\circ}C$ , it would be appropriate to assume a positive polarity of a thundercloud with a partial positive pocket at the cloud base brought about by liquid precipitation. In other words, it may be considered that negative and positive charge centers in the cloud volume are located around the  $-15^{\circ}$ C level and the  $-25^{\circ}$   $\sim -30^{\circ}$ C level, respectively. Takahashi observed positive charged particles at a temperature range between -18° and -39°C and negative charged particles at a range between  $0^{\circ}$  and  $-18^{\circ}$ C by using an electric charge These temperature ranges coincide with the growth ranges of sonde. the above described shapes of snow crystals. A positive pocket may will be explained by a melting effect.

Although the reason for the difference of polarity by the shapes of snow crystals is not clear at this time, this finding seems to be of considerable importance in the discussion of charge generation and separation in thunderclouds and other electrified clouds.

A numerical study of the airflow and cloud-drop

trajectories past thin oblate spheroids of ice

by

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Only very few theoretical and experimental investigations have been reported in literature on the collision efficiency of ice crystals colliding with cloud drops. Most of these are theoretically unsatisfactory or only qualitative in nature 1,2. In the present paper a theoretical study is presented which assumes that the shape of a simple ice plate and a graupel particle in its initial stages of growth can be approximated by an oblate spheroid. Extending the method outlined by Masliyah and Epstein<sup>3</sup> to a wider range of Reynolds numbers and smaller aspect ratios  $(AR = b'_s / a'_s, where a'_s is the semi-major axis and b'_s the semi-minor axis of the spheroid) we solved$ the complete, steady state Navier-Stokes equation of motion for viscious flow past an oblate spheroid with 0.5  $\leq$  AR  $\leq$  0.05 and for Reynolds numbers (N<sub>Re</sub> = 2a'<sub>s</sub> V'<sub>w</sub>/v') 0.1  $\leq$  N<sub>Re</sub> $\leq$  100. The oblate spheroidal coordinates are chosen with  $\xi = \xi_0$  on the surface of the spheroid, yielding  $tanh \xi_{\alpha} = b'/a'$  and  $c' = a' \operatorname{sech} \xi_{\alpha}$ . The radial distance from the axis of symmetry,  $\widetilde{\omega}'$ ,  $(\widetilde{\omega} = \widetilde{\omega}'/a')$ and vertical distance z' are found by  $\widetilde{\omega}'=c'\cosh\xi\sin\eta$ ,  $z'=c'\sinh\xi\cos\eta$ . In these coordinates the Navier-Stokes equations of motion can be written in non-dimensionalized form as  $E^{a}\psi=G \operatorname{sech}^{a} \xi_{O} \quad (1), \quad E^{a}G=\frac{N_{Re}}{2} \quad \frac{\widetilde{\omega}}{\sinh^{a}\xi+\cos^{a}\eta} J_{\xi,\eta}(\psi,F) \quad (2), \quad \text{with } G=\widetilde{\omega}\zeta, \quad F=\zeta/\widetilde{\omega},$  $J_{\xi,\eta}(\psi,\zeta) = \frac{\partial \psi}{\partial \xi} \frac{\partial \zeta}{\partial \eta} - \frac{\partial \psi}{\partial \eta} \frac{\partial \zeta}{\partial \xi}, \text{ and } E^2 = \frac{1}{\sinh^2 \xi + \cos^2 \eta} \left[ \frac{\partial^2}{\partial \xi^2} - \tanh \xi \frac{\partial}{\partial \xi} - \frac{\partial^3}{\partial \eta^2} - \cot \eta \frac{\partial}{\partial \eta} \right].$  The stream function  $\psi'$  and vorticity  $\zeta'$  were nondimensionalized by setting  $\psi = \psi' / V_{\infty}' (a_s')^3$ ,  $\zeta = \zeta' a_s' / V_{\infty}'$ , where  $V_{\infty}'$  is the terminal velocity of the spheroid,  $\rho_a'$  and  $\nu'$  are the density and kinematic viscosity of the medium. Equations (1) and (2) were solved with the same boundary conditions as those proposed by Masliyah and Epstein but with numerical techniques perfected by Woo and Hamielec<sup>4</sup>. The computations yielded flow and vorticity fields from which the pressure distribution at the surface of the spheroid, and the drag force coefficients were computed. The values for  $c_{D}$ , the pressure distribution and the flow and vorticity fields agreed well with the results of Masliyah and Epstein for  $1 \le N_{Re} \le 100$  and for AR = 0.5 and 0.2. As a further B6.1

check, we determined  $c_D$  experimentally from the fall velocity of oblate spheroids with AR = 0.5 and 0.3, a' = 0.635 cm, and of disks with a' =0.635 cm and a thickness h' =0.05 cm, giving h'/2a' = 0.04. High precision spheroids and disks were manufactured from nylon, aluminum, brass and lead and were allowed to fall in a plexiglass tank of 5 m in height and 30 cm in diameter filled with oils of various viscosities. The drag force coefficient was determined by the method described by Pruppacher and Steinberger<sup>5</sup>. The experimental results agreed well with our numerical computations. Encouraged by this agreement we used our numerical flow fields to determine the trajectory and collision efficiency of cloud drops, assumed to be water spheres and lying in the path of a thin oblate ice spheroid of AR = 0.05 falling in air. For this computation the superposition method was assumed to be adequate since the size ratios of the chosen colliding drops and ice spheroids were small. The equations of motion for the ice spheroid and cloud drop are then given by  $M_s' \frac{d\vec{V}'}{dt'} = M_s'\vec{g}^{*'} - \frac{c_{DS}N_{Re}}{24}6\pi a_s'\eta'(\vec{V}_s' - \vec{U}_d')$  and  $M'_{d} \frac{dV_{d}}{dt'} = M'_{d} \vec{g}^{*'} - \frac{c_{Dd} n_{Re}}{24} 6\pi a'_{d} \eta' (\vec{V}'_{d} - \vec{U}'_{s}) \text{ where } g'^{*} = g' (\frac{\rho' - \rho_{a}}{\rho'}), \vec{V}' \text{ and } M' \text{ are the}$ velocity and mass of the colliding particles,  $\vec{U}$  is the velocity of the air, t' is time,  $n_{Re}$  is the Reynolds number of the drop,  $\eta'$  is the viscosity of air and ho' is the density of ice. Trajectories were computed for ice spheroid sizes corresponding to the available flow fields. For given  $N_{Re}$ , AR and  $c_{DS}$  the size of the spheroid was computed from  $c_{D}N_{Re}^{a} = \frac{32}{3}(a')^{3}b'g'(\rho'-\rho_{a}')\rho_{a}'/\eta^{3}$ . Two

examples of computed trajectories are given in Figure 1 for an air temperature and pressure of

-19°C and 500 mb,  $a'_{s} = 229\mu$  and drops of  $a'_{d} = 5\mu(\bullet)$  and  $30\mu$  (x).

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### AN AERODYNAMIC MODEL FOR THE TUMBLING OF SPHEROIDAL HAILSTONES ABOUT A HORIZONTAL MAJOR AXIS

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Wind tunnel measurements of the static drag, lift and torque acting on oblate spheroids, are incorporated into a model of the aerodynamics of freely falling hailstones. A dynamic damping torque proportional to the angular velocity is also included. The hailstones are allowed three degrees of freedom; namely, translation in a vertical plane and rotation about a major axis which is normal to this plane. Two cases, I and II, are considered, corresponding to axis ratios of 0.50 and 0.67, terminal Reynolds numbers of 3.25x10<sup>4</sup> and 4.19x10<sup>4</sup>, and Best numbers of 0.85x10<sup>9</sup> and 1.14x10<sup>9</sup> respectively. Initially, the spheroid is falling at terminal speed with the minor axis vertical, when the motion is perturbed by a finite angular displacement. The equations of motion are then integrated numerically using a fourth order Runge Kutta method. The resulting amplitudes and phase relations of the angular and horizontal oscillations, as well as the trajectory, are illustrated in Fig. 1 for Case II. Depending upon the value of the non-dimensional damping coefficient, K, and the initial perturbation angle, the angular oscillation may either damp or amplify. Resonant coupling between the horizontal and angular oscillations provides an energy source for the angular oscillations. If this exceeds the energy sink due to damping, the oscillation amplifies; if it is less, it damps. The boundaries between these regimes are shown in Fig. 2. For K sufficiently large, all perturbations are damped. For smaller K, small perturbations amplify and large ones damp, leading to neutral oscillations of constant amplitude. For still smaller K, all perturbations

amplify until the spheroid flips over, and rotation (tumbling) begins. Thereafter, the rotational velocity increases until, after several hundred revolutions, a steady rotation occurs with a rate as high as .02 rotations per diameter fallen; the mean fall speed is about 25 per cent higher than the terminal speed with the minor axis vertical. Horizontal translation is negligible. K factors have not been measured yet; however, tumbling about a horizontal major axis has been observed in a vertical wind tunnel.





Figure 1.



### THE AERODYNAMICS OF SPINNING HAILSTONE MODELS

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The stability of the three-dimensional angular motion of freely falling, spinning spheroids was studied experimentally and theoretically. Three oblate spheroids of axis ratios, .5, .67, and .79, were used as hailstone models. Aerodynamic torques were measured in a wind tunnel for each spheroid, spinning about its minor axis, which was inclined to the mean air flow. The torques, which were measured as a function of Reynolds number, spin rate, and inclination angle, were used in an analysis of the general motion. The general equations of motion for a freely falling, spinning spheroid were derived in terms of the coefficients of aerodynamic drag, lift and torque. Numerical solutions to the equations of motion were obtained by approximating the non-static coefficients of lift and drag by static values as was done by the Toronto group for non-spinning spheroids. Criteria for the occurrence of possible motions under this quasistatic approximation have been developed.

A special case of motion is described by the minor axis precessing (in an oscillatory fashion) and nutating about a mean position in the horizontal. Such a motion was also observed when spinning spheroids were allowed to fall freely for 10 m. If natural hailstones would grow wet with such a motion, the liquid surface would be shaped by centrifugal effects rather than by aerodynamic molding. The quasistatic approximation for the aerodynamic torques neglects aerodynamic damping. However, this may be justified since amplification of angular oscillations about a major diameter of a spheriod was observed for Reynolds numbers greater than  $10^5$ .

Under certain conditions the gyroscopic properties of a spinning spheroid were found to add stability to its angular motion and in particular to motion which would tend to preserve the symmetry of the spheroid if it were growing by accretion of particles. This type of motion affects factors such as aerodynamic molding and heat and mass transfer. These effects must be determined before a complete understanding of the growth of spheroidal hailstones can be achieved.

### COMPUTATIONS OF HAILSTONE GROWTH AND TRAJECTORIES IN MODEL CLOUDS

By

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The histories of hailstones forming under a wide range of conditions have been studied with the aid of a numerical hailstone model based upon a consideration of hailstone mass and heat budgets. The hailstone embryos are derived from the freezing of raindrops or drizzle droplets. Physical processes simulated include raindrop breakup, drop freezing, dry hailstone growth, spongy hailstone growth, water shedding from hailstones, and hailstone melting.

The hailstone growth model has been used to study the histories of hailstone embryos introduced at different times and heights into over 50 versions of a one-dimensional time-dependent cloud model. Hailstone diameters at the ground have been calculated as a function of such cloud parameters as cloud lifetime, cloud height, height of freezing level, maximum liquid water content, maximum updraft speed and the temperature at the level of maximum updraft speed.

It is found that the hailstone diameter at the ground is determined closely by the maximum updraft speed experienced by the stone and the temperature at which this maximum updraft is experienced. Updraft maxima of 40 m sec<sup>-1</sup> or more occurring well above the freezing level are required to produce hailstones with diameters in excess of 5 cm at the ground. Our results differ from some previously announced, principally by Soviet authors, in that very large liquid water contents exert little influence on hailstone diameters at the ground and that updraft maxima near the -30C level are more favorable to the production of large hail than are updraft maxima nearer the freezing level.

The model has been used to predict effects of artificially induced cloud glaciation upon the sizes of hailstones reaching the ground. In the "seeded" model cloud, the transformation of cloud water to cloud ice occurs in a non-linear fashion between -5 and -25C; the comparable figures for an unseeded cloud are -20 and -40C. Hailstone diameters are generally reduced by artificial glaciation when the updraft maximum occurs at a low temperature. For cases where the updraft maximum is not far above the freezing level, the model indicates that artificial glaciation could result in larger hailstones at the ground. This occurs in cases where hailstones grow from an ice-water mixture on their final descent through the seeded cloud, so that they have less difficulty in disposing of latent heat than they would have if growing upon supercooled water alone.

Features of the large scale flow within baroclinic

disturbances which influence the distribution of precipitation

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An analysis of the flow within baroclinic disturbances over or near the British Isles is made assuming that the wet bulb potential temperature ( $\Theta_{w}$ ) is a conserved property of the flow. Conservation of mass and, when the flow is dry, of mixing ratio, are also used as constraints in the analysis. A model of the flows significant to the production and distribution of precipitation is derived. It is shown that most of the precipitation reaching the surface initially forms within a single well defined flow labelled the 'ascending conveyor belt' which is typically a few hundred km wide and a km deep and flows parallel to and immediately ahead of the surface cold front (stippled in the diagram). It ascends above the warm frontal zone at the order of 10 cm s<sup>-1</sup>. Ascent often begins within the warm sector so that there is no well marked discontinuity between the precipitation within the warm sector and in advance of the surface warm front. This flow has been identified as the most significant flow producing widespread precipitation in a variety of synoptic situations over the British Isles, including some analysed conventionally as non-frontal.

Potential instability is continually generated by differential advection in advance of the cold front as a result of the low level flow being over-run by a mid-tropospheric flow of lower  $\Theta_w$ . The instability is gradually released as the low level flow ascends. The leading edge of the warm frontal precipitation is eroded by the evaporation within the descending flow beneath the warm frontal zone. When the melting level is low, or the air entering the system on its forward side is dry, the evaporation decreases the width of the surface precipitation by several hundred km compared to that aloft. Some of the moisture which

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is evaporated is eventually re-precipitated further north as the flow ascends close to the surface warm front, probably as a result of frictional convergence. At some occlusions, in particular 'bent back occlusions, all of the precipitation is produced within this flow.

It is considered that this model provides a useful framework within which studies of the smaller scale variability of the precipitation can be interpreted. Also it could provide a valuable forecasting model, being more directly related to the precipitation than is the conventional Norwegian Polar Front Model.



The figure depicts the three flows considered to be of most significance in the development of baroclinic precipitation. + signifies regions of growth and - regions of evaporation. Stippling denotes the extent of the surface precipitation.

### PERSISTENT FLOW PATTERNS WITHIN CONVECTIVE STORMS

#### INFERRED THROUGH DOPPLER RADAR OBSERVATIONS

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Convective storms profoundly alter the environmental wind field. A full description of the convective process requires some knowledge of at least the main features of this flow. Doppler radar offers an excellent opportunity to acquire detailed information on winds within precipitating regions of storms. By directing the radar at nearly horizontal elevation angles, errors introduced by variable fall speeds of precipitation may be minimized.

Doppler radar does not solve all the problems in specifying flow patterns within storms, however. Motion components normal to the radar beam are not sensed by a Doppler radar. Complete measurement of motions within a volume would require a suitable array of three Doppler radars. Useful clues may, however, be had with just one Doppler radar, and when disturbances in the flow pattern are intense, of relatively simple form, vertically extensive, and persistent, it may be possible to make reasonable inferences on the nature of the major flow features.

An extensive surveillance of motions in the horizontal plane within thunderstorms has been conducted for several years with the AFCRL meteorological Doppler radar located at Sudbury, Massachusetts, U.S.A. This radar operates at a relatively attenuation-free wavelength of 5.4 cm and has a half-power conical beamwidth of 0.9°. Doppler velocities are processed in real time and displayed on a Plan Shear Indicator, or PSI. The key feature of the PSI display is a sharing of range and velocity information on the same radial coordinate. As the radar antenna rotates, the location of any precipitation echo is marked by a series of narrow concentric arcs, representing consecutive range intervals of 855 meters. Wrinkled arcs, irregularly

spaced, represent disturbances in the motion field. An example of a disturbed PSI pattern is illustrated in the figure.

Several thunderstorms which inflicted severe damage at the ground displayed arc patterns suggestive of a mesoscale vortex, with diameters between velocity extrema of up to 5 km, maximum vorticities near the center approaching 0.1 sec<sup>-1</sup>, persisting for periods of 30 to 60 minutes. Development of a mesoscale velocity structure of anticyclonic sense was observed in a storm of 20 June 1969 which passed through Boston. Early in its history the pattern was consistent with a clockwise wortex located at heights between 4 and 7 km above the ground. The base of the vortex descended to around 2 km as the upper part intensified somewhat, and finally the pattern appeared to elongate toward a shear line, still with anticyclonic sense. Interestingly, the echo observed with a conventional radar displayed no unusual shape during the initial stages of this disturbance, but eventually developed a beautiful mirror image of the classical tornado hook which trailed behind the major precipitation area.



PSI display of Boston storm of 20 June 1969. North is toward top; radar is located at center of arcs. Disturbed pattern to WSW suggests an anticyclonic vortex, centered at height of 2.7 km for this scan elevated 6° above horizon.

The Use of Geosynchronous Satellite (ATS III) Data for Estimation of the Structure and Dynamics of Thunderstorms and Severe Local Storms.

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An University of Wisconsin experiment: a spin scan camera on board the ATS-III (Applications Technology Satellite), has provided time-lapse information on the brightness and area of clouds and cloud systems associated with squall line thunderstorms and severe local storms. This remote sensing technique has been shown to be useful in (1) arriving at estimates of the convective transport of mass and energy in such storms (Sikdar, Suomi, and Anderson 1970); (2) revealing the circulation characteristics of such storms (Auvine and Anderson 1972); and (3) determining rainfall patterns and rainfall rates at the meso-scale (Sikdar, 1972). This paper describes the basic measurement techniques, the various types of data available, the underlying theories and assumptions involved, and offers

examples which illustrate how these estimates are obtained.

The authors shall present evidence which indicates that conclusions based on these remotely-sensed data fit very well with conclusions about the structure and dynamics of such storms obtained by conventional ground based and aircraft sensors. In addition, the authors shall suggest that unusually destructive storms (tornadoes, hail, etc) involve synergistic influences that are detectable in advance by remote sensing satellites such as ATS (III).

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# STUDIES OF THE FORMATION OF PRECIPITATION IN CUMULUS CLOUDS BY STATISTICAL METHODS

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A better understanding of the processes governing the formation of precipitation can be achieved not only with the use of theoretical and numerical cloud models but also by applying the statistical methods. The statistical models of precipitation can be useful in determining the principal factors governing the formation of precipitation and for solving some problems of practical value.

In the paper the results of statistical studies of cumulus clouds over the Ukraine are presented. Using the data about 83 cumulus clouds which were obtained with the use of an aircraft, the discriminant function was constructed. This function divides all the clouds investigated into classes: raining and notraining. It was found that the formation of precipitation mainly depends on cloud vertical depth and out of cloud relative humidity in the layer of cloud development.

For clouds with rain the equation of multivariable regression, which connects the rain amount with cloud and atmosphere parameters was constructed; it was chosen 7 different parameters and their products. It was found that the multitative correlation coefficient is equal to 0,72 and the highest significance has vertical cloud depth.

The second regression equation was constructed for well developed cumulonimbus clouds. In addition to the earlier used parameters the radar data were introduced to this equation. The multitative correlation coefficient proves out to be equal to 0,87. The correlation coefficient between real and calculated with the use of the deduced equation amounts of precipitation for some number of control clouds is equal to 0,91. The principal contribution to the amount of rain gives vertical cloud depth and the magnitude of its radar echo volume.

The results obtained in the paper allow to determine the most important rainforming factors and can be also used for the estimation of the results of cloud modification experiments.

## STRATIFICATION OF LIQUID WATER CONTENT IN CUMULUS CLOUDS I.P.Mazin, F.Ya.Voit Central Aerological Observatory, Ukr. Res. Inst. for Hydrometeorology, U.S.S.R.

The results of LWC measurements carried out by aircraft in cumulus clouds above Ukraine area in 1959-69 are summarized. The data obtained in more than 1000 crossings of Cu clouds are analysed. Experimental data were subdivided according to cloud thickness  $\triangle$  H and temperature at cloud base  $t_{g}$ .

The mean vertical LWC profile is found to fit well the Beta--distribution, i.e.

$$\mathcal{N} = \frac{\overline{W}}{\overline{W}_{m}} = \frac{\overline{\xi}^{m}(1-\overline{\xi})^{n}}{\overline{\xi}^{m}(1-\overline{\xi})^{n}}, \text{ where } \overline{W} - LWC, \overline{\xi} = \frac{h}{\Delta H}$$

A being height above cloud base and  $\xi_0$  - relative height of  $W_m$  - maximum LWC (see Fig. 1). If  $\xi_0$  and  $\xi_1$  are given ( $\xi_1$ ) corresponds to  $\mathcal{N}(\xi_1) = 0.5$ ) one can determine the  $\mathcal{M}$  and  $\mathcal{N}$  values. For 58 cases analysed the modal values  $\mathcal{M}_{mod} = 2.8$ ;  $\mathcal{N}_{mod} = 0.38$ . The mean values  $\xi_0 = 0.83$ ;  $\xi_1 = 0.55$ . The  $\xi_1$  value is extremely stable -  $\delta_{\xi_0} = \frac{\sigma_{\xi_0}}{\xi_0} = 0.08$  as compared with  $\delta_{\xi_1} \approx 0.2$ .

With  $0 \neq \approx 0, 2$ . The near LWC of cloud's column is  $\overline{W}^* = \frac{B(m+1, n+1)}{\xi^m(1-\xi_0)^n} \overline{W}_m$ and the total amount of liquid water in such column is  $Q = \overline{W}^* A H$ Here  $B(m+1, n+1) = \int \xi^m(1-\xi)^n d\xi = \frac{\Gamma(m+1)\Gamma(n+1)}{\Gamma(m+n+2)}; \frac{1}{\xi^m(1-\xi_0)^n} = \frac{(m+n)^{m+n}}{m^m n^n}$ 

The easy-to-use nomogram is given to determine  $\overline{W}_m$  as accurately as  $\pm 30\%$  if  $\triangle$  H and  $\overline{t}_{\mathcal{B}}$  are known.

Our results that W is much smaller than adiabatic value  $W_{\alpha}$  coincide with obtained by other authors but in contrast with the widespread opinion (see e.g. Warner, 1970, Journ. Atm. Sci., 27, 1035-1040) the ratio  $\overline{W}/W_{\alpha}$  is found to increase with height at least till  $\xi = 0.5 - 0.66$ .

The measurements in and under clouds proved that for continental clouds the small value of  $\overline{W}_{W_{\alpha}}$  can not be explained by existence of precipitation as it is proposed for tropical clouds (**Y**.Simpson,1971, Journ. atm. sci, 28, 449-455). We explain the observed  $\overline{W}_{W_{\alpha}}$  value and its change with height by more intensive lateral entrainment in lower cloud parts.

There are regions at any levels in Cu-clouds with dimensions

of 100-200 m where ratio  $\overline{w}/w_{c\iota}$  is much more than the mean values and often close to 1. It might be that processes in these very regions govern the cloud thickness. If this is correct the discrepancy between numerical models and observational results pointed out by Warner would be removed.



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Johannesburg, Republic of South Africa Drop-size distributions in rainfall were measured continuously using an 8.6 mm high resolution meteorological Doppler radar. The radar set, located in Johannesburg (Republic of South Africa), was pointed vertically and a digital computer was used to reduce the data obtained.

A typical case was selected from many recordings obtained in different types of rainfall during the period March to October 1971. In the particular case presented the target range was 280 metres while the range resolution was 15 metres.

The drop-size distributions were computed at one minute intervals from the beginning to the end of the storm which lasted approximately 17 minutes. Each drop-size spectrum was obtained by averaging 15 spectra over a period of 5 seconds. Two very interesting features, which were observed in other recordings as well, are:

- (1) There is a higher abundance, as compared with the Marshall-Palmer distribution, of drops having a diameter less than 1 mm.
- (2) The drop-size distribution changes quite rapidly at the beginning and the end of the storm.

In the central part the spectrum is approximately constant and compares closely with the Marshall-Palmer distribution for drops greater than 1 mm in diameter.

The above two points are shown very clearly in Fig. 1 where the drop-size distribution compares closely with the Marshall-Palmer distribution in the period from 5 minutes after the beginning of the storm to 6 minutes before the end.



### RAINDROP SPECTRA AND THE CELLULAR STRUCTURE OF PRECIPITATION

by A. Waldvogel and J. Joss Osservatorio Ticinese, Locarno-Monti, Switzerland

Measurements of size distributions of raindrops at the ground and of radar reflectivity profiles were made continuously during precipitation. The variations of the raindrop spectra and of the reflectivity profiles can be explained easily with recent findings of the cellular structure of precipitation (AUSTIN 1968, BROWNING and HARROLD 1969, BROWNING 1971). Two kinds of time intervals can be distinguished in the variations: long ones of the order of an hour and short ones of the order of minutes. The first ones may be attributed to meso-scale phenomena (areas of tens of kilometers), the second to the small scale convective cells within the precipitation.

Raindrop spectra were measured with an electromechanical distrometer (JOSS and WALDVOGEL 1967). Parameters of the spectra were calculated and recorded in real time. The parameters chosen permit a quick and easy examination of the distributions and thus enable the investigation of variations due to small and meso-scale effects. Reflectivity profiles, measured up to 15 km with a 5 cm radar pointing vertically, were recorded in analog form. Eight different levels of reflectivity were displayed, the difference between two successive levels corresponding to a factor of 3 of the rainfall rate.

The curves in the following figure were measured during a wide spread rain. The reflectivity profile is shown at the top, the parameters  $N_0$  and  $\Lambda$  of the spectra, the rainfall rates R are shown below and the median volume diameter  $D_W$  and the median reflectivity diameter  $D_Z$  are shown at the bottom. On the abscissa the time is indicated in hours.  $N_0$  and  $\Lambda$  correspond to an assumed exponential distribution with the same liquid water content W and the same reflectivity factor Z as the measured spectrum (each one measured during 30 sec.). Two curves of the rainfall rate are shown: the first (solid curve) is calculated directly from the distribution, the second (dashed curve) is derived from the calculated reflectivity factor Z. Time intervals of different precipitation character (A, B, C and D) lasting ca. 1/2 - 1 hour follow in turns. The variations are probably due to the influence of different meso-scale aereas. Each interval shows compared to the following one drastic changes in the reflectivity profile and in the drop spectra. Small scale convection is illustrated by the streamers of the radar profile and the short term variations (ca. 5 minutes) of the parameters of the distributions.

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## "Condensed Water in Hail Clouds" by T. G. Kyle NCAR, U.S.A.

Measurements of the condensed water in severe convective storm clouds will be described. These measurements are in regions of the cloud from -20C to -5C. The instruments being used are of two types. One is an evaporative instrument which measures the condensed water content by the temperature rise of air flowing through a heated cavity interspersed with screens. The temperature rise is a measure of the heat required to vaporize the condensed water. The second instrument measures the optical extinction of a laser beam by individual drops and cloud droplets. It will provide both water content and size spectra information.

The instruments are carried into the storm by a T-28 aircraft belonging to the South Dakota School of Mines and Technology. The aircraft has been provided with armor to permit flights into cloud regions containing hailstones as large as 7 cm diameter. The aircraft is directed into regions of high radar reflectivity from the ground with the information of cloud structure obtained from a ground-based 3 cm and 10 cm radar.

The measurements to be discussed will be obtained during the summer of 1973 as part of the National Hail Research Experiment in northern Colorado. Additional measurements of radar reflectivity, vertical wind velocities at cloud base and other meteorological observations will be available for discussing with the data. The results will establish whether there is an accumulation zone in hail clouds, and permit the comparison between radar reflectivities and measured cloud drop distributions.

### CELLS IN A CANADIAN STORM

### by C. WARNER

### McGill University, Montreal

A large convective storm occurred on 29 July 1967. It was observed with the AN/FPS-502 radar of the Alberta Hail Studies Project, which operates at wavelength 10.4 cm with a  $1^{\circ}$  conical beam. Stereo measurements were made from photographs exposed simultaneously every 3s, the cameras being at the ends of a north-south baseline 5.6 km long, with the north site near the radar.

Plan Position Indicator diagrams are shown in Fig. 1 for the 12 min interval 1439 to 1451 MST (2151 GMT). To these have been added the locations of cloud tops seen at the higher levels.



Figure 1. RADAR PPI's. Time from left to right. Height above ground from bottom to top. Loci defining positions are given at top left: the radar/ north camera site is to the right.

Contours (reflectivity thresholds) at about 25, 35, 45 and 55 dB above  $1 \text{ mm}^6 \text{ m}^{-3}$ . Echo tops 17, B and A identified. Cloud tops indicated schematically by circles commensurate with measured widths. Beside each circle is: cloud number, height. Clouds 12Y, 15, 16, 17, 17A and 18 are shown.

At 1439, the small cloud elements 17 and 17A appeared directly over an echo at 9.3 km at  $6^{\circ}$  elevation, small tops on a large rising tower. With drifting tops 15 and 16, a small corresponding echo appeared at 7.4 km. Top 12Y was well downstream.

Cloud 17 began to drift downwind at 11 km at about  $37 \text{ m s}^{-1}$  from  $260^{\circ}$ , a velocity similar to that of a nearby radiosonde balloon at the same height. An extension of echo downwind was associated with this cloud up to 1451.

Cloud 18 was a southward protrusion of spreading cloud rather than a main top. It is associated with echo B, which appeared at 1445 (5<sup>o</sup>).

On the western side of the storm, cloud corresponding to echo A was obscured by cloud 17. As it reached its peak, echo A moved across the left rear side of the storm.

Two regions of cell development have been found: on the south, and on the west side of the storm. Turning now to the lower troposphere, it is interesting to look in Fig. 1 from low levels at late times towards high levels at early times. The patterns show remarkable similarity along these diagonals, up as far as  $5^{\circ}$  elevation: 7 or 8 km. This seemed to be the generating level, and lower PPI's seemed to be dominated by fall-out. The fall velocity suggested is 8 to 10 m s<sup>-1</sup>.

The PPI's at  $1^{\circ}$  extended further towards the northeast than those higher up. This region was overlain by the cloud anvil, and the area increases occurred near the freezing level. It seems likely that precipitation became detectable to the radar as it fell. The PPI's at  $0^{\circ}$ , 400 m above ground, all were smaller than those at  $1^{\circ}$ , probably because of evaporation.

## A NUMERICAL STUDY OF SHALLOW CONVECTION by : Jiin -lang Lee Pennsylvania State University, U.S.A.

A numerical model of cumulus convection emphasizing the dynamic framework is developed and discussed. The model is set into an axisymmetric coordinate system and consists of equations of motion, thermodynamics and continuity. The system is treated as shallow convection thereby justifying the application of Boussinesq approximation. With functional profile assumptions in the horizontal for each variable, all equations are integrated in the radial direction and then solved numerically in the vertical direction as a function of time.

The numerical experiments simulate both dry and moist convection. In the experiment with moist convection, additional equations for the continuity of water substance are included by using Kessler's microphysical parameterization. The distribution of pressure perturbation associated with the evolution of the convective cell is shown. Furthermore, the effects of the perturbed pressure on the development of a cumulus cloud are studied and discussed in detail. A parameterized representation of eddy diffusivity which is expressed as a function of the characteristic size of eddies and the turbulent velocity is used in the centered differencing computation scheme.

It was found that the perturbation pressure gradient term in the momentum equation acts to reduce the intensities of most fields in the model solutions. Two case studies are made and the model is proved to be more physically realistic than earlier one-dimensional models and to be compatible with two-dimensional models.

## THE THREE-DIMENSIONAL SIMULATION OF A THUNDERSTORM CELL

### R. Wilhelmson

### University of Illinois, U.S.A. The numerical simulation of the life cycle of a thunderstorm

cell including both dynamical and physical aspects is to be carried out in three-dimensions. The need for such an investigation is indicated by Murray (1970) who noted differences in using an axisymmetric and a rectilinear cloud model. The axisymmetric model resulted in a stronger updraft and a smaller ratio of maximum downdraft to maximum updraft, and seems to give more realistic results. The former relationship is also indicated in a preliminary working three-dimensional simulation in comparison to a corresponding two-dimensional model. One reason for prefering to work with a three-dimensional model over a two-dimensional axisymmetric one is that recent studies indicate a difference in the micro-scale energy cascade between two- and three-dimensional motions. Another is that wind shear can be readily introduced in the former. The numerical experiments underway are intended to give an introductory look at three-dimensional moist convective simulations including the differences between similar two- and three-dimensional experiments. The first experiment is a look at the life cycle of thunderstorms in an environment having no wind shear. The deep convective equations of motion in both cases (two- and three-dimensional) are used in primitive form as several two-dimensional experiments indicate that a similar form

of the equations of motion for comparison is important, since differences were noted in similar primitive and vorticity cases. A second and third experiment include a wind shear that is linear with height and a wind shear with a low level jet. Results of interest include the growth and shape of the cell, the amount of precipitation, the life cycle time of the cell, the relation between updraft and downdraft, regions of convergence and divergence of air, and a look at the pressure field. The prognostic variables include the three components of velocity, the potential temperature, and the mixing ratios of water vapor, cloud droplets, and raindrops. The diagnostic variables include temperature and pressure. The conversion from cloud droplets to raindrops is given by the autoconversion expression of Berry (1968) and the collection expression of Kessler (1969). Ice is not included although the size of the initial perturbation is more typical of the middle latitudes than the tropics. It is intended that these studies will help check our understanding of two- and three-dimensional simulation and stimulate further investigation.

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### ON THE ROLE OF PRECIPITATION MICROPHYSICS IN CUMULUS DYNAMICS

Phanindramohan Das, Texas A & M University College Station, Texas, U. S. A. <u>and</u> Maddukuri China Subbarao, Atmospheric Environment Service Downsview, Ontario, Canada

That microphysical processes play a decisive role throughout the dynamical life of the cumulus has been established by the U. S. Thunderstorm Project which observed an unmistakable association between precipitation formation and downdraft development in thunderstorms. Later, theoretical-model studies have established this association on a firm physical basis. A further demonstration of the importance of microphysical processes comes from a theoretical explanation of the unsaturated downdraft, in which it is shown that a strong downdraft carrying all its liquid-water content in a small number of larger drops will not have enough evaporation for sustaining the moist-adiabatic process.

The question as to what happens if the <u>condensation</u> in an <u>updraft</u> is dominated by large drops has been explored by the present writers in a numerical-model study under the extreme condition of all the cloud water being divided into precipitation-size drops. The result shows a very rapid cooling off of the updraft together with the development of fantastic magnitudes of supersaturation. While these results are obviously unrealistic and only of academic interest, one should not ignore the observed fact that in large cumuli precipitation progressively dominates the cloud processes. Indeed, an understanding of the larger cumuli can and should be sought in terms of their "microdynamics" -- a term presently being used for the dynamical changes that both influence and are influenced by the microphysical processes.

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In the first of a series of theoretical studies on the microdynamics of the large cumulus the initial conditions are set by an adiabatic, steady-state updraft in which the water content is represented by droplets having concentrations characteristic of continental clouds. In the subsequent march of events the conversion from cloud water to precipitation is parameterized by a relation implying stochastic processes of collisioncoalescence. The precipitation generated in this conversion lies in the lowest three (40-µm, 56-µm and 80-µm) sizes of a drop spectrum described by thirteen classes which have their mean radii lying on a logarithmic The precipitation grows by continuous collection -- of cloud dropscale. lets by drops as well as of smaller drops by the larger ones. The largest drops are made to break up into smaller ones. The condensation (evaporation) proceeds on (from) droplets and drops and on nuclei brought in from the layers above and below the cloud. No unactivated nuclei are allowed for in the cloud. The study shows the following results:

(1) As precipitation develops, supersaturation in the updraft increases slowly at first. Later, when precipitation dominates the cloud processes, supersaturation attains unrealistic magnitudes in some areas.

(2) Precipitation causes an accumulation of rain water, which, in the lower parts of the cloud, drives a downdraft and the accompanying rain gush.

(3) In the downdraft below the cloud evaporational cooling is small: the moist-adiabatic process is not sustained.

In the subsequent studies some of the restrictions on the above model are relaxed. The results of the first model are changed most when inactive nuclei carried in the updraft are activated by the higher supersaturations resulting from precipitation development.

A COMPARATIVE STUDY OF THE SENSITIVITY OF PREDICTED CLOUD PARAMETERS TO VARIOUS FORMS OF THE EDDY EXCHANGE COEFFICIENT

### bу

# William R. Cotton

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It is well known that one of the most difficult problems in numerical modelling is a proper treatment of sub-grid scale processes. The purpose of this paper is to demonstrate the sensitivity of various cloud parameters to the formulation of eddy exchange coefficients.

A sequence of numerical experiments were performed with a onedimensional Eulerian time-dependent model employing horizontal turbulent entrainment as discussed by Cotton (1971).

The model is vertically staggered between vertical velocity and the heat and moisture variables. The nonlinear terms in all equations are solved with the Crowley [Crowley (1968)] 4th order differencing scheme on a 50-meter grid.

Numerical experiments were performed on a case study discussed by Saunders (1965) while assuming that the area of updraft to total domain area is 1%. The precipitation process is parameterized on the assumption that the initial cloud droplet concentration is 100/cm<sup>3</sup> and the radius dispersion is 0.25.

Four different experiments were performed with the eddy exchange coefficients defined as follows:

> (A)  $K_{\rm m} = K_{\rm H} = K_{\rm qv} = K_{\rm Qc} = K_{\rm QH} = 60 \text{ m}^2/\text{sec}$ (B)  $K_{\rm m} = k' (Dz)^2 |D_{zz}|$ , where  $|D_{zz}| = |\frac{\partial w}{\partial z}|$  and k' = 0.028,  $K_{\rm H} = K_{\rm qv} = K_{\rm Qc} = K_{\rm QH} = 60 \text{ m}^2/\text{sec}$

(C) 
$$K_{m} = k' (Dz)^{2} |D_{zz}|$$
, where  
 $k' = 0.7$   
 $\frac{K_{H}}{K_{m}} = 1.3$ ,  $K_{qv} = K_{Qc} = K_{QH} = K_{m}$   
(D)  $K_{m} = /k' (Dz)^{2} |D_{zz}| + \alpha (Dz)^{2} \{ |\nabla\beta_{zz}| \}^{1/2}$ , where  
 $k' = 0.08$ ,  
 $\frac{K_{H}}{K_{m}} = 1.3$ ,  $K_{qv} = K_{Qc} = K_{QH} = K_{m}$ ;  $\alpha = 1.0$ .

The latter term in (D) is introduced to account for local bouyancy fluctuations in the cloud field with

$$|\nabla \beta_{zz}| = \left[ \left( \frac{\partial}{\partial z} \quad g \left( \frac{T_v}{T_{vo}} - Q_T \right) \right)^2 \right]^{1/2}$$

It was found that some coupling between the momentum field and the local thermodynamic field was essential. The latter term was deduced by combined dimensional analysis and logical deduction.

The results of the numerical experiments (A) through (D) demonstrated that the precipitation field is the most sensitive parameter to the nature of the eddy exchange coefficient. Experiment (C) demonstrated that if all the exchange coefficients are formulated only as a function of the deformation of the momentum field, significant errors may occur in the heat and moisture fluxes. Finally, Experiment (D) demonstrated that the formulation of the exchange coefficient as in (D) is qualitatively reasonable. Further testing must be done, however, to determine its quantitative validity.

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- Crowley, W. P., 1968: Numerical advection experiments. Mon. Wea. Rev., Vol. 96:1, 1-11.
- Saunders, P. M., 1965: Some characteristics of tropical marine showers. J. Atmos. Sci., 22, 167-175.
Numerical Simulation of the Life-cycle of Cumulus Clouds

# by : R. E. Lopez Colorado State University, USA

A numerical time-dependent model has been developed that can simulate the entire life-cycle of different types of cumuli under different environmental conditions. The model has been designed to provide a dynamical framework for studying natural and artificial microphysical processes, and to be a tool in investigating the interaction of cumulus convection with larger-scale flows.

The model is basically one-dimensional and parametric. In this type of formulation the complex turbulent and microphysical phenomena are expressed in terms of cloud-scale variables, while the more straightforward dynamical and thermodynamical processes are treated in detail in a prognostic fashion. Although the model is fundamentally one-dimensional, the clouds are assumed to consist of two regions: a protected core and an exposed surrounding shell. The entire depth of the cloud is numerically simulated for each of these mutually interacting regions.

The mixing between cloud and environment is parameterized in the model in terms of the turbulence intensity of the interior and exterior of the cloud. Thus, the velocity of entrainment is made proportional to the r.m.s. of the velocity fluctuations inside the cloud, while the velocity of detrainment is proportional to the r.m.s. of the velocity fluctuations in the environment. In this way, the commonly used but physically invalid assumption of similarity is avoided. Additional equations have been introduced in the model to predict the turbulence level of the cloud at all times.

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The internal circulation of the clouds, and the attending redistribution of mass between levels, is parameterized in the model in terms of the one-dimensional velocity field of the core. Laboratory and theoretical information about spherical vortices is used in the parameterization. The effect of this approach has been to prevent the unrealistic increase in the radius of the upper layers of the clouds which is unavoidable in one-dimensional models that do not take into consideration internal mass redistribution in the vertical.

The clouds simulated with the model have been used to investigate the effects of different populations of clouds on the thermodynamics of tropical cloud clusters. It has been found that the net warming of the environment due to direct diffusion of cloud mass (detrainment) is very small or in most cases negative. This results from the fact that the cooling produced by the evaporation of detrained cloud water compensates or overcompensates for the detrainment of sensible heat excess. The model is also utilized to calculate the sinking motion induced on the environment to compensate for the upward mass transport by the clouds. The warming due to the induced subsidence can be large enough, under certain large scale flow patterns, to explain the development of hurricanes from weaker tropical disturbances.

# ENERGY TRANSFORMATION AND REARRANGEMENT CAUSED BY CUMULUS CONVECTION

L. Randall Koenig and Francis W. Murray

A numerical model of cumulus convection including warm microphysics (1) has been used to examine the production, decay, rearrangement, and flux of energy of various kinds that result from the growth and decay of small cumulus clouds in initially static environments. The work is directed toward the understanding of how clouds interact with and modify their environment, the goal being to aid in the formulation of realistic parameterization of cumulus convection in large-scale atmospheric models. The forms of energy examined are:

 static energy (and its individual components: enthalpies of (a) dry air, (b) water vapor, and (c) condensed water; latent heat of water vapor; and potential energy); and

2. kinetic energy.

Figure 1 summarizes the net redistribution of static energy caused by the growth of one simulated cloud in a cylindrical computational domain 9 km in height and 6 km in radius. The net energy change in horizontal slabs 200 m thick with radii equal to 6 km over a period of 60 minutes from the cloud's inception until after its decay is shown. The change in energy associated with the latent heat of water vapor is shaded. In accordance with expecuation, results indicate that cumulus convection changes the vertical distribution of static energy by causing a net loss in regions near cloud base and a net gain near the summit. Almost all of the redistributed energy is associated with the movement of water vapor. This conflicts with the viewpoint that cumulus convection causes net upward flow of energy principally in the form of sensible heat carried aloft by convection currents after having been released as latent heat during condensation at relatively low levels [see, for example (2)].

The cloud shown in Fig. 1 was small, having a low efficiency in converting condensed water to rain on the ground. A relatively large proportion of the water carried aloft as droplets does not return to the ground as rain but evaporates causing an increase in water vapor content of the atmosphere near the cloud summit. Thus energy required to evaporate the droplets remains in the atmosphere in the form of latent heat of condensation at the expense

of sensible heat content. A similar situation occurred with a more efficient cloud approximately twice the size (in terms of thickness and amount of condensed

water). Larger clouds may conform more closely to the generally accepted concept of upward transport of sensible heat rather than latent heat. This concept is based on presumption of a moist-adiabatic process where all condensate falls out as rain, re-evaporation being excluded. We hope to examine this hypothesis with models of larger clouds.

Examination of the net change in static energy density as a function of position in the atmosphere reveals the fact that while intense energy redistribution is confined to the cloud itself, the gain in energy near the cloud summit occurs at the expense of loss in energy by a large portion of the domain, particularly at levels near the cloud base and below.



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# ON THE PRECIPITATION DRIVEN DOWNDRAFTS IN CUMULONIMBUS CLOUDS

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The paper deals with the theory of development of quasi-steady cold downdrafts in Cb clouds, in cases when the temperature lapse rate is not very close to the dry-adiabatic value. As shown by Hoockings (1965) and Ludlam and Kamburova (1966) such downdrafts cannot be maintained by evaporation of precipitation, since large droplets which prevail in heavy showers evaporate too slowly. Explanation proposed in the present paper is based on the supposition that downdraft is cooled by rapid evaporation of small droplets supplied by entrainment of cloudy air from the neighbouring updraughts. However, the question of effectiveness of this mechanism as a motor of downdrafts is not quite simple, since together with entrainment of small droplets, there is entrainment of higher temperature and opposite vertical momentum from the updraft.

The problem is quantitatively investigated on the basis of a simple model, in which updraft and downdraft are represented by steady jets with entrainment and with "top hat" cross profiles of all thermohydrodynamic parameters. Updraft is assumed to be entrained by cloudless environmental air from the front of the cloud; downdraft is entrained by cloudy air from the updraft and eventually "detrains" part of its air to the environment in the rear of the cloud (Fig.1).

The speed (w+) and temperature (T+) of the resulting downdraft at given altitude (z) is estimated in terms of environmental temperature (T), environmental lapse rate ( $\gamma = dT/dz$ ), environmental dew point deficit ( $\Delta Td$ ), entrainment coefficients for updraft ( $\mu^{\dagger}$ ) and downdraft ( $\mu^{\dagger}$ ), wet-adiabatic lapse rate ( $\Gamma\omega$ ) and dry-adiabatic lapse rate ( $\Gamma$ ).

In first approximation  $\mu^{\dagger}$ ,  $\mu^{\downarrow}$ ,  $\Delta Td$ ,  $\gamma$ , and  $\Gamma \omega$  are assumed to be height-independent. Values of  $w^{\downarrow}$  and  $T-T^{\downarrow}$  are then estimated and optimal (in sense of maximalizing  $w^{\downarrow}$  and  $T-T^{\downarrow}$ ) values of  $\gamma$ ,  $\Delta T_{d}$ ,  $\mu^{\uparrow}$ and  $\mu^{\downarrow}$  are found. The most striking feature of this case is that downdraft critically depends on the entrainment of dry, environmental air to the updraft; for  $\Delta T_{d} = 0$ , or  $\mu^{\uparrow} = 0$  downdraft cannot exist. Direct application of these results to real conditions must be made with caution, but investigation of the constant-parameters case give clear insight into the physics of the process and can be used as a convenient base for further extensions of the theory to more sophisticated models. Some extensions of such sort are undertaken and yield

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among others the following results:

1) Variation of ratio  $(\Gamma - \gamma)/(\Gamma - \Gamma \omega)$  with altitude is an important factor for development of downdrafts. Decrease of this ratio with height intensifies downdrafts, increase - attenuates them; case of constant value of this ratio can be reduced to that with constant parameters.

2) Accumulation of liquid water in the updraft is favourable for the downdrafts.

3) Additional entrainment of environmental air to the downdraft attenuates it.

4) In most cases optimal values of  $\gamma$  only slightly exceed  $\Gamma \omega$ ; for  $\gamma = \Gamma$  or  $\gamma = \Gamma \omega$  cold downdrafts of discussed sort are impossible.

5) Optimal values of  $\mu \downarrow$  correspond to the downdrafts which are at the edge of saturation.

6) Estimated values of w+, T-T+ and  $\mu$ + are in good agreement with this, what is actually observed, provided that the altitude is not too close to the cloud base, where another mechanism of maintaining the downdrafts becomes dominating. Theory requires relatively large values of  $\mu$ + but this can be easily explained by the fact, that even in weak wind shear, the column of precipitation is inclined with respect to the air current and thus additionally ventilated from the lateral side. (Fig.2). Estimation of this effect shows that resulting  $\mu$ + might be of correct magnitude.

<u>Conclusions</u>: The presented theory gives satisfactory explanation of quasi-steady, cold downdrafts down to ca 1-2 km above the cloud base. Precipitation maintains the downdraft mainly by inducing suitable entrainment of cloudy air from the updraft; drag of precipitation is of secondary importance, while its evaporation can be practically neglected.

References: Hoockings C.A., J.Appl.Met.4, 1965; Ludlam, F.H., Kamburova P.L., Q.Jl.R.Met.Soc., 92, 1966.



Fig.l. Scheme of drafts and entrainments



Fig.2. Scheme of entrainment in wind shear.

## Micro- and Macrostructures of

Numerically Simulated Convective Clouds

by

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and

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This paper describes four numerical experiments conducted by means of a two-dimensional dynamic model. The model, simulating a warm trade-wind cumulus, predicts a space-and-time-variable vertical velocity and non-turbulent entrainment consistent with continuity requirements. In three out of the four experiments, it includes explicitly the microphysics of condensation. Coalescence, however, is simulated in the simpler framework of an isolated parcel model. Convection is in all cases initiated by means of a bubble which is buoyant because of excess temperature and, in one case, has additionally excess momentum. Condensation takes place on a specified population of sodium chloride nuclei representing maritime conditions.

The main results may be summarized as follows:

- Supersaturation is in the range 0 to 0.5 percent, the highest values occurring in the cores of the main updrafts.
- 2. The effects of supersaturation on the macrophysical events is noticeable. Comparison with conventional experiments, which allow no supersaturation, shows that storage of vapor as supersaturation reduces the amount of latent heat released through condensation. This leads to diminished temperature excess, less vertical motion, less conversion of vapor to liquid, and consequently an overall reduction in the intensity of convection.

- 3. Near the peak of its development, the cloud is cooler by a few tenths of °C than its environment. But it is still buoyant owing to a small but intense core of positive temperature excess which provides a driving force for further convection.
- 4. Evaporation taking place near the edge of the cloud leads to cooling and thereby produces negative buoyancy and kinetic energy which maintains the cloud despite decreasing intensity of the main updraft.
- 5. Liquid water content in the cores of the main updrafts range from 1.0 to 1.8 gm kgm<sup>-1</sup>. In most parts of the cloud much lower values prevail, and in areas of downdrafts near-zero values develop that make for "holes" in the cloud.
- 6. Cloud droplet spectra for selected grid points show, in most cases, a development of a single maximum which are centered around a radius anywhere from 6 to 24µ. In some cases, multiple maxima occur; the appearance of such maxima is here ascribed to advection.
- 7. Computation of coefficients of dispersion show good agreement with observed values reported by Twomey (1966), who gives a range of 0.15 to 0.50, and Warner (1969), who reports the most common value to be 0.2.
- 8. Results of experiments for a single parcel model indicate that had coalescence been incorporated in the convection model, at least some of the simulated clouds would have produced rain within the span of 20 to 30 minutes.

LIGHTNING FLASHES

A J Illingworth

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Precise measurements of the variation of the electric field, or 'recovery curve', after close lightning flashes to ground show that these variations are not exponential, but that there are consistent deviations from the exponential form.

It has previously been established that the 'recovery curves' after distant discharges become more rapid with increasing distance, and that this is due to re-arrangement of the space charge existing in the fair weather conductivity gradient which at large distances screens the cloud charge. Such space charge re-arrangements have much less influence on close recovery curves which might therefore be expected to reflect the charge regenerated within the cloud more accurately. Because of reversal distance complications which lead to irregular recoveries after close cloud discharges, this study was restricted to discharges to ground with field changes above 1,000 V m<sup>-1</sup>. So far all the ground discharges with a signal to noise ratio of more than 200 recorded at a four-station network in Socorro, New Mexico for one day in 1971 have been analysed.

Previous workers observing that the rate of recovery is a maximum directly after the discharge and decreases monotonically thereafter, have assumed the recoveries are of an exponential form decaying to the pre-flash level. Measurements of the initial gradient or the time to decay to a certain level could then be expressed as a time constant. In this study least squares fits of the field E with time t

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to the polynomial

 $\mathbf{E} = \mathbf{A} + \mathbf{Bt} + \mathbf{Ct}^2 + \mathbf{Dt}^3 \dots$ 

are presented. If the curves are exponential of the form  $E = E_0 \exp(-t/\tau)$  then, without assuming the field is decaying to its pre-flash level,  $\tau$  should be given by B/2C and C/3D. The results obtained show that the apparent value of  $\tau$ obtained increases as increasing lengths of the recovery are fitted. This increase is not explicable by the truncation of the infinite polynomial series.

Assuming the curve to be exponential and differentiating it we see that a fit of  $\ln(\frac{dE}{dt})$  against t should also give an apparent value of  $\tau$ . Again the apparent value of  $\tau$ increase as longer sections of the recovery curve are included. Typically the time constant is 10 seconds for the first ten seconds of recovery, but doubles after twenty seconds.

Close recovery curves need not only reflect the Coulomb field of the regenerated charge in the cloud but may also be influenced by point discharge at the ground and by rearrangement of any charged layers at the cloud/air conductivity gradient. However early workers interpreted the supposed exponential recovery as reflecting a constant generation current and an ohmic dissipation current proportional to the charge regenerated, and attention has been drawn to the high ohmic conductivity implied. If we still assume a constant generation current, then these new measurements can be interpreted as a non-ohmic dissipation current, that is a field dependent conductivity. Work is continuing on an analytic expression for this dissipative term as a function of field.

ON THE CONVECTIVE ELECTRIFICATION OF A SIMPLIFIED MODEL CLOUD

by: R. F. D. Perret, Florida State University, U.S.A.

The low mobility of natural ions and charged droplets allows significant convective transport of charge against an adverse potential gradient in cumulus clouds. This conversion of mechanical energy to electrical energy forms the basis for the earlier schemes of convective electrification. The absence of solid conducting boundaries and the unsteady nonuniform air motion has made difficult the quantitative study of convective electrification of natural clouds. Under the constraints of a steady uniform updraft, however, the problem has been found tractable.

Three areas of the cloud and its surroundings (the subcloud region of clear convection, the cloudy region, and the clear stationary cloud environment) have been analyzed and coupled at their common boundaries to determine the steady state operating point for the generating system. The operating point determines the self-consistent boundary electric field and the consequent vertical profiles of electrical parameters in terms of the initial values of convection-speed, net space charge and conductivity, and a parameterized load, the non-convecting environment of the cloud.

Ionization and Thomson recombination are included in the charge conservation equations in both the clear and cloudy regions. Additional terms are included in the cloudy region accounting for charge exchange between the droplets and ions according to Wilson's ion-scavenging mechanism. 234

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Three distinct modes of behaviour of the cloudy region have been discovered and the transition values of the initial conditions determined. Two modes are asymptotic solutions while the third is a non-equilibrium solution leading to the development of large fields in the cloud. Exemples of the electrical parameters for the three modes against distance from cloud base are shown below in order of their appearance with increasing system current density. The normalized variables shown are the electric field, E', the droplet charge, q', the ionic space charge,  $\rho'$ , and the conductivity,  $\rho_{\rm T}$ '. Additional studies detailing the transformation of ionic space charge into drop charge and rough comparison of the model results with existing data have been carried out.









The effect of electric charge on the freezing temperature of the drop

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On the freezing of the water drop in consideration of the electrical effect, many investigator had made the experiments by the various methods. The author made the experiments on the freezing of the charged drop.

In a cold box, a charge of order of  $10^{-1}$ esu. was given to a water drop which was suspended and insulated from the box. The size of the drop was ranged from 1.9 to 2.2 mm. The seeding of freezing nuclei by AgI was made in the box. And the concentration of the nuclei was measured by the sugar solution method. The concentration was of order of 10 per liter.

When the seeding was not made, the charged drop and the uncharged drop were frozen at  $-17^{\circ}$ C on the average. It is found that the electric charge alone is not effective on the freezing temperature. While the freezing temperature of the charged drop rose up about 5 to  $10^{\circ}$ C compared with that of the uncharged drop, When seeded. Some of the charged drops froze at near  $0^{\circ}$ C as shown in Fig. 1.

Concerning with the result, it is considered that the charged drop collect the freezing nuclei at high collection efficiency by the image force, although the nuclei are not charged. The effect of image force is effective when the size of freezing nucleus is small. It is considered from this experiment that the charged droplets in a cloud collect aerosol particles in a higher efficiency, then they freeze at a temperature higher than the uncharged droplets. If the surface nucleation by some part of aerosol occurs on the cloud droplets as the report by Gohkale and Goold, the freezing temperature of the cloud droplets will rise up considerably.

# Reference

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# ELECTRIC CHARGE OF SMALL PARTICLES $(1\mu - 40\mu)$

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The electric charge measurement of this size range is relatively missing although there are extensive observations of the electric charge of raindrops, snow crystals and ion range.

The electric charge of small particles (above  $1 \times 10^{-6}$ e.s.u.) was measured by the impactor method (first stage,  $40\mu - 10\mu$ ; second stage,  $10\mu - 1\mu$ ) and continuous observations were carried out at various stations and in various weather conditions in Hawaii for half a year.

In fine weather, positively charged particles were predominant. Charged particles increased in number with strong wind (20 miles/hour). Negatively charged particles were frequently observed under conditions of high humidity (80%).

Strongly electrified, negatively charged particles were observed during rain showers. The space charge due to the first stage particles was higher than that of the space charge due to other drop sizes. Negatively charged particles were also observed at the "apparent shower" where the wind record indicated changes characteristic of the passage of rain showers, even though no precipitation was measured.

The electric charge of cloud droplets was observed in cloud along Mauna Loa-Mauna Kea saddle road by this impactor method. Positively charged cloud droplets were predominant at the upper part 238

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of warm cloud and negatively charged cloud droplets were predominant at the lower part of warm clouds.

The surface potential of condensing drops may be an important factor to determine the charge separation of droplets.



# Three Stages of Massive Fragmentation of

Hydrometeors and Electrification in the Atmosphere

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Photomicroscopic investigation revealed three stages of massive fragmentation of hydrometeors:

1. A large percentage of the total mass of water in the form of numerous microdroplets, ranging from less than a micron to 20 microns in diameter, was separated from a freezing supercooled water drop (Fig. 1). The duration of generation of these microdroplets was 50 seconds for a 1 mm water drop.

2. Ice crystal fragments were produced by sublimation, governed by the curvature effect, from the frosty surface of a frozen particle (Fig. 2), such as, an ice pellet, hail or graupel. High concentrations of ice fragments were observed near two temperature zones:  $-5^{\circ}$  C and  $-15^{\circ}$  C.

3. Microdroplets were ejected by the bursting of air bubbles released from a melting ice pellet or hail. More than 500 air bubbles with an average diameter of 50 microns were observed to be released from a half-melted 1 mm ice pellet (Fig. 3).

High electrical charges carried by these fragments were measured, and their polarities were determined by the temperature gradient within each hydrometeor in accordance with the thermoelectric effect.

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This observed microscopic phenomenon of the fragmentation of hydrometeors with their accompanying electrical charges suggests a possible mechanism of generation of electricity in the atmosphere.







Fig. 2



Fig. 3

The Electrification of a T.N.T. Explosion Cloud

# by: D. R. Lane-Smith University of Western Ontario, Canada

The explosion of 500 tons of T.N.T. in July 1970 at Suffield, Alberta, Canada, produced a cloud which rose to 8,000 ft. in 2 minutes.

16 mm and 35 mm cameras at different sites recorded the cloud growth to give reliable figures for the cloud height and position. Field mills in the ground at positions ranging from 800 m to 6 km from ground zero recorded the changing pattern of electric field as the cloud became electrically charged. Later, the cloud drifted down the line of field mills depositing a fine dust as it went.

Numerical analysis of the results clearly shows a) the development of an electric dipole in the cloud with charge magnitudes of the order of  $10^{-1}$  coulomb b) the screening of that dipole by conduction with a relaxation time of the order of 5 mins., and c) the later precipitation of negatively charged dust in a vertical column about 1 km in diameter of very low conductivity surrounded by a cylindrical screening charge.

Some ideas regarding the physical processes occurring to produce the phenomena observed and the magnitudes measured will be put forward.

The presentation may be introduced by a 16 mm film lasting not more than 4 minutes.

# MEASUREMENT OF ELECTRIC FIELD CHANGES DUE TO INTRA-CLOUD LIGHTNING FLASHES FROM THE TROPICS OVER THE INDIAN REGION

By M.V. Sivaramakrishnan Meteorological Dept.,India

The point discharge apparatus used at Poona Meteorological Office affords a very convenient means of recording the rapid fluctuations of field which occur in thunderstorms. Some good examples of these fluctuations are shown in suitable figures in the The effect of lightning discharge in the neighbourhood is to paper. cause a sudden change in the point discharge current followed by a more or less gradual recovery. It is seen that for intra-cloud lightning flashes at less than 5 km distance the potential gradient changes as shown by the point discharge record are found to show an increase while for flashes at distances greater than 7 km to 10 km the potential gradient changes are found to be negative. At intermediate distance between 5 to 7 km the potential gradient changes show both positive and negative values. These are also found by various workers before, like Malan and Schonland (1950), Wilson (1916, 1920), Wormell (1939), Whipple and Scrase (1936), Tamura (1954).

The time interval between lightning phenomena and thunder heard multiplied by the velocity of sound had been used to give the distance of the thunderstorm from the point of observation.

At Poona (1967) which is representative of the Tropics there are more intra-cloud discharges than discharges from cloud to earth. During the past three years of study 1966-68, no discharge from cloud to earth (ground) has been noticed near the observatory site at Poona but lightning strike reports over the whole of Poona district shows an average of 6.6 flashes to earth per year based upon the disaster reports received at Poona observatory from 1962-1966 (33 reports). An Intra-cloud lightning flash generally occurs every 31 secs. at Poona (tropics).

# OBSERVATIONAL DATA ON THE CHARACTERISTICS OF PRECIPITATING CLOUDS IN THE TROPICS OVER THE INDIAN REGION

By M.V. Sivaramakrishnan, Meteorological Dept., India

Three parameters which will have an important bearing on which particular mechanism is likely to be dominant in the release of precipitation from a particular cloud type are the temperature of the cloud base, the cloud thickness and the most important the temperature of the cloud summit. The base temperature will largely control the liquid water content of the cloud which, together with the cloud thickness, will largely determine whether or not rain drops may be produced by coalescence, while the cloud summit temperature will indicate the likely presence or absence of ice crystals. These three parameters can be readily measured from Tephigrams prepared from radiosonde ascents. Unfortunately, reliable observational data of this kind are rather scanty in the tropics, but those which are available now will be summarised as follows for the following types of clouds Stratiform clouds (code figures 4, 5, 7), Cumulonimbus observed; clouds, the summits not in the form of an anvil (code figure 3), Cumulonimbus clouds with the summits in the form of an anvil (code figure 9). Clouds of thickness less than the height of the cloud base produce no precipitation, while clouds of thickness greater than the base plus 3,500 ft. or greater precipitated. The reports of rain from tropical clouds during monsoon season and thunderstorm seasons during the years 1969, 1970 are also included in the paper giving the location of the stations, cloud base temperature, cloud top temperature, cloud thickness and remarks on rain.

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## THE ELECTRICAL BALANCE SHEET OF EARTH FROM THE TROPICS BASED ON THE EVALUATION OF PRECIPITATION AND CLOUD ELECTRICITY

By M.V. Sivaramakrishnan, Meteorological Dept., India

If the negative charge on the earth is to be maintained, there must be for the earth as a whole a balance between the different processes bringing charge to the earth. The four known processes by which charge can reach the earth are (1) air earth conduction and convection currents, (2) point discharge currents, (3) precipitation currents and (4) lightning discharges. As no attempt has been made in India so far about the measurement of the electrical balance sheet of the earth, this paper describes the measurements made at Poona Meteorological Office for the electrical balance sheet of the earth by making continuous measurements of air-earth conduction currents, point discharge currents, precipitation currents and lightning discharge currents for the years 1966, 1967.

For the continuous measurement of the positive and negative component of the conduction current two pieces of duraluminium sheets about 1 metre square separated by a distance of 20 cms are connected to the two insulated halves of a quadrant electrometer, the deflections being recorded photographically. Full description of the method has already been given in another paper submitted for the Conference. For the measurement of the rain current and intensity of rainfall, a rain electrograph using a quadrant electrometer and tilting type raingauge has been used. For the measurement of point discharge current a sensitive mool galvanometer and a freely exposed insulated platinum point is used. For the measurement of the electrical field a radioactive collector has been used.

Since cloud to earth lightning discharges are rare in the

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tropics, especially at Poona, the charge brought down by lightning to earth is negligible. The number of disasters due to lightning strikes in Poona District for the past six years (1962 to 1966) has been noted to get an idea of the number of lightning discharges to earth.

Wormell (1930) first discussed the possibility of working out such a balance for a small area of earth  $(1 \text{ km}^2)$  and used available data to give results for Cambridge. Since then more accurate data have become available for some places and attempts have been made to determine the balance sheet for the earth as a whole using the various data.

# MONSOON DYNAMICS AS INFERRED FROM RADIATIVE EQUILIBRIUM TEMPERATURES

BY

R. R. KELKAR (Met. Dept. India)

# Abstract

Vertical distributions of radiative equilibrium temperature have been simulated along the 80°E meridian for four representative months, and compared with the temperatures actually observed. The configuration of heat sources and sinks so obtained lends support to Koteswaram's theory of the genesis of the Indian summer monsoon.

# Meteorological Parameters Effecting the Success of Warm Fog Dissipation by Hygroscopic Particle Seeding

by

# Alan I. Weinstein

Air Force Cambridge Research Laboratories Bedford, Mass.

A numerical model of the evolution of warm fog in response to hygroscopic particle seeding is used to determine the effects of several meteorological parameters on the success of airborne seeding for clearing warm fog.

The model is a two dimensional (height & downwind distance), time dependent, Eulerian solution of equations simulating 1) condensation on the hygroscopic particles and evaporation of the fog droplets, 2) turbulent diffusion, 3) horizontal advection and 4) drop fallout. Details of the model structure will be presented in order to allow realistic assessments of the validity of the study conclusions.

Among the parameters found to have important effects on the fog clearing capability of urea are wind shear, turbulence, and fog liquid water content and droplet size distribution. Whereas it had been previously shown that the first two were important for fog with relatively small droplets (& thus extremely low initial visibility), the new calculations show how important these parameters are in less dense and consequently more frequent fog conditions.

Where applicable, the details of the time evolution of particle size, liquid water content, relative humidity, visibility and other parameters will be presented in order to better understand interaction of the various processes simulated in the model and the bearing of these interactions on the conclusions of the study. Thermal Warm Fog Dissipation by Controlled Merging of Heat Plumes

## Bruce A. Kunkel Bernard A. Silverman

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A thermal warm fog dissipation system for airports that is efficient, meets air quality standards, and is safe for aircraft operations is under development. The system is based on the controlled merging of heat plumes from a strategically-placed array of burners to produce a uniformly warmed mass of clear air over the airport runway. The required heat is generated by the combustion of clean burning fuels. An interactive program of model computations and field experiments to determine the burner spacings and burner intensities that are required to achieve fog clearing with this system under various combinations of wind, stability, and fog conditions is described.

The numerical model simulates the trajectories and interaction of the heat plumes, and their effect on the fog environment, by obtaining time-dependent solutions of the primitive equations of motion, the equation of mass continuity, the first law of thermodynamics, and the continuity equations for water vapor and fog droplets. The model is framed in an Eulerian, three-dimensional system of coordinates.

Field tests of a pilot-scale configuration of the thermal fog dissipation system will be conducted at Vandenberg AFB, California during July 1972 to verify and improve the predictability of the model and to demonstrate the feasibility of the controlled merging heat plune concept. The heating system will consist of an array of liquid propane burners that are arranged in four lines perpendicular to the prevailing wind. It is designed to clear the fog in a cross-sectional area normal to the wind that is 400 feet long and 200 feet high. A systematic series of tests will be conducted to determine the effects produced by single and multiple line sources of heat of various intensities over a range of wind and stability conditions. The effects of the heat on the fog environment will be monitored by meteorological instrumentation mounted on a 200 foot tower that is located downwind of the burner array. Measurements will include visibility, liquid water content, drop size, wind direction, the three components of wind speed, temperature and dew point. In addition, a lidar and an accoustic radar will be used to obtain two-dimensional cross-sections of the visibility and temperature structure of the affected volume as a function of time. Results of the field tests and their comparison with the model predictions will be presented.

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# ON EXTINGUISHING FOREST FIRES BY ARTIFICIAL PRECIPITATION Yu.P.Sumin, D.D.Stalevich, T.S.Uchevatkina, N.S.Shishkin

## Main Geophysical Observatory, Leningrad, U.S.S.R.

1. At the A.I.Voeikov Main Geophysical Observatory there has been developed a method for artificial initiation of precipitation with the help of powder cartridges with lead iodide that are shot into the supercooled part of a cloud. The agent becomes active at the temperature -7°C and lower, at -10°C it gives 10<sup>12</sup> active particles per gram. The method allows us to initiate precipitation reaching the surface with cloud thickness of 2.2 km and more.

2. The method has found practical application in extinguishing forest fires by artificially initiated precipitation in sparsely populated regions of Siberia and Far East. Work on extinguishing forest fire using this method was begun in 1968. Weather conditions in the above-mentioned regions prove to be favourable for initiating precipitation from convective clouds during 40 to 50% of the days with fire danger.

3. In 1970 artificially initiated precipitation put out (with the following extinguishing of remaining locations with fire by surface means) 36 forest fires over an area of 15.9 thousand ha; and in 1971 35 forest fires over an area of 103.2 thousand ha. The optimal norms of agent release have been worked out and tested for artificial initiation of precipitation.

#### ITS USEFULNESS AND LIMITATIONS

Robert I. Sax, HQ Air Weather Service, Scott AFB, Illinois

The results and conclusions drawn by Sax and Cress (1971) from Project COLD RAIN, a joint Air Weather Service-Bureau of Reclamation rain-augmentation program conducted in Texas during June of 1971, serve to point out the usefulness and limitations of convective cloud modification as an operational tool. When the flow of tropical maritime air provided copious moisture through an atmospheric layer deep enough to sustain a strong coupling between turret and root, the selective dynamical seeding of supercooled convective clouds was found to result in turret growths of up to three kilometers in the vertical in the first fifteen minutes following seeding. More importantly, the horizontal dimensions of the seeded towers were found to increase significantly enough to make it possible to create a sub-mesoscale convective cloud mass by merging together several of the seeded elements. It is not unreasonable to expect that the resulting larger and deeper seeded cloud mass will have a much longer lifetime and be so much more efficient at processing the available moisture that rainfall will be increased over what would occur from a series of relatively shallow, poorly organized, unseeded clouds. The concept of seeding to artificially induce cloud mergers was used operationally by St. Amand et al. (1971) in the Philippine Islands and by Woodley et al. (1971) in Florida. The COLD RAIN results suggest the exportability of such a technique to more continental locations.

Dynamical seeding is, however, a highly selective procedure and not all clouds can be expected to respond in an identical manner. For optimum results a stable layer, strong enough to cap natural cloud growth but sufficiently weak to enable penetration by the more buoyant seeded clouds, should exist somewhere between the -5C and -15C isotherm levels. Also, the atmosphere must be moist enough to prevent entrainment from separating the seeded tower from its cloud body. During the first week of the COLD RAIN operations, the synoptic conditions were such as to suppress convection throughout the entire target area.

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Those clouds which managed to grow to a "seedable" height possessed high bases and contained little liquid water and very weak updrafts. It was not until upper level (500 mb) troughing triggered convective activity that seeding was able to enhance localized rainfall within widespread portions of the target area.

Therefore, the paradox of dynamical cloud seeding is that the only occasions suitable for its <u>widespread</u> operational implementation are those when synoptic conditions favor convection. The use of such a seeding technique to alleviate already occurring severe drought, the common cause of which is largescale subsidence, is very obviously going to fall short of its goal. However, a program of convective cloud seeding during normally wet seasons of the year, combined with effective watershed management, could offer great relief from periodic water shortages which occur in certain regions of the world.

Even during predominantly dry conditions a dynamical seeding program could provide beneficial rainfall over a specified, very limited area. Opportunities for useful application of the technique will be rare, but it has been shown from the operations in Florida (Woodley et al., 1971) that it is possible, <u>on a localized basis</u>, to extinguish brush fires and to halt temporarily the fall of the water table. Although the total rainfall gained from an operationally oriented cloud modification program is just a proverbial "drop in the bucket" compared to what nature can produce on a disturbed day, depending upon the value placed on the additional water such a convective seeding program can still prove to be cost effective.

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### ARCTIC FOG DROP SIZE DISTRIBUTION AND THE LIGHT ATTENUATION

#### Motoi Kumai

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The sampling and analysis of arctic fog at Barrow, Alaska, during summer of 1971 is described. The object of the research was to obtain the' mean concentration and the size distribution of fog droplets between a laser site and its targets, and from these data to calculate the attenuation coefficient for wavelengths of 0.571  $\mu$  and 1.06  $\mu$ .

For the first test at Barrow we used a 100-cm<sup>3</sup> two-stage impactor containing slides precoated with silicone oil or gelatin to collect fog drops or replicas. Since a thousand drops were needed to determine size distribution over the desired distances, the two-stage impactor was too inefficient for use in a fog of low concentration.

During preliminary sampling trials a good collection of fog drops was found on the upwind side of the shaft of the anemometer being used to measure the wind speed. This led to the adoption of the following method. A slide precoated with a gelatin film was cut into 5-mm-wide, 30-mm-long strips which were attached to the upwind side of the anemometer shaft. The volume of sampled air containing fog drops and the length of the airstream sampled were calculated from the exposure time of the film and the wind speed. The fog drop collection efficiency was determined for different wind speeds, drop radii, the density and viscosity of the air, and the film width. A gelatin reagent film containing colloidally dispersed red silver dichromate was used for drop replication and also for chloride identification of sea salt nuclei larger than  $10^{-12}$ g. A droplet impacting on the gelatin film dissolved some of the gelatin and a trace of the drop was left after the drop evaporated. The evaporation of a water drop on the film was observed under an optical microscope. The diameter of the drop on the film was found to coincide with the diameter of the replica. The contact angle of the drops on this film, measured from photomicrographs of side views of the drops, was found to average 35°. The diameter of the drop before contact with the film was calculated to be one-half the diameter of the replica at the contact angle of 35.7°.

Optical attenuation coefficients were computed for the observed Barrow fog for optical wavelengths of 0.571  $\mu$  and 1.06  $\mu$  using the Mie theory. The attenuation coefficients for 0.571  $\mu$  wavelength were found to be smaller than those for 1.06  $\mu$  wavelength. The values of the visual range calculated at a threshold contrast of 5% were closer to the observed visibility than those at a threshold contrast of 2%.

## MICROPHYSICAL STRUCTURE OF WARM FOG

by

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During the months of August through October 1971, observations were made in warm fogs at the Arcata-Eureka Airport, Humboldt County, California, USA. The airport is adjacent to the coastline and therefore has a high frequency of warm fogs. The purpose of the observations was to define the microphysical structure common to the fogs.

The observations consisted of data taken at the surface and data taken from an instrumented aircraft. At the ground, high resolution temperature, dew-point, and wind data in the vertical were obtained from a GMD-1 radiosonde unit. Drop-size distribution data were obtained by capturing settling fog droplets on hand-held slides coated with gelatine. The liquid water content at the surface was computed from these data.

The airborne observations consisted of temperature, dew-point, and altitude data. A novel optical cloud particle spectrometer was employed to determine the drop-size distribution at different levels through the depth of the fog. The liquid water contents corresponding to these data were computed. This instrument continuously samples fog droplets and computes the size distribution at five second intervals (325 m intervals). This small length of interval demonstrates the spacial variations of the distribution in exceeding detail. Examples of these data will be presented.

The vertical wind data were used to estimate variation of wind speed through the depth of the fog (shear) and the variation of wind direction through the depth (turning). This information will define the structure of the flow within the layer of fog. Shear deforms the flow in the vertical plane and turning deforms the flow in the horizontal plane. The preliminary results from analysis of the data have shown the average shear to be  $0.0100 \pm 0.003 \text{ sec}^{-1}$  and the average turning to be  $63 \pm 30$  deg. The average shear result suggests a small amount of deformation and the turning results suggest a moderate amount of deformation.

The drop-size distribution data at the surface were incorporated with the similar data obtained from the aircraft through the depth of the fog. This procedure produced data on the vertical structure of the drop-size distribution and the liquid water content. The preliminary results from the liquid water content data illustrate that the highest liquid water content value occurs at the top of the fog and the lowest value occurs at the bottom. A typical variation is from 0.42 g m<sup>-3</sup> at the top to 0.18 g m<sup>-3</sup> at the bottom. The preliminary results from the drop-size data show a similar median volume diameter at the top and bottom of the fog (~ 22 µm). An explanation of these curious results will be presented. The results should be useful inputs to the microphysical fog models presently used to simulate the improvement in visibility in fog through artificial means.

# OPTICAL EXTINCTION PROPERTIES OF ICE FOG AT 6328A

R. H. Munis and A. Delaney

An experimental investigation was conducted on the extinction properties of ice fog at 6328A using a HeNe laser. Ice fog was generated in a 4 meter chamber at -40°C by injecting a spray of steam into the chamber. Particle sampling was carried out using an impactor while the laser beam attenuation was continuously recorded. Immediately upon sampling the fog, the glass slide was removed from the impactor, placed under a microscope and photographed. These photographs were used to generate particle size spectra. The experimental extinction coefficient was then calculated at each wavelength and for each particle concentration using the classical exponential law of absorption. Theoretical extinction coefficients were obtained by using the measured particle spectra in the classical expression for Mie scattering. The theoretical transmission was then calculated using the Mie extinction coefficients. Comparison of theoretical and experimental transmission coefficients indicates that for low particle concentrations the agreement between the two is fairly good. However, as the particle concentration increases the disagreement also widens. This probably can be attributed to the fact that as the optical depth through the fog is increased there is an onset of multiple scattering. Since the Mie scattering equation is only valid for the case of single scattering when the particle concentration reaches a certain value Mie theory will no longer provide a valid comparison with experimental results. The results presented here indicate that the divergence between theory and experiment seems to occur at rather low particle concentrations.

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## A PILOT FIELD INVESTIGATION OF RADIATION FOG

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This paper describes observations made at Cardington, Bedford of the meteorological and structural properties of a radiation fog, obtained during the first phase of a continuing project designed to study the physics of fogs.

The fog began to form at about 0400 GMT and dispersed rapidly at about 1030 GMT on 7 Dec. 1971. The analysis of the results is not yet complete, but a preliminary assessment is possible.

#### Meteorological

The fog appeared to pass through two principal phases. In the first (lasting until about 0630 GMT) a strong inversion developed in the lowest 10 m. The fog thinned rapidly with height and its top was generally ill defined.

Radiative cooling rates were  $0.5 - 1^{\circ}$ C/hr in the 9-37 m (upper) layer and about 2°C/hr in the 2-9 m (lower) layer.

Between about 0630 and 0730, the ground inversion lifted to 15-20 m above ground, carrying with it a now well-defined fog top. Radiative cooling rates increased to  $2-3^{\circ}$ C/hr in the upper layer and persisted as such until the fog dispersed. In the lower layer, radiative cooling rates of nearly  $5^{\circ}$ C/hr persisted for about an hour, but rapidly decreased after sunrise (at 0755 GMT) - becoming positive at about 0930 GMT.

Observed cooling rates were in general less than the radiative cooling rates, especially in the second phase of the fog when discrepancies of  $3-5^{\circ}$ C/hr were observed.

Wind speeds in the fog were  $1 \stackrel{+}{=} 0.5 \text{m sec}^{-1}$  throughout the period and gradient Richardson numbers in general > 0.25.

There was evidence of oscillations of a predominant period of 10-12 mins in the wind, temperature and radiation records.
# ZERO-GRAVITY CLOUD PHYSICS

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Although gravity plays a dominant role in the physics of clouds, there are microphysics experiments for which one would like to turn off gravity. For these experiments, motion of droplets or ice crystals due to gravity restricts observation time, whereas vertical wind tunnels possess other limitations, and artificial suspension of droplets, etc. by electrical, acoustical or mechanical means introduces unacceptable effects. In zero-gravity there is no such suspension problem, but, of course, the aerodynamic effects due to the falling motion of the droplets are lacking. However, for certain experiments these aerodynamic effects can be accounted for separately.

An experiment definition and feasibility study is underway for NASA by McDonnell Douglas in order to establish the scientific merits and goals of a zero-gravity cloud physics experiment capability on a manned earth orbiting platform. The study has two major objectives as follows: a) solicitation, discussion and study of cloud physics experiments that can be done under zero-gravity conditions and to evaluate approaches to specific problems envisioned in these experi-

ments; b) development of preliminary design concepts for a zero-g laboratory including the cloud chamber and supporting apparatus and equipment.

The original solicitation was oriented primarily to universities, government laboratories and private meteorological firms in the United States. The solicitation will continue to expand into other countries involved in cloud physics research. Experiment suggestions were received from scientists at the agencies that are involved in cloud physics laboratory research.

The present study involves consultant cloud physicists in the form of an advisory board, and as members of a team which will suggest and evaluate approaches to specific problems envisioned in the various experiments. Selection of the experiments from the list of suggestions was made by a panel of cloud physicists.

The selection criteria for the cloud physics experiments was 1) experiments must be relevant to cloud behavior and weather modification, 2) have clear cut scientific merit, and 3) require zero-gravity for their execution.

This paper presents a discussion of the experiments that were identified as meeting the criteria and being potential candidates for space flight opportunities. The paper also includes a discussion of the special problems associated with zero-gravity cloud physics research and the approaches planned for the solution of these problems. The Use of Chemical and Biological Tracers in Cloud Physics

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### Introduction

Laboratory and field experiments using tracer materials are being conducted to infer cloud processes. The laboratory studies concern use of biological spores as an aerosol for the determination of collection efficiencies of various drop sizes. The field experiments utilize chemical tracers released into convective storms to estimate the scavenging efficiency of clouds.

## Laboratory Experiments

The spore <u>Bacillus subtilus</u> is used to represent an aerosol of uniform size equivalent to 1  $\mu$ m diameter. A dilute aqueous solution is prepared and injected into a closed 1.5 m<sup>3</sup> chamber. The chamber forms the bottom section of a tower 12.2 m high and 1 m diameter.

Drops are formed at various levels in the tower and allowed to fall toward the aerosol-filled chamber. Entrance and exit ports are opened and closed to permit a drop to be exposed to the aerosol during the final 2 m of the fall. The drop is then caught on a petri dish containing an appropriate nutrient for the growth of spore colonies.

It was found that the technique was quite repeatable and relatively simple. The aerosol concentration was monitored throughout a set of experiments to determine the viability of the spores. The number of spores collected by an individual drop were easily counted under low-power magnification. The results from a number of experiments are shown in Fig. 1.

It is readily noted that the collection efficiency is rather insignificant for the drop sizes investigated. The addition of electrical charge to the simulated raindrops increases the collection efficiency dramatically, but the results still reflect the insignificance of the removal of aerosol by washout. Fig. 2 illustrates the increase of collection efficiency as a result of the addition of electrical charge on the drops.

# Field Experiments

A trace element has been used to study the rainout of atmospheric aerosol by convective storms. A network of 60 total water collectors were installed in the downwind area of the city of St. Louis, Mo. This network

formed a subset of a larger recording raingage network surrounding the entire St. Louis urban-industrial complex.

An aerosol of lithium was produced by burning a solution of acetone and lithium perchlorate. A nominal particle size of 0.4  $\mu$ m of lithium chloride is generated in this way and released from an aircraft flying in the updraft region of a selected storm.

The "treated" storm is tracked across the sampling network by the aircraft and 10-cm radar. The precipitation samples are subsequently collected and chemically analyzed. The deposition of the tracer chemical is computed for the storm. An example is shown in Fig. 3.

The results from such experiments illustrate the interaction between the dynamics of various storms which compose a thunderstorm line. The tracer material does not deposit entirely beneath the treated cell, but is horizontally mixed into adjacent cells and deposited over a rather broad area. The implications of these experiments to cloud seeding research are evident.







Figure 3. Lithium deposition on August 14, in pg/cm<sup>2</sup>



Figure 2. The effects of drop electrical charge on the collection efficiency of the spore aerosol

The Time Lapse of the Size Distribution of Raindrops in Various Storms - A Movie

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The time lapse of the size distribution of raindrops in various storms are shown in a movie. The movie is photographed from a computerized CRT display.

The size distributions are measured with a ground-based electro-mechanical disdrometer (Joss and Waldvogel, 1967). A small digital computer processes the disdrometer data and produces a CRT display of the time lapse of the distribution. Simultaneously, it overlays this animated display with a steady picture of the evolution of some of the parameters of the distribution during the whole storm.

A keyboard monitor allows for various quick changes in the presentation of the data on the CRT display, such as slow and quick motion, choice of different parameters and scales, etc. Thus, the user gets an excellent view of the drop-size data of a whole storm without loosing the possibility to study simultaneously the detailed features of an individual size distribution. For example, effects of drop sorting due to wind and/or due to different fall velocities of the drops can be investigated.

A sample frame of the movie is shown in Fig. 1. On this frame a single size distribution of raindrops, which has been measured during 60 sec, is displayed together with the time lapse of the parameters W, Z and  $N_0$  during the whole storm. W is the liquid water content, Z the radar reflectivity factor and  $N_0$  a parameter of the exponentially approximated drop-size distribution. An arrow indicates the time at which the distribution has been measured.

### Reference:

Joss J. and Waldvogel A., 1967, "Ein Spektrograph für Niederschlagstropfen mit automatischer Auswertung" <u>Pure and Applied Geophysics</u> Vol. 68, 240-246



D [mm]

Fig. 1 A sample frame of the movie, showing a histogram of a size distribution N (D) of the rain drops, and the time lapse of the liquid water content W (t), the radar reflectivity factor Z (t) and the distribution parameter N<sub>0</sub> (t). The arrow indicates the time at which the displayed distribution has been measured (during 60 sec).

Compared values in the upper atmosphere of saturation vapor pressure over ice and barometric pressure in relation with the occurrence of natural and artificial noctilucent clouds.

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According to the hypothesis that noctilucent clouds (NLC) are explained as consisting of ice crystals formed on dust nuclei of cosmic origin, the occurrence of these clouds implies that saturation vapor pressure over ice does not exceed ambient pressure at the cloud level. Considering the atmospheric models up to 110 kilometers (U.S. Standard Atmosphere Supplements, 1966), it thus appears that some layers are "regions of exclusion" for water drops and ice particles; curves on Fig.1 and 2 point out these regions of exclusion for  $60^{\circ}$ N latitude where NLC are generally observed; the occurrence of NLC well agree with these curves : NLC are only observed between March and October, best in June through August, and the altitude determinations vary between 74 - 92 km.

This agreement strengthens the ice crystals hypothesis, and indicates the possibility of observing NLC in other latitudes. In France, the occurrence of NLC is quite possible in July, and not impossible in winter inside a thin layer; this is in accordance with the display between 75 and 80 km of an artificial NLC created on February 23 rd, 1971 by a rocket releasing water vapor between 150 and 60 km. As this rocket released A1 and A1<sub>2</sub>0<sub>3</sub> particles but also hygroscopic material, ice crystals may have formed by the direct sublimation of water vapor on nuclei, but more probably, as seen on photographs, by condensation and freezing at a temperature of about  $-40^{\circ}$ C.

The most surprising feature of this french artificial NLC was its optical density and horizontal extent, suggesting that man-made clouds in the upper atmosphere might be considered.



Fig 1. Pressure P and saturation vapor pressure e<sub>i</sub> over ice (over water for 0 - 5 km), 60°N. January.

Fig 2. Pressure P and saturation vapor pressure e<sub>i</sub> over ice (over water for 50 - 55 km), 60°N. July.

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