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Weather Modification by Cloud Seeding

Arnett S. Dennis

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Contents

Preface

It is an understatement to say that people are confused about cloud seeding. While it has been called "the crime of the century" and outlawed in Pennsylvania, the governments of the dry, western part of the United States continue to spend tax revenues on cloud seeding to increase water supplies.

During the past five years, I have talked with officials responsible for decisions regarding cloud seeding programs in about 15 states of the U.S.A., in a dozen other countries, and in the World Meteorological Organization. Some of the officials involved are experts in atmospheric physics, but most are not. Members of the latter group have expressed a great need for reliable, concise information about how cloud seeding works and what it does to the weather. This book is addressed to them and to all students of the environmental sciences in general and of the atmospheric sciences in particular who need answers to those questions.

The book should be useful to graduates of a university program in any of the physical sciences or engineering, and to upperclassmen in those subjects. A review of basic cloud physics is provided to make the presentation meaningful to persons with only minimal preparation in meteorology. The presentation of statistical methods used in the evaluation of cloud seeding projects presupposes no special knowledge of statistics on the part of the reader.

The subject of weather modification has been studied by meteorologists, physicists, chemists, statisticians, lawyers, ecologists, economists, and sociologists, each from a different point of view. Disagreements among these groups sometimes obscure the larger issues. For example, the debate that followed the publication of the Final Report of the Advisory Committee on Weather Control in 1957 was dominated by statisticians. The real question of whether or not the Advisory Committee was correct in concluding that cloud seeding had increased precipitation in mountains of the western U.S.A. was obscured by debate over such details as the relative merits of square root and gamma transformations for normalizing rainfall statistics.

This book attempts to present a general view of the subject but with the emphasis on physical rather than legal, economic, or sociological aspects. The current state of the art is described, and some promising areas for additional research are pointed out. Underlying the entire book is the author's conviction that the evidence on the effects of cloud seeding can be sifted to yield a coherent picture that is consistent with the laws of atmospheric physics.

I am indebted to literally hundreds of people for the ideas expressed in this book. Many of these people are listed in the References, but some are not. To all of them I extend my thanks.

I am especially appreciative of useful discussions with Drs. Briant Davis, Paul Mielke, and Harold Orville regarding choice of material and diagrams. I also wish to thank Mr. Thomas Henderson and Professor R. R. Rogers for the photographs used in the book and Mr. Melvin Flannagan for drafting the figures.

The work of Miss Carol Vande Bossche in preparation of the manuscript deserves unqualified praise.

I am grateful to my wife Maralee Dennis for constant support and encouragement during the writing of this book.

List of Principal Symbols

B	Mobility; buoyancy
C	Capacity of an ice crystal
C_D	Drag coefficient
C'	Dimensionless constant used in studies of turbulence
D	Diffusivity of water vapor; generalized eddy diffusivity
D_B, D_1, D_2	Brownian diffusion coefficients
D_x, D_y, D_z	Coefficient of turbulent diffusion in x, y, and z directions
E	Collection efficiency; expected value of a function; turbulent energy distribution over wave numbers
E_1	Collision efficiency
E_2	Coalescence efficiency
E_γ	Energy of surface tension
E_i	Collection efficiency of a hailstone with respect to cloud ice
E_l	Collection efficiency of a hailstone with respect to cloud liquid water
$\mathbf{F}$	Drag force vector
F_v	Ventilation factor
H	Collection kernel
K	Coagulation kernel
K_B	Brownian coagulation kernel
K_*	Coagulation kernel for microscale turbulence
L	Latent heat
L_s	Latent heat of sublimation
L_v	Latent heat of vaporization
M_s	Molecular weight of a solute
M_w	Molecular weight of water
N	Number concentration; number of observations in a sample

N_a	Concentration of ice nuclei active at a given temperature or degree of supercooling
N_{Be}	Best Number
N_c	Cloud droplet concentration
N_0	A reference concentration
N_{Re}	Reynolds Number
P	Proportion of cases reserved as no-seed
Q	Source term
R	Gas constant; multiple correlation coefficient
R_w	Specific gas constant for water vapor
S	Saturation ratio with respect to water
S_{eq}	Equilibrium saturation ratio
S_i	Saturation ratio with respect to ice
T	Temperature (°K); test statistic
T_o	Reference temperature (°K)
T_v	Virtual Temperature
U	Wind speed
U, V	Generalized response variable in seeded and unseeded storms respectively
X	Control area rainfall
Y	Target area rainfall
Y_E	Estimated or predicted target area rainfall
a	Intercept in linear regression equation
a'	Empirical constant in Cunningham mobility formula; empirical constant in equation describing aerosol size distribution
b	Slope of regression line
d	Diameter of a particle
d_c	Critical diameter for a CCN
d_g	Diameter of cylinder below a drop which defines a grazing trajectory for a drop–droplet pair
$\bar{d}_G$	Geometric mean diameter
d_o	Dispersion of cloud droplet spectrum
e	Vapor pressure
e_i	Saturation vapor pressure over plane ice surface
e_s	Saturation vapor pressure over plane water surface
$e_{s,d}$	Saturation vapor pressure over liquid droplet of diameter d
e'_s	Saturation vapor pressure over a plane surface of a solution
$e'_{s,d}$	Saturation vapor pressure over a solution droplet of diameter d
g	Gravitational acceleration
i	Ionic dissociation factor
k	Boltzmann's constant; thermal conductivity
l	Molecular mean free path
m	Mass; number of seed cases in an experiment
m_B	Mass of a "bubble" in a cloud
m_s	Mass of solute in a droplet
n	Number of molecules of dissolved solute; number density function; number of no-seed cases in an experiment
n'	Unit distance normal to a streamline; number of water molecules in a solution
n_d	Raindrop number density function
n_o	Raindrop size distribution parameter

p	Pressure
q_1, q_2, q_3, q_4	Heat transfer terms for a hailstone
r	Radius of a particle; correlation coefficient
r_u	Radius of an updraft
s	Standard deviation
s_E	Standard error of estimate
t	Time
$\mathbf{u}$	Velocity vector
u_T	Terminal fall speed
v	Volume
w	Vertical speed of air; mixing ratio
x	Measure of position in horizontal plane
y	Measure of position in horizontal plane; radius of a turbulent plume
y_0	Initial radius of a turbulent plume
z	Height
α	Significance level
α'	Empirical constant introduced into cloud model to adjust entrainment
β	Coagulation factor; power of a statistical test
γ	Surface tension coefficient
γ'	Shape factor of a gamma distribution
Δ	Combination factor controlling power of an experiment
ε	Rate of turbulent energy dissipation
θ	Factor in empirical equation describing conversion of cloud water into rainwater; effect of seeding upon precipitation (assumed constant multiplier)
κ	Wave number (turbulence)
Λ	Raindrop size distribution parameter
μ	Dynamic viscosity
ν	Frequency of collisions in an aerosol
ρ	Density
ρ_a	Density of air
ρ_L	Density of liquid water
σ	Standard deviation
σ_G	Geometric standard deviation
σ_y, σ_z	Standard deviations of a Gaussian plume in horizontal and vertical displacements, respectively
τ	Noncentrality parameter
ϕ	Factor in empirical equation describing conversion of cloud water into rainwater
χ	Concentration
χ_l	Cloud (liquid) water concentration
χ_{l0}	Threshold of χ_l for autoconversion to rainwater
χ_I	Cloud ice concentration
χ_R	Rainwater concentration
Ω	Parameter relating activity of ice nuclei to temperature or degree of supercooling

Table of Physical Constants

Boltzmann's constant (k)	1.380×10^{-23} J K^{-1}
Gravitational acceleration (g) at sea level at 45° latitude	9.806 m s^{-2}
Latent heat of vaporization (L_v) of water at 0°C	2501 J g^{-1}
Latent heat of sublimation (L_s) of ice at 0°C	2835 J g^{-1}
Melting point of ice under pressure of 1 atm (101.325 kPa)	273.15 K (0.00°C)
Molecular weight of dry air	28.964
Molecular weight of silver iodide	234.77
Molecular weight of water (M_w)	18.015
Specific gas constant of dry air	287.0 J kg^{-1} K^{-1}
Specific gas constant of water vapor (R_w)	461.5 J kg^{-1} K^{-1}

CHAPTER

I

Introduction

1.1 GREAT EXPECTATIONS

Pioneering Cloud Seeding Flights

On 13 November 1946 Vincent J. Schaefer dropped about 1.5 kg of dry ice pellets (solid CO_2) from a light aircraft into a supercooled lenticular stratocumulus cloud near the Berkshire Mountains of western Massachusetts. Within about 5 minutes the cloud had turned into snowflakes, which subsequently penetrated about 600 m into the dry air below the base before subliming completely away (Langmuir *et al.*, 1948; Schaefer, 1953). With this spectacular achievement the ancient dream of weather control seemed to come within the grasp of mankind. In the long run 13 November 1946 may prove almost as fateful a day as 16 July 1945, when the first nuclear explosion was set off in New Mexico.

Schaefer's cloud seeding mission was not an isolated event. It was the culmination of half a century of research by brilliant scientists into the physics of clouds and precipitation. These scientists were mostly Europeans; the best known among them were perhaps A. Wegener, Tor Bergeron, and Walter Findeisen. Their studies gradually led them to realize that important atmospheric processes, including precipitation, sometimes occur or fail to occur because of the abundance or scarcity, respectively, of ice-forming nuclei in the atmosphere, and that the ice-forming nuclei might be supplied artificially. As early as 1932, the U.S.S.R. had established an Institute of Artificial Rain to consider the possibility of weather modification (Fedorov, 1974).

During World War II Findeisen served as a meteorologist with the German Luftwaffe. He reportedly made a cloud seeding flight in a Luftwaffe plane over German-occupied Czechoslovakia in 1942 (Schaefer, 1951). His seeding agent, sand, apparently proved ineffective as an ice nucleant. Findeisen disappeared during the closing days of the war and is presumed dead.

In the United States, World War II led to the formation of a group at the General Electric Research Laboratories in Schenectady, New York, headed by Irving Langmuir and dedicated to the study of such problems as the generation of smoke screens and ways to combat aircraft icing and radio static while flying through storms. The group continued its research after the war ended.

In 1946 Schaefer, a member of Langmuir's group, was conducting experiments on supercooled clouds in a cold box. In a hurry to cool the box to a temperature sufficiently low for his experiments, he dropped a pellet of dry ice into the cold box. Immediately a trail of tiny ice crystals appeared along the path of the piece of dry ice (Schaefer, 1946).

Schaefer quickly realized that the extremely low temperature near the surface of the dry ice pellet (−78°C) had caused the droplets along its path to freeze. We cannot say whether or not he also realized at that moment that the tremendous temporary supersaturation produced by the cooling had activated many aerosol particles as condensation nuclei, and that the droplets formed around them had also frozen. In any case, Schaefer was quick to see that he had accidentally discovered a way to glaciate supercooled clouds. He immediately laid plans to test the method in the free atmosphere, and did so with the results already noted.

Discovery of Silver Iodide as an Ice Nucleant

Despite the discovery of the glaciogenic (ice-forming) properties of dry ice, other scientists pressed on with the search for an artificial ice nucleant. Bernard Vonnegut (1947), also of the General Electric Research Laboratories, searched the chemical tables for solid substances with crystalline structures similar to that of ice. In particular, he studied the lattice constants, the measure of the spacing between atoms in the various crystals, reasoning that the inefficiency of most natural ice nuclei was due to the large misfit between the spacing of atoms in the surface layer of the nucleus and in the first layer of water molecules laid down in the ice structure. On the basis of his literature search, Vonnegut decided that silver iodide (AgI) was the most promising compound. In its hexagonal crystal form, its atoms assume an arrangement identical to the positioning of the oxygen atoms in ice, and the difference in spacing is small.

In cloud chamber experiments, Vonnegut found that AgI crystals acted as ice nuclei at temperatures as high as $-3°C$, which agreed quite well with his theoretical predictions. He immediately tackled the problem of generating large numbers of ice nuclei from a given quantity of AgI. He decided that the most efficient approach would be to vaporize the AgI and then quench the vapor, that is, to cool it suddenly, causing the AgI molecules to form very large numbers of minute solid particles. The first attempts yielded up to 10^{19} particles per kilogram of AgI, and cloud chamber tests showed some of them to be active as ice nuclei at temperatures up to around $-5°C$ (Vonnegut, 1947).

The laboratory tests were followed by tests on supercooled clouds in the free atmosphere using airborne generators. The results duplicated Schaefer's dry ice experiment, with holes being cut in decks of supercooled clouds and snowflakes observed falling from the cleared areas. Enthusiasm ran high. Some of the experimenters, notably Langmuir, spoke of the possibility of modifying the weather over the entire United States with only a few dollars worth of AgI.

Vonnegut's discovery of the ice nucleating properties of AgI particles made the modification of large cloud volumes economically feasible. The fact that the particles could be released from generators on the ground to seed promising cloud formations reduced the cost of operations remarkably. The cost of operating a generator on the ground was only \$2.00 to \$3.00 per hour in 1948, while the cost of operating an airplane was of the order of \$25.00 per hour.

Following the release of the covering patents by the General Electric Company numerous individuals and corporations went into business as cloud seeders. Their customers included associations of farmers and ranchers, utility and lumber companies, irrigation districts, and municipalities. By 1950 some 10% of the land surface of the United States was under contract to cloud seeder firms, and operations were under way in several foreign countries as well.

1.2 OPPOSITION AND CONTROVERSY

Although cloud seeding has spread all over the world since 1946, its acceptance is far from unanimous, as the following example shows.

On the afternoon of 7 December 1974 approximately 200 persons gathered in a church basement in the small town of Chamberlain, South Dakota. They were mostly people from farms and ranches who had gathered to voice their opposition to the weather modification program which had been instituted by action of the South Dakota Legislature in 1972. Two dry summers, 1973 and 1974, had eroded confidence on the part of the general public that the program was achieving its stated purposes of stimulating precipitation and simultaneously suppressing severe hailstorms.

The meeting was addressed by E. J. Workman, who had previously been engaged in weather modification research in New Mexico and had served as a member of a National Academy of Sciences panel reviewing weather modification (Panel, 1973). Describing himself as a "reformed cloud seeder," Dr. Workman spoke in opposition to the South Dakota program. He presented a theory which he had developed some 12 years earlier (Workman, 1962) that cloud seeding could lead to the formation of more but smaller raindrops, hence to increased evaporation between cloud base and the ground, and to a suppression of rainfall. Although several persons in the audience, including the present author, questioned whether this theory was applicable to South Dakota clouds and the methods of seeding being employed, Dr. Workman was unshaken.

The outcome of the meeting was the organization of a formal opposition body called Citizens Against Cloud Seeding, which subsequently staged opposition rallies throughout the state. Their pressure was so effective that no state funds were appropriated for support of cloud seeding in South Dakota beyond 30 June 1976.

The response of the Citizens Against Cloud Seeding in South Dakota to the state's weather modification program was relatively restrained, based principally upon the failure of the cloud seeding program to prevent drought in the summers of 1973 through 1975. Some of them thought that the program was merely ineffective, and therefore a waste of public money, while others argued that the cloud seeding, especially for hail suppression, actually reduced rainfall.

Opposition groups in other states, such as the Tri-State Natural Weather Association of St. Thomas, Pennsylvania, have blamed cloud seeding for a variety of calamities, including diseases of plants and animals. They have been successful in having legislation passed banning all weather modification activities in some states (Howell, 1965). Nevertheless, some members of such groups are convinced that clandestine cloud seeding is continuing in their areas. They scoff at denials by government officials. They sometimes attribute the cloud seeding to groups with sinister economic motives, such as forcing small farmers into bankruptcy, or to secret research projects of the military-industrial complex. The fact that the U.S. Defense Department finally revealed that cloud seeding had been used as part of the American arsenal in Vietnam, after refusing to comment on earlier reports to that effect, no doubt has contributed to the distrust of federal agencies with a potential stake in weather modification.

1.3 WHY CLOUD SEEDING?

Obviously we did not get from Schaefer's achievement of 13 November 1946 to the public indignation meeting at Chamberlain, South Dakota, on 7 December 1974 without a great number of intervening events. It is not our purpose to review those events, which have been covered from the points of view of a university scientist and of a cloud seeder by Byers (1974) and Elliott (1974) respectively. Rather, it is to provide reliable answers to the questions raised at Chamberlain (and elsewhere), and which evidently will not go away.

The conclusions presented in this book are based principally upon numerical cloud modeling studies and statistical analyses of field experiments, and so must be stated in probabilistic rather than absolute terms. Although uncertainties exist and will undoubtedly persist, they do not preclude the establishment of reasonable expectations concerning the effects of weather modification by cloud seeding. As Wallace Howell has put it, there is uncertainty when a person rolls a pair of dice, but we know that no one will ever roll a 13. Similarly, the laws of physics must establish limits on the weather modification effects that can be produced by cloud seeding. We therefore complete this introductory chapter by taking note of some features of the atmosphere which must dominate the results of any human attempts to modify the weather, and which explain why nearly all serious attempts to modify the weather to date have involved the seeding of clouds.

The Atmospheric Heat Engine

The earth's atmosphere functions as a giant heat engine. It uses energy from the sun to drive the winds and lift water vapor from the ocean surfaces to deposit it as rain or snow on the continents. Very little solar energy is absorbed directly by the atmosphere. Rather, the solar energy is absorbed at the earth's surface and then enters the lowest layers of the atmosphere as sensible heat or as latent heat associated with evaporation of water.

The bulk of the solar energy absorbed by the earth enters only the earth's radiation balance, offsetting the energy radiated away from the earth into space, either from the atmosphere or from the surface itself. Furthermore, only a small part of the solar energy which is relayed to the atmosphere ever appears as kinetic energy associated with the winds.

The energy of the winds is derived principally from the differential heating of different parts of the earth's surface. As we have noted, some solar energy is stored for a time as latent heat of vaporization, to be released later when the vapor condenses into clouds. Certain storms, including local thunderstorms and tropical hurricanes, are powered by the release of latent heat of vaporization.

Although the kinetic energy of the winds is small compared to the total amount of energy reaching the earth from the sun each day, it is still so vast that any attempt to change the weather by direct applications of energy has no chance of success. For example, air conditioning a coastal city like Los Angeles by giant fans on a day with a Santa Ana wind blowing from the interior deserts would be prohibitively expensive. On the contrary, plans are being made to extract energy from the wind by giant windmill farms in Wyoming and elsewhere, with no thought that the energy extraction would slow down the wind or otherwise influence weather conditions.

Modifying the weather must involve the modulation of *the release of energy already stored in the atmosphere* or the *modification of radiative energy transfer*. A possible way to accomplish the latter would be to change the reflecting characteristics of parts of the earth's surface.

Cloud Modification

Because clouds contain stored energy, they constitute a promising place for attempts at weather modification. Of course, the fact that clouds yield practically all of the atmospheric water deposited on the surface of the earth makes their modification a matter of great practical importance. These considerations make it clear why nearly all deliberate weather modification attempts so far have been directed at clouds.

Persons have attempted to modify clouds by spreading finely divided carbon black particles in them in an attempt to influence the radiation balance and hence the temperature of individual cloud droplets. Others have attempted to create clouds artificially or to intensify existing clouds by releasing large quantities of heat from the ground in giant oil burning devices called meteotrons (Dessens and Dessens, 1964). These experiments, while very interesting theoretically, have not produced any results of practical importance so far.

Cloud droplets form in the atmosphere by condensation on existing particles (*heterogeneous* nucleation), rather than by the gathering together of water molecules from the vapor state to form pure water droplets (*homogeneous* nucleation). The particles involved are called cloud condensation nuclei (CCN). Within the last few decades it has been recognized that the CCN population varies from place to place, and from day to day at a given location, and that these variations have an impact on the microphysical characteristics of the clouds. The differences in microphysical characteristics influence in turn the ease with which clouds convert their water into rain.

The most successful attempts to modify clouds have involved some modification of the population of CCN on which the droplets form, or of the ice nuclei (IN) which are responsible for the appearance of ice and important in the formation of precipitation in some clouds.

Just as the formation of water droplets normally requires the presence of CCN, the formation of an ice crystal normally requires the presence of an ice nucleus (IN). The homogeneous nucleation of the ice phase from supercooled water is generally considered to require a temperature just below $-40°C$.

The very expression *cloud seeding* owes its origin to the fact that pieces of solid material placed in a supersaturated solution to promote precipitation of the dissolved solute or in a supercooled solution to cause it to freeze are known to physical chemists as *seeds*. Artificial IN introduced into supercooled cloud droplets to induce freezing are an example of such seeds. The term "cloud seeding," invented to describe attempts to modify clouds by artificial nuclei, is sometimes used today in a more general sense to mean the release of any material designed to modify cloud properties. It is still not synonymous with "weather modification," although the two terms are used almost interchangeably in some documents.

An understanding of the possible outcomes of cloud seeding experiments must be based upon knowledge of the ways in which nature's cloud seeds, the natural CCN and IN, perform their roles. This knowledge, in turn, is based upon the results of laboratory experiments and observational data about aerosol particles in the atmosphere. Therefore, we begin with a description of the atmospheric aerosol and the factors which control its characteristics.

CHAPTER

II

The Atmospheric Aerosol

2.1 OBSERVED CHARACTERISTICS

The small particles which make up the atmospheric aerosol are known to us through a variety of effects (Fig. 2.1). They scatter light, so they are perceivable as haze or through the brilliant colors they sometimes impart to the rising or setting sun. The larger ones make themselves known as they settle out of the atmosphere as fine dust, soiling exposed surfaces. In some cases the particles cause physical distress, particularly to persons with respiratory ailments.

A number of instruments have been developed to sample atmospheric aerosols, taking advantage of the various effects which they produce [e.g., Husar (1974)]. Mass volume samplers draw air through filters which capture the larger particles for analysis, while the air itself and the smallest particles escape through the filter pores. In other cases the aerosol particles are allowed to settle out on slides coated with a sticky material, so that the particles can be examined later by electron microscopy or in other ways. Other devices make use of the attenuation of light beams and observations of scattered light intensity at various scattering angles to deduce the number concentration and size spectrum of the particles. In recent years electric analyzers have been developed. The particles are charged by diffusion of ions and then introduced into a chamber with a charged electrode which collects particles of various sizes depending upon its electrical potential.

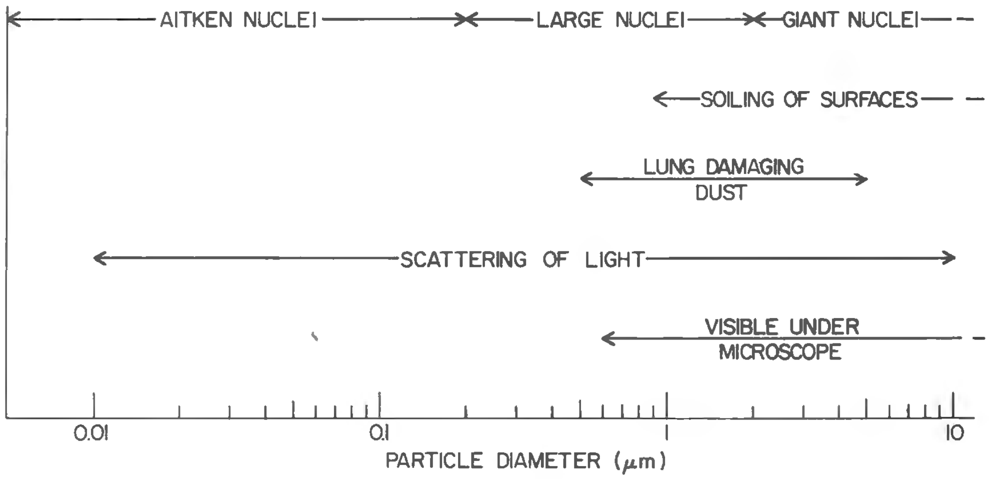


Fig. 2.1. Physical effects produced by atmospheric aerosol particles in different size ranges.

Particularly important are those instruments containing chambers in which the larger particles are activated as CCN. The resultant droplets, even though they may exist only briefly in the cloud chambers before settling out or collecting on the walls, are easier to observe than the particles themselves. The first such instruments are due to Aitken; the Gardner counter is a more recent example of them. Other important devices, to be discussed later, are the ice nucleus counters, in which certain particles are exposed by their ability to cause ice crystals to form.

Atmospheric aerosols cover a size range[1] from below 0.01 μm to over 10 μm in diameter. Those particles with diameters of less than 0.2 μm are called Aitken particles; those with diameters between 0.2 and 2 μm are called large particles; and those with diameters in excess of 2 μm are called giant particles.[2]

The total aerosol concentration varies widely, from as low as 10^9 m^{-3} in clean country air to over 10^{11} m^{-3} in heavily polluted areas. Figure 2.2 presents a comparison of aerosol observations in two different situations. Curve (a) is based upon data presented by Junge (1955) for polluted industrial areas in western Europe. The shaded area (b) is the envelope of about one year's observations taken near ground level at various rural locations in the northern Great Plains region of the United States (Davis *et al.*, 1978). Not only is the total aerosol concentration higher in the polluted air, but the particle size distribution is different. The pollution consists

[1] In this book, particle "size" refers to particle diameter or equivalent diameter unless otherwise noted.

[2] These definitions were established at a time when it was customary to specify particle sizes in terms of radius rather than diameter.

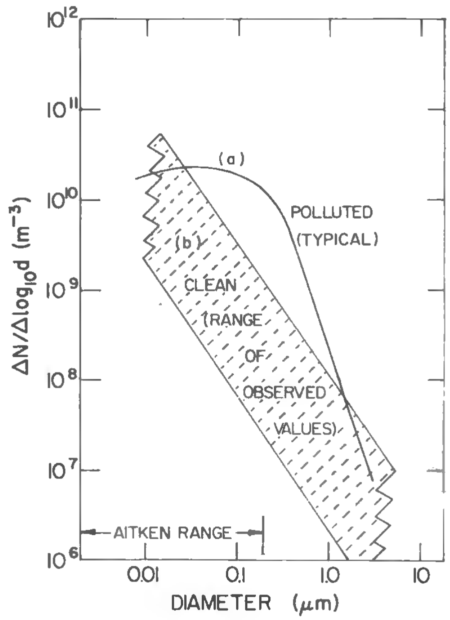


Fig. 2.2. Size distributions of atmospheric aerosols in polluted continental air (a) and in clean continental air (b) remote from large pollution sources.

mainly of large particles and particles near the upper limit of the Aitken range. Section 2.2 investigates the factors responsible for such differences and for the general characteristics of the atmospheric aerosol.

2.2 THE MECHANICS OF AEROSOLS

The concentration and size distribution of the atmospheric aerosol are controlled by the size distribution of the particles at the time of their introduction into the atmosphere, by the tendency of the particles to collide and coagulate, and by the processes removing the aerosol particles from the atmosphere. The coagulation process and the removal processes are strongly size dependent.

Fall Speeds

The atmospheric aerosol particles are sufficiently small so that Stokes law applies to them. Stokes law can be expressed in the form

$$\mathbf{F} = -3\pi\mu d\mathbf{u}, \tag{2.1}$$

where **F** denotes a drag force exerted on a particle of diameter d traveling at a velocity **u** through a fluid medium, and μ is the dynamic viscosity of the medium. Note that this is a vector equation and that the force is directed opposite to the direction of particle motion.

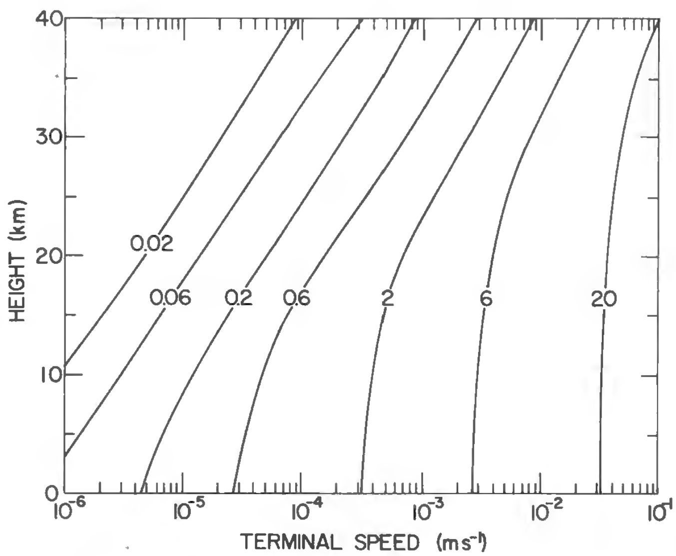


Fig. 2.3. Terminal fall speeds of spherical particles of density 2 Mg m^{-3} as a function of diameter and height in a standard atmosphere. Curves are labeled according to particle diameter in micrometers. [After C. E. Junge, C. W. Changnon, and J. E. Manson (1961). *J. Meteorol.* **18,** 81, by permission of American Meteorological Society and senior author.]

An aerosol particle in free fall quickly assumes its terminal speed, u_T, at which point the downward force exerted on it by gravity, suitably corrected for the buoyancy of the medium, is exactly balanced by the drag force. Equating the two forces we find for an aerosol particle in air

$$u_T = (\rho - \rho_a)gd^2/18\mu, \tag{2.2}$$

where g is the acceleration due to gravity, ρ is the density of the particle, and ρ_a is the density of the air.

Junge *et al.* (1961) have used an equation such as (2.2) to calculate the terminal speeds of spherical particles of density 2 Mg m^{-3} (specific gravity = 2) as a function of particle size at different heights in a standard atmosphere (Fig. 2.3). The density chosen is appropriate for many atmospheric aerosols. Even though they may be composed of substances with bulk specific gravities greater than 2, they are often irregularly shaped, consisting of coglomerations of small particles stuck together and with air in the interstitial spaces.

Brownian Motion and Diffusion

In addition to their downward drift under the influence of gravity, aerosol particles undergo erratic motions known collectively as Brownian motion.

Brownian motion is due to bombardment by air molecules, and is strongly size dependent. Large particles are being struck almost continuously on all sides by molecules traveling in different directions. Furthermore, the mass of a large particle is so great compared to that of a molecule that the net acceleration in any direction tends to be very small. A sufficiently small particle can receive a significant acceleration through a collision with a single molecule traveling at its root mean square speed and, furthermore, the collisions are so infrequent that the acceleration is not necessarily cancelled immediately by collisions with molecules traveling in other directions. Out of five collisions, say, in a given time interval, it is quite possible for four of them to produce positive accelerations in the x direction and only one a negative acceleration. However, for a large particle struck by 50 molecules in the same time period, the probability of 40 collisions imparting positive acceleration along the x axis and only 10 imparting negative acceleration is infinitesimally small.

The mathematical development of the ideas just outlined is an elaboration of the random walk theory. It is covered in several textbooks, the most authoritative treatment probably being that of Fuchs (1964). [See also Byers (1965); Pruppacher and Klett (1978, Chap. 12)]. Table 2.1, based on Fuchs, shows the instantaneous speeds and the average net distances traveled by aerosol particles of various sizes in one second. The table shows that Brownian motion diminishes greatly as one proceeds from fine to coarse aerosols.

The Brownian motion of an individual aerosol particle is isotropic, that is, motions in all directions are equally likely. Where a gradient exists in the number concentration of aerosol particles, there is a tendency for the particles to diffuse into the region of lower density. The Brownian diffusion coefficient of aerosol particles was shown by Einstein to be

$$D_{\mathrm{B}} = kTB, \tag{2.3}$$

where k is Boltzmann's constant, T is the absolute temperature, and B is the mobility of the particles. The mobility of a particle obeying Stokes's law is given by

$$B = 1/3\pi\mu d, \tag{2.4}$$

where all terms are as previously defined. Comparing (2.1) and (2.4) shows that, for a particle moving at velocity u, the drag force is inversely proportional to mobility, that is,

$$\mathbf{F} = -\mathbf{u}/B. \tag{2.5}$$

Examination of (2.3) and (2.4) shows that Brownian diffusion is more rapid for the small aerosol particles (Aitken particles) than for the large

TABLE 2.1

Average Brownian Speeds and Displacement of Aerosol Particles Compared to Terminal Speeds near Sea Level

Particle diameter (μm)	Terminal speed (μm s^{-1})	Average speed (3-dim) of Brownian motion (μm s^{-1})	Average of absolute displacement along a given axis in 1 s (μm)
10	3000	1.4×10^2	1.7
1	35	4.4×10^3	5.9
0.1	0.86	1.4×10^5	30
0.01	0.066	4.4×10^6	260

and giant particles (see also Table 2.1). This tendency is reinforced by the fact that particles with radii smaller than the mean-free path of the air molecules experience less drag than one would calculate for a simple application of (2.1). This complication can be handled by introducing into the formula for the mobility a *slip factor* originally due to Cunningham. The formula for the mobility becomes

$$B = [1 + (a'l/d)]/3\pi\mu d, \tag{2.6}$$

where l is the mean-free path of the air molecules (about 0.1 μm) and a' has been determined experimentally to be about 1.8.

For still smaller particles, in the nanometer size range, drag increases again due to gas kinetic effects.

Coagulation of Aerosols

The Brownian motions which are responsible for the diffusion of aerosol particles also lead to collisions among them and to coagulation.

It might be assumed on the basis of root-mean-square speeds (Table 2.1) that a collision between two small particles of the same size is much more likely than a collision between two large particles. However, the collision cross sections are larger for the large particles. The most likely collisions of all are those between small and large particles. Viewed in simplest terms, the small aerosol particles provide the high speeds needed to promote collision while the large particles provide the large collision cross sections required.

It can be shown using a very simple diffusion model that the number of small aerosol particles collected per unit time by a single much larger particle is given by

$$\nu = 2\pi d_2 D_1 N_1, \tag{2.7}$$

where d_2 is the diameter of the large particle, D_1 is the Brownian diffusion coefficient for the small particles, and N_1 is their concentration.

A formula for the probability of collisions between two aerosol particles is generally built on the assumption that the probability of collision is proportional to the sums of the diameters and to the sums of their diffusion coefficients. We therefore write the *kernel* for Brownian coagulation for particles of diameters d_1 and d_2 as

$$K_B(d_1, d_2) = \pi(d_1 + d_2)(D_1 + D_2)\beta, \tag{2.8}$$

where D_1 and D_2 are the diffusion coefficients for particles of diameters d_1 and d_2, respectively, and β is a coagulation factor which allows for the collisions that do not lead to coagulation and for multiple collisions between particles that rebound initially.[3] If there exist in a unit volume one particle of diameter d_1 and one particle of diameter d_2, $2\,K_B(d_1, d_2)$ is the probability that they will collide and coagulate in unit time due to Brownian motion.

The frequency of collisions with coagulation per unit volume between particles of diameter d_1 in concentration N_1 and particles of diameter d_2 in concentration N_2 is given by[4]

$$\nu = 2\,K_B(d_1, d_2)N_1N_2. \tag{2.9}$$

For a monodisperse aerosol of concentration N, the frequency of collisions with coagulation per unit volume is given by

$$\nu = K_B(d, d)N^2. \tag{2.10}$$

Each collision with coagulation results in a net reduction of 1 in particle concentration, so

$$dN/dt = -K_B N^2. \tag{2.11}$$

Using (2.9) and (2.11) and published values of K_B (Table 2.2), one can predict the rate at which an aerosol will "age," that is, change into a coarser aerosol with lower values of N.

In the Stokes law region K_B for monodisperse aerosols is independent of d, because it varies as $(D_B d)$ and D_B varies as $1/d$. It increases in the slip region ($d < 1$ μm), and later decreases again for $d < 10$ nm.

Calculations based upon (2.11) using values of K_B from Table 2.2 for monodisperse aerosols show that in any aerosol the value of N must de-

[3] Equation (2.8) is equivalent to that given on p. 294 of Fuchs (1964).

[4] Some authors [e.g., Byers (1965)] use a K_B that is twice that of Fuchs to characterize coagulation between particles of unequal size and so can write (2.9) without the factor of two.

TABLE 2.2

Coagulation Kernels for Brownian Diffusion (K_B) and Microscale Turbulence (K_*)[a,b]

d_2 (μm)	d_1 (μm) = 0.01	0.1	1.0	10	
20	7	7	8	22	K_*
20	3000	43	1.9	0.33	K_B
10	1600	22	1.0	0.30	K_B
1.0	160	2.4	0.34		K_B
0.1	12	0.7			K_B
0.01	0.9				K_B

[a] Values of K_B are based on Fuchs (1964).
[b] Units: 10^{-15} m^3 s^{-1}.

crease to about 3×10^{12} m^{-3} in 15 min, *regardless of the initial concentration* (Fig. 2.4). For lower concentrations, coagulation proceeds more slowly, but Junge (1955) has shown that over a week the modal point on a log–log plot like Fig. 2.2 would move from about 40 to 200 nm, if no new particles were added.

Examination of the coagulation kernels in Table 2.2 also makes it clear why the normal atmospheric aerosol contains virtually no particles smaller than 1 nm in diameter. In the presence of a normal aerosol with N being, say, 10^{10} m^{-3} and the representative value of d being 50 nm, the most probable lifetime of a 1 nm particle is of the order of 15 min.

The role of Brownian diffusion in causing coagulation is supplemented in some cases by gravitational settling, by phoretic effects, and by microscale turbulence. Turbulence is most important for the larger aerosol particles. The kernels for coagulation due to turbulence (K_*) included in Table 2.2 were calculated using a formula suggested by Smoluchowski, which is

$$K_* = \frac{1}{6}\frac{\partial u}{\partial n'}(d_1 + d_2)^3, \tag{2.12}$$

where $\partial u/\partial n'$ is the gradient of air velocity normal to the streamlines [assumed 5 s^{-1} in this case: compare Byers (1965, p. 73)]. The results shown are for $d_2 = 20$ μm, and indicate that microscale turbulence in the presence of fog or cloud particles is effective as a scavenger of giant aerosol particles.

The phoretic effects include thermophoresis, diffusiophoresis, and

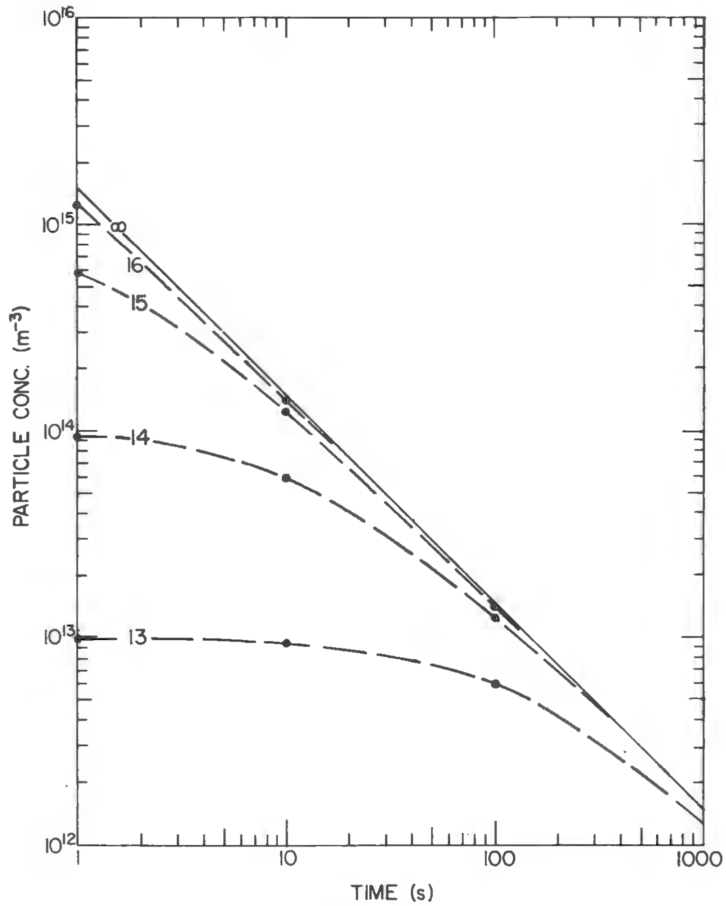


Fig. 2.4. Aerosol particle concentration as a function of time for various initial concentrations. Curves are labeled by logarithm (base 10) of concentration at time zero.

electrophoresis. Thermophoresis and diffusiophoresis are important in the case of growing and evaporating droplets. They will be discussed in Chapter V in connection with collisions between cloud droplets and ice nuclei.

2.3 TURBULENT MIXING OF AEROSOLS

The tendency for aerosol particles to coagulate into larger ones provides a mechanism for cleansing the atmosphere. The larger particles produced by coagulation have appreciable fall speeds (Fig. 2.3). However, the particles do not fall out of the atmosphere as rapidly as a simple consideration of their fall speeds would suggest. This is because turbulence, winds, and convective currents distribute the particles widely in the vertical as well as horizontally.

The role of turbulence in spreading aerosol particles and other impurities can be described by assigning to the atmosphere at each point coefficients of eddy diffusivity. Three coefficients, D_x, D_y, and D_z, are required to allow for the fact that turbulence is often anisotropic. It should be understood that these coefficients of eddy diffusivity, although they have the same units and can be used in a somewhat similar fashion to the coefficients describing molecular and Brownian diffusion processes, operate on a much larger scale and arise from different processes.

The theory of turbulence visualizes the turbulent energy in a given volume of air as being distributed systematically among eddies of different sizes [e.g., Batchelor (1960)]. Expressing the eddy sizes as wave numbers, that is, by the reciprocals of their diameters, and making the simplifying assumption that the turbulence is isotropic, one can write the turbulent energy distribution over the wave number spectrum as $E(\kappa)$, where κ is the wave number. It has been shown that over the inertial subrange of the spectrum, where energy dissipation is small,

$$E(\kappa) \sim \varepsilon^{2/3}\kappa^{-5/3}, \tag{2.13}$$

where ε expresses the rate of energy dissipation. Equation (2.13) is well known as the inverse five-thirds law of turbulence.

The turbulent energy in the inertial subrange cascades to higher wave numbers. The turbulent eddies continually break up into smaller ones, until the kinetic energy is finally dissipated as heat through viscosity. According to Batchelor (1960) the energy dissipation ε is concentrated in the high wave numbers (small eddies) according to

$$\varepsilon = \frac{2\mu}{\rho_a} \int_0^\infty \kappa^2 E(\kappa)\, d\kappa, \tag{2.14}$$

where μ is the viscosity, ρ_a is the air density, and κ is the wave number. [The quantity (μ/ρ_a) is the so-called kinematic viscosity.]

The intensity of the turbulence at any point can be characterized by the energy dissipation parameter ε. It is apparent that if energy were not constantly being fed into the system at the largest eddy sizes, that is, at low wave numbers, the turbulent motions would soon cease. Energy to drive turbulent eddies in the atmosphere is sometimes released in regions of pronounced wind shear or where the winds blow over rough terrain. The vertical currents in convective clouds are another source of turbulent energy.

Turbulent mixing of aerosols is produced by eddies of all sizes.[5] The turbulent energy at any point can be used to estimate D_x, D_y, and D_z,

[5] Only the very smallest eddies contribute to the microscale turbulence that affects aerosol coagulation (Table 2.2).

provided one knows the turbulence spectrum, including the sizes of the largest eddies present, along the x, y, and z axes. A word of caution is needed, however, because the appropriate values of D_x, D_y, and D_z also depend upon the *time interval* being considered. In considering the dispersion of an aerosol from a point source over 1 or 2 min, one cannot include the effect of large eddies with periods of 10–20 min, which would move the entire aerosol cloud in the same direction. Their effects would be modeled as local wind variations, rather than as turbulence.

2.4 VERTICAL DISTRIBUTION AND SEDIMENTATION OF AEROSOL PARTICLES

In the normal atmosphere the concentration of aerosol particles of a given size decreases with height. The large scale distribution of aerosol particles in the atmosphere is maintained indefinitely and involves vertical motions for all scales up to tens of kilometers. An appropriate value of D_z in this situation, averaging over both stormy and quiet regions, is 5 $m^2\ s^{-1}$. The masses of air passing upward and downward through a given horizontal plane as a result of turbulent motions are equal by definition. Therefore, turbulent mixing leads to a net upward transport of aerosol particles.

It is instructive to determine the vertical gradient in concentration of a monodisperse aerosol required to cause the upward flux by turbulent mixing to equal the downward flux due to sedimentation. The resultant calculation, assuming a constant value of D_z, shows that the mixing ratio, that is, the mass of aerosol particles per unit mass of air obeys an exponential relationship of the form

$$w/w_0 = \exp\{-u_T(z - z_0)/D_z\}, \tag{2.15}$$

where w is the aerosol mixing ratio, z is height, w_0 is the mixing ratio at some reference height z_0, and u_T is the terminal fall speed of the particles.

It should be noted that (2.15) is only approximate because both u_T and D_z are functions of height, u_T consistently increasing with height and D_z varying in a manner depending on wind shear and atmospheric stability. Substitution of numerical values for terminal speeds for aerosol particles of different sizes suggests that the scale height, over which w decreases by a factor of e (2.718 . . .), would vary from a few hundred meters for the giant particles to hundreds of kilometers for small Aitken particles in a typical stratified atmosphere ($D_z = 5\ m^2\ s^{-1}$). This prediction is in agreement with the observation that the giant aerosol particles are concen-

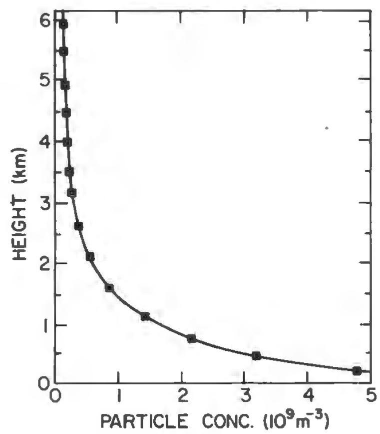


Fig. 2.5. Concentration of aerosol particles as a function of height, based on average of observations at several sites in the U.S.S.R. [After E. S. Selezneva (1966). *Tellus* **18**, 525, by permission of Tellus Editorial Office.]

trated in the lowest 2 or 3 km, while the Aitken particles are widely distributed throughout the troposphere and stratosphere.

Because air parcels expand as they rise, N, the number concentration of aerosol particles per unit volume, decreases more rapidly than the mixing ratio. Indeed, for the Aitken particles, with scale heights of many kilometers, the decrease in N with height is accounted for almost entirely by the decrease in ρ_a with height, which also follows an approximately exponential relationship. The relationship of (2.15) is greatly distorted in regions with organized up and downdrafts, for example, in the vicinity of convective clouds. Nevertheless, some observational data [e.g., Selezneva (1966)] show a surprising conformance to the ideal exponential relationship indicated by (2.15) (Fig. 2.5).

The final stage in the removal of an aerosol particle from the atmosphere is its impaction upon some collector surface. Whether the aerosol concentration in noncloudy air increases or decreases with time depends on whether aerosol particles are being added at the earth's surface or elsewhere fast enough to offset removal at the surface. Where aerosol particles are being removed rather than added at the surface, the vertical profile of the concentration shows a drop very near the surface of the earth, and there turbulent mixing leads to a *downward* flux. Some surfaces are more effective than others in removing the particles. For example, forests are very effective scavengers of aerosol particles; the term "green area effect" has been coined to describe their cleansing action.

2.5 REMOVAL OF AEROSOL PARTICLES FROM THE ATMOSPHERE BY WASHOUT

The two processes of coagulation and sedimentation provide a mechanism by which the concentration of aerosol particles is held within bounds and through which the atmosphere cleanses itself of aerosols. However, aerosol particles, particularly the hygroscopic ones, are removed from the atmosphere so effectively by cloud and precipitation processes that direct deposition on the earth's surface is often of secondary importance. The processes of removal by clouds and precipitation are referred to collectively as *washout*. The term *fallout* is also used, particularly with reference to radioactive debris from nuclear explosions.

We begin an examination of washout by noting that every cloud droplet contains a hygroscopic CCN. As a typical raindrop consists of approximately 1,000,000 collected cloud droplets, each raindrop brings to the earth's surface approximately 1,000,000 aerosol particles, as a minimum.

Cloud droplets that exist for any appreciable time collect additional aerosol particles. This process is called *scavenging*. On the basis of the values of K_B in Table 2.2, one can estimate that the existence of a cloud in a given volume of air can clean it almost completely of Aitken particles over a period of a few hours.

Experimental verification of this prediction has been obtained in various places. Observations of Aitken particles during an episode of fog and drizzle in South Dakota showed the concentration dropping from 10^9 m^{-3} to an extremely low value of 5×10^7 m^{-3} over a 1 hr period (Davis *et al.*, 1978).

Further consideration of the sum of $(K_B + K_*)$ in Table 2.2 indicates that there should be a minimum in scavenging efficiency for aerosol particles near 1 μm in diameter. However, these are the particles most likely to be incorporated into cloud droplets as CCN, so the cloud formation and scavenging processes complement each other in aerosol removal.

Even if the cloud droplets are not collected by falling raindrops and subsequently evaporate, the net result of a cloud's existence is still a marked reduction in the number of aerosol particles. Each evaporating cloud droplet leaves behind a large composite particle consisting of one or more CCN and scavenged particles.

2.6 ORIGINS OF CLOUD CONDENSATION NUCLEI

Having established in general terms the concentrations of atmospheric aerosols and the mechanics of the processes controlling their concentrations and size distributions, we turn now to a more specific examination of the sources and nature of those particles which control the formation of both water and ice clouds, that is, of the natural cloud seeds.

Aerosol particles are introduced into the atmosphere from a wide variety of sources including volcanos, meteor trails, forest fires, industrial smoke stacks, and other anthropogenic sources. Some of the aerosol particles are dust particles raised by the wind from exposed soil surfaces, while others are the remnants of spray droplets from the surface of the sea. In addition, there are many particles of organic origin including pollen particles, spores, and clusters of organic molecules such as terpenes released by vegetation. Nevertheless, the best evidence is that most aerosol particles are not introduced as solid particles at all, but are produced in the atmosphere itself by gas-to-particle conversion processes. In other words, they are formed by chemical reactions and precipitated into the atmosphere. As we have seen, the newly formed particles swiftly coagulate with the existing aerosol so that the total aerosol distribution remains approximately as indicated in Fig. 2.2.

It will be shown in Chapter III that only the largest of the atmospheric aerosol particles function as CCN under normal atmospheric conditions. It is necessary to distinguish between the *hygroscopic* particles, which readily take on water, and the *hydrophobic* ones, which do not. The search for the origin of CCN narrows down to a search for the origin of large and giant hygroscopic particles.

Leaving aside for the moment such anthropogenic sources as industrial smoke stacks, the principal source for the giant nuclei appears to be the surface of the sea. Indeed, the term giant *salt* nuclei is sometimes used, as chemical analysis of the giant particles often shows sodium chloride ($NaCl$) to be their principal constituent. The giant particles are derived from the sea surface through the evaporation of droplets of spray. However, the important spray for production of giant salt nuclei is not that from the wave crests, with which most persons with nautical experiences are familiar, but rather a very fine spray associated with the breaking of air bubbles through the foaming water surface. Mason (1971) has reviewed the observations relating production of giant salt nuclei to wind speeds over the ocean and estimates particle production rates as high as $2 \times 10^5\ m^{-2}\ s^{-1}$ in areas of breaking waves and an overall rate of $10^4\ m^{-2}\ s^{-1}$ for the ocean as a whole.

As the giant particles have significant fall speeds, of the order of a few hundred meters per day, corresponding to scale heights of a kilometer or less, they are normally found close to the surface of the ocean and do not penetrate more than a few hundred kilometers inland in appreciable concentrations. It must be emphasized that even over the oceans their concentrations are small, of the order of 10^6–$10^7\ m^{-3}$, and completely inadequate to explain the observed droplet concentrations in natural clouds. For that, we must turn to the large particles.

Because the large particles are formed by coagulation of numerous smaller ones, they are not chemically pure. Chemical analysis of the large nuclei ($0.2\ \mu m < d < 2.0\ \mu m$), which make up the vast majority of CCN, shows that NaCl is not predominant. Although a variety of compounds occur, the most common constituent of the CCN is ammonium sulfate [$(NH_4)_2SO_4$]. Observations of the concentration of the predominantly $(NH_4)_2SO_4$ particles indicate their source to be over or near land rather than over the oceans. Although some $(NH_4)_2SO_4$ is likely released by human activities, anthropogenic sources are not a satisfactory explanation for the ubiquity of that compound in the free atmosphere. It must have a natural source. Since no known natural source injects it directly into the atmosphere, it is thought that it is produced by gas-to-particle conversions.

The essential ingredients for $(NH_4)_2SO_4$ manufacture are of course nitrogen and sulfur. One can postulate a number of reactions. If the nitrogen is introduced as ammonia (NH_3) from decaying vegetation and the sulfur as sulfur dioxide (SO_2), say from forest fires, the subsequent reactions are straightforward. However, it is probable that most of the sulfur is introduced as hydrogen sulfide (H_2S) gas by decaying seaweed along coastlines and other vegetation. The multistage oxidation process required to transform H_2S to SO_3 and finally sulfates would require the action of a catalyst or of a powerful oxidant such as ozone. It is highly significant that a secondary maximum of large $(NH_4)_2SO_4$ particles has been observed about 20 km above sea level, which coincides with the height of the ozone layer. It may be too that more complex processes than oxidation of H_2S and reactions with NH_3 are involved. Observations over forests have shown that significant quantities of terpenes and other organic compounds are released to the atmosphere by trees, and some of them may contribute nitrogen for the $(NH_4)_2SO_4$.

Although the most common type of CCN, the large ammonium sulfate particle, has its origin over or close to land, the concentrations over the ocean are appreciable. These particles do not fall out rapidly as do the giant salt nuclei. Therefore, a continental air mass moving out to sea retains the characteristics of a continental aerosol for several days. Only over the most remote ocean areas do the measured concentrations of large nuclei fall much below $10^8\ m^{-3}$. On the other hand, a maritime air mass moving inland seems to replenish its supply of these particles quite rapidly, say within a day.

2.7 ORIGINS OF NATURAL ICE NUCLEI

The insoluble particles in the atmospheric aerosol, which play an important role in the nucleation of ice, have completely different origins from the hygroscopic particles. While meteoric dust, spores, bacteria, and industrial emissions have all been shown to act on occasion as ice nuclei, *most natural ice nuclei are insoluble clay particles picked up from the ground by the wind.*

Various researchers have examined the nuclei of ice crystals collected from the atmosphere or from snow samples gathered at the ground, while others have tested in cloud chambers the ice nucleating ability of various types of soil and mineral particles (Mason, 1971, pp. 194–199). The results obtained by the two approaches are similar and can be stated briefly. The most common substances, like sand, are usually not effective as ice nucleants, but some clay minerals have been observed to freeze water droplets at temperatures as high as $-5°C$. That should be considered a threshold temperature (say one particle active among 10^4). Considering activity and abundance together, Mason concluded that the kaolin minerals (complex silicates), with a threshold temperature around $-9°C$, are among the most important natural ice nuclei.

The wind speed required to raise dust particles varies, as would be expected, with the nature of the soil, especially its dryness. On cultivated fields the required wind speed varies with the angle between the wind direction and the orientation of the furrows. Recent data from west Texas suggest that wind speeds of 12–15 m s^{-1}, depending on direction, are sufficient to raise dust clouds (Porch and Gillette, 1977).

The size spectrum of particles raised by the wind also depends on the wind speeds. Winds of 25 m s^{-1} can raise particles as large as 100–200 μm into the atmosphere. However, the large fall speeds of these particles (several meters per second) imply that they can be supported only if they are entrained into strong updrafts induced by convection or by rough terrain. A more typical size spectrum for dust particles raised by the wind is shown in Fig. 2.6, which is based on several spectra observed by Gillette *et al.* (1972). The most numerous particles in their samples are in the range of 2 μm to 15 or 20 μm, whereas the hygroscopic particles derived from gas-to-particle conversions are most numerous in the Aitken size range.

The finer dust particles remain suspended for several days in some cases and travel long distances. Dust particles from the Sahara Desert have been identified in the atmosphere over Barbados and other islands in the West Indies.

Modes of Activation

An understanding of the behavior of ice nuclei and of the difficulties involved in sampling them requires consideration of their various modes of activation. It is now generally agreed that there are four modes, as follows:

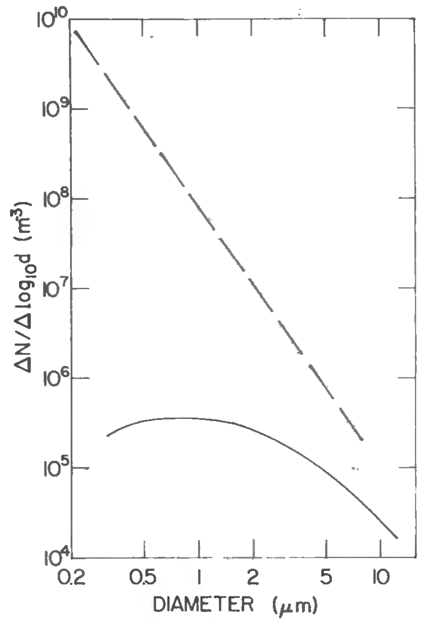


Fig. 2.6. Typical size spectrum for dust particles raised by the wind at heights of 1.5 – 6 m. Dashed line suggests size distribution of typical atmospheric aerosol for comparison purposes.

1. deposition (sometimes called sublimation),
2. condensation-freezing (sometimes called sorption),[6]
3. contact nucleation, and
4. bulk freezing.

The activation of a nucleus by deposition involves the laying down of water molecules into an ice lattice on the particle directly from the vapor state. In condensation-freezing, the particle is activated first as a CCN. Once a thin film of supercooled water has been deposited, either enveloping the particle or as a patch on its surface, a second nucleation event (freezing) takes place within the water film. There is some laboratory evidence that this process is actually more common than direct deposition from vapor to ice. Once the ice phase has been initiated, the entire water film freezes almost instantaneously and further growth is by deposition.

The other two nucleation mechanisms involve the freezing of a preexisting water droplet. Contact freezing usually results from a collision between an ice nucleus and a cloud droplet as a result of the Brownian motion of the nucleus. In such cases the nucleus often remains stuck in the surface layer of the droplet, especially if it is made of hydrophobic materials. The ice phase may be initiated preferentially at the junction of

[6] The term "sorption" will not be used again in this book, because Mason (1971) and Young (1974a) have used it to denote different processes.

the solid versus liquid and solid versus air interfaces. Bulk freezing involves the freezing of a droplet around a completely embedded nucleus.

The various modes of activation are not completely distinct. For example, condensation-freezing with the freezing long delayed merges into bulk freezing.

While it is convenient to distinguish among the four modes of freezing, it would be misleading to speak of four distinct types of ice nuclei (IN). Depending on ambient conditions, the same particle may initiate freezing in different ways. The various modes of activation are discussed again in Chapter V in connection with the activity of artificial IN.

Sampling of Ice Nuclei

A number of devices have been used to measure concentrations of IN in the free atmosphere. Among them are cold boxes in which ice crystals resulting from exposure of the IN to cold, moist air are observed visually, and the NCAR counter, which sucks the crystals through a small capillary and produces a pulse of sound. The problem of counting IN has proven very difficult, in part because of the difficulty of duplicating in a cold box the *modes* of activation which the nuclei follow in the atmosphere. A number of national and international workshops have been held for intercomparisons of various sensors [e.g., Vali (1975)]. While absolute measurements are only good to within a factor of 10, some statements are possible.

Activity as a Function of Temperature

Laboratory investigations of heterogeneous bulk freezing of supercooled drops have shown repeatedly that the probability of freezing increases with the extent of supercooling [e.g., Hobbs (1974); Mason (1971)]. This agrees with observations in the atmosphere, even though it appears that natural IN, being hydrophobic, are often activated as deposition nuclei rather than as bulk freezing nuclei.

The number of active IN per unit volume in the free atmosphere increases almost exponentially as the degree of supercooling increases. Thus

$$N_a(T') = N_0 \exp(\Omega T'), \tag{2.16}$$

where T' is the supercooling in degrees Celsius, $N_a(T')$ is the concentration of active nuclei, and N_0 and Ω are adjustable parameters (Fletcher, 1962). A typical value for N_0 is 10^{-2} m^{-3}. While Ω varies from 0.4 to 0.8, it is generally near 0.6, which means that the concentration of active IN increases by about a factor of 10 for each drop in temperature of 3.5 to 4°C (Fig. 2.7).

As the concentration of natural IN active in the free atmosphere increases by one order of magnitude for each drop in temperature of 3.5–4°C, it is necessary to specify the temperature at which a given observation is supposed to apply. For convenience, observations are often referred to the standard temperature of −20°C. The cold boxes and other sensing chambers are usually operated at −20°C, but sometimes they are not and assumed adjustment factors are applied.

Concentrations

The available observations indicate that natural IN are relatively scarce. In some regions with clean air the concentration of natural IN able to cause freezing of a droplet at −20°C is as low as 10^3 m^{-3}. Comparing this with a typical droplet concentration of 5×10^8 m^{-3} in a continental

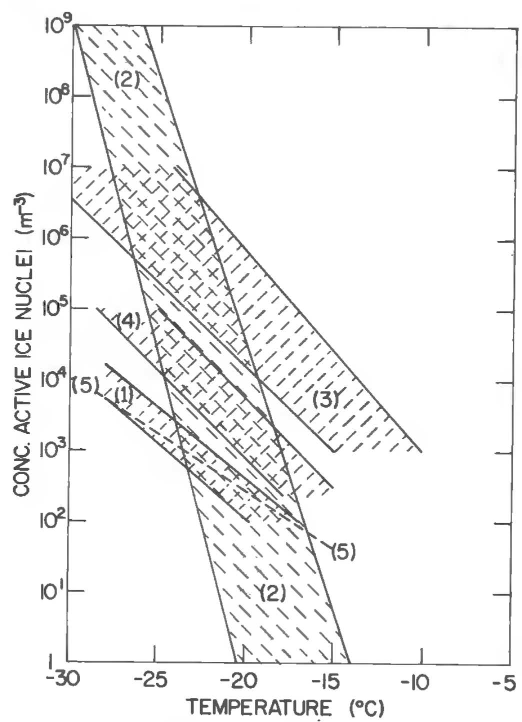


Fig. 2.7. Natural ice-nucleus spectra exhibiting simple exponential form: (1) Palmer (1949); (2) Workman and Reynolds (1949); (3) Aufm Kampe and Weickmann (1951); (4) thermal precipitator; (5) Warner (1957). [After N. H. Fletcher (1962). "The Physics of Rain Clouds," p. 241, by permission of Cambridge Univ. Press, London and New York, and the author.]

cloud shows that the probability of a cloud droplet being frozen by direct action of an IN at $-20°C$ is extremely small in some areas. The low concentration of IN explains why supercooled clouds are frequently observed in the free atmosphere, and why particle concentrations in ice crystal clouds, when they do occur, tend to be low.

Trained Nuclei

There is some evidence that, if an IN is activated and the resultant ice crystal sublimes away, the nucleus will be more effective than it was originally. This may be due to residual ice bound in crevices or other irregularities in the particle surface. Such particles are called "trained" or "preactivated" IN.

CHAPTER

III

The Formation of Clouds and Precipitation

3.1 THE MICROPHYSICS OF A CLOUD DROPLET

Release of Excess Water Vapor by Cooling

Clouds form wherever air is cooled below its dew point, whether by radiation, by mixing with cooler air, or by ascent in the atmosphere with resultant decompression.

The amount of water vapor which can exist in a given volume in equilibrium with a plane surface of pure water is a function of temperature only. Air (or space) containing this amount of water vapor is said to be *saturated*. As water vapor closely obeys the ideal gas law, it is convenient to express the amount present in terms of the pressure exerted by it (e), and we shall speak in this section of the *saturation vapor pressure, e_s* (T). This quantity is sometimes called the *equilibrium* vapor pressure.

Any water vapor in excess of the amount required for saturation is theoretically available for formation of a water cloud. The rate at which water vapor is made available for formation of cloud droplets by a known rate of cooling can be calculated from the Clausius–Clapeyron equation, which relates the saturation vapor pressure to T, the temperature in de grees Kelvin, as follows:

$$e_s(T) = e_s(T_0) \exp\left\{\frac{L_v}{R_w T_0} - \frac{L_v}{R_w T}\right\}, \tag{3.1}$$

where $e_s(T)$ is the saturation or equilibrium vapor pressure at the temperature T, $e_s(T_0)$ is the saturation vapor pressure at some reference temperature T_0, L_v is the latent heat of vaporization, and R_w is the specific gas constant for water vapor.

Equation (3.1) can be used to calculate $e_s(T)$ over supercooled water surfaces ($T < 273.15$ K) as well as over water at temperatures above freezing. In order to find $e_i(T)$, the saturation vapor pressure over ice at temperature T, it is necessary to use (3.1) with L_s, the latent heat of sublimation, in place of L_v.

The saturation vapor pressures for ice and water are exactly equal at the triple point (0.0098°C). For our purposes, they can be considered equal at 0°C and assigned the experimental value of 0.611 kPa (6.11 mb). Using 0°C (273.15 K) as the reference temperature T_0, one can write for water

$$e_s(T) = (0.611 \text{ kPa}) \exp\left\{\frac{L_v}{R_w}\left(\frac{1}{273.15} - \frac{1}{T}\right)\right\} \tag{3.2}$$

and for ice

$$e_i(T) = (0.611 \text{ kPa}) \exp\left\{\frac{L_s}{R_w}\left(\frac{1}{273.15} - \frac{1}{T}\right)\right\}. \tag{3.3}$$

The results are shown graphically in Fig. 3.1.

If the water vapor present at temperature T exerts a partial pressure e, one defines the *saturation ratio* as

$$S = e/e_s(T). \tag{3.4}$$

For temperatures below 0°C, one also defines an *ice saturation ratio* as

$$S_i = e/e_i(T). \tag{3.5}$$

The Clausius–Clapeyron equation permits one to determine for a parcel of cooling air containing a known quantity of water vapor the increasing water or ice saturation ratio as appropriate. Supersaturations, values of S greater than 1, indicate excess water vapor available for the formation of cloud. However, to determine the characteristics of the cloud which results, it is necessary to consider the microphysical processes which control the formation and growth of the individual cloud droplets. These processes involve such factors as diffusion of water vapor, heat conduction, release of latent heat, and the effects of surface tension and dissolved solutes in individual droplets.

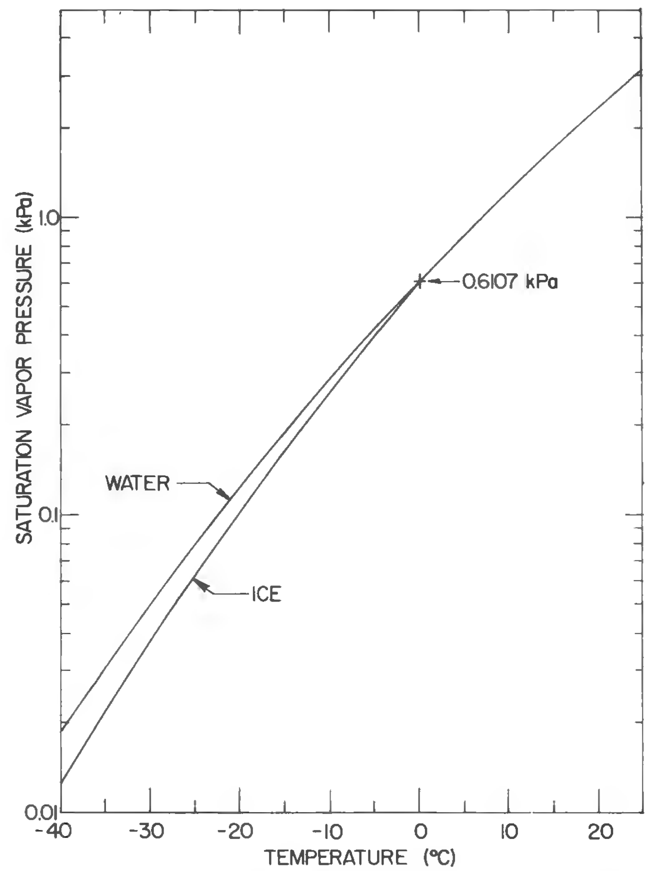


Fig. 3.1. Saturation (equilibrium) vapor pressure over a plane surface of pure water for the temperature range −40 to +25°C and over a plane ice surface for the temperature range −25 to 0°C [plotted from *Smithsonian Meteorological Tables* (List, 1958)].

The Surface Tension (Kelvin) Effect

Because of surface tension effects, there is energy stored in all water surfaces. The energy stored in the surface of a water droplet of diameter d is

$$E_\gamma = \pi d^2 \gamma, \tag{3.6}$$

where γ, the coefficient of surface tension, is equal to about 0.075 J m^{-2} for an air–water interface at 0°C.

Lord Kelvin was the first to show that because of surface tension the vapor pressure required to maintain a small water droplet in equilibrium with its environment is greater than that required to maintain equilibrium above a plane surface of pure water at the same temperature. The saturation (equilibrium) vapor pressure $e_{s,d}(T)$ above a pure water droplet of diameter d is given by

$$e_{s,d}(T) = e_s \exp\left(\frac{4\gamma}{\rho_L R_w T d}\right), \tag{3.7}$$

where e_s is the saturation vapor pressure above a plane surface of pure water, γ is the surface tension coefficient, ρ_L is the density of liquid water, R_w is the specific gas constant for water vapor, and T is the temperature in degrees Kelvin.

Role of Dissolved Solute

Consideration of (3.7) shows that as d approaches zero the value of $e_{s,d}$ for a pure water droplet approaches infinity. For this reason, the homogeneous nucleation of a pure water droplet from the vapor state requires very large saturation ratios. Only for values of S around 4.5 (350% supersaturation) do the droplet embryos resulting from chance assemblages of water molecules and large enough to grow further by capture of additional molecules have an appreciable probability of appearing in a reasonable period of time, say 1 s [e.g., Pruppacher and Klett (1978, Chap. 7)]. That is why clouds normally form by heterogeneous nucleation upon cloud condensation nuclei (CCN). The role of a CCN is to overcome the energy barrier that must be surmounted to form a new cloud droplet.

Experimentally, it is known that the presence of dissolved solute in water reduces the saturation vapor pressure according to Raoult's law, which is given by

$$(e_s - e_s')/e_s = n/(n + n'), \tag{3.8}$$

where e_s' is the saturation vapor pressure over the solution, n is the number of molecules of dissolved solute, and n' is the number of water molecules. In practice, it is necessary to take account of the dissociation of the solute molecules into ions by introducing a dissociation factor.

Combining the Kelvin (curvature) and the Raoult (solution) effects, it is possible to express the saturation vapor pressure over a solution droplet of diameter d as

$$e_{s,d}' = e_s\left[1 + \frac{4\gamma}{\rho_L R_w T d} - \frac{6iM_w m_s}{\pi \rho_L M_s d^3}\right], \tag{3.9}$$

where M_w is the molecular weight of water, M_s is the molecular weight of solute, m_s is the mass of solute, i is the dissociation factor, and the other symbols are as previously defined.

For a droplet of specified size containing a know mass of solute, one can compute the *equilibrium saturation ratio* S_{eq}, according to

$$S_{eq} = \frac{e'_{s,d}}{e_s} = 1 + \frac{4\gamma}{\rho_L R_w T d} - \frac{6iM_w m_s}{\pi \rho_L M_s d^3}. \tag{3.10}$$

Solutions to (3.10) for various sized NaCl particles are shown in Fig. 3.2. Each curve shows the equilibrium saturation ratio as a function of d for one size of NaCl particles. Such curves are called *Kohler* curves.

Figure 3.2 shows that for each nucleus there exists a *critical diameter* for which the supersaturation required to prevent the droplet from evaporating is a maximum. The larger the nucleus, the larger is the critical diameter d_c, but the smaller the value of $e'_{s,d}$ at $d = d_c$.

Kohler curves show that hygroscopic particles take on water for values of S considerably less than 1. This tendency accounts for the increasing opacity of haze at night, for example. The droplets enlarge in response to the increasing values of S resulting from radiational cooling. However, they can never surpass d_c and become real cloud droplets until S exceeds 1.

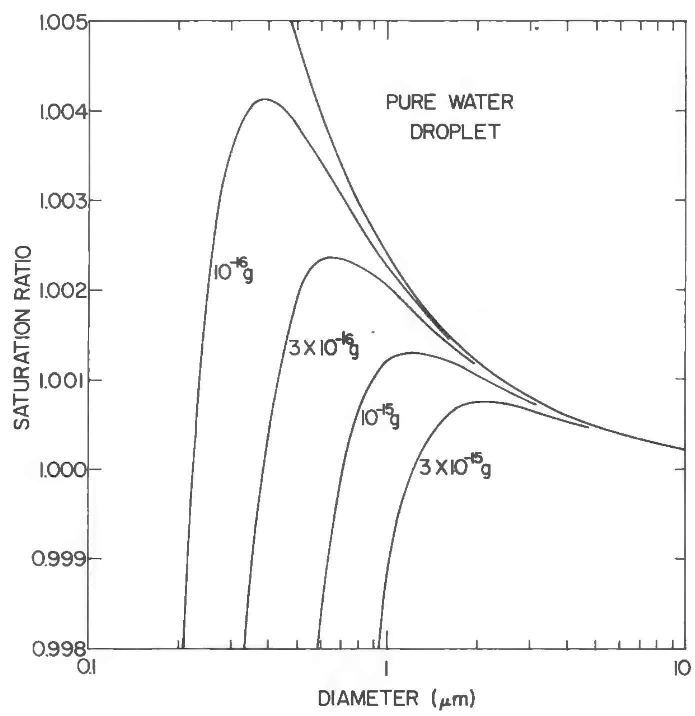


Fig. 3.2. Equilibrium saturation ratio (S_{eq}) over the surfaces of solution droplets as a function of droplet diameter for various masses of dissolved sodium chloride (NaCl).

Growth Rate Equation

In considering the growth or evaporation of cloud droplets, it is necessary to consider not only the saturation or equilibrium vapor pressure required to maintain a given droplet, but the rate at which the size of the droplet changes in response to an excess or deficit of vapor pressure in its immediate environment.

For a pure water droplet of a size sufficient that the surface tension effects can be ignored, the growth rate is controlled by the rate at which water vapor diffuses to the droplet surface and the rate at which the latent heat of vaporization released at the droplet surface is conducted away from the droplet. The release of the latent heat warms the droplet slightly, by about 0.01°C, thereby raising the equilibrium vapor pressure over the droplet surface and acting as a brake on the droplet growth rate.

The equation governing growth or evaporation of a solution droplet derived by combining the major controlling factors is

$$\frac{dd}{dt} = \frac{4}{d}\left(\frac{(S - S_{eq})F_v}{(L_v^2\rho_L/kR_wT^2) + (\rho_L R_w T/De_s(T))}\right), \tag{3.11}$$

where k is the thermal conductivity of the air, D is the diffusivity of water vapor in air, F_v is the ventilation factor (discussed below), and all other symbols are as previously defined.

Equation (3.11) has important implications for cloud formation. For two droplets of different sizes, the rate of increase in diameter of the smaller droplet exceeds the rate of increase in the diameter of the larger. Therefore, condensation upon a population of cloud droplets tends to bring the diameters of the small and large droplets closer together, even while all the droplet diameters increase.

Further consideration of (3.11) shows that the surface area of a growing droplet varies linearly with time once the initial formation stage is past. Therefore, the diameter varies with the square root of the time, and the mass of the growing droplet as $t^{3/2}$. For simulations in a computer, it is generally more convenient to keep track of the mass of a growing droplet rather than its radius. The form of (3.11) giving the rate of increase in droplet mass is

$$\frac{dm}{dt} = \frac{2\pi d(S - S_{eq})F_v}{(L_v^2/kR_wT^2) + [R_wT/De_s(T)]}. \tag{3.12}$$

Equations (3.11) and (3.12) predict that the growth rate will vary with the temperature for a given value of S. The thermal conductivity of air k and the diffusivity D are themselves functions of the temperature and density of the air (Table 3.1). It turns out that the most rapid growth occurs

TABLE 3.1

Dynamic Viscosity and Thermal Conductivity of Air and Diffusivity of Water Vapor in Air[a,b]

Temperature (°C)	Dynamic viscosity (μ) (10^{-6} kg $m^{-1}s^{-1}$)	Thermal conductivity (k) (10^{-3} J m^{-1} s^{-1} K^{-1})	Diffusivity of water vapor (D) (10^{-6} m^2 s^{-1})
30	18.7	26.4	27.3
20	18.2	25.7	25.7
10	17.7	25.0	24.1
0	17.2	24.3	22.6
−10	16.7	23.6	21.1
−20	16.2	22.8	19.7

[a] From *Smithsonian Meteorological Tables* (List, 1958).
[b] Values of k and D apply at 100 kPa.

under conditions of low density at relatively high temperatures, such as in the upper portions of tropical clouds.

The ventilation factor F_v is equal to one for the growth or evaporation of a droplet small enough for its terminal fall speed to be ignored. The growth or evaporation of a larger droplet is enhanced, because the diffusion processes which transport heat and vapor to or from the droplet are supplemented by the droplet's fall into air previously unaffected by it.

A commonly quoted formula for the ventilation factor is

$$F_v = 1 + 0.22(N_{Re})^{1/2}, \tag{3.13}$$

where N_{Re}, the Reynolds number of a droplet in air, is given in turn by

$$N_{Re} = \rho_a du/\mu \tag{3.14}$$

with all symbols as previously defined.

Equations (3.11) and (3.12) can be modified further to take account of molecular processes at the droplet surface [e.g., Pruppacher and Klett (1978, p. 133)]. As stated here, they overestimate growth rates, especially for droplets in the 1 μm size range, but they are satisfactory for many applications.

Evaporation of Droplets

Equation (3.12) predicts droplet evaporation when $S < S_{eq}$. For a given $S < 1$, the rate of decrease of a droplet's surface area is a constant, so the time required to evaporate a droplet varies as d^2. Considering that u_T varies as d^2 for droplets less than about 75 μm, it is found that the distance a droplet falls while evaporating varies as d^4. For larger drops the distance fallen varies with some power of d around 3–3.5. Therefore, in a given situation, there is a sharp cutoff between drops able to reach the ground without evaporating (raindrops) and those which cannot. Typically, the cutoff is around 0.3 mm, but by convention all drops greater than 0.2 mm in diameter are considered as precipitation drops. We shall adopt the common practice of referring to them as raindrops. [Some authorities prefer to limit the term "raindrop" to drops of 0.5 mm or more, and call drops in the 0.2–0.5 mm range "drizzle drops."]

3.2 FORMATION OF WATER CLOUDS

Simulation of Cloud Formation

Use of (3.12) or an equivalent expression permits one to simulate the formation of a cloud, provided the ambient conditions, including the rate of cooling and the size spectrum and chemical composition of the CCN present, are specified. A simplified flow chart of the problem is shown as Fig. 3.3. If the cooling is caused by ascent and adiabatic expansion of the air parcel, a number of adjustments must be introduced at each time step.

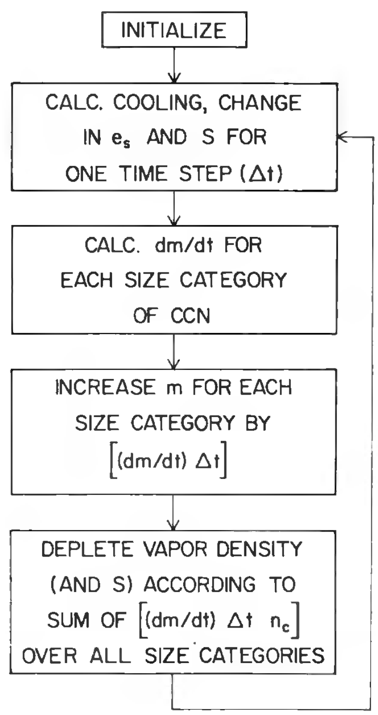


Fig. 3.3. **Flow chart of a computer simulation of cloud formation allowing for competition for available water vapor.**

They cover changes in the number concentrations of CCN, the coefficients of thermal conductivity and diffusivity, and in the concentration of the excess water vapor.

Figure 3.4, based upon Mordy (1959), shows the history of nuclei of different sizes as a parcel of air is cooled by an upward motion of 0.15 m s^{-1}. As the air cools and the supersaturation increases, the largest nuclei are activated and the droplets forming on them quickly exceed their critical diameters. Although the droplets are growing, they are unable to take up the available water vapor quickly enough to keep the supersaturation from increasing further. The increasing supersaturation activates more and more of the smaller nuclei, until finally the rate at which water vapor is being condensed exceeds the rate at which excess water vapor is made available by the cooling. The supersaturation then begins to fall and the smallest droplets, which have not reached their critical diameters, begin to evaporate. If the supersaturation falls very rapidly, even some of the droplets which have passed their critical diameters are evaporated.

The saturation ratio in a forming cloud normally does not exceed 1.005. Once a cloud has formed, the saturation ratio in the cloudy air returns to a value close to 1.000, and no more CCN are activated. At the base of a cumulus cloud, the layer of air with appreciable supersaturation is usually only about 50 m thick. Further condensation as the air parcels continue their ascent increases the sizes of the existing droplets, rather than the droplet concentration.

Returning to Fig. 3.4 we note that the diameters of the droplets which

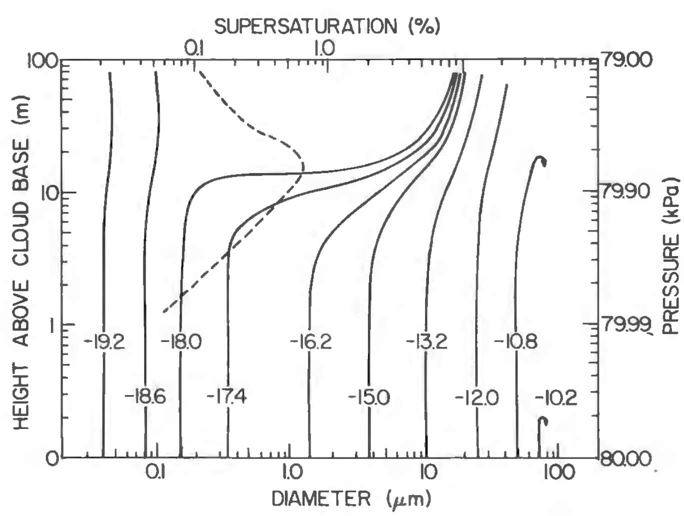


Fig. 3.4. The growth of cloud droplets on NaCl particles of different sizes in the initial formation of a water cloud. Curves are labeled with logarithm of CCN mass in moles. The updraft in this case was assumed to be 0.15 m s^{-1}. [After W. Mordy (1959). *Tellus* **11,** 16, by permission of the Tellus Editorial Office.]

survive the selection process do not differ by as large a factor as the diameters of the nuclei around which they form. Cloud droplet populations have narrow size distributions; in some cases all of the droplets in newly formed cloud are in the 10–20 μm size range, and further condensational growth tends to narrow the differences further.

Numerical simulations indicate that the cloud droplet concentration is a function of the rate of cooling and of the CCN spectrum [e.g., Howell (1949); Mordy (1959)]. Rapid cooling produces larger temporary supersaturations and activates more CCN than does slow cooling. Therefore, the concentration of droplets at the base of a cloud with a strong updraft exceeds that at the base of a cloud with a weak updraft, other things being equal. However, the concentration of droplets turns out to be more strongly influenced by the concentration and size distribution of CCN present. An abundance of large hygroscopic particles to act as CCN generally results in a high value of N_c, the number concentration of cloud droplets.

Twomey and Wojciechowski (1969), for example, have derived empirical formulas which quantify the three-way relationship among updraft speed, the maximum supersaturation during cloud formation, and the concentration of CCN activated.

Comparison with Observations

While the numerical simulations are in general agreement with observations, the situations in real clouds tend to be more complicated than the simulations would indicate.

Real clouds are generally "patchy." The cloud liquid water concentration χ_l may vary in a heavy cumulus from less than 1 g m^{-3} to 4 g m^{-3} or more over a few hundred meters distance. Furthermore, drop size distributions are often broader and more irregular than the simulations would indicate. Warner (1969) has observed bimodal spectra. Turbulent mixing of cloudy and clear air, or of cloudy air parcels with different histories, and pulsating updrafts have been invoked to explain the observations. The only thing clear at present is that there are influences at work which tend to broaden cloud droplet spectra with time.

Continental versus Maritime Clouds

Despite the complications just noted, there are detectable differences between the microphysical structure of clouds over the oceans and those over land [e.g., Squires and Twomey (1958)]. In general, droplet concen-

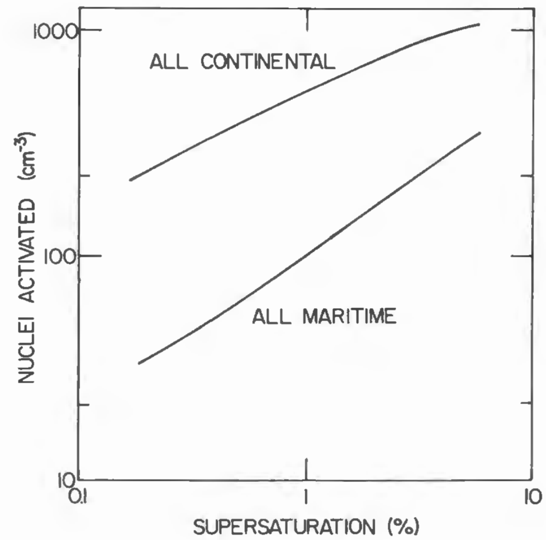


Fig. 3.5. Concentrations of CCN activated in typical continental and maritime air as a function of supersaturation. [After S. Twomey and T. A. Wojciechowski (1969). *J. Atmos. Sci.* **26**, 684, by permission of American Meteorological Society and senior author.]

trations are lower in maritime than in continental clouds, with typical values of N_c being 50 and 500 cm^{-3} for the two cases, respectively. Quite wide variations occur in each case. Cumulus clouds forming with strong updrafts in heavily polluted air have exhibited droplet concentrations as high as 1500 cm^{-3}. Nevertheless, the general rule appears to hold.

While clouds over the oceans do not usually exhibit the strong updrafts characteristic of continental cumulus clouds, the main reason for the difference in cloud droplet concentrations between maritime and continental clouds appears to lie in the differences between aerosol distributions in maritime and continental air masses. The aerosol over the oceans generally has fewer of the large hygroscopic nuclei than the continental aerosol contains. Figure 3.5 compares the number of CCN activated as a function of supersaturation for the median continental and maritime air masses sampled by the U.S. Naval Research Laboratory (Twomey and Wojciechowski, 1969). The result agrees in general with the cloud droplet counts.

3.3 GROWTH OF ICE CRYSTALS FROM WATER VAPOR

Crystal Habits

While ice particles may appear in the atmosphere either as new particles or through freezing of supercooled cloud droplets, their next stage of growth always involves direct deposition of water vapor.

Natural scientists have long known that ice crystals in snowflakes take on a wide variety of forms. All of these forms, or habits, are basically hexagonal structures (Fig. 3.6). Among the common crystal forms are flat hexagonal plates; hexagonal columns, which can be thought of as plates which have grown along the *C* axis rather than the *a* axes; and dendrites, finely branched six-pointed stars. There are flat dendritic crystals, which can be thought of as having grown from hexagonal plates as a result of preferential deposition of water vapor at the corners, and true spatial dendrites, with branches extending out in three dimensions. In addition to the well defined crystal forms, there are irregularly shaped needles. All of the crystals may undergo varying degrees of riming due to collisions with supercooled water drops, which freeze to the crystals.

Laboratory investigations by Nakaya (1951) were among the first to delineate the conditions favoring growth of the various ice crystal forms. The ice crystal habit is apparently controlled jointly by the temperature and the excess vapor pressure over that required for ice saturation in the atmosphere. Hobbs (1974) reviewed the available data and concluded that the effect of temperature tends to be the more important. Furthermore, in a cloud of water droplets, where many ice crystals actually grow, the temperature determines the vapor pressure.

Figure 3.7 shows ice crystal habit as a function of temperature and ice saturation ratio. While very simplified, it is consistent with practically all of the data available to the present author. The transitions among the do-

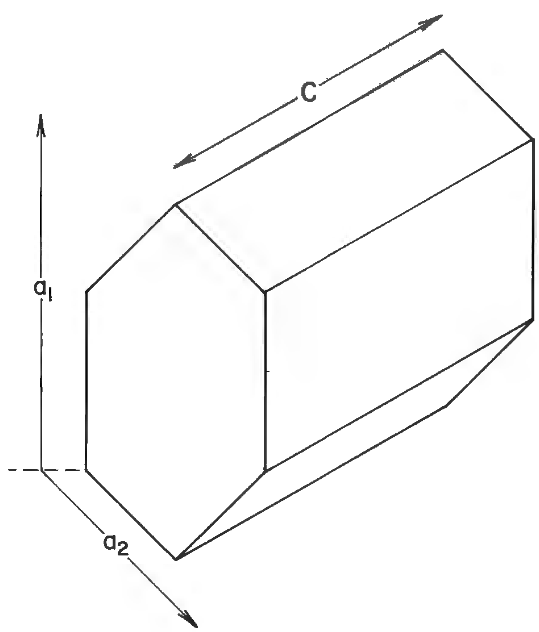


Fig. 3.6. A columnar ice crystal. Orientations of axes of basic cells making up the crystal are indicated.

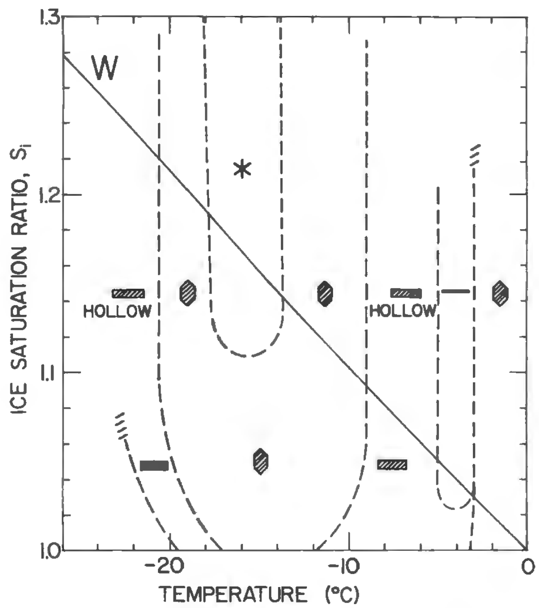


Fig. 3.7. Predominant ice crystal habits as a function of temperature and ice saturation ratio. Line W indicates water saturation, ——— indicates needles, ▨ indicates columns, ⬢ indicates plates, * indicates dendrites and stellars.

mains as indicated are not abrupt. For example, sector plates (with large extensions on the six corners) are observed near the transition between the plate and dendrite domains [e.g., Hobbs (1974)].

If a growing ice crystal in a laboratory cold chamber is subjected to a temperature change, it responds by changing its crystal habit. Thus, it is possible to cause dendritic growths to appear on the corners of a plate, or to cause plates to appear at both ends of a column, creating a capped column. Examination of a snowflake at the ground sometimes reveals variations in crystal habit, which can be related to temperature variations experienced by the snowflake in its fall through the clouds. Hobbs (1974) has noted that the results of such studies by several investigators conform fairly well to laboratory findings.

Growth Rate Equation for Ice Crystals

Equation (3.12), which gives the growth rate or evaporation rate of a cloud droplet as a function of the saturation ratio in the environment, can be adapted to the case of growing ice crystals once they have grown to a few micrometers. The initial growth process ordinarily occupies some tens of seconds, even if ice nucleation takes place on a very small nucleus.

For ice growth, the saturation vapor pressure over ice at the ambient temperature T, $e_i(T)$, must be substituted for the saturation vapor pressure over water $e_s(T)$. This substitution is made at the appropriate point in the denominator of (3.12).

The saturation ratio with respect to ice S_i is substituted for S in the numerator. It is also necessary to replace the latent heat of vaporization L_v by the latent heat of sublimation L_s.

One final adjustment is required. Equation (3.12) for a growing water droplet contains the diameter of the droplet in the numerator. As ice crystals have irregular shapes, it is necessary to substitute *for the radius* of the droplet the capacity of the crystal C, which measures the ability or capacity of the ice crystal to grow by extracting water vapor from its surroundings. The final equation is

$$\frac{dm}{dt} = \frac{4\pi C(S_i - 1)}{(L_s^2/kR_wT^2) + (R_wT/De_i(T))}. \tag{3.15}$$

This equation applies once the ice crystals have reached a diameter of 1 μm or so, where the equilibrium value of S_i is 1, the same as over a plane ice surface.

Theoretical values for C have been derived for a few simple crystal shapes by analogy with electrostatics. For a sphere, $C = d/2$, while for a circular disk $C = d/\pi$. Hexagonal plates, a common crystal form, have about the same value of C as thin disks of equal area, while dendritic plates again have the same values as plates of the same outside dimensions.

In addition to considering differences in the forms of (3.12) and (3.15), it is important to note that the values of S_i are often much greater than the maximum values of S ever observed in the atmosphere. The reason is that IN, unlike CCN, are scarce. The ice crystals formed are often unable to use up the available water vapor, and the only limit to the increasing values of S_i is imposed by condensation once S reaches one. In many cases, therefore, ice crystals grow in a situation where the value of S is equal to one.

The supersaturation with respect to ice S_i, which corresponds to water saturation ($S = 1$), increases steadily as the temperature decreases from 0°C. However, the most rapid growth of an ice crystal at water saturation does not occur at very low temperatures.

The absolute difference in vapor pressure between water and ice saturation reaches its greatest value near −12°C (Fig. 3.1). After allowance for variations in k and D with temperature (Table 3.1), it turns out that the most rapid depositional growth of ice crystals for a given value of C at water saturation occurs at slightly lower temperatures, say −15 to −17°C, depending on the air density. However, there is evidence that C is

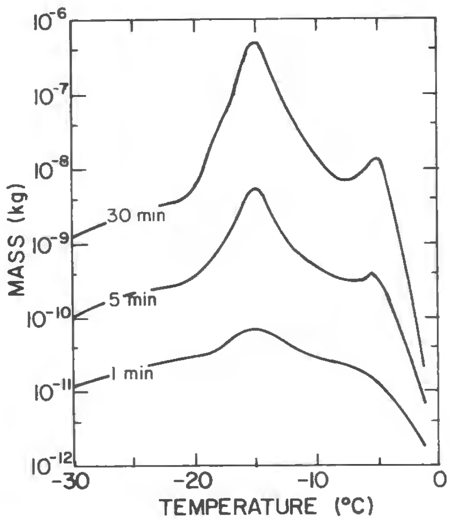


Fig. 3.8. Masses of growing ice crystals at various times as a function of temperature. [After E. E. Hindman II and D. B. Johnson (1972). *J. Atmos. Sci.* **29**, 1313, by permission of American Meteorological Society and senior author.]

strongly temperature dependent because of crystal habit effects. Rapid growth has been observed around $-5°C$ [e.g., Hobbs (1974)]. Figure 3.8 shows masses of ice crystals after different growth periods as a function of temperature as calculated by Hindman and Johnson (1972). Maxima in growth rates are evident near $-5°C$ and $-15°C$.

The growth of ice crystals in mixed ice-water clouds is of great importance to the formation of precipitation, particularly over the colder parts of the earth. Discussion of this aspect of ice growth is deferred to Section 3.5.

Ice Multiplication

While measurements indicate that IN are very scarce in the atmosphere, there have been some observations of ice crystal concentrations that exceed IN concentrations measured simultaneously by as much as a factor of 10 to 100. This situation was brought out clearly by observations taken in the tops of cumulus cogestus clouds near the $-10°C$ level in Missouri during Project Whitetop. Ice crystal concentrations there were often observed to run as high as 5000 m^{-3} near $-5°C$ (Braham, 1964). These observations led quickly to suggestions that some mutiplication process was at work, so that the activation of a single IN eventually produced many ice particles.

Actually, there are a number of possibilities to consider in trying to reconcile observations of IN and ice particles. For one thing, the measurements do not refer to the same thing. The IN measurements inside a cloud, if done correctly, refer to those nuclei which have escaped capture up to the moment the observation is made. Ice particles, on the other hand, are indicative of the activation of IN at some earlier time. Furthermore, the observation of an ice particle at $-10°C$, say, does not prove that it formed by activation of an IN at $-10°C$. This consideration is especially important in convective clouds, where there is little doubt that downdrafts cause the appearance at low levels of ice crystals or ice particles that originated higher in the cloud at lower temperatures.

After accounting for all the above considerations, it still appears that some form of ice multiplication process occurs on occasion (Mossop, 1970). Of course, the ideal IN is a piece of ice, which causes water to freeze at temperatures up to 0°C. Possibilities considered for an ice multiplication process included the breaking of fragments from starlike dendritic crystals as a result of collisions with other hydrometeors, and the expulsion of minute ice fragments when a drop freezes. It was thought that a drop would freeze from the outside forming a shell of ice around a water core. The core, upon freezing itself, would expand, shattering the shell and releasing a number of minute fragments.

The present thinking on ice multiplication is that it is associated with the riming of graupel particles collecting supercooled water droplets in their paths. Laboratory evidence suggests that the collection of large cloud droplets with diameters between 50 and 70 μm by a graupel particle results in the release of one small ice particle for roughly every 250 captures, provided the captures take place in the temperature range between -5 and $-12°C$ [e.g., Mossop and Hallett (1974)]. This evidence is consistent with the observation that multiplication seems most common in the tops of cumulus congestus clouds with large droplets and tops around $-10°C$ after the appearance of precipitation particles. However, the question of ice multiplication is still under active investigation in several laboratories, and new results may force another revision in the currently accepted explanation.

3.4 THE FORMATION OF RAIN BY COALESCENCE

Nearly all of the precipitation that falls on the earth is made up of cloud droplets which have been accreted, that is, swept up and captured by larger hydrometeors falling through the clouds. *Accretion* is a general term referring to the capture of any hydrometeor by one with a greater terminal fall speed. The special case of cloud droplets and raindrops being captured by larger droplets or raindrops is called *coalescence*. The capture of supercooled cloud droplets by solid precipitation particles to which they immediately freeze is called *riming*. Small ice crystals can be captured by larger ice crystals or snowflakes; this process is sometimes called *aggregation*. All of these processes are covered by the general term accretion.

Continuous Collection

We begin with the simplest situation, that of a large drop of diameter d falling at terminal speed u_T through a uniform cloud of water droplets. The falling drop sweeps out a volume of $(\pi d^2 u_T/4)$ per unit time. If it accretes all of the cloud droplets with centers lying within this volume, the raindrop increases its mass at a rate given by

$$dm/dt = \pi d^2 u_T \chi_l/4, \tag{3.16}$$

where χ_l denotes the (liquid) cloud water concentration.

Terminal Speeds of Water Drops

Numerical solutions of (3.16) require that u_T be known as a function of d.

Experimental data on u_T are quite abundant, but the theoretical development is complex. Cloud droplets are spherical, but raindrops are flattened by the aerodynamic and other forces acting upon them [e.g., Pruppacher and Klett (1978)]. It is customary to express u_T as a function of the *equivalent diameter*, that is, the diameter of a sphere containing the same volume of water as the drop in question.

The terminal speed of a drop is strongly dependent on the air density, and hence upon pressure, temperature, and relative humidity. Figure 3.9 shows u_T over the entire size range from small cloud droplets to large raindrops at one standard atmosphere (101.325 kPa) and +20°C. Figure 3.10, replotted from Beard (1976), emphasizes the variations in u_T for raindrops due to pressure and temperature variations.

One can handle u_T in computer simulations by tabulating data such as those in Fig. 3.10 in memory, but analytic formulas are more convenient. The following paragraphs outline one useful approach.

The retarding force on a falling drop is given by the product of the excess pressure at the stagnation point and of its horizontal cross section, modified by a drag coefficient C_D. The stagnation point is the point on the bottom of the drop where the air is at rest with respect to the drop. Invoking Bernoulli's principle shows that the excess pressure there is $(\rho_a u_T^2/2)$.

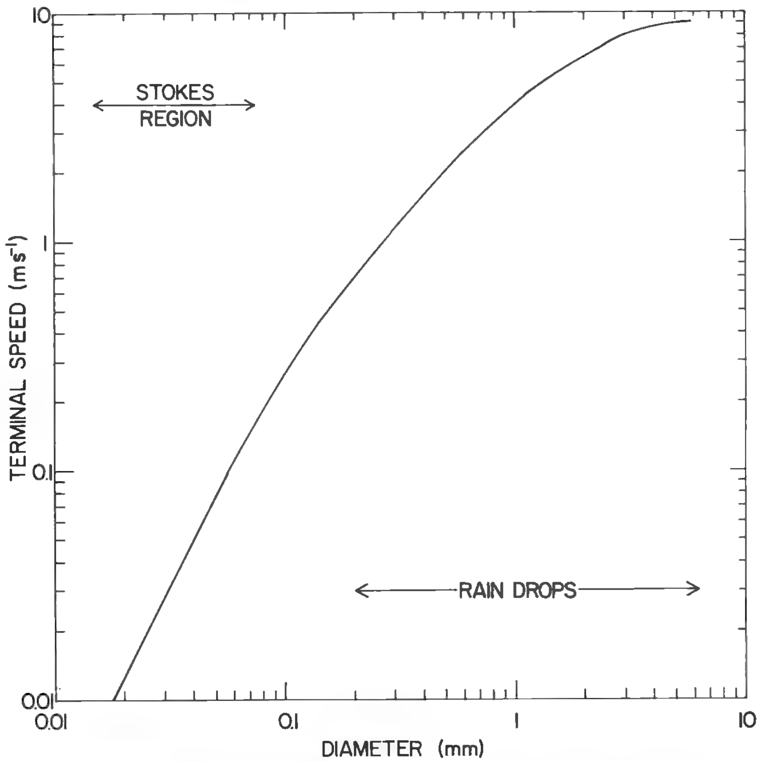


Fig. 3.9. Terminal fall speeds of water drops in air at one standard atmosphere pressure (101.325 kPa) and +20°C. [Values beyond Stokes region plotted from *Smithsonian Meteorological Tables* (List, 1958).]

Neglecting buoyancy effects,

$$mg = (\tfrac{1}{2}\rho_a u_T^2)(\pi d^2/4)C_D. \tag{3.17}$$

In general,

$$C_D = 8mg/\pi\rho_a u_T^2 d^2. \tag{3.18}$$

Recalling (2.2) and (3.14), we see that in the Stokes region,

$$C_D = 24/N_{Re}. \tag{3.19}$$

The basic problem in calculating terminal speeds beyond the Stokes region ($d < 80$ μm) is that the terminal speed u_T is a function of Reynolds number N_{Re}, which in turn depends on u_T. As several authors have shown, a way out is provided by noting that

$$C_D \propto 1/u_T^2,$$

while the Reynolds Number for a drop falling at terminal speed varies as u_T (3.14). By defining the Best number N_{Be} as

$$N_{Be} = C_D N_{Re}^2, \tag{3.20}$$

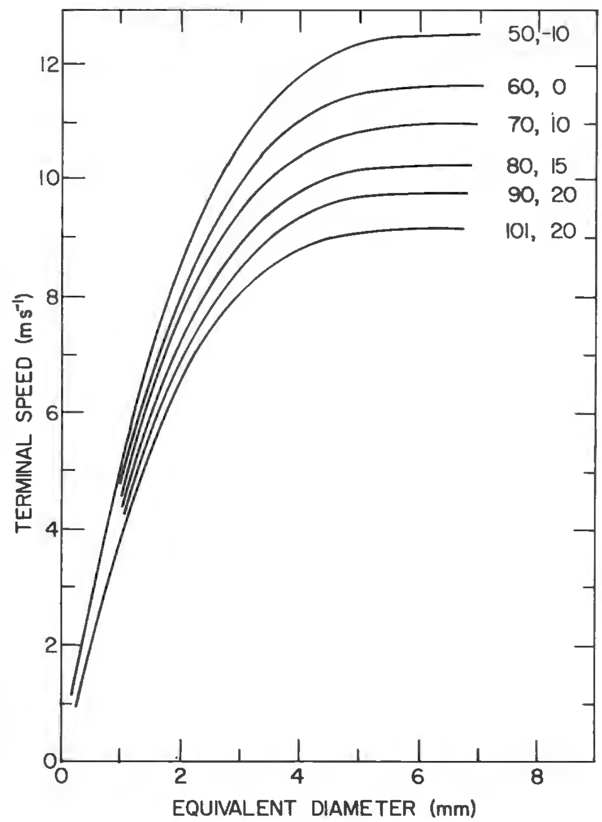


Fig. 3.10. **Terminal speed of raindrops as a function of equivalent diameter at various temperatures and pressures. Curves are labeled in kilopascals, degrees in Celsius. [After K. V. Beard (1976). *J. Atmos. Sci.* 33, 851, by permission of American Meteorological Society and the author.]**

one escapes the circular argument to find

$$N_{\mathrm{Be}} = 8\rho_a mg/\pi\mu^2. \tag{3.21}$$

Beard (1976) gives formulas for N_{Re} as a function of N_{Be} (which he called the Davies Number) on the basis of experimental results. Once N_{Re} is known, the calculation of u_{T} follows immediately from (3.14). This approach can be applied over a wide range of conditions and is useful in computer modeling of accretion.

Precipitation Embryos

Making use of (3.16) and the fall speeds of Fig. 3.9, and assuming a typical value of χ_l of 1 g m^{-3}, we find that a collector drop of say 100 μm diameter can grow into a typical raindrop size (d = 1 mm) in approximately 5 min. This calculation shows that precipitation forms readily in

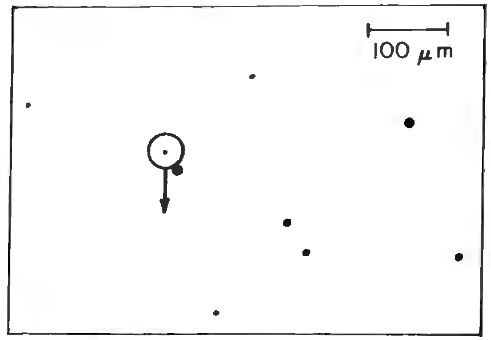


Fig. 3.11. A raindrop embryo in a water cloud growing by gravitational capture of cloud droplets.

clouds with appreciable liquid water concentration, provided some large drops or other collector particles are present. The required collector particles are called *precipitation embryos.*

Explaining the formation of precipitation is essentially a matter of explaining the origin or embryos. Large drop embryos do not form by condensation alone. Use of (3.12) shows that a 40 μm cloud droplet would require several hours to reach a diameter of 200 μm by condensation alone in air with a saturation ratio of 1.01. This is too slow to explain the formation of rain in many convective clouds some 20–40 min after initial cloud formation.

There are two sources of precipitation embryos. The first is the coalescence of liquid cloud droplets, and the second is the growth of large ice crystals by deposition of water vapor in mixed clouds because of the large supersaturations with respect to ice which occur in such clouds. As the coalescence process operates over a wider range of conditions than does the other, we shall consider it first (Fig. 3.11).

Precipitation Embryos through Coalescence of Liquid Droplets

Although a number of factors have been suggested to account for the coalescence of cloud droplets into raindrop embryos, gravitational capture appears to be by far the most important. Turbulence or electrical effects may hasten the process in a few special situations [e.g., Moore and Vonnegut (1960)].

Consideration of the coalescence of cloud droplets requires an elaboration of (3.16). The terminal speed of the smaller drop is no longer negligible in comparison with the terminal fall speed of the larger drop overtaking it. Neither is the size of the smaller droplet; a collision occurs if the distance between droplet centers decreases to the sum of the radii of the droplets.

A greater complication is introduced by the fact that the drop and droplet influence the air motions around them. The basic geometry of the coalescence process is shown in Fig. 3.12, where a relatively large droplet of diameter d_2 (hereafter called the drop) is shown overtaking a smaller droplet of diameter d_1 lying in its path. As both particles are falling at terminal speed, it is convenient to think of the drop as being stationary and the droplet as moving upward toward it. Viewed in this framework, the air moves upward with a speed equal to the terminal speed of the drop, parting to go around it. The droplet, as it moves upward toward the drop, tends to be swept around the drop by this air motion. One can consider the droplet as subject to two effects: (1) the inertial effects which tend to resist changes in its motion and (2) the viscous effects which tend to make it follow the streamlines of the air around the drop. The droplet actually follows an intermediate path, being deflected somewhat by the viscous forces but crossing some of the air streamlines.

In considering any drop–droplet pair, it is convenient to introduce the concept of their *collision efficiency*. The collision efficiency of a drop–droplet pair is defined in various ways. The following is now most commonly used and convenient. Define d_g as the diameter of an infinitely long cylinder with its axis passing vertically through the center of the drop, such that the droplet will collide with the drop if the droplet center is located within the cylinder while still at a long distance below the drop, but will not collide otherwise. In other words, placing the droplet a distance $d_g/2$ from the vertical axis at a long distance below the drop leads to a *grazing* trajectory. The collision efficiency is then defined by

$$E_1 = [d_g/(d_1 + d_2)]^2. \tag{3.22}$$

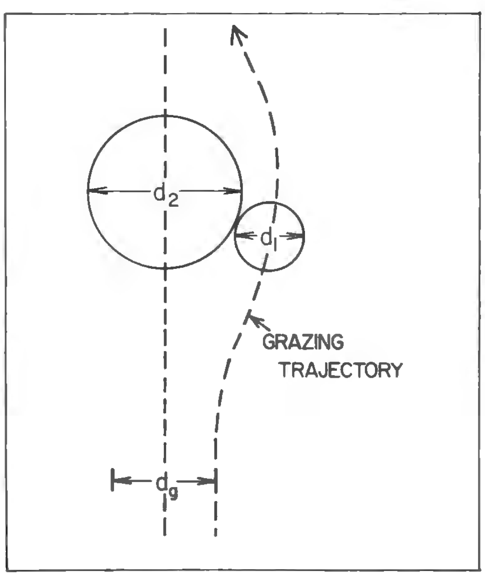


Fig. 3.12. Basic geometry of coalescence process for liquid drop and droplet. A grazing trajectory is indicated. The position of the droplet center at a long distance from the drop for a grazing trajectory defines d_g, which in turn defines the collision efficiency for the drop–droplet pair.

A *coalescence efficiency* E_2 is introduced to allow for the fact that droplet collisions may not result in coalescence of the droplets. The *collection efficiency* E is then defined by

$$E = E_1 E_2. \tag{3.23}$$

There is evidence that collisions may result in bounces rather than coalescence, especially in the presence of electric fields, and this point will be mentioned again. Unfortunately, the accuracy of laboratory experiments is not sufficient to permit good estimations of E_2.

The exact process by which the surfaces of droplets merge is not well understood. It is difficult to distinguish between grazing collisions followed by a rebound and those cases in which a very thin film of air always separates the drop–droplet surfaces. However, the coalescence efficiency is close to one in normal situations, so the search for collection efficiencies amounts in practical terms to a search for collision efficiencies.

The collision efficiencies for drop–droplet pairs have been studied in the laboratory and also mathematically. The mathematical treatments, all of which contain approximations, vary according to the sizes of the droplets, as these sizes determine which simplifying assumptions are possible in a given case. The pertinent results have been summarized by Pruppacher and Klett (1978, Chaps. 14, 15).

In the mathematical treatment, computer simulations of the trajectories of the drop and droplet have proven an extremely powerful analytical tool. Drop–droplet pairs are started with the droplet assumed to be a long distance below the drop and with its center displaced a certain distance from the vertical line through the center of the drop. The trajectory of the droplet with respect to the drop is then worked out in short time steps. Actual collisions in the computer simulations or (in some cases) approaches of the drop and droplet surfaces to within 1 nm of each other are accepted as evidence that a collision and, presumably, coalescence will occur. By working out the trajectories for various initial displacements from the vertical line through the drop center, the value of d_g appropriate to the drop–droplet pair can be determined. The calculation of E_1 is then straightforward.

One of the earliest studies of the drop–droplet coalescence problem indicated that no collisions at all would occur if the drop diameter were less than 38 μm (Hocking, 1959). The so-called Hocking limit of 19 μm (referring to the drop radius) played a large role in studies of coalescence from the time it was announced. Later studies indicated that E_1 is not exactly zero for d less than 38 μm. Although E_1 is very close to zero in such cases, the vast number of drop–droplet pairs in even a few cubic meters of cloud insures that some collisions will occur in a typical cloud.

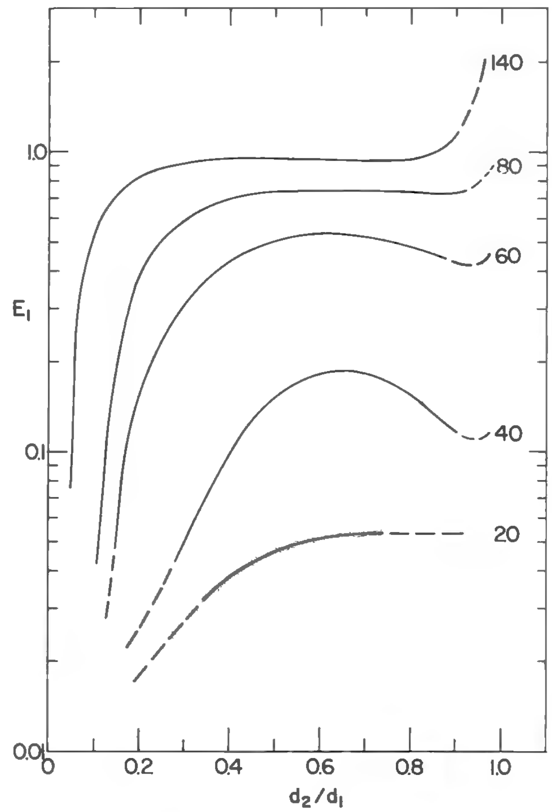


Fig. 3.13. Collision efficiency E_1 as a function of ratio of drop diameter d_2 to droplet diameter d_1, as deduced from computer simulations. Curves are labeled by drop diameter in micrometers. [After J. D. Klett and M. H. Davis (1973). *J. Atmos. Sci.* **30,** 107, by permission of American Meteorological Society and senior author.]

Values of E_1 as calculated by Klett and Davis are presented in Fig. 3.13 as an example. In their paper Klett and Davis (1973) compare their results with previous theoretical and experimental results and discuss possible reasons for differences. There is a considerable variation in results for those cases where d_1 and d_2 are almost equal. Some authors have reported $E_1 > 1$ in such cases, and attributed this to *wake effects*.

All the calculated collision efficiencies confirm that E_1 is strongly dependent on the sizes of both the drop and the droplet. Collision efficiencies are high when both the drop and droplet are large, a result in line with expectations based on elementary reasoning concerning the relative importance of inertial and viscous effects (Fig. 3.12).

Evolution of Cloud Droplet Spectrum by Coalescence

In the case of a raindrop collecting cloud droplets, a simple application of (3.16) is perfectly acceptable. A typical raindrop cloud collects hundreds of cloud droplets every second and, furthermore, each collision changes the characteristics of the raindrop by a negligible amount.

It might appear that (3.16) could be adapted to the coalescence of cloud droplets by introducing additional terms, notably E_1 and E_2. This approach has been tried by several authors, but it is not satisfactory. Applying a continuous collection model (3.16) is equivalent to assuming that each droplet captured is divided equally among collector drops in some larger size range. This would impart to each collector a slightly greater fall speed and improve slightly its chance of capturing another droplet.

The actual coalescence process is stochastic, and more favorable to droplet growth than is the continuous process (Telford, 1955). Obviously only one drop captures each droplet. The mass and fall speed of the collector drop are increased dramatically, while those of the other drops are unchanged. The "lucky" drop is therefore much more likely than the others to make a second capture, further increasing its competitive advantage. The ideas are illustrated in Fig. 3.14, which is taken from Berry (1967).

Further consideration shows that one must abandon the simple concept of collector and collected droplets, and treat the evolution of the entire cloud droplet spectrum. The evolution of an entire cloud droplet spectrum due to coalescence is complicated, because the drops of intermediate sizes are able to capture all droplets smaller than themselves while being subject to capture by still larger drops.

The evolution of a drop size spectrum by coalescence is handled best by a computer simulation. The cloud droplet spectrum is expressed by

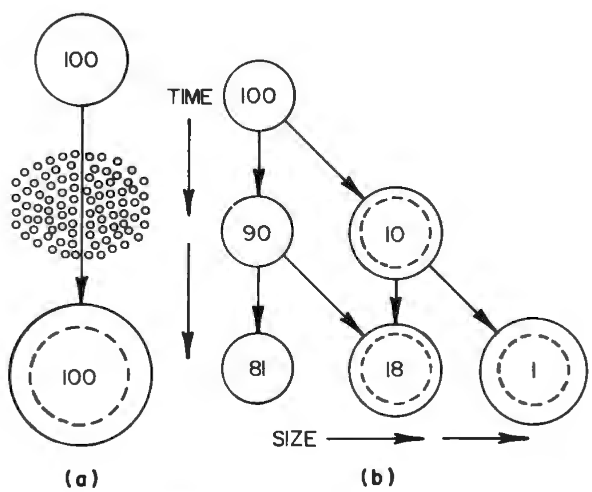


Fig. 3.14. The stochastic nature of the droplet coalescence process is illustrated here. Mass increases to the collector droplets are not distributed uniformly, as in the continuous model (a), but go to a few "fortunate" droplets (b), whose probability of further collision is greatly enhanced. [After E. X Berry (1967). *J. Atmos. Sci.* **24,** 688, by permission of American Meteorological Society and the author.]

specifying the number of droplets in each size interval within some specified cloud volume, say 1 m^3. Because volume (or mass) rather than the sum of the droplet diameters is conserved in coalescence, it is convenient to express the droplet spectrum in terms of droplet volume.

Let $n(v)$ be the density function for the cloud droplets, that is, $n(v)dv$ be the number of droplets per unit volume of space with droplet volumes between v and $(v + dv)$ [Fig. 3.15]. We define a *collection kernel* for droplets of volumes v and v', corresponding diameters d and d', and corresponding terminal speeds u_T and u'_T, respectively, as follows:

$$H(v,v') = \tfrac{1}{4}\pi(d + d')^2|u_T - u'_T|E(v,v'). \tag{3.24}$$

Consider now the change in n at a particular size v_0. Droplets of this size disappear by collisions with both larger and smaller droplets, but are created any time a droplet of volume v' (smaller than v_0) collides with a droplet of volume $(v_0 - v')$.

Mathematically, then,

$$\frac{\partial n(v_0)}{\partial t} = \frac{1}{2}\int_0^v H(v', v_0 - v')n(v')n(v_0 - v')\, dv' \\ -n(v_0)\int_0^\infty H(v_0, v')n(v')\, dv'. \tag{3.25}$$

The factor $\frac{1}{2}$ prevents collisions from being counted twice.

For a computer simulation it is necessary to break up the continuous size distribution into discrete increments of Δv. The collision efficiencies of Fig. 3.13 (or other reliable source) can be entered into the computer as a table. Several authors [e.g., Berry (1967); Long (1974)] have noted the computational advantages of analytic expressions as collection kernels and have suggested various ones, but without any conspicuous success.

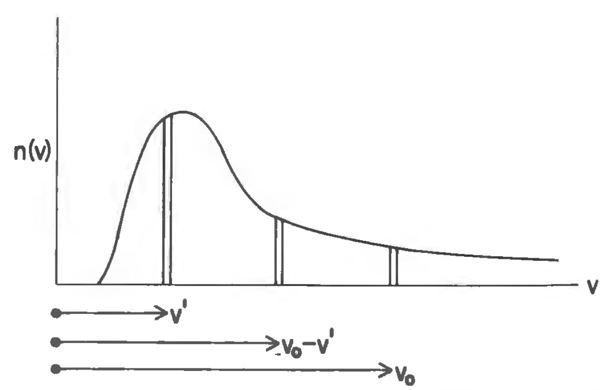


Fig. 3.15. **A hypothetical, continuous drop size distribution to illustrate the evolution by collision and coalescence.**

In the simulations, it is necessary to use sufficiently short time steps to avoid the problem of multiple captures by a single drop in one time step, or else complicate the problem further by invoking Poisson statistics to take account of the possibility of multiple captures by a single drop.

An example of the results of a computer simulation using the above ideas is presented in Fig. 3.16. Other simulations have been published giving results which differ in detail but not too much overall. The simulations show that the coalescence of droplets in a typical convective cloud can lead to the appearance of raindrops in 20–30 min.

The simulations confirm the importance of the initial cloud droplet spectrum. In general, coalescence for clouds of a given liquid water concentration is favored by having the water concentrated in relatively few droplets and by a broad droplet size spectrum. The typical continental cumulus cloud with its high droplet concentration and narrow drop size distribution, with no drops larger than, say, 15 or 20 μm, is unlikely to produce rain by coalescence in less than 40–60 min, according to some published results.

We noted earlier that condensation alone cannot lead to the formation of raindrop embryos. However, continued condensation as the particles in a convective cloud rise from the cloud base can cause sufficient growth to increase the collision efficiencies significantly. Therefore, conditions higher up in a convective cloud, particularly in its unmixed core, are normally much more conducive to rain formation by coalescence than condi-

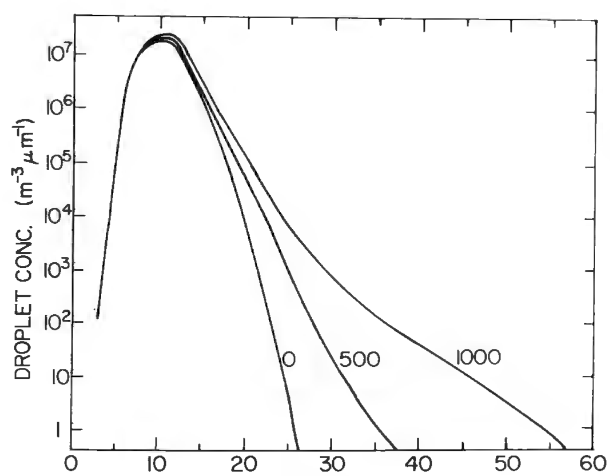


Fig. 3.16. Evolution of a cloud droplet spectrum with time as indicated by a computer simulation of the coalescence process assuming collection efficiencies of Klett and Davis (1973). Note the tendency for water mass to be transferred to larger droplet sizes. Curves are labeled in seconds. [After B. F. Ryan (1974). *J. Atmos. Sci.* 31, 1942, by permission of American Meteorological Society and the author.]

tions close to the base. The initiation of rain by coalescence in a typical cumulus cloud is dominated by events taking place in only a few percent of the total cloud volume containing the most favorable conditions for coalescence.

Although (3.25) is sometimes called the stochastic collection equation, there has been considerable debate in the literature over whether or not it *completely* captures the advantages of the stochastic process as compared to the continuous collection model [e.g., Gillespie (1972)]. The present thinking is that it does not. A straightforward application of (3.25) gives the *expected* result (in a statistical sense), but a given situation in nature may progress faster or slower depending upon the positions occupied by the cloud droplets at the starting time.

In comparing model predictions to developments in natural clouds, it is nearly impossible to distinguish these random variations from those related to local variations in liquid water concentration.

Evolution of Raindrop Spectra

Raindrops which form by the coalescence process in a cloud of liquid water droplets do not continue to grow indefinitely. Langmuir (1948) hypothesized that large drops of more than 5 mm diameter become hydrodynamically unstable and break into smaller drops. He theorized that these fragments of raindrops would serve as raindrop embryos to hasten the conversion of more of the cloud water droplets into raindrops. This process would appear to be particularly important in clouds with updraft speeds in excess of 6–8 m s^{-1}, which would be strong enough to support raindrops of such sizes.

Later laboratory studies by Brazier-Smith *et al.* (1973) and others have indicated that raindrop collisions are responsible for most if not all raindrop breakups (Gillespie and List, 1976; List, 1977). The collision of two raindrops leads to coalescence, a rebound, or breakup into several smaller drops, depending on the angle of impact as well as on the drop sizes and speed of impact (Fig. 3.17). Collisions lead to the breakup of many raindrops in the 2–4 mm size range, so that very few, if any, reach the size required to break up through their own hydrodynamic instability.

Raindrop size distributions at the ground are often found to obey an exponential law of the form

$$n_d \, dd = n_0 \exp(-\Lambda d) \, dd, \tag{3.26}$$

where $n_d \, dd$ is the number of raindrops with diameters in the range between d and $(d + dd)$ and n_0 and Λ are size distribution parameters

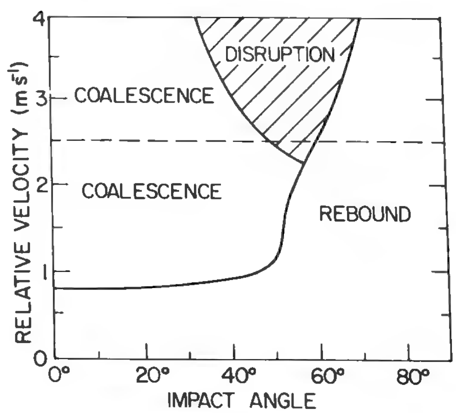


Fig. 3.17. **Expected result of collision between 0.9 and 0.3 mm drops as a function of impact velocity and impact angle. Horizontal dashed line shows difference in terminal speeds of the two drops. [Figure is adapted from one by Pruppacher and Klett (1978) based on work by Park at the University of Wisconsin. Replotted by permission of D. Reidel Publishing Company and senior author.]**

(Marshall and Palmer, 1948). In most observational samples n_0 is of the order of 80 m^{-3} mm^{-1} and Λ is around 2.5 mm^{-1}, with a slight tendency to decrease in heavy rain.

Computer simulations of the evolution of a population of falling raindrops growing by coalescence with cloud droplets and with one another, but subject to breakup by collisions, have shown that these processes lead to an exponential distribution, but the indicated values of Λ do not always agree with observations.

3.5 THE BERGERON PROCESS

The second principal source of precipitation embryos is mixed clouds (clouds containing ice and supercooled water). Bergeron (1935) was the first person to give a reasonably complete explanation of how mixed clouds give rise to solid precipitation embryos and hence to precipitation. The process described below is therefore called the Bergeron process, a name we shall use in this book, although the names of Wegener and Findeisen are also associated with it. It is sometimes called the cold cloud process, but it cannot occur in very cold clouds consisting entirely of ice crystals.

Cloud droplets often remain supercooled at temperatures down to almost $-40°C$ due to the scarcity of natural freezing nuclei. A single ice particle introduced in an otherwise supercooled water cloud tends to grow rapidly by deposition, because the ambient vapor pressure is that equiva-

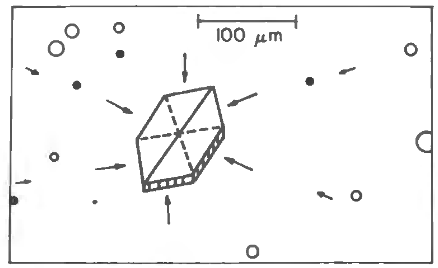


Fig. 3.18. An ice crystal in a supercooled cloud grows to embryo size (greater than 250 μm) in less than 10 min by deposition of water vapor. Droplets evaporate to maintain vapor pressure near saturation for water surfaces. Arrows indicate net diffusional drift of water molecules.

lent to water saturation. For example, a 10 μm spherical droplet frozen at −15°C in a water cloud will grow into a dendrite with diameter exceeding 250 μm in less than 10 min. By that time it qualifies as a precipitation embryo, and further growth occurs by both deposition and accretion (Fig. 3.18).

A simulation of the Bergeron process requires consideration of the depositional growth rates of ice crystals along their different axes, and the fall speeds and the *riming efficiencies* of the crystals.

Terminal Speeds of Ice Particles

The fall speeds of ice crystals and aggregates cannot be expressed as neatly as the fall speeds of water droplets because of the variety of shapes involved. Snowflakes typically fall with their shortest axes vertical, but undergo marked oscillations. Terminal speeds range upward to 1.5 m s^{-1} for aggregates. Wake effects are important; a crystal overtaking another tends to be drawn in behind it and increase its own speed. A number of suggested empirical formulas are noted by Hobbs (1974) and by Rogers (1976); Fig. 3.19 suggests the range of terminal speeds reported as a function of size by several investigators, notably Heymsfield (1972) and Locatelli and Hobbs (1974).

Riming

The riming efficiency of a crystal is a function of its size and shape, air density, and of the sizes of the cloud droplets to be collected. The problem has been investigated by aircraft sampling of clouds [e.g., Ono (1969)] and by computer simulations analagous to those used in studying droplet collisions [e.g., Pitter and Pruppacher (1974)]. A useful concept is the *riming threshold,* the crystal size below which no riming occurs. For plate

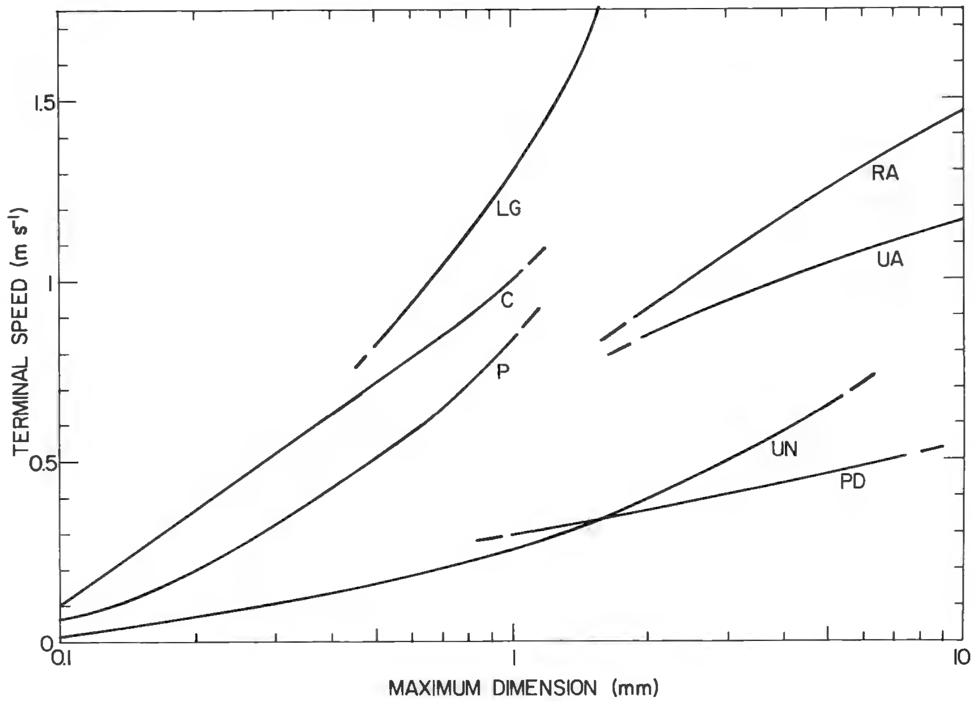


Fig. 3.19. Terminal speeds of ice crystals and snowflakes. Hydrometeor types and sources of information are as follows: LG—lump graupel of density 0.1-0.2 Mg m^{-3} at the ground (Locatelli and Hobbs, 1974). C—columns at 40 kPa, $-20°C$ (Heymsfield, 1972). P—plates at 40 kPa (Heymsfield, 1972). RA—rimed aggregates at the ground (Locatelli and Hobbs, 1974). UA—unrimed aggregates at the ground (Locatelli and Hobbs, 1974). UN—unrimed needles at 40 kPa (Heymsfield, 1972). PD—plane dendrites at 85 kPa (Heymsfield, 1972).

crystals, riming apparently begins when the *a* axes reach 0.3 mm. For columns, the *a* axes (the *thickness*) must reach 90 or 100 μm for riming to start, the exact value depending upon air density and cloud droplet size. Recent results are summarized by D'Errico and Auer (1978).

The riming of an ice crystal raises its temperature by release of latent heat of fusion, thereby inhibiting deposition. Therefore, the total growth of a crystal cannot be calculated simply by adding the growths due to deposition and accretion separately. This point is covered in a review by Neiburger and Weickmann (1974), who quote growth equations by Cotton corrected for this effect.

It should also be noted that the onset of riming (or of snow crystal aggregation) is a stochastic process, just like droplet coalescence. The "lucky" ice crystals that make the first few dozen captures in each cubic meter of cloud tend to dominate the subsequent evolution of the snowflake spectrum.

The liquid water in a cloud is obviously depleted by the presence of ice crystals. Unless continued cooling of the air, for example by ascent in a convective cloud updraft, continues to make fresh supplies of supercooled water available, mixed ice and water clouds tend to become ice clouds as the water droplets evaporate and the water vapor is deposited upon the ice crystals.

Simulations of the Bergeron process taking account of both depositional and accretional growth of the ice particles show that introduction of 10^5 ice particles per liter dries up a typical cloud in only 1 or 2 min. However, introduction of one to ten ice particles per liter allows growth of the ice particles to continue for some 20 min before the cloud water is completely used up [e.g., Beheng (1978)]. By that time the ice particles usually have grown into snowflakes or small hailstones.

Rain Formation by Melting Snow

The Bergeron process is not limited to situations yielding snow at the earth's surface. In many situations precipitation initiation through the Bergeron process leads to rain at the ground as snowflakes, graupel, or hail melt after falling through the 0°C isotherm.

One point worth noting is that wet snowflakes near the 0°C level tend to form large aggregates. The raindrops that result from the melting of solid hydrometeors tend to be larger than those forming by coalescence [e.g., Gillespie and List (1976)]. Observations by Doppler radar[1] in Illinois suggest that raindrops formed by melting snowflakes require about 1 km of fall before collision-induced breakups and coalescence of fragments produce the typical exponential raindrop distribution. A modeling study by Gillespie and List (1976) suggesting that the equilibrium drop size distribution would be achieved in heavy rain in less than 2 km of fall is in reasonable agreement.

Geographic Extent

The Bergeron process is predominant at middle and high latitudes and takes place all over the earth. Even in the tropics many storm clouds tower above the 0°C isotherm. However, it is thought that most tropical

[1] Weather radar is a vital tool in nearly all research into precipitation formation. Because the energy scattered back to the radar antenna by a hydrometeor varies to a first approximation as the sixth power of its diameter, radar sets can readily distinguish regions of precipitation from regions of clear air or regions containing cloud particles only. Doppler radars measure frequency shifts in the backscattered radiation, permitting estimates of particle speeds toward or away from the antenna [see Battan (1973)].

clouds form rain first by coalescence and that the Bergeron process is active in only the later stages of the clouds' lifetimes.

The relative importance of the Bergeron and coalescence processes is a function of season. Over the eastern United States, for example, rain often forms by coalescence in the convective clouds of summer, with their warm bases and high liquid water concentrations, but the Bergeron process appears to be the more important mechanism during the winter.

The melting of snow to rain in the widespread stratiform systems characteristic of eastern North America, for example, is well known to radar meteorologists, as the melting snowflakes give rise to a characteristic intensification of radar returns called the bright band [e.g., Battan (1973)] (Fig. 3.20).

Observations in extensive frontal cloud systems tend to show abundant ice particles at the upper (cirrus) levels, with lower concentrations further down in the cloud mass. In some situations the cloud decks are not continuous from the surface to the cirrus level, and one observes ice crystals falling from the cirrus level to seed the lower cloud layers. Some authors have called the upper cloud layers in such a case "seeder clouds" and various names have been proposed for the lower cloud layers. Hall (1957) and others have examined these problems using cloud physics data collected from aircraft. Their data emphasize the wide range of hydrome-

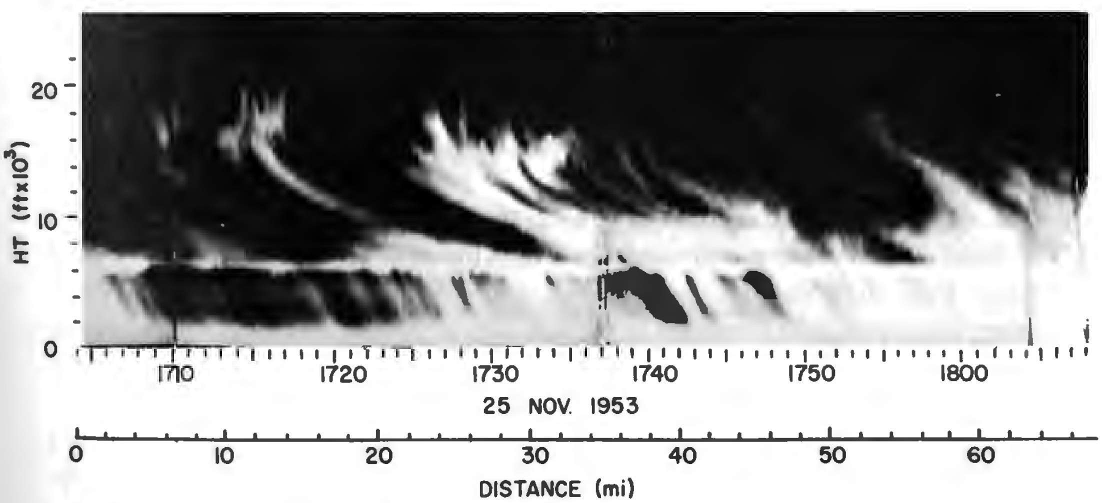


Fig. 3.20. The Bergeron process in action. This time–height section obtained by a vertically pointing 3 cm radar at Montreal in November 1953 shows snow generating cells 5–6 km (16–19,000 ft) above ground, the trails of falling snow stretched in the horizontal by the wind shear, the bright band at 2 km (6500 ft) indicative of wet snow aggregates near the 0°C isotherm, and the raindrops below the 0°C level falling to earth at higher speeds (and hence more nearly vertically) than the snowflakes above it [photo courtesy of R. R. Rogers and McGill University].

teor types and the varied interactions which take place among them in a complex cloud system straddling the 0°C isotherm.

3.6 HAILSTONE GROWTH

Solid hydrometeors growing by accretion in regions with updrafts in excess of 10 m s^{-1} and an abundance of supercooled cloud water or rainwater rapidly take on the characteristics of hailstones rather than rimed snowflakes. There is a continuum of situations with rimed snowflakes giving way in turn to spherical snow pellets, to graupel, and finally to hail as the water concentration increases. Hydrometeor densities may be as low as 0.2 Mg m^{-3} for snow pellets, but range upward to 0.9 Mg m^{-3} for clear hailstones. By convention, only hydrometeors with diameters exceeding 5 mm are called hailstones.

The typical hailstone is an ellipsoid. It falls with the shortest axis oriented vertically or precessing about the vertical. The shortest axis is typically 0.5–0.8 times the length of the longer axes (English, 1973; Macklin, 1977).

Examination of a hailstone usually shows a distinct inner part called the hail embryo. The hail embryo can be up to 5 mm in diameter. It can usually be identified as a frozen raindrop or as a rimed snowflake, often a dendritic crystal [e.g., Knight and Knight (1970)].

Observational and theoretical work on the subject of hailstone growth goes back at least to Schumann (1938). He identified the basic problem: How does the hailstone dispose of the latent heat of fusion released as accreted water freezes? In the 1950s List used a wind tunnel for hail growth experiments at Davos in Switzerland, while additional theoretical and field studies were pursued by Macklin (1963), Ludlam (1958), and others. Canadian researchers modeled hailstone growth in regions containing supercooled rainwater [e.g., Douglas (1963)], and their ideas were amplified in extensive hail research in the U.S.S.R. [e.g., Sulakvelidze *et al.* (1967)].

The increasing availability of computers led to attempts by Musil (1970), English (1973), and others to model hailstone growth. As hailstones grow by accretion, it is necessary to build into the models what is known about hailstone fall speeds. The fall speeds can be calculated if the drag coefficients (C_D) are known. Figure 3.21 presents an estimate of the drag coefficient of a hailstone as a function of diameter. The drag coefficients of typical stones are near 0.6 over a considerable size range, so to a first approximation the fall speed of a hailstone varies as the square root

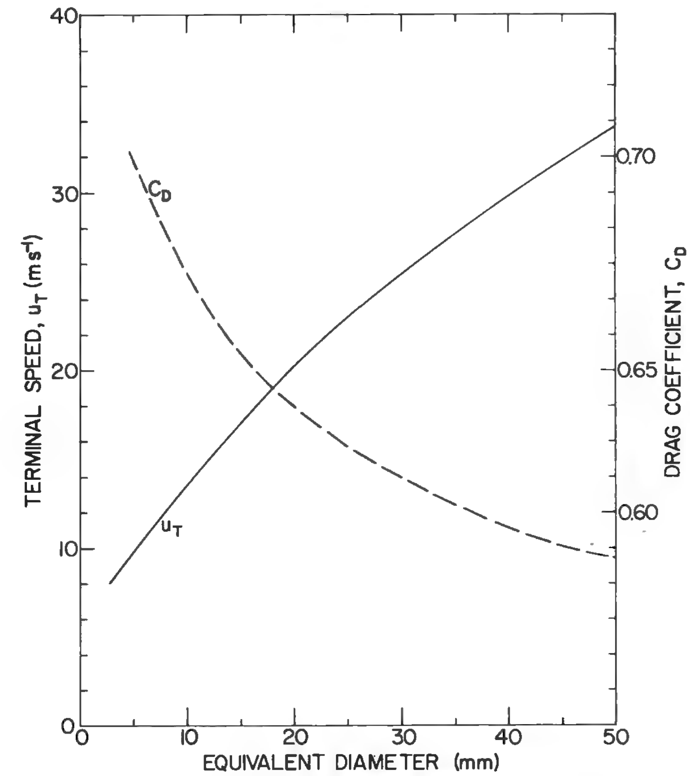


Fig. 3.21. Drag coefficients and fall speeds of hailstones as a function of diameter. Experimental values of C_D for stones with axis ratios of 1:1:0.7 are shown based on W. C. Macklin and F. H. Ludlam (1961). *Q. J. R. Meteorol. Soc.* **87,** 72. Terminal speeds were calculated for given values of C_D at 70 kPa and 0°C.

of the diameter (Fig. 3.21). An individual stone may depart widely from the indicated values of C_D because of irregularities in shape and variations in surface roughness.

A Model of Hailstone Growth

The Musil (1970) model of hailstone growth calculates a dry growth rate, assuming all accreted water retained, and a wet growth rate, assuming only the *frozen* water retained, for each time step, and assumes that the *smaller* of the two rates actually applies.

The dry growth rate, in the absence of supercooled raindrops, is

$$\frac{dm}{dt} = \frac{\pi d^2 u_T}{4} [\chi_l E_l + \chi_i E_i], \tag{3.27}$$

where d is hailstone diameter, u_T is its terminal fall speed, χ_l and χ_i are the concentrations of cloud liquid and cloud ice, respectively, and E_l and E_i

are the collection efficiencies for cloud liquid and cloud ice respectively. E_l was assumed to be 1.0 for all embryos and hailstones more than 200 μm in diameter, although experimental work by Macklin (1977) indicates that it would fall off as the hailstones grew beyond 10 mm. E_i is not well known for dry hailstones; most model runs assumed 0.25 on the assumption that most of the ice crystals encountered would simply rebound.

The wet growth equation is more complex and is not shown here. It is derived by considering the heat balance of the stone, assumed to be at 0°C, and requiring the total heat exchange to be zero. That is,

$$q_1 + q_2 + q_3 + q_4 = 0, \tag{3.28}$$

where q_1 is the heat conducted away from the stone, q_2 is the heat loss by evaporation, q_3 is the heat added to the stone by accreted cloud water (latent heat of fusion less the heat required to warm the accreted cloud water, both that retained and that shed, to 0°C), and q_4 is the heat required to heat the accreted ice crystals to 0°C. q_4 proved to be an important part of the heat balance. Because all intercepted ice particles would presumably stick to a wet hailstone, E_i was set at 1.0.

For the model results to be meaningful, realistic ambient conditions must be assumed. The model was run to predict hailstone growth in a variety of one-dimensional cloud models considered representative of clouds in the north central United States.

An important outcome of Musil's calculations was the indication that hailstones in the 20–30 mm size range can be in the wet growth state at temperatures as low as −30°C, and that the most rapid growth occurs at such temperatures, rather than just below 0°C as some previous authors had supposed.

Use of the model showed that particles of 50–100 μm diameter introduced at various places in the cloud models could sometimes grow into hailstones of up to 10 mm diameter in a single ascent in the updraft. The results emphasized that a "tuning" process was at work, with some embryos falling out from the lower part of the cloud, many being ejected into the anvil, and only a few achieving maximum growth (Musil, 1970).

As noted, the Musil program assumes that all excess water is shed from the stone. Some model calculations were run to check the suggestion that hailstones can grow as spongy structures with some of the excess water retained (Dennis and Musil, 1973). The model simulations indicated that such stones in regions of high supercooled water concentrations would grow rapidly into "blobs" of 50 or even 100 mm diameter containing only 1 or 2% ice. It may safely be concluded that such objects, like large raindrops, would break up as a result of collisions.

It has been noted that hailstones spin and oscillate about their vertical axes as they fall, which would tend to throw off excess water. Quite apart from the irregularities in hailstone motions, Chong and Chen (1974) examined the hydrodynamic forces on falling (nonrotating) ice spheres surrounded by water shells, and concluded that the water would be torn away, except for a very thin film. This conclusion is in line with hailstone observations at the ground. Although "slushy" hailstones have been seen occasionally, the calorimetric evidence [e.g., Gitlin *et al.* (1968)] says that most hailstones are almost solid ice. The shedding of water from hailstones in wet growth has been suggested by Dennis (1977) as an important source of new precipitation embryos in hailstorms.

The modeling of hailstone growth has been extended to two-dimensional cloud models by English (1973) and by Musil *et al.* (1975), and to a three-dimensional cloud model by Paluch (1978). These models permit one to simulate such special situations as hailstone growth along trajectories passing over a region of weak radar echo[2] in a supercell hailstorm [e.g., Browning and Foote (1976)]. The models indicate that recycling in a complex storm is likely responsible for most hailstones at the ground with diameters exceeding 25 mm. This result is in line with surmises of earlier investigators and indications from recent field projects involving storm scans with multiple Doppler radars.

[2] The absence of radar echo (and hence of precipitation) is attributed to updrafts so strong (~20–30 m s^{-1}) that the cloud droplets have insufficient time to coalesce into precipitation before reaching the cloud top and being ejected into the thunderstorm anvil.

CHAPTER

IV

Concepts and Models for Cloud Modification

4.1 INTRODUCTION

Now that we have described the factors controlling the formation of water clouds, the appearance of ice particles, and the formation of precipitation, we are in a position to assess the chances of success for the various ways that have been proposed to modify clouds artificially.

Once the basic discoveries of Langmuir and Schaefer were announced, many persons developed ideas on how clouds might be modified to obtain specific operational objectives, for example, the suppression of hail. The important ones were summarized by Howell (1966) under the title "Conceptual Models That Guide Applied Cloud Seeding."

Every conceptual model has the grave weakness that it tends to follow a chain of hypothesized cloud seeding effects to a desired end result without considering side effects. A determination of whether a particular conceptual model applies to the atmosphere requires quantitative investigations including, in some cases, field experiments. However, a selection of the more promising ones for further investigation can be carried out using basic physical reasoning, a process which is attempted in this chapter.

4.2 MODIFICATION OF CCN SPECTRUM

Figure 4.1 shows the principal conceptual models advanced for modifying the microphysics of clouds by artificially altering the CCN spectrum. The models are discussed below.

Artificial Formation of Clouds

We begin by examining the question of whether or not it is possible to produce a cloud artificially by supplying CCN. This would require a situation where air is supersaturated but no clouds form because of the absence of CCN. This situation occurs, but only rarely. Above the hot pools of Yellowstone National Park, for example, supersaturation ratios sometimes reach 3 or 4. The atmosphere there contains very large amounts of water vapor drawn from the hot pools, while the constant diffusion of water vapor upward and away from the surfaces of the hot pools drives

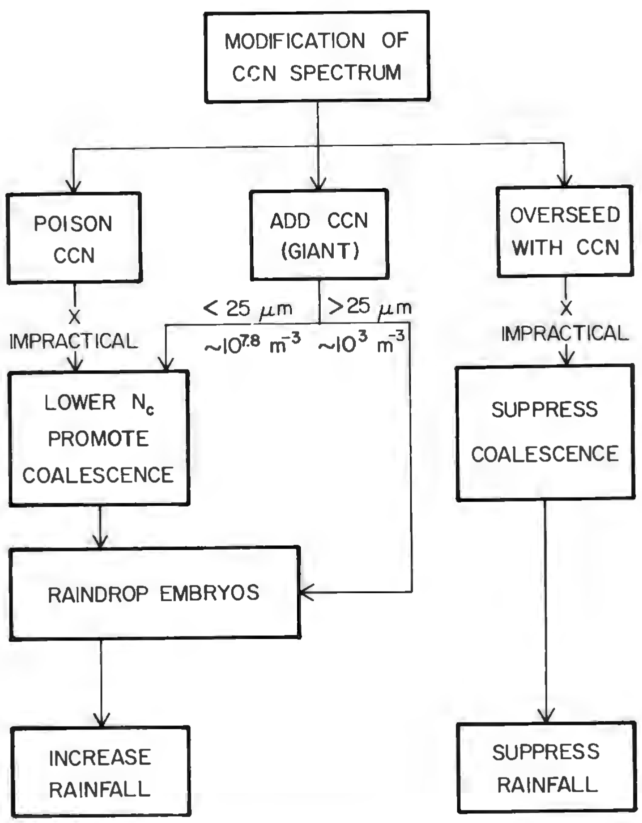


Fig. 4.1. Conceptual models for modifying the microphysics of clouds by altering CCN spectrum.

away most of the CCN by diffusiophoresis. Experimenters on the Yellowstone field research expeditions organized by Schaefer in the 1960s demonstrated many times that clouds could be produced over the hot pools simply by burning a match or otherwise introducing hygroscopic particles capable of acting as CCN (Fig. 4.2).

Any attempt to extend this approach to the atmosphere in general is doomed to failure. The atmosphere ordinarily contains large numbers of particles capable of acting as CCN. Even if the most active CCN were removed, there would still be myriads of additional particles which could be activated at slightly larger than normal saturation ratios, say around 1.02 or 1.03. We conclude, therefore, that the occurrences of high super-

Fig. 4.2. Formation of a cloud by burning a match. (a) A thermal pond in Yellowstone Park where the water temperature is about 85°C and the ambient air temperature above the pond is approximately − 15°C. Even though the air immediately above the pond is supersaturated, no cloud is formed because there are so few CCN in the clean air. Clouds are forming over the warm ponds in the background because the air in that area is contaminated with particles from an unknown source. (b) Cloud condensation nuclei are supplied by burning a single match above the pond. The combustion products from this match provide an abundant supply of CCN, and a small cloud is formed when nature would not otherwise produce one [photos by Atmospherics Incorporated, Fresno, California].

Fig. 4.2. Continued.

saturation ratios which can be exploited simply by adding artificial CCN are very limited indeed and of no practical importance.

Woodcock and Spencer (1967) attempted to heat an *almost* saturated atmosphere by releasing finely powdered sodium chloride (NaCl) in concentrations of 40 mg per kilogram of air. They reasoned that the heat of vaporization released might provide sufficient added buoyancy to produce a convective cloud. Calculations of the temperature increases which could be achieved in this way suggest that they would amount to a fraction of 1°C.

Woodcock and Spencer conducted their experiments in the warm, moist marine layer over the Pacific Ocean off Hawaii. Salt powder was released from aircraft in layers where the relative humidity was in the range of 80–90%. Although visible clouds were produced, it appears that they consisted of solution droplets in equilibrium with the ambient vapor pressure, and that none of the droplets were driven past their critical diameters. Aircraft observations gave 0.4°C as the modal point of the distribution of resultant temperature increases. As updrafts in vigorous cumulus are often 1–3°C warmer than the ambient air, it does not appear that the experiments had much chance of producing significant convective clouds, especially as the marine layer is ordinarily capped by a subsidence inversion.

Although one must be alert to the possibility of success in some future experiments, it appears safe to say for the time being that none of the attempts to produce clouds artificially by adding artificial CCN has had any pronounced success. The topic of producing clouds artificially will come up again under the headings of glaciogenic seeding and inadvertent weather modification.

Changing the Microphysics of a Cloud

This could be a matter of practical importance for improving visibility in fog and for modifying precipitation. The reader will recall that maritime clouds produce rain more readily than do continental clouds, and that the difference is due principally to the differences in the maritime and continental aerosols which provide their CCN.

It has been suggested that one could cause continental clouds to assume the drop size distribution characteristic of maritime clouds and thereby increase their chances of producing precipitation. The "problem" with continental clouds is a cloud droplet concentration (N_c) higher than optimum for rain formation by coalescence. It might appear that a lower N_c could be obtained by nucleus *poisoning,* that is, the addition of some chemical which would deactivate many of the CCN. However, one can raise many objections to this approach. No chemical is readily available which has been demonstrated to produce the desired nuclei poisoning effect. A more basic objection is that, no matter how many CCN are poisoned, there would still be numerous particles near the upper limit of the Aitken size range which would be capable of serving as CCN if the more suitable particles were removed.

A more promising approach, in theory, would be to add artificial CCN of a sufficient size and in sufficient quantity to prevent the activation of the natural CCN. Computer simulations of the problem indicate that the shape of the tail of the size spectrum is the crucial factor in determining how many of the CCN present, whether natural or artificial, will be activated. By introducing particles with diameters of the order of 1–3 μm in concentrations of 25–100 cm^{-3} one could, in certain cases, ensure the formation of a "maritime" cloud even in the most heavily polluted continental air mass. The large artificial CCN would simply capture the available moisture and prevent the more numerous, but smaller, natural CCN from participating in the cloud formation process.

While the concept just outlined is theoretically feasible, it runs into great difficulties from a practical standpoint. To begin with, let us compute the mass of artificial CCN required to seed an entire cloud. Assuming that 2 μm NaCl particles are used and that the desired particle concentration is 50 cm^{-3}, we find the required seeding rate for a vigorous cumulus cloud ingesting 10^6 m^3 of air s^{-1} to be 30 kg min^{-1}. The logistics involved in applying this seeding concept to an entire cloud are obviously formidable.

One could ease the problem somewhat by noting that precipitation acts as an *infection*. That is, once precipitation gets underway anywhere in a cloud, it spreads as raindrops and fragments are carried about by the cloud's internal circulations and by turbulence. Therefore, a "oneshot" seeding of, say, 10^6 m^3 of air might work. However, there is another compromise approach available, which will now be described.

Artificial Raindrop Embryos

A second compromise approach to the promotion of coalescence in water clouds is the introduction of artificial raindrop embryos. Instead of changing the entire cloud droplet spectrum to speed the coalescence process, one simply bypasses the initial phases by introducing particles big enough to function as raindrop embryos right away.

A direct method of introducing precipitation embryos into a cloud is to spray it with fine water droplets. Experiments along these lines over the Caribbean Sea in 1954 by the University of Chicago showed conclusively that this procedure increased the probability that a cumulus cloud would produce a radar echo, indicative of the appearance of rain inside the cloud, and that the time required for the appearance of an echo was reduced by about 5 min (Braham *et al.,* 1957).

The water spray seeding method has the disadvantage that very large quantities of water must be transported to a cloud by aircraft to produce any detectable effect. It is not an economically attractive proposition.[1] One way to reduce the logistical problem is to treat the cloud with hygroscopic agents, either as dry particles or spray droplets, which form raindrop embryos by their own hygroscopic action. This is roughly equivalent to providing the giant CCN that play a role in the formation of some rain showers over the sea.

For the artificial raindrop embryos to be effective, the giant particles introduced must be some tens of micrometers in diameter. The particle must yield an embryo well beyond the Hocking limit (d = 38 μm), and

[1] Todd (private communication) has estimated the return from water spray seeding of cumulus congestus at 10^6–10^7 m^3 of rainfall per cubic meter of spray.

one cannot assume that there will be enough time for the seed to take on enough water vapor to assume its equilibrium radius according to (3.10) [see Woodcock and Spencer (1967)] before it has to start working as a raindrop embryo.

Calculations by Biswas and Dennis (1972), for example, tracing the history of individual raindrop embryos through a one-dimensional cloud model, suggest that the use of NaCl particles more than 120 μm in diameter can lead to rain formation in certain convective clouds some 10–12 min after introduction of the particles at cloud base. If smaller particles are used, the solution droplets will grow very slowly and will often be ejected at cloud top without ever growing large enough to start falling back down against the updraft.

A recent treatment by Klazura and Todd (1978) follows Biswas and Dennis (1972), but considers condensational as well as accretional growth of the embryos. Klazura and Todd (1978) have run their growth model in a 1-dimensional cloud model with updrafts invariant in both space and time to investigate effects of particle size, updraft speed, cloud depth, and cloud base temperature (Fig. 4.3). They conclude that hygroscopic seeding is promising in continental clouds with cloud base temperatures above 0°C, and very promising for cloud bases warmer than 10°C.

The appropriate size for the seeds increases with updraft speed. For a cumulus of moderate depth (5 km) and with a moderate updraft (12 m s^{-1}), NaCl particles introduced at the base must be more than 40 μm in diameter for the resultant droplets to avoid ejection at the cloud top.

The requirement for hygroscopic seeds of 50–100 μm poses a serious logistical problem. If it is assumed that the embryo concentration required

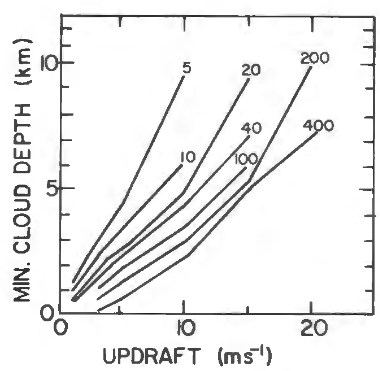


Fig. 4.3. Results of a computer simulation of releases of hygroscopic seeds at base of a convective cloud near 3 km and + 10°C. Minimum cloud depth required for a particle to grow large enough to fall against the updraft is shown as a function of updraft speed for various particle diameters (labeled in micrometers). [After G. E. Klazura and C. J. Todd (1978). *J. Appl. Meteorol.* **17,** 1758, by permission of American Meteorological Society and senior author.]

to speed the coalescence process is 1000 m^{-3}, the number of particles required to seed an entire convective cloud is of the order of 10^{15}. A requirement for 10^{15} particles of 100 μm diameter translates to a need for tons of seeding agent per cloud.

The situation becomes more promising if one assumes that the raindrops formed about the large artificial embryos would undergo breakup and thereby create more raindrop embryos. For this to occur, raindrops of 2–3 mm diameter would have to be produced and retarded sufficiently in their fall through the cloud to undergo the necessary collisions and breakup. The process is unlikely in clouds where updrafts do not exceed 5 m s^{-1}.

Biswas and Dennis (1972) reported that each cycle of growth, breakup, and growth would occupy 3–4 min, in the cloud they modeled, and Klazura and Todd (1978) estimated 2 min for some of their model runs. Both pairs of authors assumed breakup at d = 5 mm due to hydrodynamic instability (Langmuir, 1948), rather than collision-induced breakup at d = 3–4 mm [e.g., Brazier-Smith *et al.* (1973)], so their estimates of the time required are a little high. Nevertheless, it remains problematical whether the breakup and recycling process could infect an entire cloud before the natural precipitation processes would achieve the same result.

Hygroscopic seeding has also been modeled in more advanced field-of-motion cloud models [e.g., Farley and Chen (1975)]. The results so far have not resolved the issue.

A reasonable compromise appears to be the use of 25–50 kg of NaCl or other hygroscopic powder per cloud for experiments on clouds of moderate size. This is within reason, and has been realized in several experiments. The experiments are described in Chapter VII.

Overseeding with CCN

To this point we have been considering implicitly that it would be a desirable objective to increase the droplet sizes in water clouds to facilitate rain formation. If one wished to pursue the opposite objective (delay of onset of rainfall), there are again concepts worth exploring. To convert maritime clouds to continental clouds, at least in terms of their microphysics, it would be necessary to introduce additional CCN. Such an action would be an example of *overseeding,* which is loosely defined as suppression of precipitation by interference with accretional processes (Fig. 4.1).

While overseeding with CCN could delay the formation of rain in principle, it would not likely work in practice. The artificial CCN would have to be of a size comparable to the natural CCN present and be provided in concentrations of hundreds per cubic centimeter. The logistical problems would be greater than those involved in hygroscopic seeding due to the larger CCN concentrations required. Furthermore, in this case the fact that precipitation acts like an infection would work *against* the cloud seeder. It would be necessary to treat *all* portions of a target cloud because, once precipitation appeared anywhere in it, the raindrops and raindrop fragments would be circulated throughout the cloud by its organized internal motions and by turbulence. We conclude that overseeding with CCN to modify precipitation is not a practical concept.[2]

4.3 GLACIOGENIC SEEDING FOR MICROPHYSICAL EFFECTS

Glaciogenic seeding is defined as seeding designed to add ice particles to clouds or to portions of the clear atmosphere. This objective can be accomplished in two ways, (1) by chilling the air to temperatures below −40°C, where homogeneous ice nucleation takes place, or sufficiently to activate the natural ice nuclei present, and (2) by adding artificial ice nuclei capable of producing ice particles by deposition, condensation-freezing, or freezing existing cloud droplets or raindrops. The mechanics of generating and applying artificial ice nuclei are treated in Chapter V.

Artificial Formation of Ice Clouds

It will be recalled that the production of a water cloud by adding CCN to the atmosphere works only on rare occasions, because CCN are abundant. Natural ice nuclei (IN) are quite scarce, so glaciogenic seeding can produce demonstrable effects over a much wider range of conditions than can seeding with artificial CCN.

It happens quite frequently in the atmosphere that the ambient vapor pressure is intermediate between that required for ice saturation and that required for water saturation. However, the ice clouds which could be supported sometimes fail to appear because of a lack of suitable IN. In such a case, glaciogenic seeding can form an ice crystal cloud, which may prove very persistent [e.g., Bigg and Meade (1971)].

Aircraft contrails which persist for a long time apparently form in air

[2] The U.S. Air Force did make a practical use of overseeding with CCN to suppress contrails. The water vapor in the exhausts was condensed into droplets so fine as to be invisible for practical purposes.

supersaturated with respect to ice. There are also short lived contrails which depend on water vapor in the aircraft engine exhaust. They disappear in a matter of minutes as they mix with the ambient air and the vapor pressure drops below that required for ice saturation.

Artificial Snowflake Embryos

Glaciogenic seeding of certain supercooled clouds leads to the formation of precipitation particles through the Bergeron process. That this should happen is fairly obvious. Whether it leads to a net increase in precipitation at the ground is a more complex question. Ice particles produced artificially in a supercooled cloud grow at a rate depending upon their concentration and upon the temperature, air density, and the concentration of supercooled water available to support their growth. They are governed by the same growth equations (Chapters II and III) as the natural ice particles.

For glaciogenic seeding to have much impact on the total supply of precipitation embryos, there must exist parcels of supercooled cloud which can remain in that state for at least 5 or 10 min before being glaciated by natural IN. The principal opening to be exploited in glaciogenic seeding is that very few natural IN are active at temperatures higher than -12 to -15°C. This fact is of particular importance because the time required for an ice crystal to reach its riming threshold apparently is minimized if it can grow in the temperature range from -5 to -15°C (Chapter III).

In certain types of clouds, particularly short lived orographic clouds with tops warmer than -20°C, the required conditions for artificial intervention are met. Supercooled cloud water also exists in the upper portions of convective clouds, but there is a question concerning how long it persists before being carried higher and frozen by natural IN. In certain maritime clouds where the ice multiplication process is operative, it appears that no more than 5 or 10 min may elapse before a given cloud tower becomes seriously infected with ice particles around the -5°C level. In these situations the opportunities for producing precipitation embryos by glaciogenic seeding would appear restricted.

Effects on Crystal Habits and Fall Speeds

Ice crystal habit is a function of the temperature at which the crystal grows. This means that glaciogenic seeding to produce ice particles at higher temperatures than where they would otherwise appear changes not only the location of the appearance of ice crystals in a moving parcel of air, but the types of crystals produced. In many natural clouds ice particles are scarce at temperatures higher than -15 to -20°C, so that most of the crystals are small plates. If seeding induces ice crystal formation in the region around -12 to -15°C, it induces the formation of dendrites. If nucleation can be induced at still higher temperatures, say around -5°C, the result is the formation of needles and columns. These distinctions may be important as the various types of crystals have different depositional growth rates and fall speeds for a given mass, and also have different collection efficiencies for capture of the cloud droplets, i.e., for riming. Cooper (1978) has suggested that the production of columns near -5°C is especially favorable to formation of graupel, and investigation of this suggestion is continuing.

Changes in crystal habits by seeding implies changes in fall speeds. In addition, one must consider that the glaciogenic seeding to produce, say, 50–100 ice crystals liter^{-1} will suppress riming [e.g., Jiusto and Weickmann (1973)]. Rimed crystals fall more rapidly than unrimed ones, so the possibility of *snowfall redistribution* by seeding is very real.

It would be easy to underestimate the extent to which snowfall could be moved by seeding. Rimed snow aggregates often reach the ground as much as 50–75 km from the places where the individual crystals started to grow. Simply by eliminating riming, one could reduce the fall speeds from say 1 m s^{-1} to around 0.5–0.7 m s^{-1} (Fig. 3.19). This in turn could cause the snowflakes to drift another 20 or 30 km before reaching the ground.

Snowfall redistribution will be mentioned at appropriate points in later chapters.

Overseeding

The reader will recall that production of ice particles in a concentration of 10^8 m^{-3} can completely glaciate a parcel of cloudy air in a minute or so. In such a case, the mass of each of the resultant ice crystals will be about five times that of one of the cloud droplets originally present. Therefore, none of the particles will be large enough to capture other ice particles or droplets, much less to fall from the cloud and reach the ground without evaporating. Inducing this effect by glaciogenic seeding agents is an extreme example of overseeding.

In less extreme cases, the ice crystals will have sufficient fall speeds to aggregate into snowflakes, provided enough time is available, so overseeding is not much of a problem in long lived stratiform clouds which extend downward to near or below the 0°C level [e.g., Jiusto and Weickmann (1973)]. However, in short lived orographic or convective clouds the possibility of overseeding is real.

The result of overseeding a convective cloud may be the production of a larger than usual cirrus anvil, which may persist after the parent cloud has dissipated. It should be noted, though, that detached or "orphaned" cirrus anvils are characteristic of even natural convective clouds in their dissipating phases, and may persist for many hours. Therefore, the mere observation of detached anvils should not be taken as proof that cloud seeding has occurred, much less as proof of overseeding. The overseeding effect may be more subtle: a change in precipitation size distribution and increased evaporative losses between cloud base and the ground. This last effect appears to be important in arid regions like New Mexico, where the bases of cumulus clouds are sometimes as much as 5 or 6 km above sea level (Workman, 1962).

Hail Suppression Concepts

The formation of hail is such a complex phenomenon that much of one chapter will be devoted to the recent thinking concerning the possibility of its suppression by cloud seeding. In the framework of this chapter, it is appropriate to note two simple concepts that have been advanced for suppression of hail by glaciogenic cloud seeding. They are the *cloud glaciation* and *competing embryo* concepts [e.g., Howell 1966)].

It will be recalled that the growth of hail depends principally upon the availability of supercooled cloud droplets and/or supercooled raindrops in a cloud with updrafts strong enough to support the growing hailstones.

The cloud glaciation concept for suppression of hail through glaciogenic seeding is that one might succeed in freezing all of the supercooled cloud droplets at some specified temperature, say -5°C, thereby eliminating the possibility of hailstone growth. We shall see in Chapter VIII that the quantities of seeding agent required to do this in a strong updraft are totally unrealistic.

A more realistic hypothesis for suppression of hail through microphysical effects is the competing embryo hypothesis. The competing embryo hypothesis is based on the assumption that the amount of hail to fall from a storm is a fixed quantity determined perhaps by the vapor flux through the base of the convective cloud, the updraft speeds, the temperature at which natural glaciation occurs, and the fraction of the cloud water which can be collected by growing hail embryos before natural glaciation occurs. The competing embryo hypothesis holds that by introducing additional freezing centers in the hail formation zone the size of the resulting hailstones will be decreased.

Studies of crop damage by hail show that it would often be decreased by a reduction in hailstone size, which implies a reduction in hail impact energy, even if the total hail mass falling per unit area were unchanged [e.g., Wojtiw and Renick (1973)]. Dividing the total mass of hail among more stones also increases the fraction of the hail melted before reaching the ground.

Early analyses of the concept by Iribarne and de Pena (1962) and others suggested that this technique might have a chance of success. However, there are basic objections to the competing embryo hypothesis. There is no guarantee that the amount of hail to be produced by a convective storm is a fixed quantity. It is quite conceivable that the introduction of additional freezing centers in a region where supercooled cloud droplets are available for hail growth might only serve to increase the number and total mass of hailstones reaching the ground.

Events in a hailstorm are critically dependent on timing. The situation cannot be handled at all in simple conceptual models which view a convective storm as a relatively fixed object. The total impact of seeding on hail formation must be considered with the aid of dynamic models of how seeding affects vigorous convective clouds, and with full account taken of the mechanisms by which artificial ice nucleants intervene in cloud processes. Therefore, further discussion of hail suppression models and hail suppression results is deferred to Chapter VIII.

A Numerical Model of Seeding an Orographic Cloud

The discussion to this point should make it obvious that qualitative discussions are insufficient to resolve the effects of cloud seeding. In order to form a better estimate of the probability of success for various proposed seeding techniques, it is necessary to turn to mathematical (numerical) models. We have already made use of some simple numerical models in our discussion of cloud formation and of the effects of seeding with artificial CCN. In the simplest model we considered a hydrometeor growing while at rest in a parcel of air, which was also at rest. In the cloud formation model, we saw that growing droplets in a parcel of rising air compete for the excess water vapor made available by cooling. Under hygroscopic seeding, we considered kinematic models, in which a particle moved through a cloud, whose circulation was specified, and eventually grew to a size sufficient to permit the particle to fall from the base of the cloud. In discussing the Bergeron process, we considered the fact that the cloud water in a parcel was depleted by the growing precipitation embryos, thereby introducing the concept of *competing embryos,* and saw that competition might lead in theory to suppression of precipitation (*overseeding*).

Seeding with glaciogenic agents inevitably releases latent heat of fusion, and may thereby alter the dynamics of clouds. Nevertheless, there are many situations in which the atmosphere is so stable that the heat released by glaciogenic seeding would not be expected to have any detectable effect upon the air motion. Even in unstable situations, seeding may be conducted in such a way that no major effect upon cloud dynamics is anticipated. Seeding in this fashion is sometimes called *static* seeding to distinguish it from *dynamic* seeding, which is conducted with the conscious objective of altering the motions in and around clouds through latent heat release.

For the time being we defer consideration of the more complex question of seeding for dynamic effects, which inevitably involves microphysical effects as well, and concentrate on a simple situation where seeding would likely be conducted for microphysical effects alone. Even this situation involves choosing among a wide range of numerical models, all of which are sufficiently complex that computers are required to obtain solutions in reasonable lengths of time. These models are sometimes called kinematic models to distinguish them from the dynamic models, which will be introduced in a later subsection.

The kinematic models may be one- , two- , or three-dimensional and may be steady state or time dependent. They may assume, as in the results quoted earlier, that the cloud water is not significantly depleted by the artificial precipitation particles or they may assume cloud water depletion by the competing precipitation embryos.

We now present the results of one kinematic modeling exercise to illustrate the possible impacts of glaciogenic seeding for microphysical effects upon precipitation particles forming in a cloud. We begin with a steady-state kinematic model of an orographic cloud produced by the lifting of moist air on the windward side of a long mountain range. In such a situation the cloud is continually forming at the upwind side and dissipating on the lee side of the mountain due to descending motions. The fact that the cloud on the upwind fringes consists of newly condensed water droplets is subject to easy verification by an observer on the ground. Looking toward the sun through the upwind edge of an orographic cloud generally reveals brightly colored coronae and other optical effects, which can be shown to result from the scattering of light by monodisperse water droplets with diameters in the range of 10–20 μm. This type of distribution is characteristic of newly formed clouds with moderate liquid water concentrations.

Whether any precipitation is deposited upon the mountain range and in what locations depends on many factors including the shape of the mountain, the wind speed, the moisture contents of the clouds, the atmospheric stability, the temperature, and the colloidal stability of the cloud. Sometimes precipitation from such cloud systems is continuous and persists for many hours, leading to significant accumulations of snow, while in other situations no precipitation occurs, the amount of cloud water evaporated on the lee side being equal to the amount condensed on the upwind side.

Bergeron (1949) and Ludlam (1955) both pointed out that orographic clouds extending into regions at temperatures below 0°C would be ideal candidates for artificial seeding to increase precipitation by continuous seeding from AgI generators upwind. The models which they used to analyze the problem quantitatively were very simple. In recent years more advanced models have been developed. Chappell and Johnson (1974) used a one-dimensional model of a snow producing cloud to deduce that clouds with top temperatures above -25°C have a potential for snow augmentation. They also deduced that the concentration of snowflake embryos is not critical, but probably should be in the range of 30–200 liter^{-1}

The particular model which we shall describe in some detail was developed by Young (1974b). A similar model was developed by Plooster and Fukuta (1975).

Young (1974b) considered the case of an orographic cloud forming on the upwind side of a mountain range and the effects of ice particles introduced into the cloud in different concentrations and at different temperatures. The reader will readily appreciate that, while the problem of tracing the trajectory and growth of a single snowflake through the orographic cloud is fairly simple, the tracing of the effects of a large number of competing ice crystals is more difficult. In the latter case, it is necessary to keep water budgets for each point on a two-dimensional grid and consider the rate at which water is condensed or evaporated in response to the vertical motions, as well as the rate at which water vapor is depleted by deposition on the growing ice particles and the rate at which cloud water is removed by accretion by the falling ice crystals (snowflakes). It is necessary to consider the fall speeds of growing ice crystals as a function of size and crystal habit, the collection efficiencies of the various types of ice crystals for cloud droplets of various sizes, and the rate of deposition on different types and sizes (capacities) of ice crystals. It is also necessary to incorporate in the model some provision for the aggregation of ice crystals into larger snowflakes.

Snowflakes may follow intersecting trajectories. Ice crystals introduced into the cloud near its upwind edge would eventually begin to fall toward the base of the cloud, even in the presence of the upslope winds. On their way down they might well pass by smaller ice crystals which were formed closer to the mountain crest and which were still rising in the updraft.

Young's (1974c) application of his model to a particular wintertime situation suggested that artificial snowfall on the total mountain barrier would be maximized by seeding near cloud top 40 km upwind with 50 IN liter^{-1} capable of acting as collision nuclei at $-4°C$. The distributions of the liquid water concentrations (LWC) for the unseeded case and for this optimum seeded case are compared in Fig. 4.4. The model results suggest that, depending on the assumptions regarding the natural cloud conditions, including the abundance or scarcity of natural IN, the addition of artificial IN might either increase or decrease the rate at which moisture is precipitated. Perhaps of equal importance is the fact that the model demonstrates clearly that the areas of precipitation can be moved or *redistributed* somewhat by varying the rate and location of glaciogenic seeding. Introduction of many artificial IN leads to the formation of smaller snowflakes, which take longer to reach the ground, so that the region of maximum precipitation is shifted upward on the upwind slope or even beyond the crest.

The type of exercise illustrated above for the relatively simple two-dimensional, orographic cloud situation can be extended to more complex systems. Orographic cloud systems themselves have a three-dimensional character because the wind seldom blows normal to a mountain range during a storm. Instead, low level winds tend to blow parallel to the mountains, and only at higher elevations are winds perpendicular to mountain ranges. Additional complications are introduced by irregularities in the terrain and the fact that many orographic storms contain convective bands or cells or other mesoscale organizations. Furthermore, time variations may be important. It is important to recognize that the model described above, even though quite complex and requiring a large computer for a solution, is still a very simplified picture of a mountain snowstorm.

Once the three-dimensional nature of the cloud system and its time variations are recognized, it is possible to run a kinematic model much like the one just described to study, for example, the effects of glaciogenic seeding upon extensive frontal storm systems. However, no realistic simulations of this type have been performed so far.

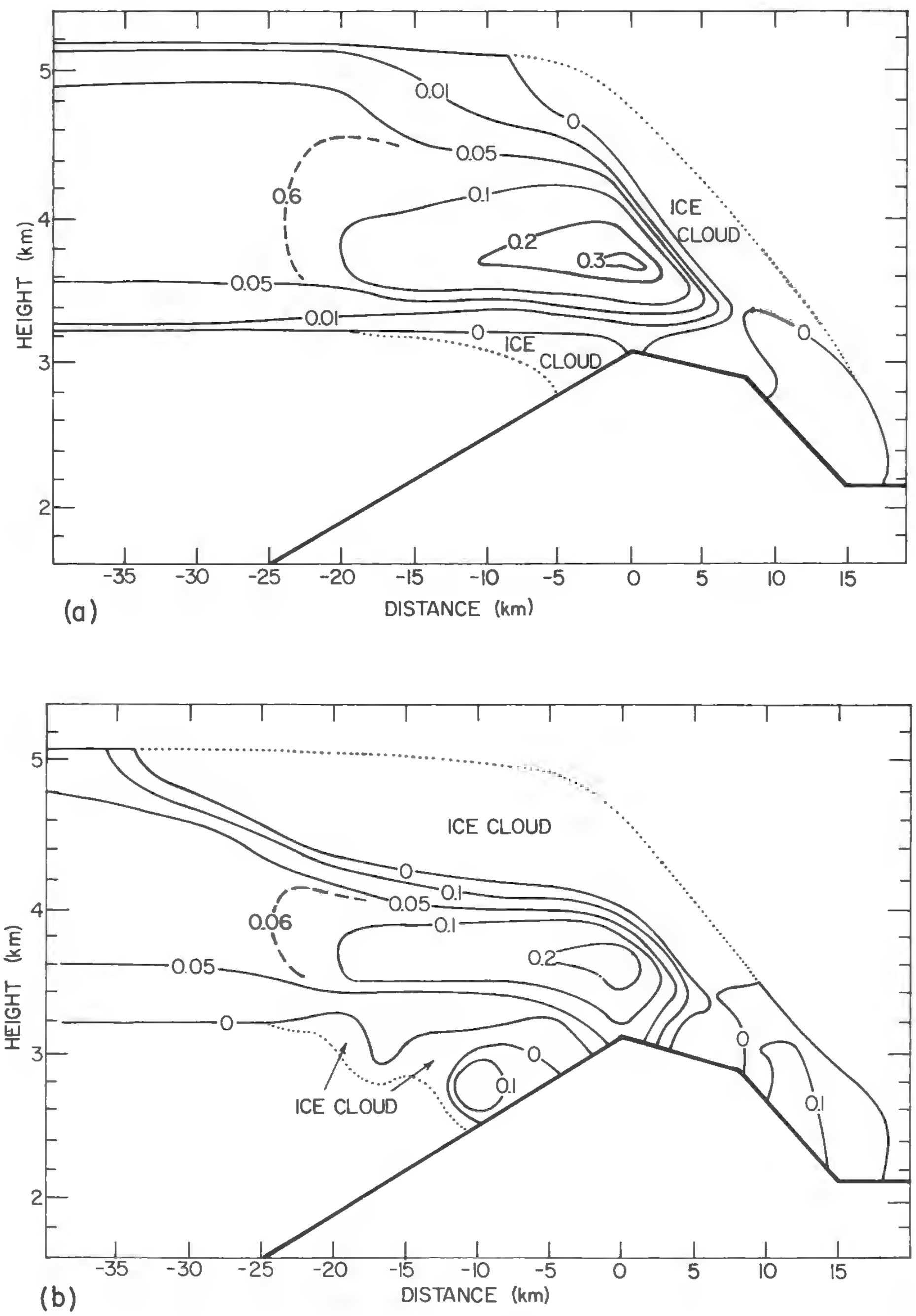


Fig. 4.4. Output from a numerical simulation of an orographic experiment with glaciogenic seeding. Liquid water concentrations in a vertical plane transverse to ridge are compared for unseeded case (a) and seeded case (b) with 50 ice nuclei per liter active at $-4°C$ released near cloud top 40 km upwind. Seeded case has lower LWC values due to enhanced growth of snowflakes. [After K. C. Young (1974c). *J. Atmos. Sci.* **31**, 1749, by permission of American Meteorological Society and the author.]

4.4 SEEDING FOR DYNAMIC EFFECTS

Background

Estimates of the total possible impact of weather modification by cloud seeding for microphysical effects alone do not lead to any indications of drastic changes in weather or climate. Clouds which fail to precipitate because of a lack of precipitation embryos are generally clouds of moderate depth and/or short lifetimes. Even though seeding can cause some such clouds to precipitate and can increase the precipitation efficiency of other, slightly larger clouds, the contribution of the affected clouds to the total annual precipitation at any point remains rather small. Therefore, estimates of the total increment to the annual precipitation possible by such methods generally run in the range of 5–15% of the amount which would have fallen naturally.

A few of the earliest experiments in cloud seeding produced more dramatic effects. Following seeding, some isolated convective clouds grew into cumulonimbus and produced moderate to heavy rain showers. A very striking case in Australia was documented by Kraus and Squires (1947). It was reasoned [e.g., Schaefer (1951)] that such growth was a dynamic effect of seeding caused by release of latent heat of fusion. The basic idea is shown with the aid of a simplified thermodynamic diagram in Fig. 4.5. During the 1960s a number of investigators began to investigate quantitatively the possibility of greatly increasing rainfall from convective clouds by seeding for dynamic effects [e.g., Davis and Hosler (1967)].

The impetus for such work can be seen in Table 4.1, which compares the total rainfall from typical small, medium-sized, and large convective clouds. Rainfall from a large cloud exceeds that from a small one by a factor of 100 or more. The small clouds do not process as much water vapor as do the large ones and, furthermore, a large fraction of the water which is condensed is lost by evaporation through mixing of cloudy air with the ambient air and by evaporation as the narrow precipitation shafts make their way to the ground.

One attractive feature of the concept of seeding for dynamic effects is that it provides a way to modify precipitation from convective clouds that already contain ice particles near the −5 to −10°C level. In theory, effects can be produced as long as an appreciable quantity of supercooled water remains which can be frozen artificially more rapidly than the natural glaciation processes, including ice multiplication, would operate.

Although it might seem obvious that release of latent heat near −5°C

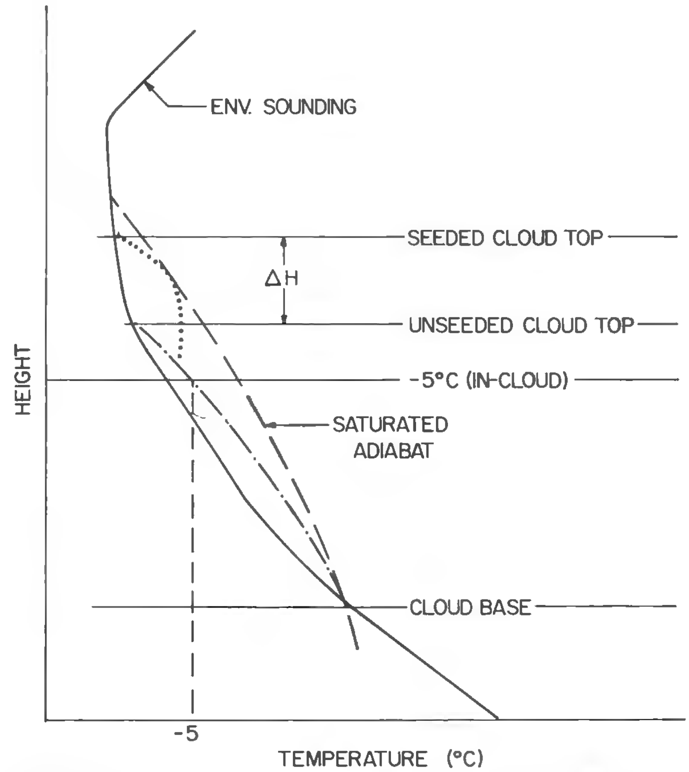


Fig. 4.5. **Simplified thermodynamic chart indicating how artificial glaciation of cumulus towers could lead to increases in cloud height. The temperature profile within an unseeded cloud is indicated by the dot-dash line. The hypothetical temperature profile within a seeded cloud tower where glaciation is accomplished near -5°C is shown by the dots. Increase in cloud top height due to seeding, sometimes called seedability, is indicated by ΔH.**

TABLE 4.1

Rain from Cumulus Clouds as Function of Precipitation Efficiency[a]

	Little cloud	Middle-sized cloud	Big cloud
Air density ρ (kg m^{-3})	~1	~1	~1
Mixing ratio (g g^{-1})	18×10^{-3}	18×10^{-3}	18×10^{-3}
Updraft area (km^2)	0.20	0.78	12.6
Updraft speed (m s^{-1})	0.5	1	2
Lifetime (s)	600	1800	3600
Rainfall (10^3 m^3)			
100% Eff.	1.1	25.3	1633
50% Eff.	0.5	12.6	816
10% Eff.	0.1	2.5	163

[a] After Simpson and Dennis (1974).

would cause an intensification of a cloud updraft and lead to a larger cloud than would otherwise occur, there are actually many questions to be raised concerning this hypothesis. Early investigators reported that on some occasions in the tropics seeding convective clouds from above by aircraft caused cloud towers to rise and separate from the main cloud mass with no resultant growth of the cloud as a whole. There are also subtle effects related to changes in the loading of precipitation upon the cloud updrafts. The complexity of the combined microphysical and dynamic effects can best be sorted out with the aid of numerical models run on computers. Obviously, the models used must provide for interactions between microphysical processes and cloud dynamics and be capable of simulating seeding treatments. The following subsections describe two types of cloud models which have been used extensively for the simulation of dynamic responses to seeding. The two types are the entity models and the field-of-motion models. As the entity models were developed first and provided useful insights into certain aspects of the potential for seeding for dynamic effects, we shall consider them first.

Entity Models

The entity models treat a growing cumulus cloud as a bubble or as an entraining jet and assume that it will obey the laws that have been found to apply to such entities in fluid mechanics. Development of the entity models was no doubt encouraged by the work of Scorer and Ludlam (1953) and others who attempted to fit observations of convective clouds into such a conceptual framework. Simulations of convection in tanks in laboratories using fluids of different densities helped in selecting the relationship most likely to apply to convective cloud towers [e.g., Simpson and Dennis (1974)].

One useful model, developed originally at Pennsylvania State University and subsequently refined elsewhere, will be described as an example. The model treats a cumulus cloud as a succession of bubbles, each rising to the same height in the atmosphere. It is a one-dimensional or "stick" model, meaning that variations in temperature, humidity, etc., within the bubbles are ignored.[3]

The usual way of running the model is to assume a bubble of air of specified radius with a slight upward speed at the height calculated for

[3] Results from the early one-dimensional models must be checked closely. Some of them involved inadvertent violations of the equations of continuity with resultant artificial changes in cloud water concentration (χ_l) and rainwater concentration (χ_R). The necessary adjustments in subsequent models have led to so-called 1.5-dimensional models.

cloud base. The upward acceleration is calculated, and the bubble moved upward through one height step, say 50 m. All processes affecting buoyancy are worked out, and a new velocity calculated, which is assumed to apply through the next height step. The process is repeated until the bubble reaches a stable layer, where its buoyancy becomes negative and the upward speed decreases to zero.

Each bubble, like a real one in the atmosphere, possesses positive buoyancy as long as its virtual temperature exceeds that in the ambient air, but is slowed by mixing with surrounding air. Mathematically,

$$\frac{dw}{dt} = w\frac{dw}{dz} = gB - \frac{1}{m_B}\frac{dm_B}{dz}w^2 = gB - \frac{2\alpha'}{r_u}w^2, \tag{4.1}$$

where w is the upward speed of the bubble, z is height, g is the acceleration due to gravity, B is the buoyancy, m_B is the mass of the bubble, α' is an empirical constant, and r_u the radius of the updraft (assumed to consist of succession of bubbles, each of radius r_u). The buoyancy B is given by

$$B = (\Delta T_v/T_v) = (T_v - T'_v)/T_v, \tag{4.2}$$

where T_v is the virtual temperature of the bubble and T'_v is the virtual temperature of the environment.

The temperature of the bubble as it ascends is controlled mainly by three factors: (1) the decompression as it moves to higher altitudes, (2) release of latent heat, and (3) entrainment of ambient air. Not only is the ambient air cooler than the in-cloud air, but it ordinarily is not saturated. Some of the cloud water evaporates to bring the relative humidity in the entrained air up to 100%, which causes further cooling. It has been established experimentally that the entrainment varies inversely with the radius of a bubble or jet, and this relationship is assumed to apply in the cloud model. Variations in the rate of entrainment under different air mass conditions are controlled by the empirical constant α'.

In some cloud models, it is assumed that all condensed water falls out immediately. Otherwise, it is necessary to adjust the buoyancy to take account of the downward drag exerted by the cloud water. In that case

$$B = (\Delta T_v/T_v) - w, \tag{4.3}$$

where w is the cloud water mixing ratio (grams of water per gram of air, say). Obviously, further refinements can be added to make the models more realistic, but in practice nobody has found it worthwhile to go much further than this with a one-dimensional, steady-state model. The artificialities, such as the neglect of temperature variations within a bubble, mean that the models can never simulate cloud behavior except in a general way.

Comparison of Entity Model Predictions with Observations

The one-dimensional entity models, both steady state and time dependent, have proven useful as predictors of the heights to which cumulus clouds of different updraft radii will rise in the atmosphere under differing lapse rate and moisture conditions. Working with a refined version of the Penn State model, Hirsch (1971) compared observed and predicted height of convective storms in western South Dakota (Fig. 4.6). The updraft radii were measured by aircraft probing the subcloud layer and the environmental sounding data were taken from rawinsondes. The cloud tops were estimated on the basis of RHI radar observations. The best agreement was found by setting the entrainment parameter α' equal to 0.20. The correlation coefficient (r) of 0.82 indicates a significant amount of skill in the computer model calculations. Similar results have been obtained in other air mass regimes.

Simulation of Seeding

Having established that the entity models have a degree of predictive skill for cloud height, it is appropriate to consider how to simulate seeding in them. The common procedure is to assume that the cloud water is gla-

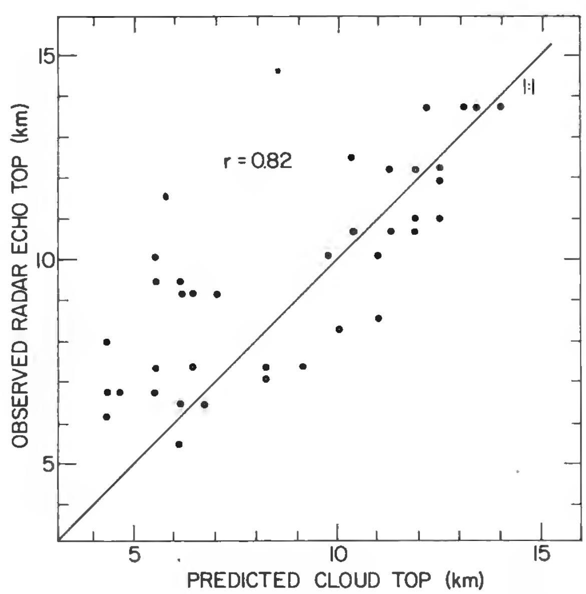


Fig. 4.6. Comparison between heights of convective showers in western South Dakota as observed by radar and those predicted from a one-dimensional, steady-state cloud model. Equivalent results have been obtained in other regions after suitable adjustment of the entrainment parameter in the model [after J. H. Hirsch (1971), by permission of the author].

ciated at relatively high temperatures. For example, Simpson *et al.* (1965) have assumed cloud water glaciation over the range of −4 to −8°C in modeling seeding of cumulus clouds in Florida. Unfortunately, cloud physics observations in seeded clouds there through 1975 showed little glaciation at temperatures above −10°C (Sax, 1976).

Comparisons of conditions inside seeded and unseeded clouds in western South Dakota led Hirsch and co-investigators to adopt nonlinear freezing curves which provided for all cloud water to be frozen between the −20°C and the −40°C level in unseeded clouds, and between −5°C and −25°C in seeded clouds (Hirsch, 1972).

The amount of heat released per unit volume by artificial freezing of cloud water and rainwater varies with χ_l and χ_R. While some authors have spoken of sudden temperature increases of 2–3°C in updrafts due to freezing, a more realistic estimate is 0.5–1.0°C (Orville and Hubbard, 1973). Not all of the latent heat released is available to warm the updraft, a point overlooked in some oversimplified treatments. Much of it is used in expanding the updraft air, doing work against the pressure exerted by the ambient air. Nevertheless, the indicated temperature rises are comparable to the commonly observed temperature excesses in the updrafts of growing cumulus clouds.

Model Predictions of Cloud Height Increases by Seeding

The entity models predict that in very stable situations cloud heights will not increase as a result of the release of latent heat by seeding. In extremely unstable situations, where convective clouds rise swiftly to the tropopause or even beyond, the models again predict no significant change in cloud top height due to seeding, although there may be minor changes in the updraft structure within the cloud. The effects are to be sought in marginal situations, where a balance exists between cloud formation and dissipative influences.

The most interesting situations are those where a natural cloud is capped by an inversion. Occasionally in such situations the model predicts that artificial glaciation can release sufficient latent heat to permit some cloud towers to penetrate the inversion and rise swiftly through the unstable layer above it. This development is referred to as *explosive growth*. Observations suggestive of explosive growth have been obtained [e.g., Simpson *et al.* (1965)]. It seems most likely in the subtropics, for example, over Florida, where subsiding air on the west side of the Bermuda High often causes weak inversions to appear, sometimes at two or three levels simultaneously.

Weinstein (1972) has used a variant of the Penn State model to estimate the potential of glaciogenic seeding to increase precipitation from cumulus clouds over the western United States. A sample of his results is given as Fig. 4.7. His results are based on rather sweeping assumptions about both microphysical and dynamic effects of seeding, and should not be accepted as final. However, his study is one of the few that have tried to combine climatological data, observed cloud size distributions, and cloud models to see how seeding response would vary with location over a season.

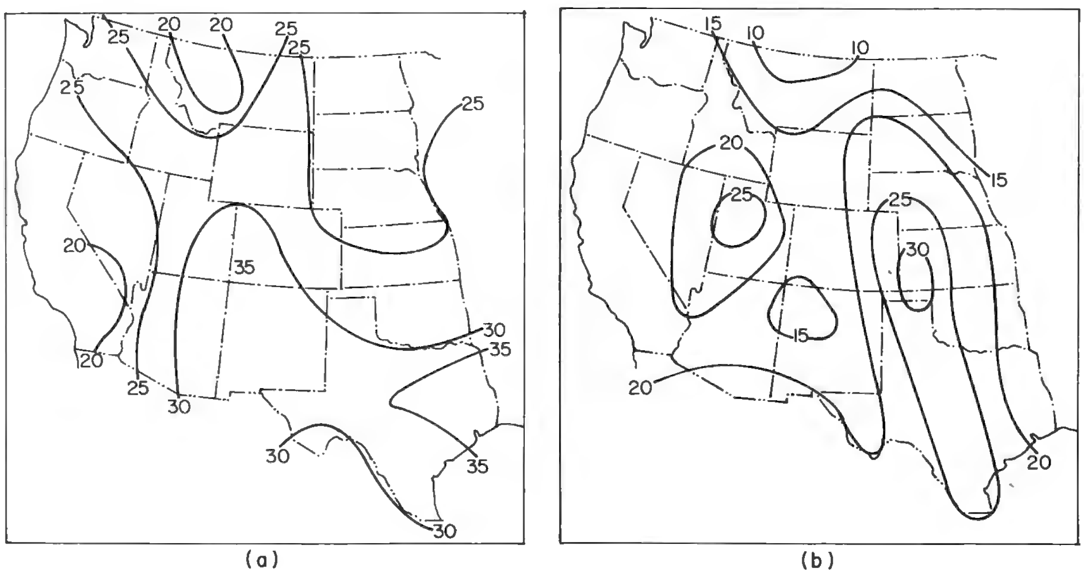


Fig. 4.7. Percentage of summer days with conditions favoring increases (a) and decreases (b) in rainfall due to ice-phase seeding of isolated cumulus clouds. Results are from a modeling study and for illustrative purposes only. [After A. I. Weinstein (1972). *J. Appl. Meteorol.* ***11***, 202, by permission of American Meteorological Society and the author.]

Field-of-Motion Models

The field-of-motion models are more powerful analytic tools than the entity models. The field-of-motion models are free of assumptions about jets, plumes, and bubbles. They are by their very nature time dependent but they may be one, two, or three dimensional.[3]

[3] Results from the early one-dimensional models must be checked closely. Some of them involved inadvertent violations of the equations of continuity with resultant artificial changes in cloud water concentration (χ_l) and rainwater concentration (χ_R). The necessary adjustments in subsequent models have led to so-called 1.5-dimensional models.

In a field-of-motion model the state of the atmosphere at a set of grid points is specified by a number of variables. The minimum number of variables for a working one-dimensional model is five. They are pressure, temperature, air motion (usually in the vertical), vapor pressure, and a measure of total condensate (usually mixing ratio).

Increasing the number of dimensions requires additional components of air motion. For a model to be of much value, the total condensate must be divided into cloud water, cloud ice, rainwater, and snow or hail. In the more primitive versions of these models, the rainwater concentration is represented by a single number. Such bulk parameterizations are derived from considerations of the commonly observed raindrop size distributions. In more advanced models the rainwater concentration may be represented by 10 or more numbers, permitting the raindrop size spectrum to be represented and to vary with place and time.

Some models include atmospheric electricity processes, and so must specifiy the concentrations of small and large ions and the components of the electrostatic field (Chiu, 1978). In those models the number of variables to be defined at a point can exceed 30.

In order for any model to simulate accurately the dynamic effects of cloud seeding, it must model at least some of the microphysical processes taking place. (One must not forget that the dynamic effects are achieved through changes in the microphysics.) Apart from heat release by freezing, which is covered in nearly all of the models, attention is usually directed first to autoconversion of cloud water to rainwater by coalescence. Obviously, the detailed calculations of the stochastic coalescence process cannot be duplicated at every grid point at every time step. Instead, parametrizations are used.

Kessler (1969) and Berry (1968) have both published empirical formulas for the rate at which cloud water is converted to rainwater by coalescence. A recent adaptation of their ideas by Orville and Kopp (1977) yields a formula equivalent to

$$\frac{d\chi_{\mathrm{R}}}{dt} = -\frac{d\chi_l}{dt} = (\chi_l - \chi_{l0})^2 \left[\phi + \frac{\theta N_{\mathrm{c}}}{d_0(\chi_l - \chi_{l0})}\right]^{-1}, \tag{4.4}$$

where χ_{R} and χ_l are the rain and cloud water concentrations, respectively, χ_{l0} is a threshold for conversion, ϕ and θ are empirical conversion parameters, N_{c} is the number concentration of cloud droplets, and d_0 is the dispersion of the droplet spectrum. The influence of the cloud droplet spectrum upon the rate at which liquid cloud water is converted to raindrops is taken into account through N_c and d_0. In general, rapid conversion to rainwater is favored by high cloud water concentration (χ_l large) consisting of a relatively few large droplets (N_c small) with widely dispersed radii (d_0 large).

Once rain has formed, accretion of cloud water droplets by raindrops begins. The freezing of raindrops to form hailstones and the growth of snow and hail from activation of nuclei by deposition or from frozen cloud droplets are also taken into account in some models.

It is seen that the field-of-motion models, developed sometimes primarily to test dynamic seeding effects, actually are better than the entity models in handling precipitation formation and microphysical seeding effects.

The one-dimensional field-of-motion models which simulate action along the vertical axis of a convective cloud have proven very useful in studies of the initial formation of precipitation and are economical in terms of computer time. Examples include the models of Wisner *et al.* (1972) and Danielsen *et al.* (1972).

However, one-dimensional models, whether entity models or field-of-motion models, are inadequate to handle the fallout of precipitation from clouds where the updraft is too strong to permit rain to fall against it. Such cases occur in nature also, but in nature the precipitation can fall outside the updraft region. Simulation of this precipitation mechanism requires recourse to a two- or three-dimensional cloud model. Such models have been developed by Murray, Orville, and Bleck, to name a few.

A Two-Dimensional Cloud Model

The IAS model is a two-dimensional model developed by Orville and his colleagues at the Institute of Atmospheric Sciences at the South Dakota School of Mines and Technology. It uses a stream function formulation to describe air motions in a vertical section of the atmosphere [e.g., Liu and Orville (1969); Orville and Sloan (1970)]. The set of equations is not given here; it is sufficient to note that such effects as buoyancy due to virtual temperature excess, precipitation loading, and latent heat effects are followed at every grid point. The microphysical processes handled in a recent version of the IAS model are shown in Fig. 4.8. The hydrometeors are divided into five classes of (1) cloud water, (2) cloud ice, (3) rain, (4) snow, and (5) graupel and hail. Each is handled through a bulk parametrization which generates mean terminal speeds for the rain, snow, and graupel, each of which is assumed to apply to all hydrometeors in those categories. Cloud water and cloud ice are assumed to have negligible fall velocities. The various processes and interactions involving the 5 types of hydrometeors can be organized into 12 distinct classes of processes as follows (numbers correspond to numbers in Fig. 4.8).

(1) Evaporation or sublimation.
(2) Condensation or deposition.
(2a) Bergeron process (evaporation from liquid droplets and deposition on cloud ice or snow).
(3) Melting.
(3a) Shedding from hailstones in wet growth.
(4) Freezing.
(5) Autoconversion and accretion involving hydrometeors of the same phase.
(6) Interactions between rain and cloud ice, which lead to the formation of graupel.

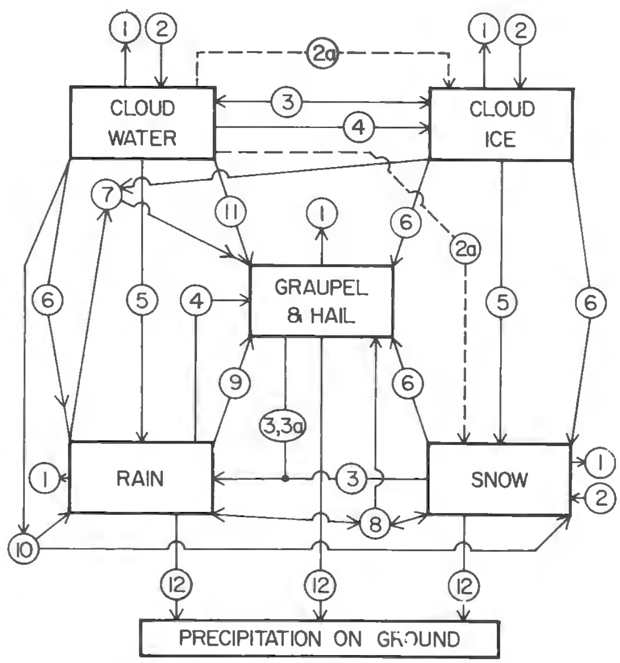


Fig. 4.8. Microphysical processes considered at every grid point and in every time step of a current version of the IAS two-dimensional cloud model. [Explanation in text.] Additional subroutines are available to simulate electrical effects and the transport and action of seeding agents.

(7) Interactions between rain and snow, which lead to formation of graupel, snow, or rain depending upon the precipitation concentrations and the temperature.

(8) Accretion of rain by graupel or hail (if the hailstones are in wet growth, much of the accreted rainwater is returned via process 3a).

(9) Accretion of snow by graupel.

(10) Interactions between snow and cloud water. These lead to increases in snow or rain content depending on the temperature.

(11) Accretion of cloud water by graupel.

(12) Fallout of precipitation to the ground.

The weakest point of the model, apart from its limitation to two dimensions, is probably the parameterization of turbulent exchanges of mass and heat between the cloud and the ambient air. However, this aspect of the model has been improved in recent years. A generalized eddy diffusivity (*D*) is now estimated at each grid point on the basis of the wind shear and the thermodynamic stability of the air.

The model output is so voluminous that it is impossible to give more than a few sample results. The output is normally provided on a cathode-ray tube display, which is photographed by a motion picture camera. By running the resultant movies, various aspects of the cloud development in the model can be viewed and "natural" cases can be compared with a variety of hypothesized seeding treatments. Of course, the specification of what is natural is itself a matter of choice, as the cloud droplet spectra, the rates of coalescence into raindrops, and the activity of the natural IN must be assumed on the basis of available observations in real clouds.

Only within the past few years has the model been elaborated to the point at which seeding treatments can be simulated in other than very crude fashion. It is now possible to "drop" dry ice and "release" AgI particles and follow their subsequent history as they move through the cloud model and interact with the cloud droplets.

Sample Results

The particular model run chosen for presentation here was made to simulate the effects of glaciogenic seeding upon one of two identical convective clouds. In Fig. 4.9 the natural cloud is shown in (a) and the seeded cloud in (b) at 24 min after the simulated AgI seeding. In time-dependent models seeding may hasten the progress of a convective cell through its life cycle, so an estimate of the total effect of seeding upon precipitation

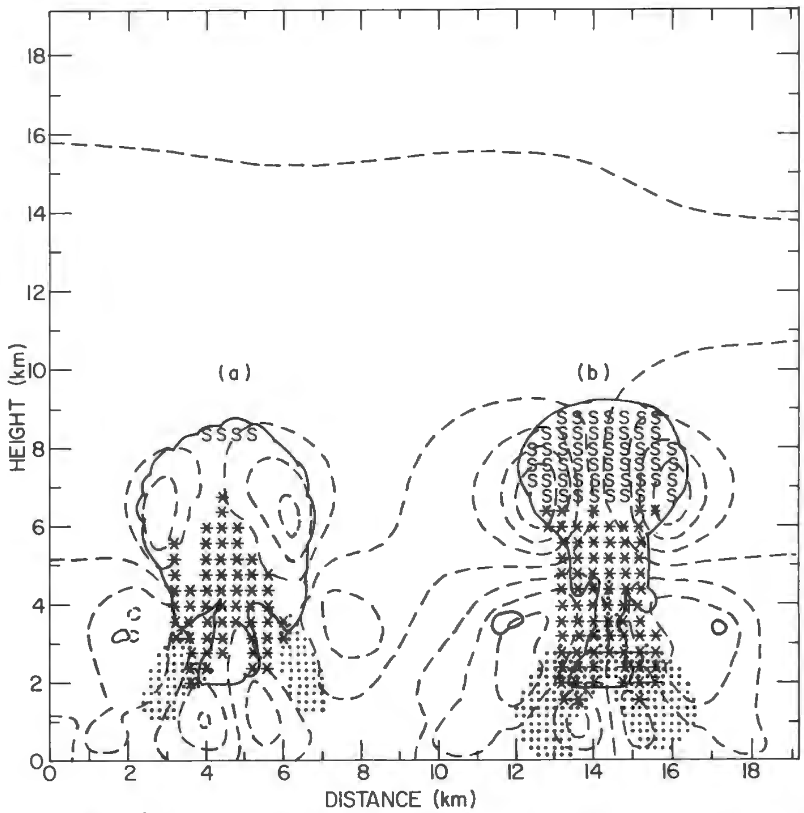


Fig. 4.9. Model predictions of seeding effects in IAS two-dimensional cloud model. Solid lines are cloud outlines and dashed lines are streamlines. *S*—cloud ice, *—snow, · · ·—rain. (a) The natural cloud and (b) the seeded cloud. With no seeding clouds developed almost identically. Here, 24 min after simulated seeding, precipitation from seeded cloud (right) is beginning at the ground, several minutes ahead of unseeded cloud, and seeded cloud is well glaciated [after C.-S. Hwang (1978), by permission of the author].

requires an integration over time. Nevertheless, the instantaneous comparison is instructive. In the natural cloud (a), the amount of precipitation is small, and precipitation is just emerging from cloud base. In the seeded cloud (b), precipitation is beginning to reach the ground. Subsequently, the seeded cloud appeared heavily glaciated, but it began to dissipate about the same time as the natural cloud (Hwang, 1978).

The artificial release of precipitation in a convective cloud earlier than would happen otherwise provides additional possibilities for dynamic effects. The invigoration of downdrafts and of the in-cloud circulation in general is a possible result of the artificial changes in water loading produced by seeding. Because of the many interacting effects that can be postulated, it is not prudent to present any conclusions based on the limited number of model runs completed to date.

Prediction of Effects on Neighboring Clouds

Having established that cloud seeding, particularly with glaciogenic agents, can modify the dynamics of a convective cloud, an atmospheric physicist immediately asks whether or not the effects could extend to neighboring clouds. It has been commonly hypothesized for years that the invigoration of a convective cloud by seeding would tend to suppress other convective clouds within 20–50 km. On the other hand, Simpson and others have argued that seeding neighboring convective clouds would tend to promote cloud mergers, thereby imitating a natural process which is important in the production of convective rainfall. A large cumulonimbus resulting from the merger of a number of individual showers may pour rain at a rate of 5×10^7 m^3 hr^{-1}, even though the individual showers considered as a group would be unlikely to produce rain at a rate much above 10^6 m^3 hr^{-1}.

The merging of convective clouds by seeding, although intuititively appealing, is very difficult to document. It is generally agreed that the merging would require enhanced convergence in the subcloud layer so as to imitate natural mesosystems, but it is not clear how the increased buoyancy of the seeded cloud towers would produce the required convergence. Investigations of the merger concept are in progress within the National Oceanic and Atmospheric Administration and elsewhere.

Dynamic Effects by Hygroscopic Seeding

It has been suggested that hygroscopic seeding could produce dynamic effects. It should be noted, though, that seeding with artificial CCN does not change the total amount of water condensed to form a cloud, but only the drop size spectrum.

We have already referred to the experiments of Woodcock and Spencer (1967), who apparently succeeded in raising the temperature of nearly saturated air by ~0.4°C by adding powdered NaCl. Calculations of the warming at a convective cloud base by seeding with NaCl to form artificial raindrop embryos in concentrations around 10^3 m^{-3} show it would amount to perhaps 0.02–0.05°C. Even this minute effect does not extend up into the cloud. Once the cloud formation zone is passed, the total condensate is the same for clouds seeded with hygroscopic agents as for unseeded clouds. Therefore, it is not surprising that the one-dimensional entity models do not predict dynamic effects from hygroscopic seeding (Hirsch, 1971).

There could be more subtle dynamic effects related to changes in precipitation loading. In addition, promotion of coalescence tends to increase the fraction of cloud water frozen at a given temperature, which would release some additional latent heat. The two-dimensional models are capable of simulating such effects, but suggest they are so small as to be near the limits of the models to detect.

Persons who have investigated the microphysical and electrical effects of hygroscopic seeding have found that they are within the limits of natural variability of convective clouds (Cunningham and Glass, 1972; Murty *et al.*, 1976). This suggests that the associated dynamic effects would be impossible to detect. In summary, modeling studies and field experiments combine to show that dynamic effects of hygroscopic seeding cannot be exploited intelligently in any program of deliberate cloud modification.

Special Hypotheses

A number of quite detailed hypotheses have been put forth to indicate how cloud seeding could modify thunderstorm winds, lightning, and other severe weather phenomena through various combinations of microphysical and dynamic effects. These have generally not been reduced to satisfactory numerical form for testing in models. They will be brought up at the appropriate points in Chapter VIII.

4.5 SEEDING EFFECTS AT LARGE DISTANCES

On the basis of even the simplest microphysical seeding concepts, there is the possibility of effects extending some tens of kilometers from the point of release of the seeding agents. Silver iodide crystals might well remain active for several hours if released in cloud or at night, when they would not be subject to photodeactivation. In such a time period they could travel as much as 100–200 km. Ice crystals produced artificially by seeding could persist for even longer periods and travel long distances to seed clouds downwind. The ice crystals in cumulonimbus anvils sometimes are detectable the day after their parent thunderstorm has dissipated. Simpson and Dennis (1974) speculated on the possibility that such orphan anvils could act as seeding agents for other clouds many kilometers away. They would also lower surface temperatures during the day and raise them at night.

The fact that seeding produces dynamic effects makes the problem of large area effects at once more complicated and more intriguing. It raises the possibility of effects not only downwind, but crosswind or even upwind from the point of seeding agent release. One could visualize, for example, artificially invigorated convective clouds rising toward the tropopause, forming a barrier against upper winds, and thereby creating a positive pressure disturbance for some distance upwind. Three-dimensional mesoscale and regional models suitable for testing this and related concepts are just now becoming available.

Agencies of the U.S. government have been conducting research on the possible modification of tropical hurricanes under the title Project Stormfury since 1962. At least two distinct conceptual models have been involved. The hypothesis currently in favor in Project Stormfury holds that the stimulation of convective clouds in the rainbands of a hurricane could divert inflowing air away from the eyewall cloud and reduce the peak hurricane wind speeds (Gentry, 1974; Rosenthal, 1974). Implementation of this concept, which will be expanded upon in Chapter VIII, would force atmospheric adjustments over an area approaching 100,000 km^2.

The wide range of possible effects which one can postulate suggests that the long range effects of cloud seeding upon precipitation, if they do exist, would be of mixed sign. In some cases they would lead to increases in precipitation, and in other cases to a decrease of precipitation, or perhaps to a redistribution with rainfall increased at some points and decreased at others. It must be emphasized that the numerical models of mesoscale and regional scale disturbances precise enough to permit a realistic assessment of the effect of the seeding in a particular situation are still being developed. Therefore, the only current information on such effects is that obtained by analysis of precipitation patterns in regions around existing cloud seeding experiments and operations. This evidence, tentative at best, is reviewed in Chapter VII.

CHAPTER

V

Generation and Application of Silver Iodide Crystals and Other Seeding Agents

5.1 INTRODUCTION

In order for any concept for cloud modification to be properly tested or put into practice, there must be developed a technology by which the required seeding agents can be produced and introduced into the clouds which one seeks to alter. If a viable weather modification technique is to result, the technology must not only be effective but be reasonable in terms of costs and pose no unacceptable environmental hazards..

In Chapter IV we discussed a wide variety of concepts for weather modification by cloud seeding. However, we also noted that the implementation of some of them would pose problems which are, for practical purposes, not soluble. Among those which can be dismissed as impossible, at least for the time being, are those involving poisoning of CCN and the artificial alteration of the entire CCN spectrum throughout appreciable volumes of air.

Technologies exist for the introduction of hygroscopic powders or sprays as ways of producing artificial precipitation embryos in clouds, but they are relatively unsophisticated. The principal problems involved in implementing them are how to carry the required volumes and masses of materials in airplanes, how to keep hygroscopic powders from caking under humid conditions, and how to dispense the materials in sufficiently fine fashion to achieve reasonable efficiencies. These are problems in mechanical engineering rather than atmospheric physics and will not be explored in detail here.

The most advanced technology that has developed in connection with cloud seeding agents is the generation of large numbers of silver iodide (AgI) particles for artificial cloud glaciation. Clouds can be glaciated, as we have seen, by intense chilling induced by dropping dry ice pellets or releasing jets of liquid air, liquid propane, and so on. Furthermore, there are many substances besides AgI which act as artificial ice nuclei. The others include other inorganic salts, such as cupric sulfide (CuS) and lead iodide (PbI_2), and many organic compounds, such as metaldehyde. Fukuta *et al.* (1966) have described a metaldehyde generator with a yield of 10^{15} nuclei per kilogram effective at $-12°C$, but note that the crystals would evaporate quickly. Fukuta *et al.* (1975) have described seeding experiments in cumulus clouds using 1,5 dihydroxynaphthalene. Nevertheless, AgI remains the most important cloud seeding chemical. The others are of scientific interest and find application in special cases, but have never replaced AgI as a practical seeding agent for field experiments in weather modification. We therefore begin the discussion of seeding devices and delivery systems by describing AgI generators and their operating principles.

5.2 SILVER IODIDE GENERATORS AND THEIR PRODUCTS

Types of Generator

Any device designed to produce small silver iodide (AgI) particles is called a silver iodide generator. Most generators work on the principle of vaporizing AgI and allowing it to solidify upon cooling into particles with diameters less than 1 μm. No other process can match the number of particles produced through the sublimation–deposition process. One of Vonnegut's (1947) first generators yielded 10^{16} particles per gram of AgI. Silver iodide particles are sometimes generated for laboratory applications by grinding AgI or spraying of AgI solutions, e.g., AgI in anhydrous ammonia, but these approaches are not economically attractive for field operations.

The melting point of AgI is 552°C and its boiling point at a pressure of 1 atm (101.325 kPa) is 1506°C. In typical generators, which often work around

1000°C, the AgI obviously sublimes from solid to vapor or evaporates from the molten state without ever actually boiling.

Generators have been built in an amazing variety of ways. Although electric arcs are sometimes used, the AgI is usually vaporized in a flame, which may be continuous or associated with an explosion. Care is required to avoid prolonged exposure to a reducing atmosphere, which can decompose AgI to metallic silver and iodine vapor (I_2), with the latter often reacting with partially burned fuel to produce hydrogen iodide (HI).

The two most common types of generators are the acetone generators and the pyrotechnic generators. A quite complete discussion of both types has been given by P. St.-Amand and his associates (Burkardt *et al.*, 1970; St.-Amand *et al.*, 1970a–c, 1971a–e; Vetter *et al.*, 1970).

The acetone generator originated with Vonnegut (1950). Because AgI is insoluble in acetone, a "carrier" or solubilizing agent is added. The most commonly used carriers are sodium iodide (NaI), potassium iodide (KI), and ammonium iodide (NH_4I). The acetone solution is sprayed into a flame and burned, either alone or with propane or gasoline, to vaporize the AgI and the carrier (Fig. 5.1). Silver iodide consumption rates have varied from as little as 5 g hr^{-1} up to almost 1 kg/hr.

In some acetone generator configurations the heat is insufficient to vaporize the AgI completely. The product in such a case includes sintered AgI, or a mixture containing AgI, with each particle representing the residue of one solution droplet. St.-Amand *et al.* (1971b, p. 33) considered this mode of operation acceptable for producing particles active as ice nuclei at temperatures close to 0°C. However, this mode is inefficient in terms of particle production, unless unrealistically fine sprays are assumed, and will not be discussed further.

Liquid-fueled generators have been developed that use fuels in which AgI dissolves readily (Davis and Steele, 1968). These generators have not been widely adopted because the fuels in question tend to be hard to handle (e.g., chilled anhydrous ammonia) or toxic (e.g., isopropylamine).

The first solid-fueled generators burned coke pellets, which had been previously soaked in an AgI solution, in a portable furnace with a forced draft. Another type fed a string impregnated with AgI at a rate controlled by a clockwork into a propane flame [e.g., Vonnegut (1957)].

The pyrotechnic generators were developed primarily for use on airplanes, where the carrying of inflammable solutions has obvious risks. The earliest ones were simply fusees, railroad warning flares, with some AgI added to the standard mix. Beginning about 1959, scientists of the U.S. Naval Weapons Center at China Lake, California began a systematic program to develop and test pyrotechnics specifically for cloud seeding

Fig. 5.1. Acetone generator in operation on an orographic cloud seeding project near foothills of Sierra Nevada with a second generator of slightly different design behind it. The propane tanks (left) provide fuel and a source of pressure to drive the AgI–NH_4I–acetone solution into the combustion chambers [photo by Atmospherics Incorporated, Fresno, California].

purposes. Some pyrotechnics were developed to be dropped from aircraft and others to be burned in place on racks (Fig. 5.2).

There are many requirements for a good pyrotechnic. Obviously, it should burn smoothly and not explode unexpectedly. Those burned in place must have sufficient structural strength not to break off and fall. (Burning them in supporting tubes is not efficient, because it promotes coagulation of the particles produced.) Droppable pyrotechnics preferably burn up completely, leaving no spent shell to fall to the ground.

In pyrotechnic generators, AgI bearing compounds are packed with others in an organic binder, often a nitrocellulose fuel with a plasticizer. One compound often used is silver iodate ($AgIO_3$); the release of O_2, as the iodate breaks down to AgI, renders the oxidation of the fuel relatively independent of air density and suppresses the tendency for the AgI to dissociate into Ag and I_2, the latter of which then forms HI in response to the presence of partially oxidized hydrogen-rich fuel. St.-Amand *et al.* (1970b) suggest that the addition of iodine-rich compounds is another way to suppress the dissociation tendency, but caution that some of them could lead to formation of explosive compounds. They suggest iodine pentoxide (I_2O_5) for some cases.

Details of many pyrotechnic mixes are given in St.-Amand *et al.* (1970b). Out of literally hundreds of mixes which were tried, and out of several that were widely adopted for various operational purposes, we list here the chemical mixture for the LW-83 flare in its configuration for a droppable flare, the EW-20. The formulation is 78% $AgIO_3$, 12% Al, 4% Mg, and 6% binder. It has been customary to rate pyrotechnic units in terms of nuclei per gram of AgI. The $AgIO_3$ in the EW-20 would yield 20 g of AgI upon reduction, so the EW-20 is described as a 20 g flare.

Another interesting "generator" is Weathercord, which is explosive Primacord prepared with an AgI mix (Goyer *et al.*, 1966). It was developed to be dropped from aircraft, and has been used on operational projects, for example in Iran.

Fig. 5.2. Pyrotechnics for cloud seeding. (a) Rack on trailing edge of wing of seeding aircraft holding eight pyrotechnics. The pyrotechnics are ignited by electrical signals from the cockpit and burn in place. (b) A rack loaded with droppable pyrotechnics being installed on mounts below the fuselage of a seeding aircraft. The individual units are simultaneously released and ignited by electrical signals [photos by Atmospherics Incorporated, Fresno, California].

Fig. 5.2. Continued.

No listing of types of generators would be complete without some mention of the rockets and artillery shells developed in the U.S.S.R., principally to seed hailstorms. Considerable information about them is given in Bibilashvili *et al.* (1974), although the chemical formulations are not always complete. Apparently for reasons of economy, many of the devices use PbI_2 rather than AgI.

Factors Controlling Particle Yields

The most common form of AgI is a yellow hexagonal crystal with density near 5.68 Mg m^{-3}. The number of solid, spherical particles of this substance produced from 1 kg of AgI is shown for a range of log–normal size distributions in Fig. 5.3. The complex particles produced in some generators, notably the AgI–NaI and AgI–KI particles, would obviously require some adjustment of mean density from the 5.68 Mg m^{-3} of hexagonal AgI.

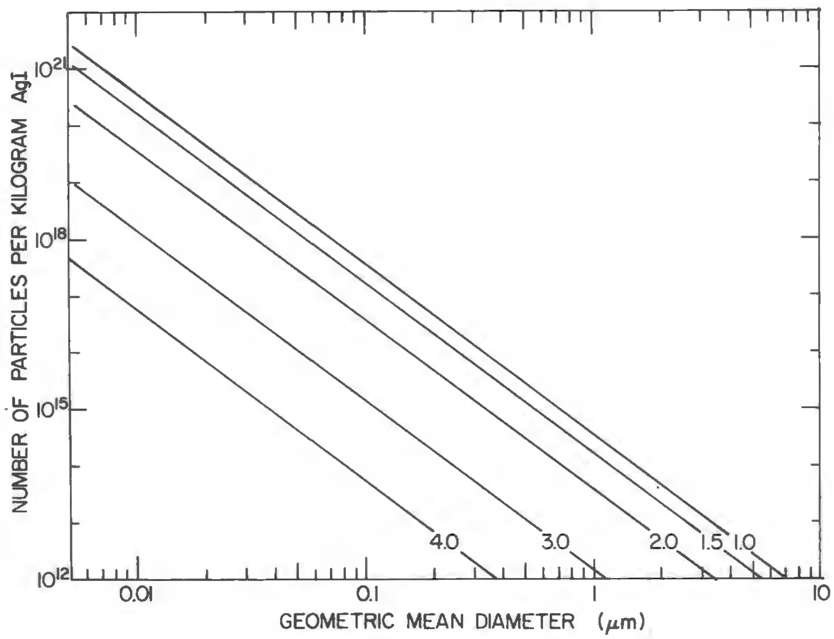


Fig. 5.3. Number of spherical particles obtainable per kilogram of AgI as a function of geometric mean diameter for log–normal aerosols with various values of σ_G, the geometric standard deviation [after Dennis and Gagin (1977)].

In many cases the number of embryonic particles formed in a generator exceeds the number released to the ambient air. The embryonic particles coagulate through Brownian motion, so a typical final product particle consists of a number of crystallites. X-ray work by Petersen and Davis (1971) suggested for some generator products an aggregate particle diameter averaging 0.15 to almost 1 μm, with the individual crystallites averaging 80 nm. The ratios of the linear dimensions suggest aggregates containing from 10 to 1000 crystallites.

Let us recall (2.10), which gives the rate of Brownian coagulation of uniform particles as

$$dN/dt = -K_B N^2 \quad . \; . \; . \; , \tag{5.1}$$

where N is the concentration of particles and K_B is a function of particle size and of the viscosity and temperature of the medium.

Use of (5.1) and Fuch's values for K_B for air at $T = 296°K$ as given in Table 2.2 leads to Table 5.1, which shows the initial rate of decrease per second in particle concentrations resulting from Brownian coagulation. Additional calculations (not shown) indicate that the coagulation rate at the exit from a flame ($T = 600°C$, say) would be larger by a factor of roughly 1.5.

Plainly, the coagulation rate is controlled mainly by the concentration.

TABLE 5.1

Initial Rate of Decrease Due to Brownian Coagulation in Concentration of a Monodisperse Aerosol as a Function of Particle Diameter d and Concentration N[a]

		N (m^{-3})		
		$10^{13.5}$	10^{14}	$10^{14.5}$
	10	1%	3%	10%
	1.0	1	3	10
d (μm)	0.1	2	7	20
	0.01	3	9	30

[a] Percent decrease per second.

Table 5.1 shows that no generator can produce more than about 10^{14} particles per cubic meter of effluent, and can produce that many only if the effluent is diluted by mixing with ambient air within seconds after the particles are formed.

Generator tests at Colorado State University have confirmed that the particle production is limited to about 10^{14} per cubic meter of effluent. Steele *et al.* (1970), on the basis of experiments with varying solution strengths and wind tunnel flow rates, recommended a dilution flow not less than 10^6 m^3 of air per kilogram of AgI consumed. They were seeking to maintain an indicated efficiency of 3×10^{19} ice nuclei active at -20°C per kilogram AgI consumed, which they found for dilute solutions with a strong wind tunnel flow.

In summary, then, *the number of particles* produced by a generator is controlled principally by the amount of air or other exhaust gas traveling through the region where the particles form from the vapor and the rate of quenching, that is, the rate at which the effluent is cooled by mixing with ambient air. Variations in the amount of AgI consumed by a generator affect the *mean particle size,* rather than the particle yield. These are general statements which apply to both pyrotechnic and liquid-fueled generators.

Coagulation obviously affects the final particle size distribution as well as the total concentration. Table 2.2 shows coagulation coefficients for collisions between small and large aerosol particles to be as much as 1000 times larger than those for uniformly size aerosols (either small or large). Hence nearly all the very small particles ($d < 10$ nm) disappear in less than a millisecond once the total number concentration of 1 μm particles reaches 10^{14} m^{-3} or so. The final size distribution is narrowed as a result.

The situation is complicated by the fact that parcels of air or exhaust gas passing through different parts of the flame have different histories.

As is often the case, where several independent factors shape a result, the particle diameters follow approximately a log–normal distribution [e.g., Mossop and Tuck-Lee (1968); Gerber and Allee (1972); Parungo *et al.* (1976)].[1]

A log–normal distribution is characterized by its geometric mean diameter $\bar{d}_G$ and geometric standard deviation σ_G, which is defined by

$$\ln^2 \sigma_G = \overline{(\ln d - \ln \bar{d}_G)^2}, \tag{5.2}$$

where ln denotes natural logarithms. For a log–normal distribution, $\bar{d}_G$ is also the median diameter.

Mossop and Tuck-Lee (1968), on the basis of electronic microscopy, reported $\bar{d}_G$ for a mixed AgI–NaI aerosol at 84 nm and σ_G at 1.47. Gerber and Allee (1972) observed an aerosol from burning of a pyrotechnic containing the widely used LW-83 formulation. St.-Amand *et al.* (1970b) considered that at least some of the particles from pyrotechnic generators would represent the melted and then frozen residues of individual fragments torn from the mix during burning. This might account for the very irregular and wide size distribution measured by Gerber and Allee, which has serious implications for generator efficiency. They reported $\bar{d}_G = 16$ nm and $\sigma_G = 5$.

In all cases, the assumption that $d = \bar{d}_G$ for all particles leads to overestimates of the particle yield per gram of AgI. According to Dennis and Gagin (1977), the mean volume of spherical particles which follow a log–normal distribution is given by

$$\bar{v} = \tfrac{1}{6}\pi \bar{d}_G^3 \exp\{4.5(\ln^2 \sigma_G)\}. \tag{5.3}$$

Application of (5.3) leads to Table 5.2 and to Fig. 5.3. Figure 5.3 shows the particle yield per kilogram of AgI as a function of $\bar{d}_G$ and σ_G. For example, the particle yield for the LW-83 aerosol described by Gerber and Allee (1972) must have been near 7×10^{15} kg^{-1}.

Table 5.2 emphasizes the incompatibility of broad distributions with large generator yields in terms of total particles per unit mass of AgI consumed. Fortunately, typical values of σ_G for generator products are in the range 1.5–2 rather than 5 [e.g., Mossop and Tuck-Lee (1968)].

A true assessment of a generator product is possible only in terms of the nucleating ability of the particles, which is a function of size and chemical composition, and in terms of the desired effects. Therefore, discussion of generator efficiency is deferred to a later subsection.

[1] Some authors prefer the expression: $n(d) = a'd^2 \exp(-b'd)$, where a' and b' are fitted constants, but the log–normal is adequate for many applications (Fuchs, 1964).

TABLE 5.2

Factor F by Which Average Particle Volume in a Log–Normal Aerosol is Underestimated by Assuming $d = \bar{d}_G$ for All Particles[a]

σ_G	$F\ [= \exp\{4.5(\ln^2\sigma_G)\}]$
1.25	1.25
1.50	2.1
1.75	4.1
2.0	8.7
2.5	44
3.0	230
4.0	5700
5.0	115,000

[a] After Dennis and Gagin (1977).

Chemical Compositions

Even when generators are designed to produce pure AgI, the particles often contain traces of impurities as well as a little metallic silver. The metallic silver exposed on the surface oxidizes to Ag_2O. Furthermore, a particle of pure AgI emitted from a generator immediately begins to undergo changes in its surface chemistry, including deactivation by exposure to UV light [e.g., Reynolds *et al.* (1952)]. Many laboratory investigators have gone to great lengths to generate fresh, pure AgI aerosols, for example, by vaporizing silver wire in the presence of iodine vapor in a controlled atmosphere. However, their results are not directly applicable to the impure aerosols produced by AgI generators in the field and will not be quoted here. A review is given in Mason (1971).

Acetone generators burning solutions of AgI with NaI or KI produce very complex particles. Laboratory studies by Mossop and Tuck-Lee (1968), Davis and Blair (1969), and others indicate that AgI–NaI particles undergo wide variations in structure.

According to Mason and Hallett (1956), Lisgarten found that smoke produced by vaporization of an AgI–KI–acetone solution did not show the electron diffraction patterns of either salt, and suggested that mixed crystals were formed. However, Petersen and Davis (1971) concluded on the basis of x-ray studies that both AgI and NaI crystallites exist on the surface of the complex particulates from AgI–NaI–acetone solutions. Their data run counter to the later suggestion of Parungo *et al.* (1976), who noted the differences in equilibrium vapor pressure between AgI and the commonly used carrier compounds, and suggested that NaI or KI would deposit from the vapor first and serve as nuclei for the AgI, which would form a shell on each particle. Burkardt and Finnegan (1970) observed that x-ray diffraction patterns from fused mixtures of AgI and NaI were not quite the same as those from simple mixtures. Their data suggest formation of a small amount of an anhydrous complex, but in general support the ideas of Petersen and Davis (1971).

The question of anhydrous complexes is not of practical importance because the particles are activated in nearly saturated air. As the particles take on water, weak chemical bonds develop and give rise to such hydrates as $AgI \cdot NaI \cdot 2H_20$ (Davis, 1972a) and $2AgI \cdot NaI \cdot 3H_20$ (Burkardt and Finnegan, 1970). With further wetting a solution envelope appears.

The aerosol from $AgI-NH_4I$ solution tends to be purer than that from AgI–NaI or AgI–KI solution, because the NH_4I is sublimed at a low temperature (around 550°C) and then broken down and partially oxidized to yield nitrogen, water vapor, I_2, or HI, and other compounds (St.-Amand *et al.*, 1971b). Some NH_4I can survive, especially if a cool flame is used (Blair, 1974). Parungo *et al.* (1976) theorized on the basis of saturation vapor pressure of AgI and NH_4I that the NH_4I should be deposited mainly on the outside of the particles after the AgI has deposited from its vapor state, but no actual evidence of such a shell structure has come to our attention.

The aerosol from pyrotechnic generators varies widely, depending on the binders, oxidants, and fuels added to the AgI-bearing compounds. The LW-83 formulation referred to above was developed in an attempt to get pure AgI particles.

5.3 ACTIVITY OF SILVER IODIDE PARTICLES

Reasons for Effectiveness of AgI as an Ice Nucleant

The simplest explanation offered for the unusual effectiveness of AgI as an ice nucleant is epitaxial. The crytalline structure of AgI closely resembles that of ice. In the hexagonal crystal form of AgI, the silver and iodide ions take on positions analogous to those of the oxygen atoms in the ice lattice, and the spacings are very similar. The separation between the oxygen atoms in the *C*-plane of an ice lattice is 0.452 nm and the corresponding spacing of silver ions and of iodide ions in the AgI crystal is 0.459 nm. The misfit is very small, 0.015 to be exact. Therefore, the first layer of water molecules to be laid down on an AgI substrate fits very closely to the AgI lattice structure, so the surface energy in the interface is small.

The cubic form of AgI is also effective as an ice nucleant. It too has spacings along certain planes which match the 0.452 nm ice lattice spacing.

Vonnegut and Chessin (1971) introduced silver bromide (AgBr) into AgI crystals to produce distorted lattices, some of which matched ice more closely than does AgI and others which matched ice less closely. The results of their experiments indicated that epitaxy is indeed important. The ice nucleating ability of the doped AgI increased as the lattice was changed to match ice more closely, and decreased as the lattice departed more and more from the ice spacings. Passarelli *et al.* (1973, 1974) found that precipitated CuI–3AgI crystals, which have essentially zero misfit, could act as ice nuclei at temperatures of −0.5 to −1.0°C.

Testing of Silver Iodide Generators

A number of facilities have been built in this country and abroad for simulating cloud processes and testing the effects of various AgI formulations (Fig. 5.4).

In principle, the testing of an AgI generator is straightforward [e.g.,

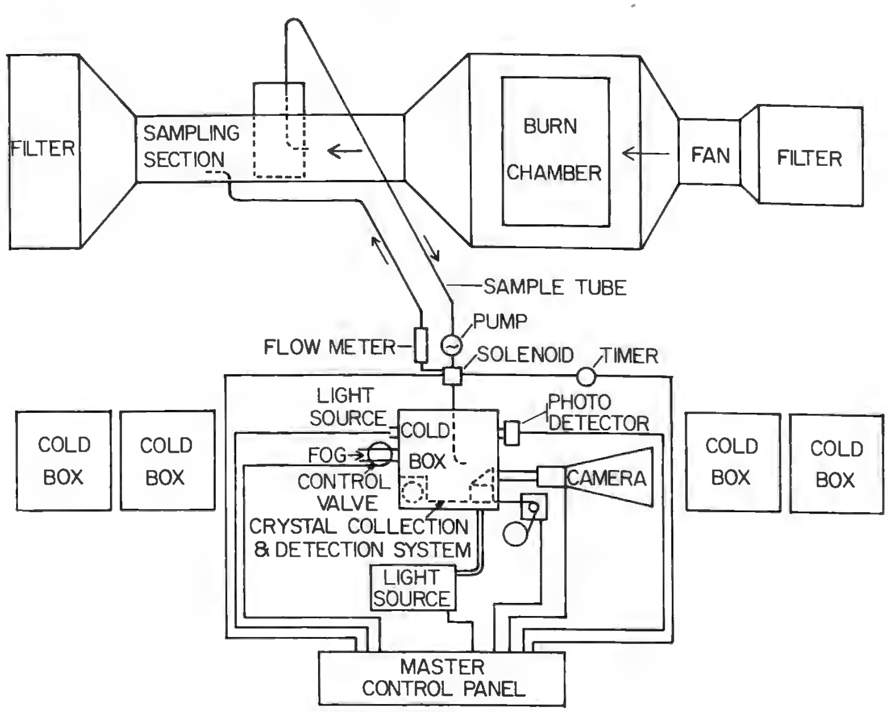


Fig. 5.4. **A schematic diagram of a wind tunnel/cloud chamber facility for testing AgI generators. Facility featured multiple cold boxes for simultaneous measurements at different temperatures and fog-density controls activated by photodetectors. [After J. A. Donnan, *et al.* (1971). *J. Weather Modification* 3, 123, by permission of Weather Modification Association and senior author.]**

Grant and Steele (1966)]. A known sample of effluent is introduced into a cloud chamber containing supercooled water droplets at a specified temperature. A count of the number of ice crystals resulting tells the experimenter how many ice nuclei are produced per unit mass of AgI consumed. In fact, the situation if very complicated. The cloud chambers do not actually count nuclei at all, but snowflakes or small ice crystals resulting from the activity of the ice nuclei.

The first problem to be considered is that of coagulation of the AgI particles. As we have seen, rapid dilution of the generator effluent is necessary to maintain generator efficiency. In collecting samples for introduction into a cloud chamber the dilution effect may be lost, say by having the effluent spend a second or more passing through a constriction forcing it into the collector. Some systems have involved collecting samples in hand-held syringes, with several seconds delay before the collected effluent is squirted into the cloud chamber. In addition to the autocoagulation of the particles for one another, such procedures introduce large losses of particles through collisions with the walls of the syringes and with the inner walls of the tubes through which the effluent is passed. These losses may cause errors by a factor of 2 to 5 in estimates of generator efficiency.

Grant and Steele (1966) have described a dynamic dilution device used on the Colorado State University cloud chamber facility to supplement the normal or "static" dilution provided by a blower. The dynamic dilution is achieved by using a venturi in the dilution tunnel exit.

Coagulation continues after the AgI particles enter the cloud chamber, but most chambers are engineered so that concentrations are down to 10^{12} m^{-3} or so by this time, and made large enough so that losses to the walls can be neglected.

It will be recalled that ice nuclei act in four distinct ways, through deposition, condensation followed by freezing, contact nucleation, and by bulk nucleation. It is generally impossible in a cloud chamber to simulate the in-cloud conditions closely enough to maintain the relative importance of the four modes of freezing. Contact nucleation plays a disproportionate role in most cloud chamber tests. In the real atmosphere, seeding often produces particle concentrations of 10^4-10^5 m^{-3}. In a cloud chamber, one uses concentrations of 10^9-10^{11} m^{-3} so that a statistical sample of ice crystals can be obtained from a very small volume in a short time. As a result, the water droplets are sometimes overwhelmed, with all of the original ones contacted by many nuclei within a few seconds of the introduction of the AgI smoke. There is little doubt that some "ice nuclei counts" have actually reflected the number of water droplets in the cloud chambers. Quite sophisticated systems have been built to maintain the density of the supercooled fog (Fig. 5.4), thereby ensuring some comparability for successive tests [e.g., Donnan *et al.* (1970).]

In some cases, experimenters have found evidence of changes with time in the characteristics of the ice particles produced. For example, Blair *et al.* (1973) reported that the rate of ice crystal production fell off rapidly at first but that production of small ice crystals, presumably by deposition, continued for more than 20 min beyond the introduction of the AgI.

In view of the above considerations, one should not accept published values on the efficiency of AgI generators as anything more than indicators of rough trends. One well known facility reported in 1975 that their previously published results should all be adjusted downward by a factor of more than 10 (Sax *et al.*, 1977a). In particular, no judgments should be made concerning the relative effectiveness of two generator systems until they have been subjected to tests *in the same facility* under the same operating conditions. Even then, the relative performance as deduced from the test results may not reflect accurately their relative performance in nucleating supercooled clouds in the free atmosphere.

It might be thought that the above remarks would not apply to sampling and testing in the free atmosphere, as when AgI plumes are followed by tracer aircraft. However, the collected ice nuclei must still be detected by passage into some kind of cloud chamber. As noted in Chapter II, the problem of sampling ice nuclei in the free atmosphere has been the subject of several international conferences and workshops, and has not yet reached a satisfactory conclusion. It is thought by some investigators that ice nuclei counts in the free atmosphere are characterized by a range of uncertainty of at least one order of magnitude.

In testing AgI generators in the free atmosphere, one must make some assumptions concerning the dispersion of the particles or else release a tracer compound with the AgI crystals. The tracer compound is then collected and sampled independently of the AgI in the collecting aircraft, thereby providing a measure of the dilution experienced by the AgI plume. The tracer compound is often zinc sulphide (ZnS), a fluorescent material which can be readily impacted on a moving adhesive strip or membrane filter and subsequently detected under UV light. Details of the fluorescent tracer method have been given by Leighton *et al.* (1965). Leighton *et al.* (1965) suggest that experimental errors of the method are less than ±20%, which is small compared to the errors in the IN counters.

With all of the above warnings in mind, we nevertheless must rely on cloud chamber tests on the ground or in the atmosphere for first hand knowledge concerning the effectiveness of AgI generators. Some sample results are given in the next subsection.

TABLE 5.3

Characteristics of Three Types of Acetone Generators[a]

Generator	Consumption ($kg\ hr^{-1}$)	Output ($nuclei\ s^{-1}$)			Particles (kg^{-1})	$\bar{d}_G$ (nm)
		$-5°C$	$-10°C$	$-20°C$		
Sprayco nozzle (Vonnegut, 1950)	0.100	~ 0	10^{11}	10^{14}	4×10^{18}	≤ 40
Skyfire (Fuquay & Wells, 1957)	0.016	~ 0	8×10^{11}	3×10^{13}	10^{19}	≤ 30
Modified "Jet-Seeder" (Blair, 1974)	0.280	1.5×10^{11}	4×10^{12}	8×10^{12}	10^{17}	≤ 150

[a] Calculation of $\bar{d}_G$ is made on the assumptions that all particles were active at $-20°C$ and that $\sigma_G = 1$, and therefore $\bar{d}_G$ as shown is an upper limit. Generators were tested with different sampling devices, and numbers are probably not strictly comparable.

Sample Results

The efficiency of an AgI generator is expressed here as the number of particles produced per kilogram AgI, which are active as ice nuclei at various temperatures. Results for three types of acetone generators are shown in Table 5.3.

The threshold of activity for typical generator products is around $-5°C$. The number of active nuclei increases by an order of magnitude for each drop in temperature 3.5–4°C down to -15 or $-20°C$ where, presumably, all of the particles are activated. This behavior obviously mimics that of the natural ice nuclei [Eq. (2.16)]. Sample curves for acetone generators charged with AgI–NaI and with $AgI–NH_4I$ solutions and for LW-83 pyrotechnic generators are shown in Fig. 5.5.

Although the activity of an AgI aerosol in a natural cloud may be quite different from its behavior in *any* cloud chamber (St.-Amand *et al.*, 1970d), the superior performance near $-5°C$ of acetone generators charged with $AgI–NH_4I$ solution, indicated in Fig. 5.5, is now generally recognized as real on the basis of both laboratory and field experiments [e.g., Blair *et al.* (1973)].[2]

[2] An early report to the contrary by Vonnegut (1957, p. 284) may have set back the whole field of weather modification, as St.-Amand has suggested. Vonnegut did warn, however, that AgI–NaI particles might "age" more rapidly than relatively pure AgI crystals from generators charged with $AgI–NH_4I$ solution.

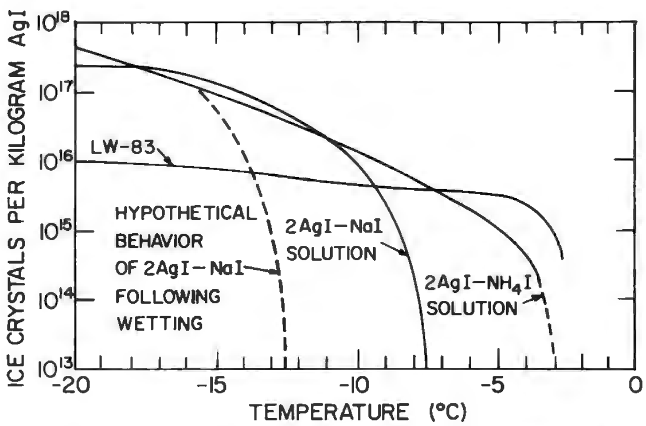


Fig. 5.5. Activity curves for AgI generator products measured in wind tunnel/cloud chamber facility at South Dakota School of Mines and Technology by J. A. Donnan. [After P. St.-Amand *et al.* (1971b). *J. Weather Modification 3*, 31, by permission of Weather Modification Association and senior author.]

Attempts to Develop a Theory

Attempts to relate empirically determined generator efficiencies to independently observed particle size distribution through basic theory have had limited success.

Fletcher (1959a) applied classical thermodynamic theory to determine the rate of ice embryo formation by AgI particles. Pure AgI is hydrophobic. Therefore, Fletcher assumed that AgI particles act as deposition nuclei. He computed on theoretical grounds the size of AgI particle required to serve as a deposition nucleus at various temperatures over the range 0 to −20°C. He found that the large particles would be capable of acting at temperatures close to 0°C, while small particles of 10 nm or so would act only at lower temperatures. From this he calculated the Fletcher curve, which showed, in theory, the maximum number of ice nuclei active at different temperatures which could be generated per unit mass of AgI consumed (Fig. 5.6).

Apparent agreement between the Fletcher curve and experimental curves such as those of Fig. 5.5 led to an uncritical acceptance of Fletcher's theory in many quarters. Furthermore, as St.-Amand *et al.* (1971a, p. 15) have pointed out, the curve has been widely misinterpreted. The Fletcher curve does *not* represent the activity performance of a single aerosol resulting from consumption of a unit mass of AgI, but rather the envelope of the optimum performance of all potential monodisperse aerosols which could be derived from a unit mass of AgI (Fletcher, 1959b). To be specific, if one obtains the 3×10^{13} particles per kilogram effective at

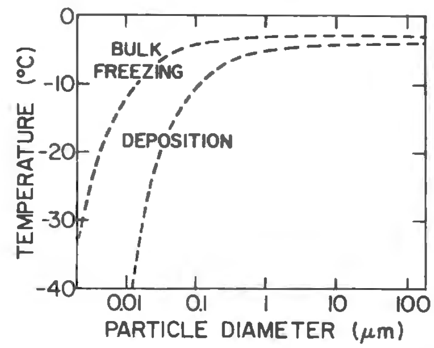


Fig. 5.6. **Temperature thresholds for activity of spherical AgI particles as deposition and freezing nuclei as a function of particle diameter, as calculated by Fletcher (1959a). The deposition curve can be used to calculate the Fletcher curve, which purports to show number of particles obtainable per unit mass of AgI active as deposition nuclei at a given temperature.**

-5°C predicted by the Fletcher curve, one cannot simultaneously obtain the 1.5×10^{17} particles effective at -10°C.

Davis (1972b) has noted that, if the Fletcher theory were correct, the activity curve for a single generator product with a typical value of σ_G would have a more pronounced bend than those actually observed. Indeed, various objections to the Fletcher theory had already been raised by other investigators before 1972.

Some investigators pointed out that the presence of hydrophilic substances such as KI in "AgI" particles rendered condensation-freezing a more likely nucleating mechanism than deposition. Fletcher (1968) elaborated his theory to allow for such effects, but Mossop and Jayaweera (1969) pointed out that it was still at variance with experimental data. In particular, Mossop and Jayaweera found, using an AgI–NaI aerosol, that there is not a 1:1 correspondence between particle size and nucleation effectiveness, as Fletcher's theory predicted, and that the nucleation effectiveness of small particles falls off more rapidly than the theory predicted. The latter point was confirmed experimentally by Gerber (1972).

Fletcher (1970) has made further attempts to reconcile theory and experiment, postulating nucleation events at surface imperfections (pits). This elaboration is in line with electron microscope observations of nucleation events [e.g., Mason (1971)], and with Davis and Blair's (1969) finding that stored strain energy enhances nucleation efficiency. Even so, several difficulties remain, as pointed out by Davis (1972b).

Modes of Activity of AgI Nuclei

It is now generally agreed that AgI particles, like other ice nuclei, act in four distinct ways:

(1) deposition,
(2) condensation-freezing,
(3) contact nucleation,
(4) bulk freezing.

The ability of a particle to act in any of these ways depends on its size, chemical composition, and surface state. We shall therefore discuss each mode of inducing freezing individually before presenting any conclusions about desirable types of generators.

The energy barrier which must be overcome to form an ice crystal by *deposition* is larger than the energy barrier to formation of an ice crystal from liquid water at the same temperature. Therefore, for nucleation at a given temperature, deposition generally requires larger nuclei than does the freezing of liquid water. Many investigators now believe that AgI generator products act as deposition nuclei only at temperatures below $-12°C$ [e.g., St.-Amand *et al.* (1971a, p. 21); Blair *et al.* (1973)].

Condensation-freezing refers to the freezing of a water shell or patch condensed upon an AgI particle. The thickness of the water shell may range from a few molecular layers up to 10 nm or so, in which case the freezing process itself begins to resemble bulk freezing. However, if an AgI particle does not act as a CCN and produce a normal cloud droplet, collide with an existing cloud droplet, or precipitate within a cloud droplet, the total process is still best described as condensation-freezing.

Consideration of the energy barriers for the four nucleation modes suggests that condensation-freezing is, under many conditions, the most promising of all. Furthermore, condensation-freezing permits activation of all nuclei in an aerosol sample in reasonable times and under conditions that might reasonably be encountered in an actual cloud.

The effectiveness of acetone generators charged with AgI and NH_4I, originally credited by Donnan *et al.* (1970) and by St.-Amand *et al.* (1971b, pp. 37–38) to the production of pure AgI particles, is now thought to be due to patches of residual NH_4I on the particle surfaces, which permit the particles to function as condensation-freezing nuclei. Blair (1974) noted that AgI–NH_4I solution gave best results when burned in a cool flame around 650°C. He found that the particulate size increased a little, but not significantly, in going from a hot to a cool flame, with about 10^{17} particles per kilogram active at temperatures of $-17°C$ or lower emanating from both hot and cool flames. [Sax *et al.* (1977a) found pyrotechnics of widely different nucleating characteristics to have almost identical size distributions.] Blair considered the main advantage of the cool flame to lie in the fact that it permits some residual NH_4I to remain on the particles, which in turn permits them to function as condensation-freezing nuclei.

Davis *et al.* (1975) later found that a complex ($3AgI \cdot NH_4I \cdot 6H_20$) with a very small misfit ratio with respect to ice is present in the generator effluent. It causes nucleation at temperatures up to $-2°C$, and may also be important to the performance of the $AgI-NH_4I$ system.

Pyrotechnic products contain a wide variety of hydrophilic compounds such as Al_20_3 and Mg0, depending on the mix used. These impurities make condensation-freezing a likely nucleation mechanism for the contaminated AgI particles, provided the impurities are limited to traces. Sax *et al.* (1977a) suspect that an *excess* of soluble components, including KI and metal oxides, was responsible for the poor performance of the flares used on the Florida Area Cumulus Experiment (FACE) up to August 1975. A change in chemical formulation in that case led to much improved results in cloud chamber tests of nucleation near $-5°C$.

In evaluating *contact nucleation,* it is necessary to consider both the probability of droplet–aerosol collisions and the probability of a collision resulting in freezing.

The most commonly considered mechanism bringing supercooled cloud droplets and AgI particles of less than 1 μm diameter together is Brownian motion, although diffusiophoresis, thermophoresis, and gravitational capture also play a part (St.-Amand *et al.,* 1971c). The roles of diffusiophoresis and thermophoresis, in particular, require careful consideration. They are always opposed for water droplets in free fall, as accompanying tabulation indicates. (The open circle indicates a cloud droplet; the dot an aerosol particle.)

	Growing droplet (updraft)	Evaporating droplet (downdraft)
Diffusiophoresis	$0 \leftarrow \cdot$	$0 \cdot \rightarrow$
Thermophoresis	$0 \cdot \rightarrow$	$0 \leftarrow \cdot$

Edwards and Evans (1961) concluded on the basis of laboratory investigations that diffusiophoresis is of no practical importance, but their experimental data did not rule out the capture of a few percent of AgI particles in an aerosol through this effect.

St.-Amand *et al.* (1971c) apparently considered diffusiophoresis to be of some importance, but Slinn and Hales (1971) and Young (1974a) calculate a different result. Young (1974a) finds thermophoresis to be a factor of 2 to 5 larger than diffusiophoresis for aerosol particles of less than 2 μm diameter, and to be *more important than Brownian motion* in controlling collisions between cloud droplets and aerosol particles with diameters greater than about 50 nm, the exact crossover point depending on pressure, temperature, and humidity. He therefore concludes that aerosol–droplet collisions are much more likely in downdrafts than in updrafts. In updrafts, thermophoresis would, for practical purposes, prevent collisions between cloud droplets and aerosol particles with diameters between 0.2 and 2 μm. Young's results add emphasis to suggestions from other authors [e.g., Parungo *et al.* (1976)] that very small particles ($d < 40$ nm) are desirable to promote *contact* with cloud droplets.

Even for very small AgI particles, the time constant of the Brownian collision process in typical clouds is a few minutes. Using (2.9) and Table 2.2, the collision rate for 20 μm cloud droplets in a concentration of 5×10^8 m^{-3} and 20 nm nuclei in a concentration of 10^6 m^{-3} is near 8×10^{-4} s^{-1}, that is, only about 0.1% of the nuclei introduced would undergo a collision in one second.

The design of some dynamic seeding experiments calls for freezing all the cloud water in a certain volume in a very short time. This has been attempted by releasing AgI particles in high concentrations to freeze the cloud droplets by contact nucleation. St.-Amand *et al.* (1971c, p. 58) have presented a table showing the nuclei concentrations needed to contact all the cloud droplets in a volume of cloud in 1 s. For crystals with $d = 20$ nm and 20 μm cloud droplets the answer is roughly 5×10^{11} m^{-3}. This kind of calculation has led to massive seeding in some cases, with several kilograms of AgI used to seed one convective cloud.

The approach is not only extravagant but leads to contamination. Over 99% of the AgI crystals released never act as ice nuclei in the region of interest in the target cloud, but may remain free to act as ice nuclei elsewhere.

If one is prepared to wait up to 5 min or so, modest rates can be used, at least in theory. If the AgI crystal concentration were 10^6 m^{-3}, about 2000 droplets would be contacted per cubic meter of cloud per second. This would be satisfactory for some applications.

We turn now to the question of the freezing of a supercooled droplet about an AgI particle embedded in the droplet's surface. Some persons have said that the probability of freezing is higher with an AgI particle embedded in the surface (contact nucleation) than with the same particle totally immersed in the interior of the droplet (bulk freezing). Solid experimental evidence on this point is scarce. Fukuta (1975) reported temperature thresholds for contact nucleation 2–3°C warmer than those for bulk freezing, but he worked with organic ice nucleants rather than AgI. Gokhale and Goold (1968) reported freezing by contact nucleation at temperatures up to −5°C, but they worked with ground AgI powder containing particles exceeding 1 μm in diameter.

The available experimental evidence indicates that freezing by contact nucleation may be long delayed after contact, especially for the very small particles most likely to make contact [e.g., Edwards and Evans (1961)]. Sax and Goldsmith (1972) found in the laboratory that AgI aerosols with size distributions peaking around 20–30 nm were almost totally ineffective at temperatures down to − 12 or − 13°C.

On the basis of the experimental evidence, we reject the suggestion that the generation of very small AgI particles is the correct way to promote contact *nucleation,* as opposed to mere *contacts.*

Bulk freezing is defined as the freezing of a water drop around a nucleus totally immersed within it. The drop and nucleus may have originally come into contact by collision, the drop may have formed around the particle, or the nucleus may have precipitated from a solution droplet.

While contact nucleation *may* be more effective than bulk freezing (Fukuta, 1975), once contact is made, bulk freezing has the advantage that the nuclei can be "stored" in the cloud droplets until the right conditions for activation occur. In seeding updrafts below summertime cumulus, for example, some of the larger AgI particles may act as CCN as they rise through cloud base, provided they are large enough ($d > 0.2$ μm, say) and contain a hydrophilic component such as KI. Some of the smaller particles would contact cloud droplets by Brownian motion during ascent to the 0°C level (St.-Amand *et al.,* 1971c), despite the offsetting effects of thermophoresis (Young, 1974a).

A point to consider in bulk nucleation is that the particles might dissolve before nucleation occurred. Complete dissolution in cloud droplets is not an important consideration for pure AgI particles exceeding 10 nm diameter at temperatures below 10°C (St.-Amand *et al.,* 1971e; Mathews *et al.,* 1972). However, the presence of NaI or KI changes the picture. St.-Amand *et al.* (1971b) have expressed strong concern about dissolution of particles containing KI or NaI, and recommend strongly against using them in situations where the aerosol must pass through warm clouds to reach supercooled clouds.

Mathews *et al.* (1972), assuming solubilities of up to 10^{-3} Mg m^{-3} for contaminated AgI, found that an AgI particle of 2 μm diameter would last only 2 s in a 60 μm drop. Davis (1972a) used previously published solubility data to calculate that, for a 2:1 mole ratio of AgI to NaI or KI, only particles with $d < 20$ nm are in danger of complete dissolution in their hygroscopically derived water envelopes. Davis concluded that AgI–NaI particles with 1:1 mole ratios could dissolve completely in their water envelopes *if enough time were available,* say some tens of minutes.

Assuming an AgI particle immersed in a cloud droplet, by whatever means, we now turn to the actual nucleation. According to Fletcher's (1959a) theory, a pure, minute AgI particle of 2 nm diameter could function as a bulk freezing nucleus at $-9°C$. Laboratory data paint a less optimistic view. Gerber (1976) found important time lags of up to an hour in freezing events involving small particles for freezing at temperatures above $-12°C$. Regardless of temperature, there seemed to be a size cutoff; freezing was not observed around particles with diameters less than about 20 nm.

If the AgI particles are aggregates containing soluble components, their efficiency is likely to be further impaired. Fletcher (1968) calculated bulk freezing thresholds for AgI·KI and $(AgI)_2$·KI much colder than those for pure AgI particles containing the same amount of AgI. Fletcher's (1968) treatment was based on the familiar fact that the freezing point of water is depressed by the presence of dissolved solute, which is only one of the possible effects. Davis (1972a) considered the aging to which the AgI surface would be subjected by the AgI–NaI or AgI–KI solution, which could, for example, etch away active sites and release the stored strain energy of the crystals.

In the complete dissolution case, the AgI sometimes reprecipitates from solution droplets if the droplets are carried upward to colder regions or as the droplets grow and the solution becomes more dilute. (This latter paradoxical possibility is due to the complex ion relationships.) The precipitate might consist of many fine particles, rather than a single particle, and certainly would not possess the strain energy of quenched AgI.

The loss of free AgI and the deterioration of nucleating ability of an aerosol from a generator charged with an AgI–NaI solution was observed in laboratory simulations of passage through warm clouds by Chen *et al.* (1972). Ice was detected subsequently only after the chamber temperature fell to $-15°C$, while for a similarly treated aerosol from an AgI–NH_4I solution, the threshold activation temperature remained near $-7°C$.

We conclude that, for bulk freezing attempts, one should carefully control the molar ratio of AgI to NaI, KI or other soluble components to avoid etching or otherwise altering the quenched AgI particles to any appreciable extent.

Deactivation of Nuclei

The fact that most AgI crystals tend to lose their ice nucleating ability with time was discovered soon after Vonnegut's basic discoveries [e.g., Reynolds *et al.* (1952)]. Laboratory experiments suggested that exposure to light, especially UV light, hastened the deactivation process.

Experiments in Australia involving airborne plume tracing generally confirmed that the nucleating ability of AgI smoke decayed exponentially

(Smith and Heffernan, 1954; Smith *et al.*, 1958, 1966). The experiments involved simultaneous releases of AgI smoke and ZnS as a tracer from aircraft or from a mountain top. The results indicated a reduction by a factor of 1000 over a 2-hour period in bright sunlight, but no significant losses over a 1-hour period at night or under an overcast. The Australian results are still a useful guide in planning field operations, even though measured deactivation rates vary considerably from one trial to the next.

Some persons engaged in the field believe that relatively pure AgI particles are less subject to deactivation by sunlight than are complex particles, but firm evidence on this point is lacking.

5.4 CONCLUSIONS ABOUT SILVER IODIDE GENERATORS

There is no "best" AgI generator or delivery technique. The choice of seeding agent and delivery system must be made in line with the changes one hopes to produce in the cloud to be treated.

The design of a seeding treatment is made more difficult by the extreme variability in cloud chamber tests and the fact that a given nucleus may initiate freezing in four different ways. Only a few models [e.g., Young (1974a)] have come to grips so far with the complexities of the nucleation process. Nevertheless, it is possible to tailor generator products to act in certain preferred modes.

Cloud models suggest maximum impact on both precipitation efficiency and cloud dynamics is obtained by production of ice particles at temperatures close to 0°C. These ice particles serve as precipitation embryos and also release latent heat as they grow into hailstones, graupel, or snowflakes. At the same time, one often wishes to avoid the production of very large numbers of small ice crystals at temperatures below -20°C; as some cloud models suggest, this might reduce the precipitation efficiency of a cloud (overseeding).

The best solution now available to the problem of getting ice nuclei active near 0°C but avoiding large concentrations active at -20°C or below is to generate an aerosol of AgI particles with a median diameter near 0.2 μm and a narrow size distribution,[3] and containing a small amount (say 1 or 2% by weight) of a soluble compound like KI or NH_4I. This implies a yield of roughly 3×10^{16} particles per kilogram AgI, but the seeding results should be better than those achieved with generators engineered to have particle yields of 10^{18} or 10^{19} kg^{-1} (Table 5.3). Following

[3] A similar recommendation on optimum size (0.1–0.4 μm) has been reached independently by Mathews and St.-Amand (1977).

the suggestion made here ensures that all particles would work promptly in the -5 to $-10°C$ range, while avoiding production of large numbers of ice crystals by deposition at temperatures below $-12°C$. The hydrophilic component is important, because it brings into play condensation-freezing or condensation followed by bulk freezing. These processes obviate the inefficient alternatives of foregoing freezing at temperatures above $-12°C$, waiting tens of minutes for collisions with cloud droplets, or using very large quantities of AgI.

The type of nucleus desired can be obtained by running an acetone generator charged with AgI–NH_4I solution with a cool flame, passing 3 g of AgI into the generator for each cubic meter of effluent. Pyrotechnics can be used also; Burkardt *et al.* (1970) have described formulations that yield nuclei active at temperatures as high as $-3°C$ and whose yields stabilize near 3×10^{16} kg^{-1} at temperatures below $-8°c$.

Recent developments in pyrotechnics have recognized the futility of trying to increase particle yields by using large devices or concentrated AgI mixes, a point made above in the discussion of coagulation. The first pyrotechnics developed for seeding hurricanes in Project Stormfury included the metal-encased Alectos, which contained 1.7 kg AgI each, and the Cyclops I and II, which contained 4.5 and 30 kg, respectively (St.-Amand *et al.*, 1970c). In contrast, the latest pyrotechnics for Stormfury measure 250 mm in length by 15 mm in diameter, and contain 17 g of AgI each. Their efficiency is vastly superior to that of the Alectos.

Other groups using large amounts of seeding agents (AgI and PbI_2) have included the hail fighters of the U.S.S.R. Their Oblako rockets have been built with up to 5.2 kg "payloads" containing up to 50 or 60% PbI_2 (Biblashvili *et al.*, 1974). Federer, who is now testing one of the Soviet rocket systems in Switzerland, reported in a private conversation in 1977 that their new formulation called Silverspare uses 2% AgI in place of the original 45% PbI_2 of the Soviet's Alazan rockets with no loss in number of ice nuclei produced. The reductions are welcome from both the economic and ecologic points of view.

5.5 ENGINEERING CONSIDERATIONS FOR VARIOUS DELIVERY SYSTEMS

The selection of an AgI generator or other seeding device solves a part of the problem of implementing a concept for cloud modification. The question of a delivery system remains. Of course, with certain types of devices, such as droppable pyrotechnics or AgI impregnated rockets, the choice of the seeding generator essentially defines the delivery system also. In other cases a wide range of possibilities exists. Acetone generators, for example, can be operated on the ground, on aircraft in updrafts below cloud base, or on aircraft flown directly through the target cloud volume. The optimum approach depends on the types of clouds being treated and the effects sought, as well as upon such engineering considerations as cost and safety.

In order to compare the merits of different delivery systems, it is necessary to examine the transport processes which control the movement and dispersion of the seeding agents after they are released. We begin with a consideration of an AgI generator located on the ground. It will be understood that the same discussion applies to any seeding agent where the particle fall speeds are negligible.

Seeding from a Ground Based Generator

Consider the problem of specifying the distribution of seeding agent emitted from a continuous source on the ground. One approach to this problem is to use a Gaussian plume model, such as those commonly employed in studies of air pollution [e.g., Perkins (1974, 182-195)]. If one considers the horizontal wind is blowing along the x axis at a mean speed $\bar{U}$, then the concentration χ downwind of a source is given by

$$\chi = \frac{Q}{\bar{U} 8\pi\sigma_y\sigma_z} \exp\left\{-\frac{y^2}{2\sigma_y^2} - \frac{z^2}{2\sigma_z^2}\right\}, \tag{5.4}$$

where Q is the strength of the source, y is the horizontal displacement from the x axis, z is the height, and σ_y and σ_z are the standard deviations of the plume. If will be noted that this equation provides for unequal horizontal and vertical dispersion of the seeding material, that is, there is no requirement that σ_y be equal to σ_z (Fig. 5.7).

Equation (5.4) states that the concentration across the plume follows a Guassian distribution. Some authors have taken the simpler view that the plume is well mixed and uniform at some distance from the generator, and that σ_y is the horizontal half-width of the plume and σ_z is the depth of the plume.

As the plume spreads downwind of the generator, it is apparent that σ_y and σ_z must be functions of distance from the generator itself. It is also reasonable to assume that they are functions of the intensity of the turbulence with respect to the mean horizontal wind speed. Empirical formulas, which have some basis in theory, are available which give σ_y and σ_z as functions of distance from a point source under different meteoro-

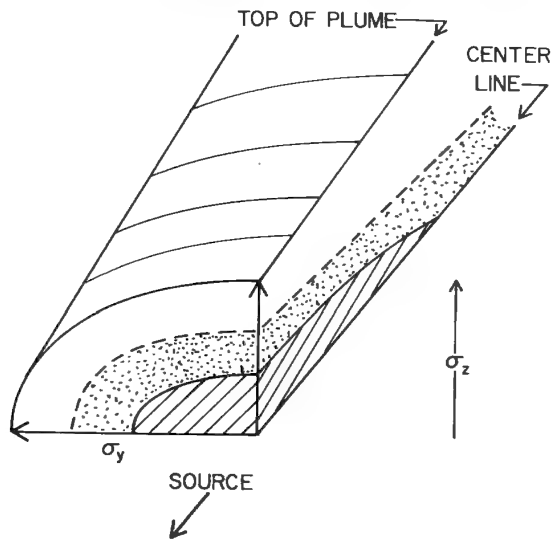


Fig. 5.7. Longitudinal and transverse sections through an idealized plume from a silver iodide generator operating continuously on the ground in flat terrain.

logical situations. The most commonly used relationships in the United States are probably those developed in the Pasquill–Gifford model (Turner, 1969), and which are reproduced here as Figs. 5.8 and 5.9. The weather conditions A–F referred to in Figs. 5.8 and 5.9 are defined in Table 5.4.

In working out concentrations, it is generally assumed that the ground "reflects" the spreading plume. For a generator on the ground, the plume theory predicts that the maximum concentrations will remain at ground level.

If one could assume a constant coefficient of eddy diffusivity in the vertical, say D_z, one could write

$$\sigma_z^2 = 2D_z t, \tag{5.5}$$

and a similar formula for σ_y.

We have already noted that the appropriate values of D_y and D_z depend on the elapsed time over which the dispersion of the material is being considered. This creates a particular problem when a generator is operated continuously on the ground, which is the normal situation, because one wishes to consider simultaneously the distribution of material released from a generator over periods of time which may vary from less than a minute to over an hour.

An alternative formula used by Smith *et al.* (1968), for example, assuming isotropic turbulence is

$$\sigma_y^2 = \sigma_z^2 = C'\varepsilon t^3, \tag{5.6}$$

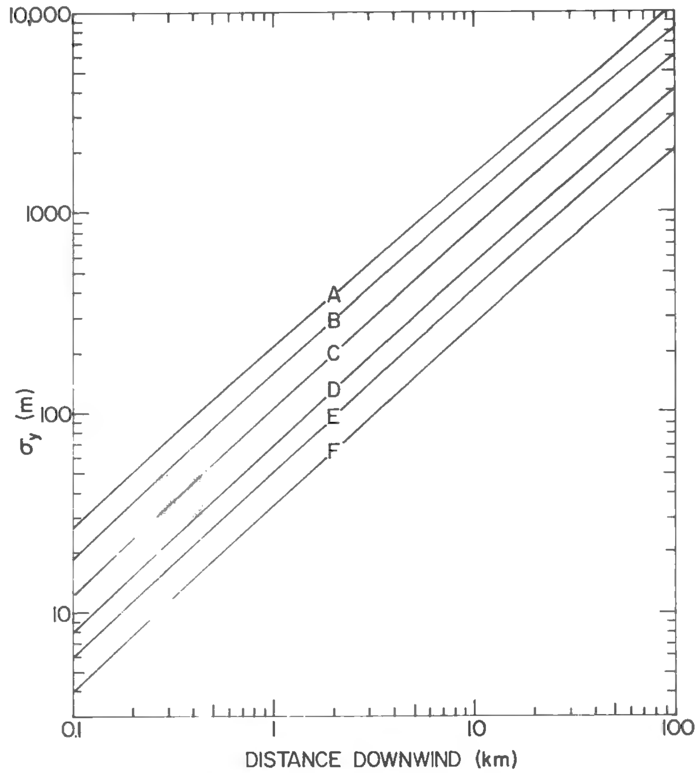


Fig. 5.8. Horizontal dispersion (half-width) of plume as a function of distance downwind from a point source. Key to meteorological conditions is in Table 5.4. [After D. B. Turner (1969). Workbook of Atmospheric Dispersion Estimates, Report 78-22, 84 pp. National Air Pollution Control Administration, Cincinnati.]

where C' is a dimensionless constant of order unity, ε is the energy dissipation parameter, and t is elapsed time $(x/\bar{U})$ downwind from the generator to the point where the observation is made. Empirically, one finds that σ^2_y varies as t^3 for small t, but changes to a quadratic and ultimately to a linear relationship. Smith *et al.* (1968) suggest that the cubic relationship should hold for 10–15 min elapsed time and that the linear relationship might be established in about an hour.

The formulas permit one to model the plume from a single generator, provided values of D_y, D_z, and ε are available. The simple model thus constructed (Fig. 5.7) will be an idealization, which can be thought of as the mean plume. Short term fluctuations in wind direction sometimes lead to a plume which meanders within the envelope of the mean plume.

Having established a model, it is necessary to see how well it conforms to nature. Sampling studies suggest that the spreading of plumes in the horizontal conforms to the models better than does the spreading in the vertical, so we start with observations of σ_y.

There is no doubt that some plumes are very elongated (narrow). Pe-

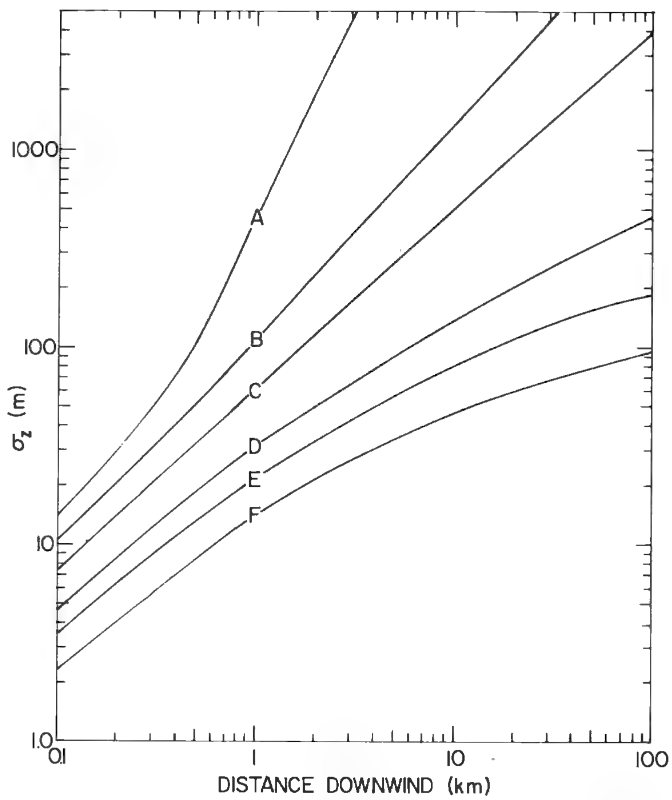


Fig. 5.9. Vertical dispersion (depth) of plume as a function of distance downwind from a point source. Key to meteorological conditions is in Table 5.4. [After D. B. Turner (1969). Workbook of Atmospheric Dispersion Estimates, Report 78-22, 84 pp. National Air Pollution Control Administration, Cincinnati.]

TABLE 5.4

Key to Stability Categories[a,b]

Surface wind speed (at 10 m), (m s^{-1})	Day (incoming solar radiation)			Night	
	strong	moderate	slight	thinly overcast or ≥4/8 low cloud	≤3/8 cloud
<2	A	A–B	B		
2–3	A–B	B	C	E	F
3–5	B	B–C	C	D	E
5–6	C	C–D	D	D	D
>6	C	D	D	D	D

[a] Based on Turner (1969).
[b] The neutral class D should be assumed for overcast conditions during day or night.

terson (1968) reported on a plume on a Category D day (Table 5.4) for which σ_y remained under 10 km for almost 10 hr of travel over the sea east of Long Island. Auer *et al.* (1970) measured plumes of ice nuclei from a mobile AgI generator passing over the Elk Mountain Observatory in Wyoming and concluded that the angular width of the plume was generally near 10°. On the basis of their own and previous measurements, they estimated that the angular width increased from 8° to over 20° as the lapse rate decreased from 12°C km^{-1} to around 6°C km^{-1}.

It might be thought that one could determine the distance upwind of target clouds that a ground generator would have to be operated by examining the relationship between σ_z and the elevation of the -5°C level above the ground. This would often lead to placing the generators one or two hours upwind of the target area. However, AgI generators are most commonly used in upslope orographic conditions or in conditions where convective currents are available to help carry the AgI crystals aloft. Experience on operational projects and experimental plume tracing exercises have suggested that under such conditions generators typically need be located only 15–60 min upwind of target areas.

Plumes in mountainous country are especially complex. Parts of a plume may be trapped under stable air layers in valleys or canyons, while parts released only minutes earlier or later are ascending a kilometer or more into the atmosphere above some exposed ridge.

A report on plume tracing in mountainous terrain in connection with tests of the NCAR ice nucleus counter has been given by Langer *et al.* (1967), who found evidence that the AgI particles are sometimes trapped in valleys.

An example comparing aircraft measurements of an AgI plume with model predictions in an orographic situation is given in Fig. 5.10. The maximum predicted concentration is always at ground level and could not be sampled by the aircraft. Interest actually centers on how quickly the top of the plume reached a sufficient height to start being entrained into the updrafts of the convective clouds which were present on that day. In this particular situation, seeding became effective about 25 km downwind of the generator site.

In placing generators in or near mountains, attention must be given to local variations in the low level winds. The winds are retarded by the mountain range itself, and so the Coriolis force cannot balance the pressure gradient force. A drift in the direction of lower pressure results. During storms with southwesterly winds over the Sierra Nevada, for instance, a low level drift from southeast to northwest across the smoothed contours which appear on most weather charts is apparent over the upwind slopes to 4 or 5 km above sea level. Elliott *et al.* (1978a) have pre-

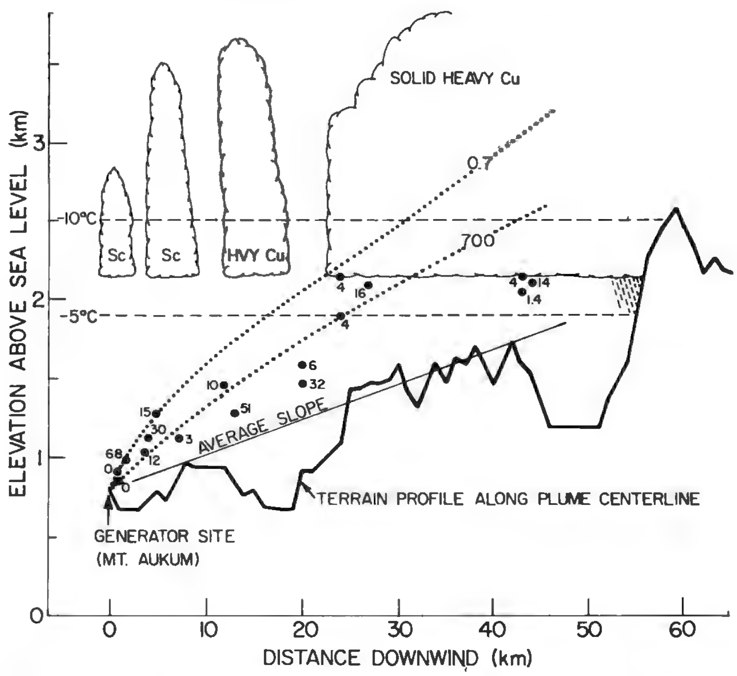


Fig. 5.10. **Vertical distribution of AgI in plume from a generator on a foothill on southwest side of Sierra Nevada on 28 February 1977. Ice nucleus concentrations are expressed in number per liter active at −24°C. Numbered dots are 30 s averages measured by sampling aircraft; dotted lines are isopleths of same quantity estimated by diffusion model for condition C. [After R. D. Elliott *et al.* (1978a). Special Report on Background and Supporting Material for the Sierra Cooperative Pilot Project Design, Report 78-22, 240 pp, by permission of North American Weather Consultants, Goleta, California.]**

sented comparisons of diffusion model and tank model predictions with aircraft plume sampling results under these complicated conditions. The "blocking-flow" situation is especially serious if the valley air is trapped under a temperature inversion. Even without an inversion, accurate targetting requires location of generators on the highest ground available.

The desire to place generators on high ground may, of course, not be compatible with the need to place the generators 20–50 km upwind of the target area to provide for the required lift of the seeding agent before it reaches the target (Fig. 5.10). Moving the generators into the mountains also reduces the amount of time avaiable for horizontal dispersion of the seeding agent. All of the evidence suggests that generators used to blanket an entire target area should be spaced rather closely across the prevailing wind upwind of the target area, certainly no more than a few kilometers apart. Comparison of this result with the densities actually used on some projects in the past shows that only portions of the target areas were seeded.

Airborne Seeding in Stable Air

In considering aircraft seeding it is convenient to distinguish between those cases where generators are operated continuously upwind of designated target areas from those cases where airplanes are operated in the updrafts under target convective clouds or are used to fly through clouds or drop seeding agents into them from above.

The operation of a continuously burning generator on an aircraft flying a horizontal path in a stratiform cloud or upwind of a designated target area can be considered as producing an expanding conical cloud of seeding agent which spreads from the flight path axis even as it drifts with the wind at flight level [e.g., Smith *et al.* (1968); Sand *et al.* (1976)]. The radius of the conical cloud expands at first roughly as $t^{3/2}$ but then at a slower rate. It is generally assumed the cone acquires a radius of about 10 m instantaneously due to turbulence induced by the aircraft itself. Thus

$$(\overline{y^2})^{1/2} = (C\varepsilon t^3)^{1/2} + y_0, \tag{5.7}$$

where y_0 is the initial radius of the cone. Substitution of typical values for a slightly unstable atmosphere where this technique would be likely to find application ($\varepsilon \simeq 10$ cm^2 s^{-3}) suggests that the radius of the cone affected by the seeding agent would reach 100–200 m within a few minutes of application but that subsequent spreading would be slow.

A recent study by Hill (1977) brings together the controlling factors in a typical airborne seeding operation on orographic clouds. Figure 5.11 shows vertical sections transverse to the plume left by a seeding aircraft passing along a fixed track once every 15 min. The assumed conditions were as follows:

Seeding altitude	3	km
Seeding temperature	−10	°C
Lapse rate	5	°C
Horizontal wind speed	15	m s^{-1}
Vertical wind speed	0.25	m s^{-1}
D_x	1000	m^2 s^{-1}
D_z	200	m^2 s^{-1}
Aircraft speed	50	m s^{-1}
Seeding rate (Olin R-15 pyrotechnic)	0.12	kg hr^{-1}

The calculations for this assumed case suggest that more frequent traverses would be desirable in order to provide more uniform coverage 30–60 min downwind of the seeding track. Alternatively, one could move the seeding track further upwind.

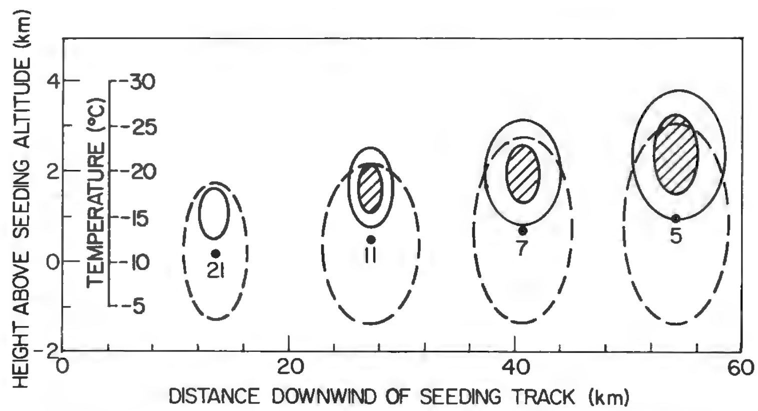


Fig. 5.11. **Artificial ice nuclei concentration as a function of distance downwind from airborne release on four passes 15, 30, 45, and 60 min earlier. Dashed lines enclose regions where AgI mass concentration exceeds 10^{-14} kg m^{-3}. Dots are at centers of high concentration and labeled in units of 10^{-14} kg m^{-3}. Solid lines enclose regions where concentration of active ice nuclei at ambient temperature exceeds 5 liter^{-1}, and hatching indicates more than 10 liter^{-1}. [After G. E. Hill (1977). "Seedability of Winter Orographic Storms in Utah," 78 pp, by permission of Utah State University, Logan, and the author.]**

Recent measurements by Hill (1979) indicate that the applicable values of D_x and D_z under neutral stability conditions are less than those assumed in Hill (1977). In that case, the coverage by the generator plumes is less complete than indicated in Fig. 5.11

Updraft Seeding

Another common mode of aircraft seeding is to fly the aircraft in updrafts below the bases of convective clouds to allow the updrafts to carry AgI or other seeding agent to the supercooled cloud regions. The key question in this type of operation is whether diffusion has sufficient time to spread the material throughout the updraft before it arrives at the -5°C level, where it is supposed to become active.

Sample calculations show that the time available for dispersion ranges from as much as 10 min in seeding clouds with moderate updrafts to as little as 1 or 2 min in seeding clouds with very intense updrafts, say 20 m s^{-1}. Referring to the values of ε shown in Table 5.5 and making estimates by use of (5.6) as to corresponding values of σ suggests that the plumes would indeed spread to fill all of the typical convective updraft (0.5 km across) in approximately 10 min, but that not even the strong turbulence in large cumulonimbus clouds would suffice to spread the material throughout an updraft 2 or 3 km across in as short a time as 2 min. Sand *et al.* (1976) have calculated from (5.7) that a single aircraft working

TABLE 5.5

Typical Values of ε Reported for Various Atmospheric Situations [a]

Situation	ε ($cm^2\ s^{-3}$)
Stable air	<1
Inside stratus cloud	1
Inside cumulonimbus	60
Inside large cumulonimbus	1000

[a] After Smith *et al.* (1968).

below the base of a large thunderstorm could affect only 1–2% of the updraft air by the time the air reached the −5°C level (Fig. 5:12).

There is a further problem. Davison and Grandia (1977) have measured turbulence spectra in and near Alberta thunderstorms and concluded that the cubic relationship of (5.6) and (5.7) would apply for only 3–5 min in intense storms. They believe that earlier calculations which assumed (5.6) to hold up to 10–15 min overestimated the diffusion for aerial seeding. If this is so, seeding strong updrafts of 20–25 m s^{-1} from just below cloud base is completely useless.

These simple calculations are supported by the results of observations and probes to determine the extent of glaciation within convective clouds following seeding. Observations in small to moderate convective clouds following seeding with AgI generators show the ice crystals produced by

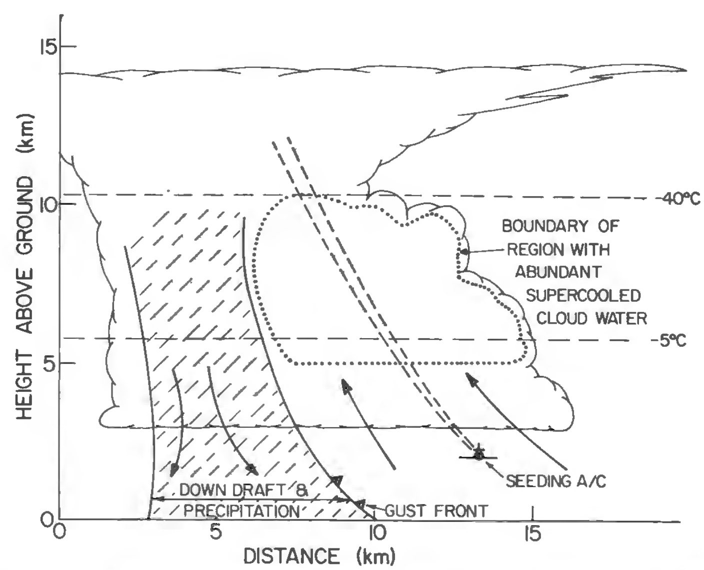


Fig. 5.12. Seeding an intense updraft from just below cloud base may affect as little as 1 or 2% of the air passing upward through −5°C level.

the nuclei to be distributed widely (e.g., MacCready and Baughman, 1968). On the other hand, the fact that experimenters have obtained pictures showing precipitation falling from one side (seeded) of a supercooled cumulus cloud while the other side remained apparently unaffected (Fig. 5.13), and the occasional evidence of thin curtains of precipitation 1 km or less in width falling from clouds along the lines followed by seeding aircraft [e.g., Takeda (1964)] show that dispersal of the seeding agent throughout a large cloud mass cannot be taken for granted. In the severe storm situation, one must find a way to seed that provides some tens of minutes for the material to disperse, or else deliver it directly to the regions of interest.

Fig. 5.13. An experiment on a single cloud was attempted in which only one-half of the cloud was treated with dry ice pellets. The seeded portion of the cloud (right) has a fuzzy appearance indicating the presence of ice crystals while the nonseeded portion (left) still has a hard outline indicating the presence of liquid water. In this particular experiment rain fell from the seeded portion of the cloud treated with dry ice while no precipitation ever occurred from the nontreated portion on the left [photo by Atmospherics Incorporated, Fresno, California].

Direct Injection

The considerations just outlined provide justification for attempts that have been made over the years to develop devices which could be distributed widely throughout a convective cloud volume in a short time to produce a rapid seeding response. We have already referred to the EW-20 and other seeding devices used in free fall through clouds. As noted earlier, Soviet scientists have developed an impressive arsenal of rockets (Fig. 5.14) and artillery shells for rapid response seeding of large convective cloud volumes [e.g., Bibilashvili *et al.* (1974)].

Substitution of numerical values for ε suggest that the downward passage of a continuously burning pyrotechnic through a convective cloud creates a vertical, cylindrical column of seeding material which quickly spreads to a few hundred meters in diameter and only slowly thereafter. Therefore, most groups following this seeding approach have used multiple drops with one pyrotechnic being dropped approximately every 300 m on a traverse over the cloud [e.g., Summers *et al.* (1972)]. For a large cumulonimbus cloud this means up to 10 pyrotechnics being dropped in a single pass. The pyrotechnics generate a curtain of seeding material intersecting the updraft. The findings of Davison and Grandia (1977) regarding the turbulence structure of large convective clouds apply to pyrotechnic seeding from cloud top as well as to updraft seeding, so one should anticipate a slower growth of the plumes after some 5 min have elapsed. It appears therefore that thorough and rapid seeding of a large cumulonimbus cloud requires repeated passes separated by, say, 300–500 m and a total expenditure in the range of 100 pyrotechnic devices.

The Soviet Oblako rocket, to take one example, burns out over 45 s and leaves an aerosol trail 8 km long. It can reach out to a range of 12 km or upward to a height of 8 km. Bibilashvili *et al.* (1974) reported that "The Oblako rocket consists of a head, an engine, a parachute compartment, and a remote control mechanism." The parachute deploys to lower the spent container to the ground. In some cases the warheads are set to explode rather than burn out (in both cases after an appropriate delay time dialed in before launch). In either case, one can model the resultant puff or the plume intersecting a cloud, and the numbers given above again apply. Multiple launches are required to seed a large cloud, with expenditures of 10 or 20 rockets in quick succession being quite common.

Direct Injection of Dry Ice

The various problems which have developed in connection with the engineering of cloud seeding by AgI generators has led to a renewed interest in the merits of dry ice as a seeding agent.

Fig. 5.14. Rocket launchers on an antihail program in U.S.S.R. The launcher at left holds short range rockets. The launcher at right holds larger, more sophisticated rockets. Azimuth and elevation launch settings and the delay time to start of payload burn (if applicable) are made on the basis of information from a radar equipped control center [photo by Atmospherics Incorporated, Fresno, California].

Dry ice was used in the first cloud seeding flights of the modern era in the United States, Canada, and Australia, and probably in several other countries. Despite its replacement by AgI as the preferred seeding agent, dry ice continued to be used on many experiments and operations [e.g., Davis and Hosler (1967)] and laboratory investigations into its effectiveness were pursued [e.g., Eadie and Mee (1963)].

The principal question regarding the use of dry ice is the number of ice crystals which can be produced per unit mass of dry ice used. Langmuir considered that a kilogram of dry ice could produce up to 10^{18} ice crystals. Braham and Sievers (1957) did not find any evidence for such effectiveness, but they treated clouds at temperatures close to 0°C.

A falling dry ice pellet not only freezes the cloud droplets along its path, but activates many aerosol particles as CCN by inducing a sudden supersaturation. As the surface of a subliming dry ice pellet is near −78°C, the newly formed cloud droplets quickly undergo homogeneous nucleation to provide a host of embryonic ice crystals. However, as the temperature returns to ambient levels most of these embryonic ice particles reevaporate and only those which surpass a critical size survive (Eadie and Mee, 1963). It appears, therefore, that the number of ice crystals surviving would depend upon the concentrations of the aerosol particles present as well as upon the speed with which the dry ice pellet falls through the cloud.

Interest in dry ice seeding has revived recently (Holroyd *et al.*, 1978). One reason for the renewed interest in dry ice is the fact that the number of ice crystals it produces is almost independent of temperature over a considerable range. It will be recalled that seeding a cloud with AgI to produce a given concentration of ice particles at −10°C, say, necessarily involves the production of much larger concentrations and possible overseeding at −20 or −30°C. The use of dry ice eliminates this problem and, furthermore, the dry ice quickly sublimes leaving no residue to contaminate other clouds.

Holroyd *et al.* (1978) have analyzed a number of experimental dry ice drops in Australia and in Montana, which were followed by penetrations of the seeded cloud volumes with instrumented aircraft to determine ice particle concentrations. The results indicate a median value for ice particle production between 10^{14} and 10^{15} per kilogram of dry ice.

The exact number of ice crystals produced per gram of dry ice is a function of the sizes of the pellets dropped and likely varies somewhat with the in-cloud conditions also, including the aerosol spectrum. The optimum pellet size can be estimated by determining the distance through which the dry ice pellets must fall to achieve their objective. The pellets are typically dropped from the −20°C level and are intended to produce ice crystals all the way down to the 0°C level. Ideally, one might wish to have the pellets sublimed completely by the time they reach the 0°C level so that no dry ice is "wasted," but in practice one prefers to have the pellets fall through the 0°C level with some finite size. Otherwise, if the pellets were retarded in their fall by updrafts, they would fail to reach the 0°C level. Holroyd *et al.* have considered these matters in some detail and suggest pellets of 7 mm diameter as being most suitable for dropping from the −10°C level.

5.6 CONCLUDING REMARKS REGARDING GENERATORS AND DELIVERY SYSTEMS

It is apparent from the above discussion that one cannot completely separate the choice of cloud seeding generators and delivery systems. In some cases the generator and the active part of the delivery system are one and the same, for example, the artillery shells used to deliver PbI_2 or other seeding agents to hailstorms in parts of the U.S.S.R. Even when the seeding device does permit a variety of delivery systems, as in the case of the acetone generator, there is still a requirement to make sure that there is an appropriate match. For example, some authors have argued that soluble complexing agents such as NaI should not be employed in seeding from the ground or from cloud base where the AgI must rise through warm cloud before reading the $-5°C$ level. While the most recent evidence on this point is somewhat more encouraging regarding the use of AgI-NaI solution in such a situation, there is little doubt that the purer AgI aerosols are most appropriate for such a method of application.

There is, of course, no best delivery system just as there is no best seeding generator. In each case one must take into account the characteristics of the cloud being treated, the path to be followed by the seeding agent, and the desired response of the cloud.

Answers to problems on delivery of seeding agents have been worked out in many parts of the world on operational programs and on applied research programs.

The principal advantages of the ground based generators are economy and the ability to provide consistent delivery of AgI or other seeding material continuously over periods of many hours or days. They are especially suited for seeding on the windward side of mountain ranges. The development of radio-controlled generators has increased their usefulness in remote regions.

Principal disadvantages of the ground based generators are the difficulty in securing rapid response to seeding treatments, uncertainties regarding the trajectories of the seeding agent, and the possibility of deactivation by sunlight or by wetting in warm cloud before arriving in the region of interest. Analysis of individual situations can show whether or not these disadvantages would rule out the use of ground based equipment.

Aircraft seeding offers the advantages of more accurate targeting, the possibility of delivering larger concentrations of seeding material to specific cloud volumes, and the ability to deliver seeding material to various clouds in relatively quick succession. Disadvantages with respect to ground generators include the greater cost and the lack of time in some updraft seeding cases for diffusion to spread the seeding agent sufficiently before it arrives in the supercooled cloud volumes.

The direct injection methods, whether by dry ice drops, free fall pyrotechnics, rockets, or artillery shells, offer further advantages in terms of rapid response and the ability to deliver material quickly to regions of interest. Disadvantages include higher costs for the high performance aircraft needed for over the top seeding, rocket launchers, or guns. Rockets and guns are ruled out in many areas because of the hazard they would pose to aircraft, and to persons on the ground as well. An additional disadvantage is that the need for spreading material widely in a short time requires the use of many separate pyrotechnic devices, sometimes as many as 100 to seed a single large convective cloud.

Further remarks on choices of seeding agents and delivery systems will be made in subsequent chapters in discussing the state of the art with regard to specific operational objectives.

CHAPTER

VI

Statistical Evaluation of the Results of Cloud Seeding

6.1 THE NEED FOR STATISTICAL EVALUATION

In 1952 a fascinating situation developed in the state of Washington. The cherry growers of Yakima, whose fragile crop is damaged by rain when the cherries are ripe, hired a weather modification expert who stated that he possessed a method for suppressing rainfall. His method involved the use of a secret chemical. At the same time, a number of wheat growers in the same general area had a firm under contract to increase rainfall by seeding with AgI generators. The question of which group obtained the better value for its money was never resolved, but the Yakima situation did bring into sharp focus the need for evaluation of cloud seeding programs.

Some weather modification experiments have produced such striking visual results that there is little doubt that precipitation was produced artificially. Trails of snow from the seeded portions of supercooled stratus cloud decks are an example of obviously artificial precipitation. However, the precipitation produced from such clouds following seeding is normally light and does not contribute significantly to the annual precipitation at any particular point. Therefore, operational cloud seeding nearly always involves the seeding of deeper or more extensive cloud systems, where the observation of visual effects is not a sufficiently accurate measure of the results.

A number of investigators have sought to determine cloud seeding effects in these more complex situations by using more sophisticated sensors than their own eyes. Devices which have been used include IN and CCN counters, devices for sampling cloud droplets and replicating ice crystal forms, radar sets for tracking packets of precipitation released by seeding, and special devices for collecting precipitation samples [e.g., Ruskin and Scott (1974)]. Precipitation samples have been analyzed, especially for their silver content, by a variety of methods, including atomic absorption analysis and neutron activation techniques. Electron microscopes have been used to photograph snowflake embryos, and show in some cases AgI particles which undoubtedly served as IN for ice crystals which grew into snowflakes.

All of the above methods are grouped under the general heading of *physical evaluation*. While physical evaluations are informative, they do not address directly the question that usually concerns people the most in a given situation, namely, "Did the seeding affect the total amount of precipitation reaching the ground?" After all, the water that came down in a snowflake formed around an AgI particle might otherwise have come down in a snowflake formed around a natural clay particle. It was probably inevitable, therefore, that persons interested in cloud seeding programs would examine precipitation records in their attempts to evaluate seeding effects. However, they quickly ran into difficulties, difficulties which are the subject matter of this chapter.

The Resort to Statistics

The principal difficulty in evaluation stems from the tremendous variability of weather phenomena, especially precipitation. The effects produced by a typical cloud seeding program are smaller than the natural background variations. In data processing terms, the search for seeding effects is a search for a weak signal in the presence of random noise, and the effects can only be estimated.

It was perhaps inevitable that evaluators of cloud seeding programs should turn to statistics, which is defined as "systematic compilation of instances for the inference of general truths." Through comparisons of seeded and unseeded clouds, the experimenters and operators tried to show that the behavior of seeded clouds was clearly outside the range of natural occurrences.

Langmuir (1953) presented statistics and physical arguments purporting to show that a cloud seeding trial in New Mexico on 21 July 1949 had produced over 10^9 m^3 of rain in a single day. On 6 December 1949 he started an experiment in which an AgI generator was to be operated every Tuesday, Wednesday, and Thursday in New Mexico and periodic effects were sought in the rainfall over the entire eastern United States. Very soon a seven-day rainfall periodicity appeared in the Ohio Valley, which Langmuir (1950) attributed to the seeding. Although this claim was greeted with skepticism by most meteorologists, the New Mexico seeding was cut back at the end of January 1950 lest it contribute to heavy rainfall at remote points (Langmuir, 1953).

6.2 EVALUATION OF OPERATIONAL PROJECTS

While the debate over the possibility of seeding in New Mexico affecting the weather over 1000 km away continued, other scientists pursued the more modest goal of evaluating the effects of cloud seeding over typical target areas of 10,000 km^2 or so. Some evaluations merely stated rainfall amounts in the targets in terms of climatic normals, but it quickly became apparent that more sensitive methods were required.

Target–Control Ratios

The most widely used approach to the evaluation of operational cloud seeding programs is the comparison of events in the target area to events in one or more control areas, which are assumed to be unaffected by the seeding.

The essence of the approach is expressed here in terms of a program designed to increase average rainfall over a designated target area, which is assumed to be adequately sampled by rain gages. Obviously, the same line of reasoning can be applied to programs designed to produce other effects. Hail damage statistics have been analyzed in France and the U.S.S.R. Streamflow data are used to evaluate some orographic projects designed to increase snowpack and runoff.

The average rainfall in the target area for each seeded event, called here for convenience a storm, is computed on the basis of observations at a number of rain gages and compared to the amount of rainfall observed at predetermined controls (Fig. 6.1).

One very simple comparison between the rainfall in the target and control area is to calculate the rainfall in each in terms of percentage of normal, where "normal" is understood to mean the average rainfall during some predetermined historical period. Percent-of-normal comparisons have been used in some evaluations of operational programs. Suppose

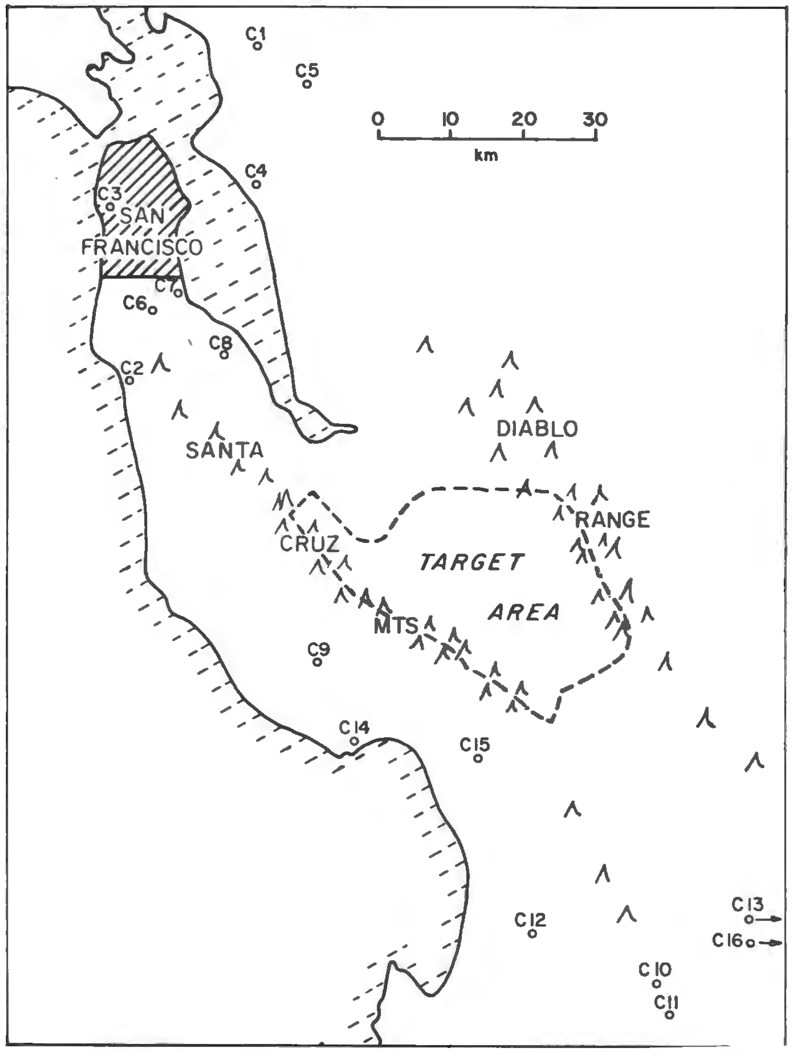


Fig. 6.1. Map of Santa Clara cloud seeding project. Twenty-five rain gages in target area and 18 control stations were selected by mutual agreement of operator and sponsor in 1955 for evaluation purposes. Two of the control stations were eventually dropped, leaving the 16 shown here (C1-C16). [After A. S. Dennis and D. F. Kriege (1966). *J. Appl. Meteorol.* **5**, 684, by permission of American Meteorological Society.]

that, during a given period of operations, the target area received 75% of its normal rainfall for that period, while the controls received only 70% of normal. The operator in such a situation would estimate that the rainfall had been increased by $\frac{5}{70}$, or approximately 7% over that which would have fallen without seeding.

The use of ratios in evaluating weather modification experiments, whether target–control ratios or seed–no-seed ratios drawn from a single target area, carries a special risk of error. A target–control ratio cannot be less than zero, but has no upper limit. Assuming that the target and control areas have similar rainfall patterns, the expected value of the target–control ratio is generally greater than one. Its exact value depends on the underlying distributions of the rainfall events in the target and control areas.

Rainfall observations, say for one hour or day at a point, tend to be highly skewed, with most observations clustered near zero and a long tail extending to large amounts (Fig. 6.2). Hydrologists and others have fitted samples of rainfall observations to a variety of parametric distributions, including the gamma, beta, and log–normal distributions (Mielke, 1979a).

The expected value of the ratio of two numbers drawn at random from a gamma distribution fitted to a data sample like that of Fig. 6.2 is in the range 1.10–1.25. One should therefore not accept a target–control ratio for a single storm greater than one, or even a set of target–control storm ratios with an average value exceeding one, as evidence of rainfall increases due to seeding without further analysis of the underlying rainfall distributions.

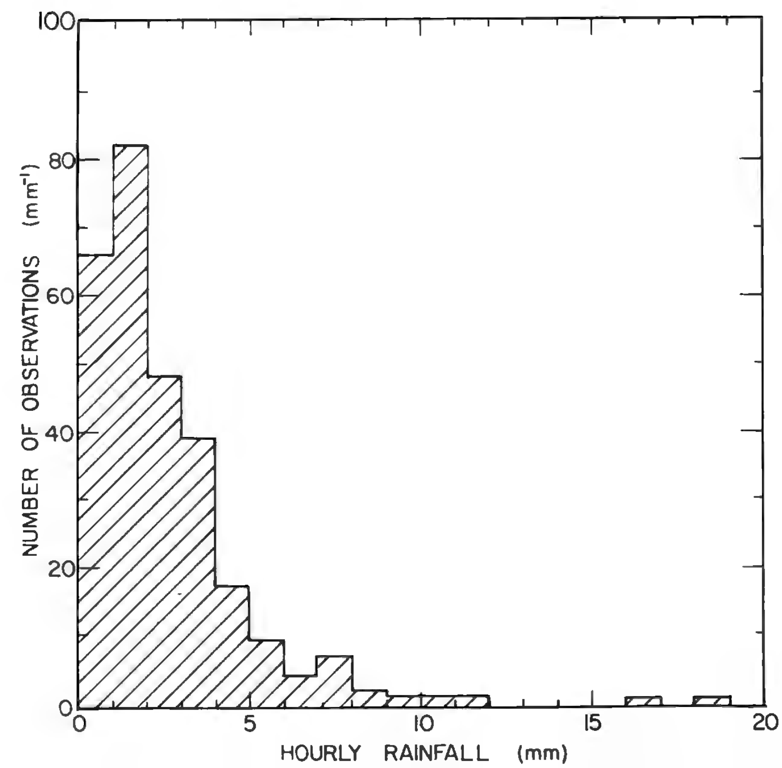


Fig. 6.2. Distribution of hourly rainfall observations collected by a network of 22 recording gages in McKenzie County, North Dakota, in summer of 1972.

Historical Regression Method

The historical regression method of comparing target and control rainfall is considerably more sophisticated than a ratio comparison. It allows for a variety of relationships between precipitation in target and control areas, and also permits an estimate of the probability that observed changes in the target–control relationship during a period of seeding are merely due to chance (Thom, 1957a).

The historical regression method consists of the following steps. After gages are selected in the target and control areas (Fig. 6.1) with reliable records extending through some period of years prior to the beginning of seeding, average rainfalls for the target area and the control area are computed for a number of historical events (preferably more than 30). For convenience we continue to call these events storms, although in some projects daily, monthly, or annual rainfall have served as the individual observations. The data are presented on a scatter diagram with the control area rainfall normally plotted along the abscissa and the target rainfall along the ordinate, as shown in Fig. 6.3.

A measure of the value of a control area is given by the correlation

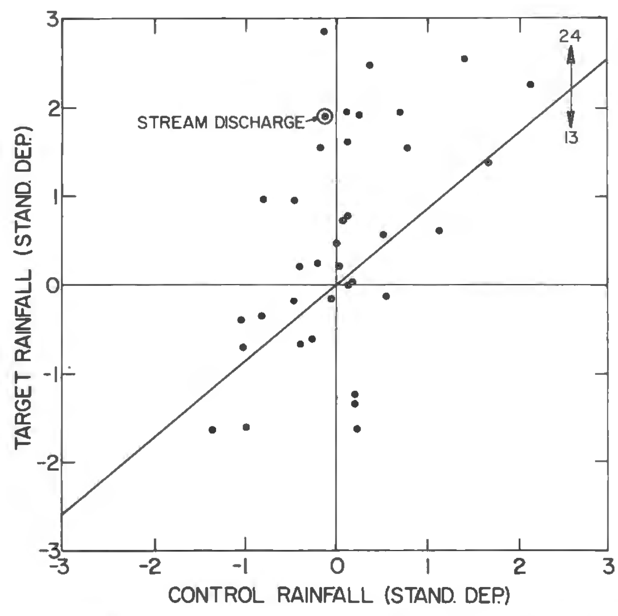


Fig. 6.3. **Scatter diagram showing normalized monthly target and control rainfall amounts for operational cloud seeding projects in western U.S. and result of one streamflow analysis. In this case, there are 24 entries above the historical regression line and 13 below it, indicating the net effect of seeding on the projects in question (mostly orographic) was to increase precipitation. [After H. C. S. Thom (1957b). *In* "Final Report of the Advisory Committee on Weather Control," Vol. II, pp. 25–50. U.S. Govt. Printing Office, Washington, D.C.]**

coefficient $r(X, Y)$ between target and control rainfall. The formulas involved, which are given in any text on elementary statistics [e.g., Draper and Smith (1966)] are as follows:

$$r(X, Y) = (\overline{XY} - \bar{X}\bar{Y})/s(X)s(Y), \tag{6.1}$$

where the bars denote averaging and s denotes a standard deviation. The regression line drawn by the least squares method is

$$Y_E = a + bX, \tag{6.2}$$

where

$$b = [s(Y)/s(X)]r(X, Y) \tag{6.3}$$

and

$$a = \bar{Y} - b\bar{X}. \tag{6.4}$$

Once the regression line is established, the success of a cloud seeding attempt is estimated by computing $(Y - Y_E)$ for the particular storm. Evidence of success is often expressed as a percentage increase in rainfall over that expected to occur naturally. The percentage increase is estimated by the historical regression method as

$$[(Y - Y_E)/Y_E] \times 10^2. \tag{6.5}$$

In addition to obtaining an estimate of the effects of seeding upon precipitation, the historical regression method provides an estimate of the probability that the departure from the predicted value for any seeded storm is entirely due to chance. If the regression equation is based on a sufficiently large sample of unseeded storms, say 30 or more, one can assume that the departures from the regression line are normally distributed with a standard error of estimate s_E given by

$$s_E = s(Y)(1 - r^2)^{0.5}, \tag{6.6}$$

where $s(Y)$ is the standard deviation of the target rainfall sample. If a smaller number of points are used to establish the regression line, it is necessary to assume that the departures in the observed sample follow a Student t distribution with (N-2) degrees of freedom, where N is the number of points used to establish the line.

The historical regression method can involve multiple control areas. One can also introduce controls based on observed meteorological variables other than precipitation. In such exercises, the aim is generally to find a multiple correlation coefficient R higher than that for any single control area. Standard statistical techniques and computer programs exist for performing multiple linear regression (MLR) analyses. The addition of new controls usually reaches a point of diminishing returns rather quickly due to intercorrelations among the controls.

Selection of Control Areas

The success of any target–control evaluation, whether by calculation of ratios or by historical regression, depends critically on a wise selection of control areas. In general, choosing control areas close to the target area tends to maximize the correlation coefficient. Therefore, one should place the control area as close to the target area as possible without running the risk of contamination. Physical contamination can be avoided by placing the control area upwind or across the wind from the target area, but there is still the possibility of dynamic interactions (Section 4.4). In the final analysis subjective judgment is required, because there are no numerical models sufficiently powerful to predict whether or not dynamic interactions would be important in a given situation.

It is also important to chose control areas similar to the target area in terms of elevation and exposure to prevailing winds. Failure to do so will result in very low correlation coefficients and an insensitive analysis.

Decker *et al.* (1957) evaluated a program in Oregon using multiple controls which were separated from the target area by the towering Cascade Mountains. The multiple target–control correlation R was 0.59. Substituting this into (6.6) shows that s_E differed from $s(Y)$ by about 20%, an almost negligible improvement in predictive power. Unsurprisingly, Decker *et al.* reported no significant evidence of a change in rainfall in the target area. On the other hand, Thom (1957b), on the basis of analyses involving control areas only 30–60 km upwind of the target, included the same project among those where he reported evidence of increases in rainfall of the order of 9–17% due to seeding of winter storms.

The above result is cited here, not to make a point about the effectiveness or ineffectiveness of seeding in a particular situation[1], but to emphasize that some evaluations published in the literature have been so insensitive as to be virtually useless. Even rainfall increases of 25–50% of the naturally occurring amounts would go undetected by such gross measurements. In contrast, situations where careful selections of control stations

[1] Thom (1957b, p. 40) did not ascribe statistical significance to the results of any individual project, but to the results for a set of 195 storms drawn from several orographic seeding projects and to a set of 299 storms from West Coast projects, in general, which included the 195 orographic storms.

have resulted in multiple correlation coefficients R of 0.8 or better have yielded better estimates of the range of possible effects. Streamflow analyses have been particularly valuable in some mountain watersheds, where values of R as high as 0.97 have been calculated [e.g., Henderson (1966)].

Normalization of Data by Transformations

Although the basic idea involved in the historical regression analysis is intuitively appealing, there are a number of difficulties with it. The results are not completely reliable unless the underlying data sets conform to the normal (Gaussian) distribution.

Rainfall data for short periods of observation are highly skewed, with most observations clustered close to zero and a long tail extending to large but rare rainfall events (Fig. 6.2). The non-normality of the underlying data calls into question probabilities deduced on the assumption that the departures from the regression line follow the normal distribution (or the Student t distribution for small data sets).

Rainfall distributions can be normalized by data transformations. It is often found that the correlation coefficients are improved and the variations in the scatter of the points about the regression line as a function of X are simultaneously suppressed by the transformations.

Thom (1957a) used a gamma transformation to normalize rainfall data in his evaluation for the Advisory Committee on Weather Control (Fig. 6.3). Other authors have used simpler transformations, such as square root and logarithmic transformations. In general, rainfall observations averaged over short periods of time or over a few gages are more skewed than rainfall data averaged over many gages or over long periods of time, and require more drastic transformations. Monthly rainfall averages at a station or average rainfall for a storm over an entire county are normalized satisfactorily by a square root transformation [e.g., Neyman *et al.* (1960)]. Average rainfall over a few hundred square kilometers during one hour responds to a cube root transformation [e.g., Dennis *et al.* (1975a)]. Rainfall from a single convective cloud may require a fourth root transformation followed by a gamma transformation for normalization [e.g., Simpson (1972)]. The gamma transformation is sufficiently flexible to take these variations into account.

Caution is necessary in interpreting the results of experiments analyzed with the aid of data transformations. An indication that the average of the square root of the rainfall is increased 10% by seeding does not translate into an increase of 10% in the actual rainfall.

In general, if Y is the target rainfall, $Z(Y)$ is the transformed value and $Y(Z)$ represents the back-transformed variable, then

$$E(Y) \neq Y[E(Z)], \tag{6.7}$$

where $E(Y)$ is the expected value of the target rainfall and $E(Z)$ is the expected value of the transformed variable.

Equation (6.7) is always applicable, even to those situations where the effect of seeding is uniform. If the effects of seeding are variable, as many authors have postulated, the situation becomes more critical. In general, the transformations tend to exaggerate the importance of the smallest and most numerous events at the expense of the larger events. Consider an example where the effect of seeding is to increase rainfall for all storms up to the 80th percentile (measuring from the smallest storms) and to decrease markedly rainfall from the remaining 20% (the wettest storms). It would be quite possible for an evaluation using transformed data to indicate an increase in total rainfall, even though the actual net effect was the opposite, unless the analyst was aware of the transformation pitfall.

Further Statistical Elaborations

A number of additional refinements have been introduced into the historical regression method by various authors.

It should be noted that the values of $s(Y)$, $s(X)$, r, a, and b that one works with in a given case are only estimates based on a finite data sample.[2] Equation (6.6) neglects errors in the estimate of b and assumes that the value of s_E is the same for all storms. Thom (1957a, p. 9) has taken note of the fact that even after normalization the uncertainties are minimized near the middle of the range of values of X and maximized for extreme values of X. He suggests the formula

$$s_E = \frac{s(Y)(1 - r^2)^{0.5}}{(N - 2)^{0.5}} \left[1 + \frac{1}{N} + \frac{(X - \bar{X})^2}{\Sigma(X - \bar{X})^2}\right]^{0.5} \tag{6.8}$$

However, rather than pursue the technical details of these statistical procedures, we turn to examine some more basic difficulties associated with the historical regression method.

[2] It should be noted that the calculated correlation coefficient r in a given situation reflects not only the lack of a perfect target–control correlation, but randomness introduced by the lack of perfect sampling networks. Experience in Australia has shown that values of r can be improved *up to a point* by simply installing more rain gages in both target and control areas (Smith, 1967). In practice, it is usually not worth going past 40 or 50 rain gages in each area, regardless of its size. Hail data are more variable and require more closely spaced sampling devices.

Sources of Bias and Uncontrolled Variance

A number of possible sources of bias and uncontrolled variance have been identified in the historical regression method. Some statisticians [e.g., Brownlee (1960); Neyman (1967); Neyman and Scott (1961)] have argued strongly that the residual uncertainties are so pervasive as to render useless all analyses of operational weather modification programs. Others [e.g., Thom (1957a); Court (1960); Panel (1966)] have pointed out ways in which biases can be eliminated and uncontrolled variance at least partially suppressed, so that useful information can be gleaned from operational projects.

A number of possible biases are dealt with fairly simply. Agreements before a project begins as to which rain gages are to be included in calculating the target and control rainfall, for example, go far toward eliminating both unconscious bias and any temptation to select data to demonstrate a desired result (Court, 1960). Starting-time and stopping-time biases can also be reduced greatly by agreements fixed well in advance on when the evaluation period should begin and end.

The most serious difficulty with the historical regression method has to do with the stability in time of the target–control relationship. This difficulty arose very early in the evaluation of operational cloud seeding projects. MacCready (1952) performed an evaluation of a winter cloud seeding project in central Arizona using the historical regression technique and reported indications of a significant increase in rainfall. Brier and Enger (1952) performed several tests of the same project using different controls and different historical periods for establishing the target–control regression line. Their results showed considerable variation in the apparent rainfall increase due to seeding.

Neyman and Scott (1961) and others have hypothesized that the lack of stability in the target–control relationship is related to the occurrence of specific storm types, some of which favor the target area and some of which favor the control area. The apparent success or failure of a rainfall stimulation experiment might then depend on the relative abundance of the two storm types during the seeded period as compared to their relative abundance during the historical period. Brownlee (1960) not only endorsed the concept of variations in storm type, but suggested that weather modification operators could take advantage of their forecasting ability to select for seeding only those situations which promised to favor the target area. Thom (1957b) had reported no evidence that operational cloud seeders were skillful enough forecasters to do such a thing, but Brownlee (1960) countered that they typically did not seed all storms crossing a target area. Court (1960) pointed out that this source of bias, whether deliberate or unconscious, could be overcome by objective typing of storms during both the seeded and historical periods, or by agreeing that all storms during the operational period should be counted in the evaluation. This latter arrangement, which is the simplest and most objective, but which involves some dilution of results by incorporating conditions unsuitable for seeding in the target area, was followed in the arrangements worked out in 1955 between sponsors and cloud seeding operators in Santa Clara County, California, and adhered to over a 10 year period (Dennis and Kriege, 1966).

Even with the above arrangement in force, there is no complete guarantee that long term climatological trends will not change a target–control relationship. The best that can be done appears to be to follow the criteria noted above for the selection of control areas and to be alert to any obvious changes in weather patterns that could distort the target–control relationship. One must not go to extremes in this regard; obviously, if one looked long enough, one could always find *something* that was different between the historical period and the operational period (Gabriel, 1979). For example, scientists of the Alberta Research Council have hypothesized that latitudinal shifts in the mean position of the jet stream over Alberta during the summer months may have been responsible for changes in hailfall characteristics which were observed in and around areas of operational hail suppression programs in that province. Their suggestion is interesting, but it does not automatically invalidate the previous analysis.

6.3 DESIGN AND EVALUATION OF RANDOMIZED EXPERIMENTS

The residual uncertainties associated with the evaluation of operational cloud seeding projects, particularly uncertainties regarding changes with time in target–control regressions, led many meteorologists and statisticians to conclude in the 1950s that reliable results could be obtained only through randomized experiments. They reasoned that the methods of randomized experimentation developed over the previous century to test effects of new drugs, fertilizers, industrial processes, and so on would provide a solution to the large uncertainties related to cloud seeding effects. Through the use of replication and certain assumptions regarding the underlying populations from which samples are drawn, these methods permit one to establish the probability that a given hypothesis is true or false. This section reviews the principles involved. For more complete discussions see Neyman and Scott (1967a), Brier (1974), and the report of the Weather Modification Advisory Board (1978).

Basic Assumptions and Definitions

In a weather modification experiment one observes a number of test cases or experimental units with and without a specified seeding treatment applied. If the decision on whether or not to apply the treatment is made by random choice, one can assume that other things are equal (in the long run) and that observed differences between the seed and no-seed samples can be ascribed to the seeding. In essence one views the experiment as the making of random draws from two infinite sets of numbers as

$$\begin{array}{llllll} U1, & U2, & U3, & U4, & \ldots, \\ V1, & V2, & V3, & V4, & \ldots, \end{array}$$

where U represents some observation characterizing the unseeded storms, V denotes the same observation for the seeded storms, and the storms are numbered 1, 2, 3, etc.

Unfortunately, one cannot observe both U and V for the same storm. Instead, one observes sample values of both U and V in order to estimate the parameters of the two populations. The effect of seeding is calculated by comparison of the two sets of parameters.

Suppose, for example, one wishes to know the effect of seeding on the average rainfall in the target area. One defines U (or V) to be the average rainfall as indicated by observations at a network of rain gages during each experimental unit, which for simplicity we shall call a storm. The standard statistical approach is to test the null hypothesis, that is, the hypothesis that there is no seeding effect and $\bar{V} = \bar{U}$, where the bars denote averaging over a number of storms. At the completion of the experiment, one calculates the probability that $\bar{V} \neq \bar{U}$ and draws an appropriate conclusion.

The experimenter can make two kinds of mistakes. The first, called the *type I statistical error,* is to reject the null hypothesis erroneously, that is, to conclude that a seeding effect exists when there is none. On the other hand, the experimenter may fail to reject the null hypothesis when it is false, that is, to conclude that there is no seeding effect when one actually exists. This is the *type II statistical error*. It is obvious that both types of error occur in historical regression analysis also.

The probability of making a type I error is the *significance level* of the test, denoted by α. One determines whether the result of an experiment is "statistically significant" or not by computing the p value, the probability that an indicated seeding effect is due to chance, and comparing it with α. If the p value is less than α, one concludes that a real seeding effect exists.

In analyzing experiments in weather modification many authors have used a significance level of 0.05 as the criterion for rejecting the null hypothesis and accepting a seeding effect as *statistically significant*. Others, noting the extreme variability of precipitation data, have argued for a significance level of 0.10 as the criterion for rejecting the null hypothesis.[3]

The more stringent the significance level one establishes, the greater the possibility of the type II error. The *power* of a test is defined as the probability that the test will lead to the correct conclusion that $(\bar{V} - \bar{U})$ is not zero when in fact it is not. The power of a test β is a function of α, of the characteristics of the population of storms being treated, of the magnitude of the seeding effect, and the duration of the experiment.

It is possible, and highly desirable, to study proposed randomized experiments in advance to determine their power. Otherwise, resources can be squandered on experiments with very little chance of achieving definitive results.

Neyman and Scott (1967a, p. 340) have defined an important "noncentrality parameter" τ, which combines the seeding effectiveness, assumed to be a constant multiplicative effect θ, the number N of experimental units in the experiment, and the proportion P of the cases reserved as no-seed cases. The resultant formula is

$$\tau = \Delta[NP(1 - P)]^{0.5} \ln \theta. \tag{6.9}$$

This relationship among α, β, and τ is graphed in Fig. 6.4.

Equation (6.9) shows among other things that the probability of detecting a seeding effect in a randomized experiment is maximized by having exactly half of the cases reserved as no-seed cases. However, the power is not greatly damaged by choosing a P slightly different from 0.5. Another point to note is that the noncentrality parameter varies as $N^{0.5}$, the square root of the number of cases, rather than as N itself. Both of these points apply generally to parametric tests and to many nonparametric tests.

The factor Δ reflects all other characteristics of the experiment, including local conditions, contemplated design, and statistical tests used. Neyman and Scott found that if an optimal test were employed without predictor variables and the target precipitation followed a gamma distribution with a shape factor γ', then

$$\Delta = (\gamma')^{1/2}. \tag{6.10}$$

Typical shape factors for sets of rainfall data are from 0.5 to 1.5.

[3] A complication arises. Some authors have tested the null hypothesis against the alternative of *either* a decrease or an increase (two-tailed test). Others have tested the null hypothesis against a change in a prespecified direction (one-tailed test). Unless otherwise specified, significance levels and *p* values quoted in this book are for one-tailed tests.

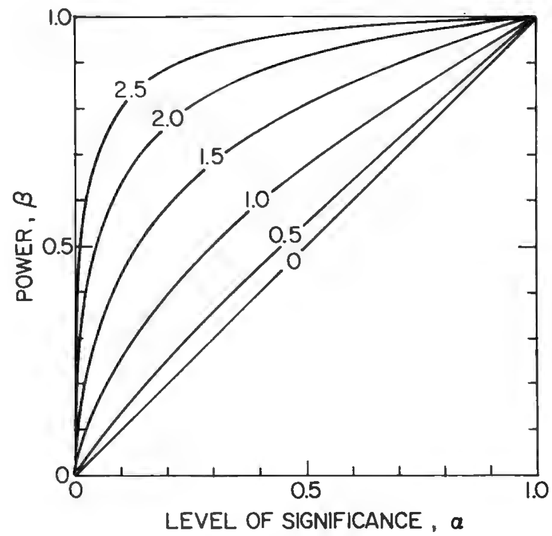


Fig. 6.4. Relation between power and level of significance for various values of the noncentrality parameter τ. [After J. Neyman and E. L. Scott (1967a). In *Proc. Berkeley Symp. Math. Statist. Probab. 5th,* Vol. V: Weather Modification Experiments (L. LeCam and J. Neyman, eds.), pp. 327–350, by permission of Univ. California Press, Berkeley.]

Examples worked out using data from Neyman and Scott for experiments in different climatological regimes are shown in Table 6.1. The first three listed are actual experiments, but the "East Central Illinois Experiment" exists only in the computer. It should be noted that the calculations assume a randomized single-area design and the use of a particular test, the $C(\alpha)$ test, which is an optimal test for detecting a constant scale factor increase in precipitation by seeding, that is, a constant percentage increase in terms of the natural precipitation. It is also assumed that the individual experimental units are statistically independent of one another.

The results in Table 6.1 were worked out assuming that the effect of

TABLE 6.1

Number of Experimental Units and Seasons Required to Detect a 40% Rainfall Increase under a Randomized Single-Area Design[a]

Experiment	No. of units	No. of units per season	Seasons
SCUD (Eastern U.S.)	178	19	9
Grossversuch IIIA (Swiss hail experiment)	390	15	26
Arizona I	460	22	21
East central Illinois (summer showers)	770	38	20

[a] Based on data from Neyman and Scott (1967a).

seeding is to give a uniform 40% increase in rainfall and setting β at 0.9 and α at 0.1. To make it very plain, a randomized experiment on convective clouds in east central Illinois producing a 40% rainfall increase from every seeded case could be run for 20 years, and the experimenters would still have a 10% chance of failing to detect the fact that any increase existed at all. This result as well as the results of Schickedanz and Huff (1971) and others show clearly that controls are as necessary in cloud seeding experiments as in the evaluation of operational programs. The requirement for controls has led to the design and adoption of a variety of experimental designs, which will now be discussed briefly.

Fixed Area Designs

Controls have the effect of reducing Δ in (6.9) and reducing the number of experimental units required in cloud seeding experiments. Controls can be drawn from surface and upper air maps or from the vertical structure of the air mass over the target area during seeding. Spar (1957) took three control variables for Project SCUD from upper air charts for eastern North America and used them as predictors of storm development. Neyman and Scott found that Spar's predictors would reduce the number of experimental units required on SCUD (Table 6.1) from 178 to only 42.

As in the evaluation of operational projects, one of the best controls on the response variable in the target area is the observed value of the same variable in some nearby control area. The target–control design and the randomized crossover design take advantage of this fact. In the *target–control* design the control area is never seeded and the decision on whether or not to seed the target during an experimental unit is made at random. It is assumed that the control area is unaffected by seeding in the target. If there is any doubt about this condition, the results of the entire experiment are called into question. Therefore, one must exercise extreme caution in the assignment of control areas to randomized experiments, despite the apparent improvement in efficiency which is thereby achieved.

The *randomized crossover* design apparently originated with Moran (1959) in Australia and has been used on several projects there (Fig. 6.5). Two similar target areas are set up and the seeding treatment is applied to one or the other in accordance with a random decision.

In theory the randomized crossover design is more effective than either the single area or the target–control design. According to Gabriel (1967, p. 93), the relative efficiency of the three designs is as given in Table 6.2. These figures indicate that the relative advantage of the ran-

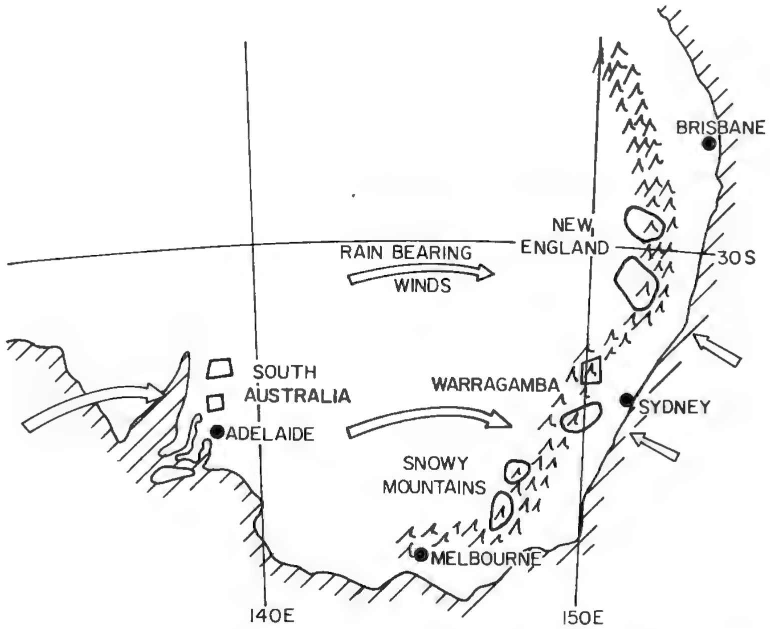


Fig. 6.5. Map showing paired target areas in four Australian experiments on stimulation of precipitation that used the randomized crossover design. Shaded area is the Great Dividing Range. [After E. J. Smith (1967). *In Proc. Berkeley Symp. Math. Statist. Probab., 5th,* Vol. V: Weather Modification Experiments (L. LeCam and J. Neyman, eds.), pp. 161–176, by permission of Univ. California Press, Berkeley, and the author.]

domized crossover experiment increases as the correlation between the two target areas increases toward one.

Enthusiasm for the randomized crossover design has waned in recent years for a number of reasons. Cross-target contamination is always a possibility although, as Gabriel has pointed out, cross-target contamination could never lead one to reject the null hypothesis erroneously. His point is that, if there is no effect, there is no contamination. However, there are other serious deficiencies. The design provides no completely unseeded cases as a check on whether conditions in and around the target

TABLE 6.2

Effectiveness of Target–Control and Randomized Crossover Designs vis-a-vis Single Area Design[a]

Type of project	Relative number of experimental units required
Single area	1
Target–control	$1 - r^2$
Crossover	$(1 - r)/2$

[a] After Gabriel (1967).

area during the experiment resembled those of past years. Furthermore, at some appreciable distance downwind, say 250 km, the distinction between seeding of target A and target B, which may be separated by only 20 km, is of no importance, so the study of extra-area effects outside of the designated target area is practically impossible.

Other Types of Experiment

It is not necessary that a randomized cloud seeding experiment involve events over a fixed target area. Some experiments have involved *floating target areas,* which permit the recording of precipitation or other variables from clouds selected as experimental units while excluding events in nearby clouds. Radar sets, especially if equipped with digital data processors, have proven very useful in recording events in moving target areas [e.g., Dennis *et al.* (1975a)]. One of the earliest randomized experiments, the University of Chicago hygroscopic seeding trials over the Caribbean Sea, utilized an airborne radar set for evaluation purposes (Braham *et al.,* 1957). In such experiments, it is vital to guard against subjectivity in deciding what events are associated with the test cloud (experimental unit) and which are not.

The use of floating targets, radar sets, and aircraft sensors vastly increases the scope of possible experiments. The U.S. Department of the Interior is presently engaged in a program (HIPLEX) in which cloud physics data collected by aircraft will be among the primary response variables (Silverman and Eddy, 1979).

Some of the response variables monitored in randomized experiments are even harder to handle than rainfall data. The mass of hail falling at a point on the ground tends to be zero during most observational periods. Cloud physics variables, such as graupel concentrations in clouds, follow no well defined statistical distribution. Possible statistical approaches to these intractable variables are mentioned in the following subsections.

Unfortunately, *steamflow,* which is very useful in evaluating operational projects by the historical regression method, is of little use in randomized experiments [e.g., Yevdjevich (1967)]. Ordinarily only one observation is available per year, so the experimental unit has to be an entire season. Storms yielding increases and decreases may be merged in the analysis and the seeding effects obscured altogether.

Use of Nonparametric Tests

We have already mentioned that rainfall data are sometimes subjected to transformations to obtain normalized distributions. The difficulties encountered in the interpretation of statistical analyses based on transformed data apply to randomized as well as to nonrandomized situations. The difficulties become acute in dealing with very highly skewed data, such as rainfall from a single cloud, or data which obey no common distribution function.

Fortunately, there is a family of statistical tests, the nonparametric tests, which make no assumptions about the distribution functions of the data being tested. The most common of these are the rank tests, the simplest of which is the Wilcoxon or rank-sum test.

In the Wilcoxon test, all of the N storms in an experiment are ranked from the driest (1) to the wettest (N). The question is asked, "Do the ranks of the seeded storms differ significantly from what would be expected by chance?" The question is answered by considering the test statistic T, which is the sum of the ranks assigned to the seeded storms.

The expected value of T, neglecting ties, is given by

$$E(T) = \tfrac{1}{2}m(m + n + 1) = \tfrac{1}{2}m(N + 1), \tag{6.11}$$

where m is the number of seed cases, n is the number of no-seed cases, and $N = (m + n)$ is the total number of cases (experimental units). The variance of T, again neglecting ties, is given by

$$\mathrm{Var}(T) = \tfrac{1}{12}mn(N + 1). \tag{6.12}$$

Elaborations to handle tied ranks and random groupings into more than two classifications are available in the standard statistical literature.

The Wilcoxon test is not particularly sensitive to the larger storms in the sample. An increase of 1 mm in precipitation from a small storm may change its rank and have as much impact on the total result as an increase of 10 or 20 mm in a large storm. Indeed, if the largest storm of all in a sample happens to fall in the seed category (by chance), any increase in rainfall from that storm would not affect the result at all.

One may have as a hypothesis concerning seeding effects the possibility that seeding increases precipitation from some types of storms and decreases precipitation from other storms. If these two general types correspond to small and large storms, one may wish to use statistical tests which emphasize such possibilities. One can emphasize the larger storms in a sample by using sums of ranks raised to some power greater than one [e.g., Taha (1964)], and can emphasize the smaller storms by using the sums of ranks raised to some power less than one. Examples of such tests are given by Duran and Mielke (1968). In addition, Mielke (1972) has experimented with noninteger exponents on the ranks to bring out various aspects of seeding effects upon large samples of storms.

Mielke (1974) has developed a special rank test for evaluating randomized crossover experiments. The differences in rainfall at target and control stations are ranked in such a way that the ranks are centered on zero and extend to positive as well as negative numbers. The absolute values of the numbers are then squared and summed with the original signs (plus or minus) being retained. This test had the characteristics of emphasizing those events in which the rainfall in the two areas differed by substantial amounts, regardless of which area had the more rain.

Comparison of Parametric and Nonparametric Tests

The nonparametric tests are less powerful than the parametric tests in the sense that they yield generally larger (weaker) p values for a given data set. This can be considered a result of the fact that they do not take advantage of what has been learned previously concerning the distribution of rainfall amounts. A statistician taking advantage of the fact that rainfall amounts tend to follow a gama distribution can, in theory, derive from a given data set a somewhat more precise estimate of the parameters of the underlying population than a statistician who looks at the data set in isolation.

Duran and Mielke (1968) have examined the power of nonparametric tests and concluded that the power of the Wilcoxon test, for example, is only about 10% less than the power of the best parametric tests for detecting a scale shift in a gamma distribution, which corresponds to a constant percentage increase in rainfall for a typical population of rainfall amounts. Furthermore, if the rainfall amounts do not actually conform to the assumed distribution, the computed p values from the parametric test will be erroneous, inspiring unjustified confidence in the experimental results.

The nonparametric tests have the advantage of being robust. That is, their ability to detect changes is not dependent on minor variations in the characteristics of the underlying population from which the particular samples are drawn, not is it sensitive to failure of the seeding effect to conform to some preconceived pattern, for example, a constant percentage increase in very storm.

Permutation and Monte Carlo Tests

While permutation procedures have been known and used since the 1930s, it was only with the advent of computers that their use became widespread.

The essence of the method is the comparison of test statistics generated on an actual experiment with test statistics generated on hypothetical experiments using the same data, but with the total set of experimental units reallocated between the "seed" and "no-seed" categories. When done after the fact, this procedure is sometimes called rerandomization.

The permutation tests set p values which are not affected by any assumptions regarding the underlying statistical distributions. Of perhaps greater importance, the p values that result automatically take account of any lack of statistical independence in the experimental units. [Thom (1957b) was forced to do separate analyses for odd- and even-numbered storms to overcome a serial correlation in the storms he studied. Otherwise, his computed p values would have been suspect.]

Sometimes all possible permutations of seed and no-seed decisions are attempted (Neyman *et al.,* 1960). Usually, only a random sample is used to avoid excessive computer usage; in that case the simulated experiments in the computer are referred to collectively as a Monte Carlo test.

A detailed description of a Monte Carlo test was given by Dennis *et al.* (1975b), who showed that 500 simulated reruns of a randomized experiment out of a possible 4^{48} fixed the p values to about ± 0.02. Although their approach was questioned privately by some statisticians, the statistical task force of the Weather Modification Advisory Board (1978) concluded that the approach did provide proper estimates of p values.

The statistical task force of the Weather Modification Advisory Board has proposed that exact significance levels can be established through permutation methods as follows: Set up a number (say, 1000) of seed–no-seed sequences *in advance* of experiment. Each sequence must be completely satisfactory in terms of avoiding long runs of seed or no-seed decisions, allocating of treatments fairly among identifiable subsets of experimental units, and so on. This is called establishing a "sample space." It need not be done by random draws. Then one of the sequences is selected *at random* to be followed in the actual experiment. Comparing the actual results with those of the 999 "phantom" experiments (examined with the same total data set) determines exactly the probability that as good or better results would have arisen through chance, and therefore sets the p values (Gabriel, 1979).

Use of Partitioning or Stratification

To this point we have implicitly assumed that all of the experimental units from an experiment enter into each statistical test. It is quite possible for an evaluation to show no seeding effect if increases and decreases are evenly distributed throughout the seeded sample. We have noted attempts to select statistical tests which would favor the detection of increases or decreases concentrated among either the small or the large storms. However, one can visualize a situation in which increases and decreases are distributed rather uniformly between small and large storms, so that the choice of such tests would be irrelevant.

If there were other criteria which would isolate the storms responding to seeding, the situation would be much improved. Many analysts have used variables believed relevant to the response of the clouds to seeding as an aid in statistical evaluation.

Gabriel (1967) and others have spoken of these additional variables as concomitant variables. They could be introduced into the problem as additional predictors in a MLR analysis, but many investigators have found it more useful and straightforward to use them as a basis for partitioning or stratification of the data sample.

In analyzing the Climax experiment in Colorado, Grant and Mielke (1967) found it advantageous to stratify the data according to the temperature at 50 kPa (500 mbars), which was considered representative of cloud top temperatures in winter storms in the mid-Colorado Rocky Mountains. Simpson and Woodley (1975) stratified experimental days in Florida as "marching" or "stationary" days, depending on whether the radar echoes from showers showed systematic motions.

Other stratification schemes proposed or used on various projects include stratification by direction of upper winds [e.g., Mooney and Lunn (1969)], the presence or absence of cloud seeding on a nearby operational project (Neyman *et al.*, 1960), and the predictions of cloud models (Section 6.4).

Exploratory versus Confirmatory Experiments

The variety of response variables, stratification schemes, and statistical tests now available makes it increasingly likely that any cloud seeding experiment will turn up at least a few statistically significant results. This is the *multiplicity* trap. The statistical task force of the Weather Modification Advisory Board (1978) suggested a distinction between *exploratory* and *confirmatory* experiments. In the former, a wide ranging search for possible effects would be conducted using all available tools. However, no effect would be accepted as proven until it had been demonstrated in a confirmatory experiment set up to test a very small number of hypotheses (ideally one) using a response variable and a statistical test specified in advance.

Uncontrolled Background Variations

A number of authors have studied randomized cloud seeding experiments which yielded significant results to see if the results might not have been due to natural or background variations in weather patterns. Gelhaus *et al.* (1974) spoke of a possible type I error on a project in South Dakota after finding that a rainfall deficiency on seed days existed in the early morning hours before seeding began.

The question of background variations has been especially troublesome to analysts of Project Whitetop, which was operated in Missouri from 1960–1964 inclusive. The latest exchange on the project (Braham, 1979) shows that divergent views still exist, with some authors seeing seeding effects where others see natural variations in rainfall.

The search for background variations to explain away significant results of randomized experiments is analagous to the search for changes in storm types to explain changes in target–control regressions during operational cloud seeding projects. In both cases, some hypothetical explanation other than cloud seeding can usually be found if one searches long enough. As Gabriel (1979) has noted, p values presented simply because they exceed a predesignated α are not valid tests of probability. Investigators following this approach will fall into the *multiplicity trap* just as surely as will their colleagues who analyze experimental data endlessly looking for evidence of favorable cloud seeding "results."

6.4 ROLE OF CLOUD MODELS IN EXPERIMENTS

The features which distinguish a cloud which responds to seeding in a given way from a cloud that responds in some other way or not at all are very subtle. They may not be identifiable on the basis of the simple stratification systems just mentioned. Sometimes the differences are not to be identified on the basis of in-cloud conditions at all, but on the basis of ambient conditions. Lapse rate, wind shear, and the degree of convergence in the subcloud layer all play a part in determining a cloud's response to seeding. These differences have already been explored in conceptual terms in Chapter IV. As we saw there, the numerical cloud models are the best tool available at this time for evaluating the various possibilities that exist. However, one cannot conclude on the basis of cloud model studies alone whether a given seeding treatment will produce desired effects or not. The models are incomplete and require verification in the field. It is therefore inevitable that cloud modeling studies and field experiments should come together. Some field experiments have been set up purely to verify cloud models, and most field experiments now make use of cloud models to some extent.

Cloud models are used in at least three ways on weather modification experiments: (1) to select experimental units which will respond to seeding in a desired fashion, (2) as a basis for stratification of experimental units, and (3) to provide predicted values of response variables against which observations can be compared. We shall now discuss each of these three uses very briefly. Further examples of how the models have helped in specific applications will be provided in Chapter VII.

Selection of Experimental Units

The cloud models are most useful in selection of experimental units when the experiment involves seeding convective clouds for dynamic effects. In fact, the desire to modify the dynamics of convective clouds in a predictable fashion provided much of the incentive for development of the entity cloud models at Pennsylvania State University and elsewhere.

In most experiments on convective clouds the model predictions are one of the factors considered in selecting experimental units. General weather forcasts are useful in determining whether or not convective clouds will develop during an experimental period. One can also make useful predictions as to whether or not clouds would reach the $-10°C$ level and the like by simply examining the sounding data. However, the use of the cloud model has the advantage of being more objective. The elimination of subjectivity makes the data resulting from an experiment more useful for statistical analysis.

The most outstanding example of the use of cloud models in selecting experimental units is in its selection of days with predicted increased in cloud height (ΔH greater than 0) for dynamic seeding experiments over the Caribbean Sea (Simpson *et al.*, 1967), in Florida (Simpson and Woodley, 1971), and Arizona (Weinstein and MacCready, 1969). Another example of the use of cloud models in selecting experimental units is on the Alberta Hail Studies Program. There a cloud model is run to predict maximum updraft speeds, and the updraft speeds are then used with nomograms to determine likely hailstone size. Experimentation is limited to days when strong updrafts are predicted which would lead to significant hailfalls (English, 1975).

Partitioning of Data

There have been some randomized experiments which were conducted without any use of cloud models in selecting experimental units, but where the introduction of cloud models during the analysis phase proved useful for stratification of the data.

An example is provided on the North Dakota Pilot Project, where the addition of an on-site rawinsonde station halfway through the four year project permitted the data for the last two years to be stratified according to model predictions of dynamic seedability, that is, of increases in cloud height due to AgI seeding (Dennis *et al.*, 1975b). The data for the past two years (1971–1972) showed precipitation per seed day to exceed that per no-seed day, but the results were not significant (one-tailed p value of 0.12). Partitioning on the basis of the presence or absence of dynamic seedability showed that on days without it rainfall per seed day was almost identical with that per no-seed day. Practically all of the apparent rainfall increases were associated with the days when the model predicted dynamic seedability. For those days the apparent rainfall increase was very marked. The p value was reduced from 0.12 for the entire two year sample to 0.07 for the dynamic seedability stratification.

Predictions of Seeding Results

In some cases one can use the cloud models in a more refined fashion than simply having them pick out favorable instances for seeding either before or after the fact. If the cloud model output is precise enough and observational data are available for comparison, one can use the cloud model predictions as controls on individual experimental units.

A well known example of such an application of cloud models is that of the Stormfury cumulus experiments. The response variable was the height reached by cloud towers. Instead of comparing the heights of seed clouds as a group to the heights of the no-seed clouds, a cloud model prediction was made in every case. The analysis is illustrated by an actual set of experimental data in Fig. 6.6, which shows the predicted increase in cloud height associated with seeding for both the seed and no-seed cases along the abscissa and the observed height change as the ordinate. One would not expect height changes other than random variations for the no-seed cases, so they should scatter along the x axis. On the other hand, if the model predictions were perfect and there were no measurement errors, the seed cases would be distributed along the 1:1 line. The actual results suggest that increases in cloud height were achieved, and that the model showed skill in selecting clouds which would grow following seeding (Simpson *et al.*, 1967).

We have already described kinematic models capable of making predictions of snowfall in orographic cloud seeding situations as a function of wind speed, cloud top temperature, and so on. These could be used to generate control figures also, but confidence in those models is not sufficient to justify their use as quantitative predictors of precipitation at a

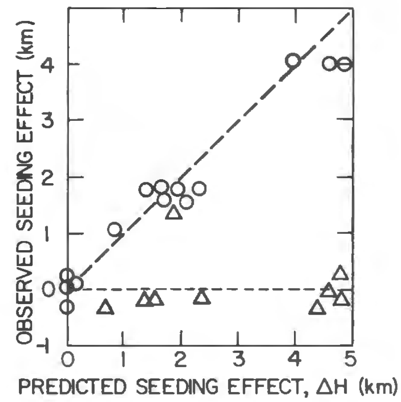


Fig. 6.6. **Observed versus predicted effect of seeding upon cloud height in a randomized experiment over the Caribbean Sea in 1965. Heavy dashed line represents perfect predictability for seeded clouds, while dashed line represents perfect predictability for control clouds, that is, zero seeding effect independent of seedability. Unfilled circle: seeded; unfilled triangle: unseeded. [After J. Simpson *et al.* (1967). *J. Atmos. Sci.* 24, 508, by permission of American Meteorological Society and senior author.]**

given point on the ground. They are, however, useful as general guides to those situations in which increases or decreases might be expected and to point out the types of snowfall redistributions that might occur under different situations.

6.5 SYNTHESIS

The evaluation of weather modification experiments is a difficult problem. Two principal responses to it have evolved, one being the straightforward use of statistics in randomized experiments, and the other the development of physical and numerical models which predict responses to seeding. Even in situations where the statistical analyses have indicated seeding effects, most scientists have been unwilling to accept them in the absence of plausible, physical models as to how the effects are produced. On the other hand, numerical models can predict extremely wide ranges of results depending on the initial assumptions about cloud droplet distributions and so on, so that one must utilize the results of field experiments in drawing conclusions about the effectiveness of weather modification technology.

The most recent experimental designs tend to emphasize the use of several response variables rather than one, so the arbitrary distinction between physical and statistical evaluations is becoming blurred. Mielke (1979b) has developed a multiresponse permutation procedure (MRPP) to calculate p values that apply to sets of response variables. It is hoped in this way to check out the steps in various physical hypotheses linking cloud seeding to changes in precipitation.

A synthesis of numerical modeling studies with field experiments seems to be the best answer available as of 1979. The numerical models should be used in all phases of experiments, including the design, the selection, and stratification of experimental units, and to provide predictions of the expected results in individual experimental units. This intensive use of cloud models should not preclude the use of other applicable data which might be available from synoptic maps, air mass soundings, meteorological satellites, and other sources.

CHAPTER

VII

The Modification of Fog, Snow, and Rain

7.1 INTRODUCTION

In this chapter and Chapter VIII we discuss weather modification in terms of operational objectives. The success which can be attained in modifying fog, precipitation, lightning, and so on by cloud seeding is to many people the ultimate test of the value of cloud seeding technology.

A successful weather modification project must combine a seeding concept or physical hypothesis appropriate to the cloud system being treated, a sound technology for implementing the concept, adequate sensors to record seeding effects, and a sensitive evaluation plan. To this point we have examined these constituents of a successful project on an individual basis. We now turn to the results achieved in the field by persons who have attempted to put the pieces together in experiments and in pursuit of operational objectives.

In view of the difficulty of evaluating cloud seeding projects, it should not be surprising that there are differences of opinion among scientists and engineers concerning the effectiveness of present methods of weather modification by cloud seeding. What is surprising, and encouraging, is the gradual emergence of a consensus shared by many scientists who have studied the field in depth. The emerging consensus is expressed in reports by successive review panels to the National Academy of Sciences (Panel, 1966; Panel, 1973) and in the report of the Weather Modification Advisory Board (1978).

This chapter and Chapter VIII present the opinions of the present author, rather than the consensus. However, the differences are not great.

As we shall see, supercooled fog at airports has been dissipated on a routine, operational basis for years, and precipitation has been increased in some cloud seeding projects and redistributed in others beyond any reasonable doubt. These limited but economically important successes continue to spur hopes that more complex systems such as tropical hurricanes will eventually prove amenable to predictable modification.

7.2 FOG MODIFICATION[1]

The principal economic incentive to fog modification lies in the fact that fog at airports is a hindrance to the takeoff and landing of aircraft. Operations are conducted to reduce the impact of fog on aircraft operations at airports in several countries, including about a dozen airports in the western and north central United States and about 15 airports in the U.S.S.R.

The controlling variable in fog episodes is the visibility. The objective in a typical fog clearing operation is to improve visibility along the final approach path of an instrument landing system and around the touchdown point. The volume of air which must be affected assuming optimum targeting is 10^5–$10^6 m^3$.

The visibility in fog varies inversely as the sum of the optical scattering cross sections of the fog droplets per unit volume. The scattering cross sections are proportional to the geometric cross sections of the fog droplets. For a given fog concentration, therefore, visibility is better in a fog of large droplets (often characteristic of advection fog) than in a fog of very fine droplets (typical of newly formed radiation fog).

Visibility in a fog can be improved by removing some of the fog droplets, by reducing the diameters of the droplets while keeping the numbers constant, or by promoting the coalescence of fog droplets while keeping the fog water concentration constant.

The more important of the many methods suggested for fog modification are set forth in the following paragraphs, which are organized according to the physical state of the fog.

[1] For a general reference on fog modification, see Silverman and Weinstein (1974).

Dissipation of Supercooled Fog

The most commonly used method to dissipate supercooled fog is based on the Bergeron process. The introduction of artificial ice particles leads to evaporation of the supercooled droplets and water vapor deposition upon the ice particles. Once the ice particles become large enough, they begin to grow by accretion also. In some cases the ice particles become large enough to fall to the ground as snow producing a complete clearing. Even if this is not accomplished, the reduction in the number of small droplets and the concentration of the water substance into a relatively small number of large particles improves visibility.

The ice particles can be introduced by seeding with AgI or other artificial ice nucleants, or by chilling the air temporarily below −40°C with dry ice, liquid propane, or liquid air. Because supercooled fog often occurs at temperatures just below 0°C, where AgI and other artificial ice nuclei are relatively ineffective, most operations have involved the chilling agents.

The desirable concentration of embryonic ice particles depends on the characteristics of the fog and the time available for clearing. If a rather complete clearing is desired, relatively low ice embryo concentrations around 10–50 liter^{-1} should be used so that the resulting ice crystals or small snowflakes will fall to the ground, thereby producing partial or even complete local clearing. The process typically takes 30 minutes or more. If one wishes to eliminate a significant fraction of the supercooled water within a short time, say five minutes, ice particle concentrations around 1,000 liter^{-1} must be employed (Jiusto and Weickmann, 1973). The likely product in such a case is an ice fog of coarse particles with visibility possibly improved enough to permit aircraft operations, but not a complete dissipation of the fog.

The engineering aspects of clearing supercooled fog can be worked out for any given delivery system. Consider, for example, a light aircraft dropping 1 kg km^{-1} of dry ice pellets from 200 m above ground with the pellets sized to just reach the ground. Assuming about 10^{14} ice crystals produced per kilogram of dry ice dropped (Holroyd *et al.,* 1978), an ice crystal curtain containing 5×10^8 ice crystals per square meter of its vertical section is the initial result of a seeding pass (Fig. 7.1).

The volume influenced by the ice crystals spreads as a result of turbulent diffusion and also drifts with the wind. The simple situation just outlined requires consideration of diffusion in only one dimension. Using (5.6) and values of ε, the energy dissipation parameter, appropriate to stratiform cloud situations suggests that the curtain would reach a thickness of 200 m in about 10 minutes, and would continue to spread thereafter at 1–2 m s^{-1} (counting both sides). Turbulent diffusion is enhanced

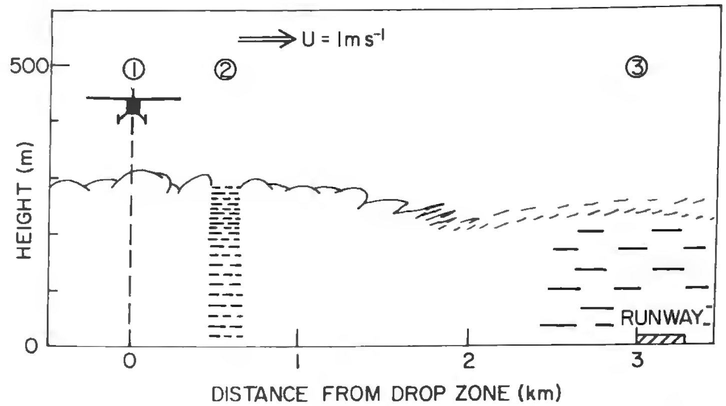


Fig. 7.1. Clearing of supercooled fog by dry ice seeding. (1) $t = 0$. Aircraft drops dry ice at 1 kg km^{-1}; ice crystal curtain forms with 5×10^8 crystals per square meter. (2) $t = 500$ s. Curtain drifts downwind while spreading due to turbulence. Ice crystal concentration has decreased to a few thousand per liter due to turbulent spreading. (3) $t = 3000$ s. Runway visibility improves as area of partially cleared cloud, now over 1 km wide, drifts across runway. Ice crystal concentration is down to about 100 liter^{-1} due to turbulent spreading, aggregation, and fallout. Very light snow near runway or upwind. (4) Cleared area fills in due to turbulent mixing with untreated air some 50–100 min after dry ice drop.

somewhat by the local heating associated with freezing of the supercooled fog.

The mean concentration of ice crystals at the end of 10 minutes would be near 2000 liter^{-1} with higher concentrations near the center. The calculated concentrations are in the range for which parcel model calculations suggest that practically all of the supercooled water would be removed by evaporation and accretion over a period of 10 minutes or less (Fig. 7.1). These predictions are verified by experience on field programs, many of which employ dry ice seeding at rates comparable to the example just quoted. A few snowflakes began to fall in 10 or 20 minutes. The smaller ice crystals continue to spread out of the narrow initial clearing, growing until they too become large and fall to the ground.

In a typical supercooled fog clearing operation at an airport, an aircraft drops 2 or 3 kg km^{-1} of dry ice on seeding passes above the fog layer parallel to the runway to be cleared. The passes are made 45–60 minutes upwind of the runway and the seeding passes are repeated frequently enough so that the resulting curtains of ice crystals are about 1 km apart.

Another method that has proven useful is the attachment of bags holding dry ice pellets to balloons which can be raised or lowered to bring the dry ice into contact with the supercooled fog at the desired elevation and temperature. In some operations the balloons and the dry ice bags are tethered to trucks or jeeps, which can be driven back and forth to simulate the action of an aircraft making seeding passes very close to the ground.

It will be recalled (Fig. 3.8) that there is laboratory evidence that ice particles in air at water saturation grow much more rapidly in the temperature range from −4 to −6°C than at either higher or lower temperatures. There is some evidence from fog modification work that the dissipation of supercooled fog is especially rapid and effective at temperatures near −5°C. If the supercooled fog is at −1 or −2°C, growth of the ice particles is slow. However, persons engaged in fog clearing operations generally consider that visibility can be improved even in those situations.

Dissipation of Supercooled Stratus

It is apparent that the aircraft seeding techniques used for supercooled fog dissipation could be used to dissipate supercooled stratiform clouds. Several authors have pointed out that a project to do so would increase the hours of sunshine experienced on the ground below, and there are reports of seeding for that express purpose in the U.S.S.R. Indeed, the first field experiment in modern weather modification involved the dissipation of part of a supercooled cloud deck.

The dissipation of supercooled stratus was investigated by the U.S. Army Signal Corps in the early 1950s. The investigators found that dry ice seeding at a rate of 0.3 kg km^{-1} or AgI seeding in clouds at rates of a few grams per kilometer resulted in clearing, and the boundary of the cleared path would spread at about 1 m s^{-1} (Aufm Kampe *et al.,* 1957; Panel, 1966, Vol. II, p. 44). Weickmann (1974) later documented the production of snowshowers and the resultant clearing of supercooled stratocumulus clouds near the Great Lakes following seeding with dry ice from aircraft.

Modification of Warm Fog by Hygroscopic Seeding

Warm fog occurs much more frequently than supercooled cloud, which is unfortunate because warm fog is more difficult to modify. The only *cloud seeding* approach to warm fog modification which has shown any promise is the spreading of hygroscopic seeding agents to dry the air by condensation upon them and to scavenge the fog droplets. This method has been explored in many numerical and field experiments. The numerical experiments which have been completed to date indicate that the accretion of the fog droplets is important only in fogs of appreciable depth, say several hundred meters, where the hygroscopic droplets have time to grow appreciably during their fall.

The action of any array of hygroscopic particles falling through a fog volume can be divided into three distinct stages as follows: (1) condensation and fallout, (2) clearing, and (3) refilling [e.g., Silverman and Weinstein (1974)]. During the condensation and fallout stage the relative humidity is depressed to say 95–98% as the hygroscopic particles take on water vapor. After the hygroscopic particles have fallen through a given volume element, the visibility improves (clearing stage) as the fog droplets partially evaporate to bring the relative humidity back up to 100%. Refilling takes place as the treated volume is invaded by fresh, foggy air through the action of turbulence. The treated volume is meanwhile advected by the wind and may be seriously distorted by wind shear. The three stages and the actions of wind shear and turbulence are shown in Fig. 7.2.

Engineering of warm fog modification requires a choice of seeding agent and delivery system, just as the modification of a cloud does. Seeding agents which have been used include hygroscopic powders, principally sodium chloride (NaCl) and urea, and sprays of hygroscopic solutions. Mixtures of urea and ammonium nitrate (NH_4NO_3) in water have been used in some fog clearing experiments. Silverman and Kunkel (1970) concluded on the basis of numerical experiments that the sprays offered little advantage, with dry particles growing large enough to acquire appreciable fall speeds while falling through only a few meters of fog.

The optimum size of seeding particles is a function of the chemistry of the seeding agent and of the fog characteristics, including the turbulence

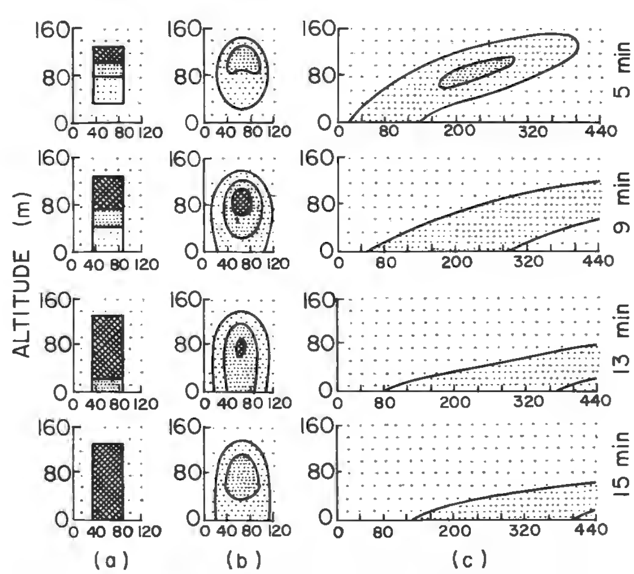


Fig. 7.2. Distortion and filling of cleared volume by wind shear and turbulence (a) $D_x = D_z = 0\ m^2 s^{-1}$, shear = 0; (b) $D_x = D_z = 1\ m^2 s^{-1}$, shear = 0; and (c) $D_x = D_z = 1\ m^2 s^{-1}$, shear = $0.01\ s^{-1}$. [After A. I. Weinstein and B. A. Silverman (1973). *J. Appl. Meteorol.* **12,** 771, by permission of American Meteorological Society and senior author.]

and wind shear. Silverman and Kunkel's (1970) simulation of seeding with powdered NaCl (neglecting turbulence) suggests that optimum results would attend use of 20 μm particles. Seeding with smaller particles can produce an overseeded condition in which the NaCl solution droplets never grow large enough to fall out within reasonable time. On the other hand, seeding with large particles, say of 50 μm diameter, yields particles which fall out of a typical fog before taking on the full amount of water which they could, in theory, extract if the fog were sufficiently thick.

Silverman and Kunkel's calculations indicate that even under ideal conditions up to 1–2 g of powdered NaCl would be required per square meter of surface area to produce significant improvements in visibility. Line seeding to clear a runway and approach zone therefore would require several hundred kilograms per operation.

As Fig. 7.2 suggests, wind shear and turbulence can destroy a cleared zone before it has time to evolve fully and settle down to a runway. Overcoming such effects requires recourse to larger seeding particles with higher initial fall speeds. Moving up to 40 μm particles would achieve the desired result in some cases. In order to reap the same benefits as when releasing the 20 μm particles under ideal conditions, it would be necessary to release almost the same number of particles, so the mass of seeding agent required would go up by a factor approaching 10.

Experience on field experiments showed quickly that targeting difficulties combined with shear and turbulence effects made line seeding impracticable in most cases. A realization that tons of NaCl powder would have to be broadcast over an airport to have an appreciable impact on aircraft operations led to a search for more effective and less corrosive seeding agents.

Kunkel and Silverman (1970) have tabulated the relative effectiveness of various hygroscopic chemicals against NaCl as a standard. Unfortunately, many of the most powerfully hygroscopic materials, e.g., potassium hydroxide (KOH), are even less acceptable from an ecological standpoint than NaCl. The most effective nontoxic, noncorrosive chemical that Kunkel and Silverman studied is urea, which they calculated to be 93% as effective as NaCl per unit mass. The optimum size for urea particles has been estimated at 60 μm or so (Weinstein and Silverman, 1973). However, urea has a tendency to shatter into fine particles of less than 20 μm diameter, so it has to be prepared and released in microencapsulated form.

Weinstein and Silverman (1973) studied the effects of microencapsulated urea upon various model fogs with the aid of a two-dimensional numerical model. Required seeding rates to bring visibility up to 800 m (0.5 miles) ranged from 1.5 to 5 g m^{-2} with an assumed cross wind of 1.5 m s^{-1}.

Limited success has attended a number of experiments in clearing advection fog by hygroscopic seeding in California (St.-Amand *et al.*, 1971f; Silverman *et al.*, 1972). [The tests described by St.-Amand *et al.* (1971f) involved spraying with solutions of urea and NH_4NO_3 in water, dispersal of powders, and pyrotechnic generation of hygroscopic powders.] On the other hand, similar operations sometimes fail to produce any marked improvement in visibility, apparently because of strong turbulence or because of difficulties in predicting how the seeded volumes of foggy air would advect through the approach zone and the volume of space above the runway (Silverman *et al.*, 1972).

Alternative Methods

Hygroscopic seeding agents to modify warm fog have fallen into disfavor in recent years except for special (e.g., military) operations. Many methods other than cloud seeding have been proposed for warm fog modification. They include (1) treating warm water surfaces in the vicinity with chemicals to suppress evaporation and reduce the possibility of fog formation, (2) mixing the fog layer with overlying warm dry air brought down by helicopter downwash, (3) promotion of fog droplet coalescence by strong sound waves, (4) evaporation of the fog droplets by heating with laser beams, (5) introduction of charged particles to collect the fog droplets, (6) the spreading of chemicals, e.g., surfactants, to modify the condensation processes (a kind of nuclei poisoning), and (7) the collection of large volumes of the fog filled air by suction devices, the removal of the fog droplets by screening or centrifuging and the exhaustion of the air, now free of fog droplets, back to the atmosphere. Several of these and other unproven concepts are listed by Weinstein and Kunkel (1976), who provide references to the original papers.

Helicopter downwash has been used successfully against shallow radiation fogs. It is useless against deep fogs.

A recent numerical model of the electric charging process indicates that it is *not* a practical approach, despite its frequent mention over the years (Tag, 1977). There are no records of successful fog clearing through electrostatic effects.

The method of warm fog modification currently enjoying the most popularity is the evaporation of the fog by warming the ambient air. This approach was first tried in England during World War II with the FIDO (Fog Improvement Dispersal Operation) system. A more sophisticated system called Turboclair was installed at Orly Airport in Paris in 1970 with eight jet engines in underground chambers along the upwind side of a runway. The Orly system was expanded in 1972, and Turboclair is now used at other airports also. A serious drawback of the FIDO system was that the warm air rose quickly from the ground and was replaced by fresh, foggy air. Turboclair uses a baffle system to mix the jet exhaust with surrounding air and discourage convective currents until the dilute exhaust reaches the runway (Sauvalle, 1976).

Although the Turboclair system is accepted as an operational tool in France, an airline official reported at the spring 1976 meeting of the Weather Modification Association that attempts to develop such a system for Los Angeles International Airport had been discontinued. This might be because Los Angeles often has advection fog drifting inland from the ocean, whereas many Orly Airport fogs form by radiation in almost calm conditions.

Modification of Ice Fog

Ice fog is sometimes a natural phenomenon, but more often a product of inadvertent weather modification. Under very cold conditions, say near −40°C, the water vapor from aircraft and/or automobile exhausts may exceed that required for ice or water saturation throughout the volumes where the vapor is spread by turbulent diffusion. Fog is a common occurrence at Fairbanks Airport, for example, where temperatures below −40°C often occur with very light winds. Furthermore, the exhausts often contain sufficient lead compounds or other impurities to cause the resultant fog droplets to freeze, producing an ice fog.

There is no satisfactory method of dealing with ice fog once it has formed. The direct application of heat is of little use because of the extent to which the ambient air must be heated to hold even an additional 0.1 g m^{-3} of water at temperatures below 0°C (see Fig. 3.1).

Problems with ice fog can be reduced in some situations by controlling local water vapor emissions (Silverman and Weinstein, 1974).

7.3 AUGMENTATION OF PRECIPITATION: IS A NET INCREASE POSSIBLE?

Of all the possible applications of weather modification technology, the augmentation of precipitation has the greatest potential economic impact. A check of field projects in the United States during 1977 shows that 76 out of 94 were for the purpose of increasing precipitation or to develop the technology for that purpose (Charak, 1978). Indeed, the entire technology of weather modification by cloud seeding is sometimes referred to loosely as "rain making."

In Chapter IV we examined a number of cloud seeding concepts which might well increase the precipitation efficiency of individual clouds or cloud systems, and which might also lead to dynamic effects with more pronounced impacts upon precipitation at the earth's surface. Before considering the results of applications of the various concepts, it is well to consider the more basic question of whether an augmentation of precipitation, as opposed to a redistribution, is theoretically possible.

The argument is sometimes offered that the total amount of water in the atmosphere is fixed, so that one cannot increase precipitation in one place without decreasing it somewhere else. This oversimplified argument confuses mass of water substance with a rate of mass transfer (rainfall), but the effect of precipitation augmentation upon the atmospheric water budget requires some analysis.

McDonald (1958) studied the potential impact of weather modification to increase precipitation upon the water content of the atmosphere (Table 7.1). The water content of the atmosphere is minute compared to that of oceans, and only a small fraction of the atmospheric water exists as visible clouds at any given moment. A comparison of the atmospheric water vapor with the annual rainfall on the globe shows that the atmosphere must turn over its water vapor inventory about once every ten days. Ten days is, of course, an average figure; an individual water molecule may exist in the vapor state for weeks if it is caught in the weak circulation around a relatively clear subtropical high pressure area; another molecule evaporated from a Kansas cornfield may be condensed and reprecipitated in a thundershower only an hour later.

Comparing the water content of the precipitating clouds in the atmosphere with the normal rainfall over the globe, we find that the precipitating clouds must have short lifetimes. Individual cloud elements form, precipitate, and dissipate in less than one hour. The precipitation process is

TABLE 7.1

Estimated Magnitudes of Quantities Controlling Effects of Worldwide Cloud Seeding[a]

Estimated fraction of total atmospheric H_2O condensed, at any one instant, into cloud liquid water (based on mean cloudiness of 0.5 and mean cloud depth of 2 km)	0.04
Same fraction for continental areas only	0.01
Estimated upper limit of fraction of all continental cloud cover amenable to treatment to induce precipitation	0.1
Generally accepted (1958) order of upper limiting magnitude of obtainable increase in precipitation by cloud modification methods	0.1
Consequent upper limit to estimate of fraction of total atmospheric H_2O which, at any instant, might be undergoing artificially stimulated precipitation over continents	0.0001

[a] Based on McDonald (1958).

not always efficient, and many clouds dissipate without precipitating more than a few percent of their total condensate. The remainder evaporates and reenters the vast supply of water vapor which is always present in the atmosphere.

An increase in the world's precipitation could be achieved by decreasing the turnover time of the water vapor inventory. This might be done by making the individual clouds more efficient or by promoting the development of larger clouds, which would process more of the water vapor on a given day. The atmosphere would compensate for either process with increased evaporation rates, which would cause some slight cooling in regions where evaporation is the dominant process, notably the tropical oceans, and slight warming in areas where the additional condensation and precipitation took place. A quite sophisticated computer model would be required to solve the total problem, but there is little doubt that the atmosphere could adapt to a faster turnover time with only minor adjustments.

There is no practical reason to seek an increase in total rainfall on the earth. Much of the earth's rain falls into the oceans, which occupy about 70% of its surface area. Most of the rainfall on the continents falls on tropical rain forests and other wet regions, where the economic incentive to increase rainfall is lacking. Even an all out effort to increase the world's food supply by increasing rainfall would be aimed at perhaps 5% of the rain that falls now. Assuming that a 10% increase could be achieved in the treated areas shows that the total increase in the earth's annual rainfall would be about 0.5%, and the average ten-day turnover time estimated by McDonald would be reduced by one hour. Of course, local water balances could be affected more markedly, so water budget studies are an important part of current research in weather modification.

The results of experiments and operations in precipitation augmentation vary considerably according to cloud types. We shall therefore consider results under the headings of *orographic* clouds, *convective* clouds, and *synoptic scale* cloud systems.

7.4 INCREASING PRECIPITATION FROM OROGRAPHIC CLOUD SYSTEMS

Results of Operational Projects

The favorable opportunity for precipitation augmentation offered by supercooled, orographic cloud systems has been noted and discussed in Chapter IV.

Operational cloud seeding programs to exploit promising orographic situations began in the western United States and in other parts of the world about 1950 and have continued ever since. Most of the projects have involved seeding from generators on the ground consuming 5–30 g hr^{-1} of AgI, but aircraft have been used in some cases to deliver dry ice, AgI crystals from airborne generators, or to drop pyrotechnic devices. The projects have been operated principally to produce increased runoff for irrigation and hydroelectric power generation. The fact that seeding is accomplished in mountainous areas during the snowy winter season to obtain water for use months later sidesteps one often cited weakness of cloud seeding technology, namely, the scarcity of seedable clouds when the need for additional moisture is greatest. A few projects have been operated for immediate results, such as increasing the snow depth at ski resorts.

Evaluations of cloud seeding projects to increase precipitation from orographic storms of the western United States have generally given indications of increases in the target areas of the order of 10–15% of the precipitation which would have occurred naturally. Estimates of increases in this range appear in reports of the Advisory Committee (1957) and in panel reports prepared under the auspices of the National Academy of Sciences (Panel, 1966, 1973). These increases were indicated by target–control analyses using the historical regression method applied to both transformed precipitation data and streamflow data (Fig. 6.3). Panel (1966, Vol II, p. 32) reported apparent streamflow increases ranging from 6% on a ten-year project on the Kings River in California to 18% on an eight-year project on the Rogue River in Oregon. Elliott and Lang (1967) applied a target–control regression analysis ($R = 0.97$) to a cloud seeding project on the San Joaquin River in California. They found the indicated streamflow increases over a 15-year period to be 8.5%. Elliott and Walser (1963) applied the less sensitive double-mass method to projects in Oregon and near Great Salt Lake and found abrupt shifts in favor of the target areas at the time orographic projects were begun. The relative consistency of results from the different projects is encouraging.

Results of Early Randomized Experiments

The objections of Brownlee (1960), Neyman and Scott (1961), and others to the utilization of data from operational projects for evaluation purposes have already been noted. As Brownlee (1960) wrote about the Advisory Committee, "It is clear that Congress is handing the committee the task of making 'a complete study and evaluation of public and private experiments,' had given them a futile assignment when in fact at that time there were virtually no experiments available to evaluate."

A series of randomized experiments on orographic clouds began in Santa Barbara County, California, in 1956 and subsequently was extended to other places in the western United States (Fig. 7.3). The experiments gave some insight into the physical mechanisms associated with the seeding of orographic clouds by glaciogenic agents, and began to identify the types of storms where seeding is most likely to cause precipitation increases.

The first Santa Barbara experiment ran for three years. The experimental design was not powerful and the entire experiment was confounded by the initiation of a nonrandomized project in adjacent Ventura County. Not surprisingly, the evaluators found no evidence of statistically significant rainfall increases (Neyman *et al.*, 1960).

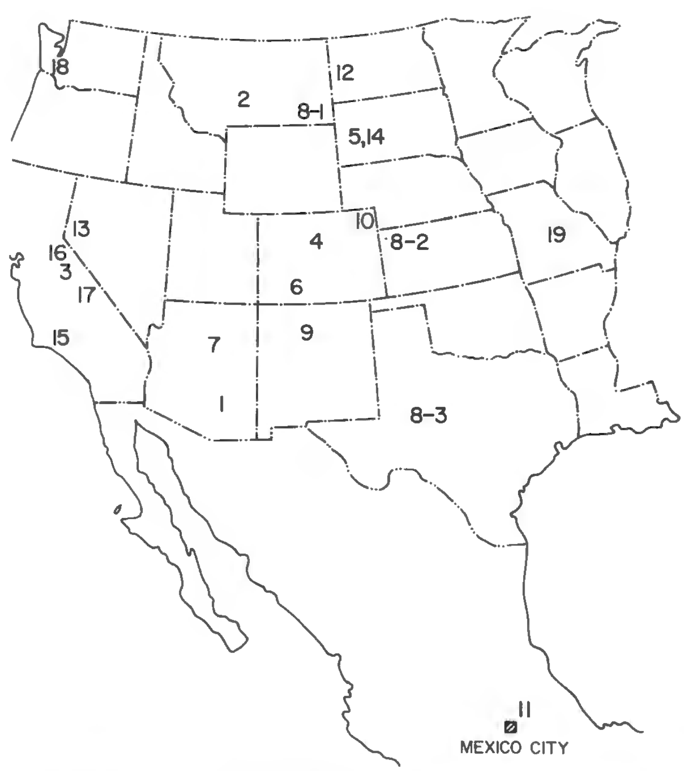


Fig. 7.3. **Map of western United States and northern Mexico showing locations of randomized cloud seeding experiments mentioned in text: (1) Arizona, (2) Bridger, (3) Central Sierra (CENSARE), (4) Climax, (5) Cloud Catcher, (6) Colorado River Basin Pilot Project (CRBPP), (7) Flagstaff, (8-1, 8-2, 8-3) High Plains (HIPLEX), (9) Jemez, (10) National Hail Research Experiment (NHRE), (11) Necaxa, (12) North Dakota Pilot Project (NDPP), (13) Pyramid Lake, (14) Rapid, (15) Santa Barbara, (16) Sierra Cooperative Pilot Project, (17) Sierra Cumulus, (18) Weather Bureau Artificial Cloud Nucleation (ACN) Project, (19) Whitetop.**

Two of the best known randomized projects were operated near Climax, Colorado (Fig. 7.3). The work there consisted of two experiments, Climax I and Climax II, encompassing about five winters each. For a time, atmospheric scientists around the world [e.g., Warner (1974) gave these experiments very high marks for credible, replicable evidence of changes in snowfall due to AgI seeding.

The Climax experiments have been described in numerous publications [e.g., Grant and Kahan (1974). The AgI generators were operated high on the western slopes of the Rocky Mountains to seed areas near the Continental Divide. An exploratory analysis of Climax I data indicated that the temperature at 50 kPa (500 mbars), which is a crude estimator of the cloud top temperature in winter storms in the central Colorado Rockies, is a useful basis for stratification (Grant and Mielke, 1967). The results indicated that AgI seeding from ground based generators probably increased snowfall in cases where the temperature at 50 kPa was −20°C or higher, and decreased snowfall when the 50 kPa temperature was −27°C or lower. Climax II tended to support the tentative conclusions of Climax I, as did some other work in the same geographic area [e.g., Grant *et al.* (1971)].

Two challenges have been offered to the Climax results within the past two or three years. The first challenge was raised by Neyman (1977) regarding the choice of days included in the analysis; it appears to have been dealt with satisfactorily by Mielke (1978). The other is more grave.

A search for large area effects downwind of Climax brought out the fact that much of Colorado showed positive precipitation anomalies on the Climax seed days, raising the possibility that the previous optimistic assessments constituted a gigantic type I statistical error. Mielke (1979b) has, therefore, attempted a more refined analysis using upwind controls. He has had some success for days with northwesterly winds, finding evidence of a seeding effect which is still positive, albeit smaller than previously estimated. A similar treatment of days with southwest winds appears impossible because of a lack of suitable control stations. As the southwest-flow days apparently produce more seeding opportunities in the Climax target area than do northwest-flow days, it is not possible to make accurate estimates at this time of the total effect of seeding on seasonal precipitation.

It must be noted too that the Colorado River Basin Pilot Project (CRBPP) in southwest Colorado, which used seeding criteria based on Climax, failed to produce any evidence of snowfall increases under its original design (Elliott *et al.*, 1978b). Elliott *et al.* (1978b) tentatively ascribe the lack of results to forecast errors and difficulties in implementing the experiment as designed.

The recent changes in thinking about Climax emphasize the desirability of basing conclusions about cloud seeding effectiveness on the total body of available evidence, rather than on whatever randomized projects enjoy the favorable attention of atmospheric physicists in a particular year. However, before attempting to state any conclusions about seeding orographic clouds, it is appropriate to broaden our view of what orographic storm systems are like and how they might be affected by glaciogenic seeding.

Orographic Clouds with Embedded Convection

The discussions so far (and the experiments described) have implicitly assumed rather stable orographic clouds, such as stratocumulus or altocumulus–altostratus decks, from which AgI seeds would extract additional snow by the Bergeron process. This picture is true sometimes, but orographic storms with abundant moisture also contain bands of convective instability. The convective bands are readily identified on radar and can also be detected by mesoscale analyses of pressure traces and of rainfall rates at recording gages. They are very prominent in storms entering the western United States from the Pacific Ocean. Graupel and small hail, rather than snow crystals grown from vapor, are typical precipitation forms in the convective bands.

Elliott (1962), analyzing results of the first randomized project in Santa Barbara County, California, found evidence that seeding the convective bands increased the rainfall from them substantially, say by 50–100%. Dennis and Kriege (1966) analyzed ten years of operational cloud seeding in Santa Clara County, which is near the central California coast (Fig. 6.1), and concluded that convectively unstable situations presented highly favorable opportunities for rainfall increases.

It is reasonable that convective bands in orographic storms near the West Coast would constitute favorable seeding opportunities. For one thing, the convective currents would transport AgI from ground based generators rapidly to the −5°C level. (The coastal ranges are not high enough to ensure such a response by orographic lifting alone, as the −5°C level is above the peaks in many cases.) Perhaps equally important, the bands represent a large supply of freshly condensed, supercooled water which can be converted to graupel or snow and may also yield appreciable dynamic effects when seeded.

A randomized experiment was run in Santa Barbara County from 1966 to 1968 to test these possibilities. Seeding was conducted with pyrotechnic generators from the ground and from aircraft. Results were checked against an area-of-effects model. The experiment indicated that precipitation from convective bands could be increased by 50–100% by seeding, thereby increasing storm totals by 25–50% (Brown *et al.*, 1976). Seeding rates which have been found effective in convective band situations range up to several hundred grams AgI per hour, somewhat higher than those applicable to stratiform orographic clouds.

While some authors have asserted that the rules quoted earlier about the influence of cloud top temperature apply in the band situations also, the present author considers that the bands might provide seeding opportunities with cloud top temperatures as low as −30 or −40°C. The updrafts in the bands (up to 10 m s^{-1}) protect the condensed cloud water from the ice crystals which would otherwise be falling from higher levels to provide natural seeding.

The idea that embedded convective cells present the best opportunity for seeding orographic cloud systems appears to have gained ground during the past two or three years. Examination of cloud physics data collected by aircraft on the CRBPP in southwest Colorado shows convective bands to be important there also. Data from the CRBPP are being reexamined to see if convective instability might not be a controlling factor on the seeding response.

Magnitude of Possible Increases

Only by combining results from many projects can one develop confidence in the reality of the seeding effects and in the validity of the physical models proposed to explain the effects. As the discussion in Section 4.3 showed, the expected effects of seeding in orographic situations would depend upon wind speed, cloud top temperature, the height of the mountain crest, and so on.

Vardiman *et al.* (1976) and Vardiman and Moore (1977, 1978) have compared the results of several randomized experiments on orographic cloud systems in a search for unifying concepts (Fig. 7.4). Their results (Table 7.2) indicate changes in precipitation at the mountain crest in these actual experiments ranging from decreases of 60% to increases of as much as 60%. The projects have been studied in terms of available cloud water, time for precipitation formation, ice nuclei supply, and degree of atmospheric mixing to determine the "seeding windows," that is, those combinations of meteorological conditions in which mountain snowfalls can be increased. Vardiman and Moore's (1978) results support the idea that the changes in precipitation are mostly explainable in terms of microphysical effects, that is, of the formation of additional snowflakes around artificial

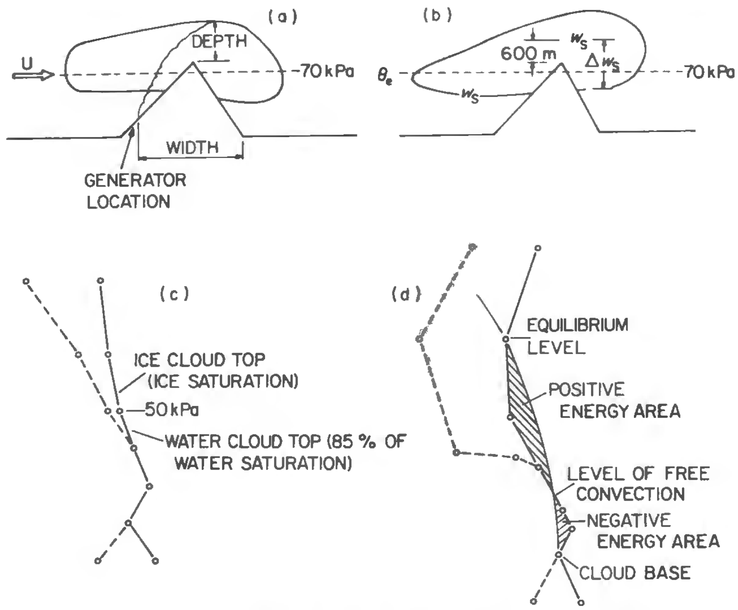


Fig. 7.4. **Vertical section through a mountain range showing some of the factors influencing response to seeding of orographic clouds. θ_e denotes equivalent potential temperature at 70 kPa; w_s denotes saturation mixing ratio. [After L. Vardiman *et al.* (1976). Papers presented at *WMO Scientific Conf. Weather Modification Boulder, 2nd,* p. 41, by permission of World Meteorological Organization and senior author.]**

TABLE 7.2

Seed/No-Seed Ratios on Randomized Orographic Projects as Function of Location and Storm Type[a]

	Region		
	Upwind	Crest	Downwind
		Stable storms	
Number cases	140	132	129
S/NS ratio	0.97	1.22	1.12
Significance level	0.11	0.02	0.10
		Unstable storms	
Number cases	194	190	187
S/NS ratio	1.38	1.57	1.58
Significance level	0.06	0.01	0.10
		Unstable "blow-over" storms	
Number cases	45	45	45
S/NS ratio	0.60	0.38	0.32
Significance level	0.002	0.001	0.002

[a] After Vardiman and Moore (1977). Projects are Bridger, Colorado River Basin (San Juan Mts.), Central Sierra, Jemez, and Pyramid Lake.

ice nuclei and their subsequent growth and fallout (e.g., Grant and Kahan, 1974).

Three of the important conclusions of Vardiman and Moore (1978) are as follows:

1. Stable orographic clouds with a crest trajectory, moderate water contents, and cloud top temperatures between −10 and −30°C showed an 18% increase in precipitation on the crest by seeding.
2. Moderately unstable orographic clouds with a crest trajectory, moderate to high water content, and cloud top temperatures between −10 and −30°C showed a 52% increase in precipitation on the crest by seeding.
3. Unstable orographic clouds with a blow-over trajectory, low water content, and cloud top temperature below −30°C showed a 54% decrease in precipitation on the crest by seeding.

As Vardiman and Moore (1978) note, these results give confidence that the seedability of winter orographic clouds depends on rather general meteorological conditions.

The net snowpack increase achievable over a winter season by seeding orographic clouds for microphysical effects depends on the frequency with which the various situations occur and how much each type of situation contributes to the total annual snowpack. An analysis by Elliott *et al.* (1978b) based on an after-the-fact analysis of the CRBPP results suggests that a flawless operational seeding program could increase winter precipitation in parts of the upper Colorado River Basin by 10–12%.

The possibility of dynamic effects on orographic cloud systems cannot be ruled out, and is under active investigation at Utah State University and elsewhere. Dynamic effects may be one way to get snowfall increases above the 15% upper limit that has characterized results in most orographic situations so far.

Snowfall Redistribution

The wide variety of results indicated in Vardiman and Moore's review emphasizes the need to study any proposed target area in detail, particularly with respect to the targeting problem. Directing snow to a predetermined target area is a special case of the redistribution question mentioned in Chapter IV.

There is no doubt that redistribution of precipitation has taken place on some projects, as well as absolute increases or decreases. (Note the "blow-over" situations in Table 7.2.) There is evidence in the CRBPP results for unstable, blow-over days of snowfall decreases over the crest in the target area and increases in the downwind valley some 50 km beyond (Elliott *et al.,* 1978b). This result could well be described as an apparent mesoscale redistribution of snowfall.

Investigators at the University of Washington have explored the possibility of shifting snowpack from the windward (western) slopes of mountain ranges in the Pacific Northwest to the eastern slopes [e.g., Hobbs (1975a,b); Hobbs and Radke (1975)]. Physical investigations there have documented changes in crystal forms and in the degree of riming of the crystals (Hobbs and Radke, 1975), but no full scale test of the concept has been attempted yet.

The Great Lakes Project of the National Oceanic and Atmospheric Administration (NOAA) considered using a similar technique to alleviate the crippling effects of heavy snowstorms upon Buffalo, New York, and other cities on the lee side of the Great Lakes (Weickmann, 1974). Preliminary tests indicated that the degree of riming and snowflakes sizes, and hence fall speeds, could be reduced. In principle, then, the heavy snowfalls along the lee shores of the lakes could be displaced inland at least 20 or 30 km and spread out. These experiments too stopped short of a full-scale demonstration.

Experiments in Australia

Australian scientists have pursued vigorously for 30 years the possibility of increasing orographic precipitation by seeding. Their results appeared confusing for a time.

The results in a series of randomized crossover experiments in the 1950s and early 1960s were promising and close to significance after two years. However, extension to four or five years showed a steadily decreasing ratio of seeded to unseeded rainfall and deteriorating p values (Smith, 1967, 1974).

The only hypothesis to explain this baffling result was advanced by Bowen (1966), who hypothesized a carry-over effect so that the distinction between seed and no-seed days became obscured after one or two years. If nothing else, this hypothesis reveals another possible defect in the crossover design. If there is any possibility of carry-over effects, a control area is needed which is *never* seeded.

Gabriel *et al.* (1967) analyzed data from the Israeli experiment, which also used the randomized crossover design, and found no evidence of carry-over effects there. This result does not disprove the existence of carry-over under the different conditions prevailing in Australia, but helps maintain the credibility of the results obtained in Israel and in randomized experiments elsewhere.

A new experiment in Tasmania which used control areas that were never seeded was operated on even numbered years from 1964 to 1970 (Smith *et al.*, 1971; Smith, 1974). It gave comparatively uniform results on each of the seeded years. There was evidence of rainfall increases of 15–20% during the autumn and winter seasons at acceptable p values, which agrees with earlier Australian results. There were no detectable increases during the summer season, which is also in accord with previous Australian results.

Summary of Techniques Effective in Orographic Situations

Various modeling results indicate the importance of seeding orographic clouds with glaciogenic agents which are effective at temperatures as close to 0°C as possible and which do not produce large quantities of ice crystals at temperatures of −20°C or lower. Seeding with ground based generators emitting 10–20 g AgI each per hour is an effective tool in many situations. As noted earlier, the generators should generally be placed at exposed sites some 20–50 km upwind of the target area, not more than 10 or 15 km apart transverse to the prevailing wind, and attention must be given to wind perturbations by the mountain range itself.

In some situations the only accurate method for targeting the seeding agent to produce precipitation on the lower parts of a windward slope is by aircraft. Flight paths parallel to the upwind edge of the target area are normally chosen, with low output seeding devices burned on the aircraft leaving a cylindrical plume at the flight altitude. One would naturally select a seeding track far enough upwind to provide time for the AgI particles or other ice nucleants to diffuse to the desired concentration and for the resultant snowflakes to grow and fall out on the target area. Careful study is needed to judge the optimum altitude for the seeding aircraft, keeping in mind that the ice nuclei will initially rise with the upward moving air but the resultant snowflakes will fall with increasing speeds as they grow. Numerical area-of-effect models which take these factors into account provide useful estimates of the appropriate seeding locations. Flying conditions during storms over or near mountains are often dangerous, so twin engine aircraft fully equipped with navigational aids and deicing equipment are normally used.

In some situations near high mountain peaks or with severe icing and turbulence, in-cloud flights near the usual levels of −5 to −15°C may be precluded. In such cases pyrotechnics or dry ice can be dropped from higher levels. This approach is costly if one wishes to seed continuously over long periods, which is the normal case for orographic cloud seeding programs dealing with stratiform clouds.

In convective band situations, generators on exposed sites can be very effective. They do not have to be as far upwind as for stratiform clouds. The use of dry ice or droppable pyrotechnics is also a reasonable approach, as the effect of a band at a given point is typically limited to less than two hours. Devices consuming small amounts of AgI (5–10 g each) are preferable to a smaller number of larger units. The rapid vertical spreading of the seeding agent in convective cells and the large terminal speeds of the resultant graupel particles are in agreement with observations on some projects of effects beginning at the ground only 10 minutes downwind of an aircraft seeding path.

7.5 INCREASING PRECIPITATION FROM CONVECTIVE CLOUD SYSTEMS[2]

The physical concepts which justify attempts to increase precipitation by seeding convective clouds have been explored in Chapter IV. They include the initiation of either liquid or solid artificial precipitation embryos by seeding with water spray, hygroscopic agents or ice forming agents, and the modification of cumulus dynamics, a process most commonly attempted through artificial glaciation of supercooled cloud water.

The extreme variability of precipitation from convective clouds in both space and time makes the evaluation of experiments and operations on convective clouds very difficult (Table 6.1). Target–control correlations are generally low, say 0.4–0.6, so operational programs on convective clouds usually do not yield clear-cut evidence of rainfall increases over conventional rain gage networks [e.g., Thom (1957b)]. We, therefore, shall rely principally upon the results of randomized experiments in discussing the possibilities for stimulating rainfall from convective clouds.

Selection of Experiments for Review

Experiments on convective clouds have yielded a wide range of results. Some scientists have been inclined to view the differences in apparent results as due to random variations and the total set of experiments as inconclusive.

[2] Section 7.5 draws heavily on material assembled by the author as a temporary employee of the National Oceanic and Atmospheric Administration (Dennis and Gagin, 1977).

The present author has reviewed the literature on randomized experiments on convective clouds, paying close attention to the types of clouds seeded, the seeding methods, and the limitations of the observational programs, and concluded that some of the differences in apparent results are real. Some projects have provided evidence of increases in precipitation due to seeding of convective clouds, while others have provided evidence of decreases. Still other experiments have been merely inconclusive due to weak statistical design, poor seeding methods, inadequate observing networks, or premature termination.

The results of eight selected projects dealing with isolated clouds or cloud groups and 12 projects involving all convective clouds over an area are shown in Tables 7.3 and 7.4, respectively. The following exclusions are noted:

1. Nonrandomized or partly randomized seeding trials, often with results quoted in terms of "percent of successes" under various conditions [e.g., Orr *et al.* (1950)].
2. Projects in which the randomization was severely compromised. For example, there was an area seeding experiment in Rhodesia in 1970–1971 in which working weeks were declared in advance to be seed or no-seed periods, and the analysis was limited to those days when clouds occurred that were considered on the basis of size and appearance to be suitable for seeding (McNaughton, 1973b). That project does not appear in Table 7.4. There are some questions about the randomization of the Flagstaff project, which paired clouds, and the Arizona project, which paired days. In both cases some test cases were chosen with advance knowledge on whether they would be seed or no-seed cases. Both projects are included because the experimenters evidently tried to avoid biases and because there are no other data available on randomized seeding of convective clouds over the arid southwestern United States.
3. Projects in which equipment was operated on a randomized basis, but in a manner very different from usual cloud seeding practices and with no physical explanation for the apparent effects revealed by statistical analysis. This category includes a famous experiment in India in which powdered salt was released in small quantities from blowers on the ground and in which the rainfall on seed days exceeded that on no-seed days over a large area (Biswas *et al.,* 1976; Panel, 1973, pp. 48–49).
4. Projects in which the principal objective was to test hypotheses about hail suppression or lightning suppression rather than the stimulation of rainfall.
5. Experiments on convective clouds embedded in orographic clouds or in widespread stratiform clouds associated with large storms.
6. Projects with strong evidence of large, uncontrolled natural differ-

TABLE 7.3

Summary of Results of Randomized Experiments on Isolated Cumulus Clouds or Cloud Clusters

Project	Dates	Main seeding agent(s) and delivery techniques	Results	p-value	References for results quoted
Central U.S.	1954	Dry ice, 5 or 15 kg/km, on penetrations above 0°C level	Greater probability of radar echoes	0.39	Braham *et al.* (1957, p. 73)
Caribbean	1954	Waterspray (1 m³/km) into clouds	(a) Greater frequency of radar echoes	0.02	Braham *et al.* (1957, p. 69)
			(b) Time to first echo reduced	0.01	Braham *et al.* (1957, p. 71)
Australia	1962–1965	AgI–NaI–acetone burners in updraft, 0.2 or 20 g per cloud	More rain at cloud base if tops colder than −10°C and 20 g treatment applied	0.02	Bethwaite *et al.* (1966); Smith (1967, pp. 161–162)
Sierra Cumulus (pair seeding)	1966–1968	Dry ice, AgI flares in updrafts	Increased probability of precipitation reaching ground	0.001	Panel (1973, p. 71)
Flagstaff, Az	1967	AgI–NaI(?)–acetone burners in updrafts 120 or 240 g per cloud	(a) Increase in height of radar tops	0.04	Weinstein and MacCready (1969)
			(b) Increase of shower duration by 10 min	0.08	
			(c) Increase in radar estimated rainfall	0.19	
Rhodesia	1968–1969	AgI–NaI solution burned in updrafts *in cloud* 4 km above sea level (∼−6°C level) 0.5 kg hr⁻¹	(a) Increase rainfall at cloud base	<0.01	McNaughton (1973a)
			(b) Increase shower duration	<0.01	
Florida	1968, 1970	AgI: about 20 50 g pyrotechnics per cloud from cloud top	(a) Increase in cloud height	0.01	Panel (1973, pp. 90–92)
			(b) Increase in radar estimated rainfall	0.005	
Cloud Catcher	1969–1970	Salt, ∼50 kg, or 1–6 120 g AgI flares in updrafts below cloud base (random choice, ⅓ cases no seed)	(a) Radar echoes closer to cloud base	0.01	Dennis and Koscielski (1972)
			(b) Increase in echo tops for AgI seed cases	0.10	Dennis *et al.* (1975a)
			(c) Increases in radar estimated rainfall	0.06 (salt) 0.01 (AgI)	
Rhodesia	1973–1975	AgI pyrotechnics fired in updrafts ∼250 m below cloud *top*	(a) No effect if top temperature > −10°C	—	McNaughton (1977)
			(b) Rainfall increase if top temperature < − 10°C	0.07[a]	
Ukraine	1973–	$AgI–NH_4I$ in acetone, 500 g hr⁻¹ in updrafts	Radar data through 1975 indicate rainfall increased over amounts predicted by regression equation	0.05	Buikov *et al.* (1976)

[a] Best result, found by measuring rain beginning 50 min after seeding.

TABLE 7.4

Effects of Seeding Convective Clouds with AgI on Rainfall over a Fixed Area as Deduced from Randomized Experiments

Project	Dates	Main seeding agent(s) and delivery technique	Apparent net results		p-value		References for results quoted
Necaxa, Mexico	1956–1968	Broadcast from generators on ground. Electric arc and string generators, ~5 g hr^{-1} each	Rainfall redistribution, (more in target, less in control)		0.001		Perez Siliceo (1970); Panel (1973, pp. 76–78)
Arizona I	1957–1960	Broadcast near −6°C level upwind, AgI–NaI–acetone generator, approximately 1 kg hr^{-1}	30% decrease		0.30		Battan and Kassander (1967, pp. 29–33)
Arizona II	1961, 1962, 1964	Broadcast near cloud base upwind, AgI–NaI–acetone generator, approximately 1 kg hr^{-1}	30% decrease		0.16		Battan and Kassander (1967, pp. 29–33)
Whitetop	1960–1964	Broadcast near cloud base level upwind, AgI–NaI–acetone generator, 2700 g hr^{-1}	Substantial decreases		<0.01		Decker and Schickedanz (1967, p. 62) (Missouri plume, seed versus no-seed)
Israel I	1961–1967	Broadcast just below cloud base in upwind area, acetone generator	15% increase in seasonal precipitation		0.02		Gagin and Neuman (1974, p. 465)
Israel II	1969–1975	Broadcast just below cloud base in upwind area, acetone generator	13% increase in seasonal precipitation		0.02		Gagin and Neuman (1976)
Climax III	1966–1969	Broadcast from ground, four AgI–NaI–acetone generators, 15 g hr^{-1} each	Possible decrease		0.26		Grant *et al.* (1974)
Climax IV	1970, 1972	Broadcast from ground, four AgI–NaI–acetone generators, 15 g hr^{-1} each	Possible decrease		0.32		Grant *et al.* (1974)
Rapid	1966–1968	Updraft seeding, AgI–NaI–acetone generators, 0.5–1 kg hr^{-1}	Number of days with $\overline{R_s} > \overline{R_{ns}}$ exceeds number expected by chance		*SW Flow*	*NW Flow*	
				Shower days	0.01	0.09	Chang (1976)
				Storm days	0.35	0.61	Chang (1976)
				All days	0.02	0.17	Chang (1976)
NDPP I	1969–1970	Updraft seeding, AgI–NaI–acetone generators, 300–600 g hr^{-1}	Increase of 2%		0.67		Dennis *et al.* (1975b)
NDPP II	1971–1972	Updraft seeding, AgI–NH_4I–acetone generators, 300–600 g hr^{-1}	Increase of ~70%		0.12		Dennis *et al.* (1975b)
FACE I	1970–1975 (intermittent)	Injection of pyrotechnics into cloud towers, ~1 kg per cloud	Decrease of 6%		0.40		Woodley *et al.* (1976)

ences between the seed and no-seed samples despite randomization. For example, Gelhaus *et al.* (1974) found evidence of natural differences on a randomized project in South Dakota which apparently resulted in a type I statistical error. The famous Whitetop Project in Missouri showed some evidence of possible type I errors (Lovasich *et al.,* 1971), but that view is disputed by the principal investigators. Whitetop is included in Table 7.4.

7. Experiments which have not been adequately analyzed for possible type I errors. This is especially important for experiments of short duration evaluated solely in terms of rainfall at the ground and using the single area design. Examples include some experiments conducted in east Africa during the 1950s [e.g., Davies (1954)].

The *p* values quoted in Tables 7.3 and 7.4 are those most representative of results for the projects according to the original design. The tabulated results do not nearly exhaust the data from the projects in question, but the complete reports from which the information was abstracted are taken into account in the following discussion.

Effects of Seeding on Isolated Clouds

It is generally agreed that seeding with water spray or hygroscopic agents sometimes hastens the formation of rain by coalescence, and that glaciogenic seeding changes supercooled water to ice crystals, which may subsequently grow into graupel or snowflakes. The graupel and snowflakes often melt during their fall to the ground. Whether these changes lead to net increase in rainfall from a cloud depends upon whether and how efficiently the seeded cloud would have rained naturally (perhaps a short time later).

Table 7.3 suggests that the most favorable situations for causing precipitation increases through microphysical effects alone are found in the largest convective clouds that do not precipitate efficiently by either the coalescence or the Bergeron process, but which respond to seeding treatments. The best candidates appear to be convective clouds with rather cool bases (<10°C) and cloud top temperatures of −10 to −30°C. The most definitive experimental results have been obtained by experimenters who selected target clouds carefully and seeded them on an individual basis with either dry ice or AgI. The amount of AgI is not critical, provided a threshold of 5 or 10 g per cloud is exceeded. The Australian experiments showed detectable effects with an expenditure of 20 g AgI per cloud but no detectable effects at 0.2 g per cloud (Table 7.3).

Seeding convective clouds, particularly glaciogenic seeding, always involves dynamic as well as microphysical effects, which suggests the possibility of large precipitation increases. The dynamic effect most commonly sought has been an increase in cloud top height. We have already provided an example (Fig. 6.6) of a randomized experiment providing evidence that AgI seeding actually does make some cumulus clouds grow taller than they would otherwise (Simpson *et al.*, 1967). Increases in cloud height have also been sought and tentatively identified with the aid of cloud models on randomized experiments at Flagstaff, Arizona, in Florida, and in South Dakota (Table 7.3).

There is a widespread impression that "massive seeding" is required to produce dynamic effects (Panel, 1973, p. 90; WMO Statement, 1976), but there is ample evidence to the contrary. For one thing, it should be noted again that the Alecto units used in the dynamic seeding experiments on Stormfury were grossly inefficient. Proper seeding devices could have achieved the same results with only a few percent of the amount of AgI expended. At Flagstaff apparent dynamic effects followed seeding with as little as 120 g AgI per cloud (Weinstein and MacCready, 1969). Later, Dennis *et al.* (1975a) reported tentative evidence of increases in the maximum height of radar echoes from convective cloud complexes averaging 600 m due to seeding with from 120 to 720 g of AgI from flares burned in updrafts below cloud base. "Explosive" growth following seeding with dry ice was observed in nonrandomized trials in Australia by Kraus and Squires (1947) and in 1963 in Pennsylvania (Davis and Hosler, 1967).

Recently St.-Amand and Elliott (1972) and Dennis *et al.* (1976) have called attention to increases in updraft speed as another predicted and potentially important dynamic consequence of AgI seeding. Other authors have spoken of downdrafts induced by artificially induced precipitation as a possible link between seeding and the cloud dynamics. This possibility has not been studied extensively, perhaps because it is not handled well in one-dimensional cloud models.

Increases in cloud height or in the speeds of updrafts and downdrafts are not the same thing as increases in rainfall. The important question for precipitation augmentation is whether or not they are followed by increased moisture fluxes through cloud base. Unless that happens, there is little chance for net precipitation increases.

While there are reports of cloud towers rising and breaking away from the main cloud mass following glaciogenic seeding from aircraft, the general tenor of reports is that seeding leads to consolidation and organization of existing clouds and showers. This could affect the subcloud layer, and also protects the interior parts of the showers from mixing with dry ambient air. The data in Table 7.3 suggest increases in shower duration, horizontal extent of radar echoes, and in total rainfall at cloud base as consequences of seeding isolated clouds for dynamic effects, and the *p* values for some experiments are satisfactory. There is no serious doubt about the existence of convective clouds which respond to artificial glaciation with increases in size and increases in precipitation.

Effects on Area Rainfall

Increases in rainfall due to seeding of single convective clouds do not necessarily represent net rainfall increases. There is always a chance that the increases from one cloud are compensated by decreases in the rainfall from other clouds. Therefore, one must consider the total rainfall on fixed target areas during experimental periods to evaluate net effects.

Attempts to modify area rainfall by hygroscopic seeding go back to at least 1952 in east Africa (Alusa, 1974) and to 1954 in Pakistan (Fournier d'Albe, 1957). There has been an extensive operational program in Thailand since about 1972. Nevertheless, most scientists are skeptical about the results, partly because of the tremendous logistical problem involved in delivering enough hygroscopic particles of, say, 10^{-9} g each to affect rain formation processes over any sizable areas. The statistically significant results of Biswas *et al.* (1967) from randomized projects in India were not generally accepted because of the lack of a credible physical model relating the quantity of salt released to the apparent rainfall increases (Panel, 1973, p. 49; Warner, 1974). Later projects in India have not attained statistical significance (Kapoor *et al.,* 1976) and so have not resolved the uncertainties.

Most of the documented area experiments on convective clouds have involved glaciogenic seeding. Although dry ice is a promising ice forming agent, most area experimenters to date have chosen AgI generators. All of the 12 projects shown in Table 7.4 involve AgI seeding. Because of the wide variety of seeding methods employed and the diversity of results, we shall discuss some of the experiments briefly on an individual basis before attempting a synthesis.

Mexico. The most prolonged randomized seeding project in the world is the one started by the Mexican Light and Power Company in the Necaxa watershed in 1956, following several years of nonrandomized seeding in the same area. Ground based generators consuming 5 g hr^{-1} of AgI each were used and the generator network was kept in the same general location through 1968 (Perez Siliceo, 1967, 1970; Panel, 1973, pp. 76–78). Some of the generators vaporized AgI powder in an electric arc, which presumably produced pure AgI particles.

The analysis of the randomized trails by Perez Siliceo (1970) showed evidence of a rainfall suppression effect in the control area on seed days and of carryover effects in the target area on no-seed days. The results varied with the synoptic situation. Perez Siliceo speaks of rainfall redistributions over the 13 years of randomized experiments as having a significance level of 0.001. The Necaxa results show conclusively that AgI seeding from the ground sometimes affects rainfall from convective clouds over mountains and that the effects can extend to control areas, even when they are upwind from target areas.

Switzerland. An interesting result not included in Table 7.4 is that of Grossversuch III, operated near the southern Alps as a hail suppression trial. AgI generators consuming 20 g hr^{-1} were operated on the ground in and around the target. There was evidence of rainfall increases on days when low level inversions trapped the AgI particles near the ground, until convective activity erupted and presumably carried them up to supercooled cloud regions (Neyman and Scott, 1967b).

Israel. Of all the experiments listed in Table 7.4, only the Israeli and Mexican experiments have given clear-cut, statistically significant evidence of net rainfall increases over a period of several years in predesignated target areas (Fig. 7.5). In the Israeli work, AgI crystals were released from an aircraft flying a track some 50 km upwind of the target area at 1500 m above sea level. Cloud bases were typically near + 10°C and near 2 km. The favorable response to broadcast AgI seeding, with increases in seasonal precipitation averaging 13–15%, probably occurs because the clouds treated (winter cumuli of moderate size with high droplet concentrations) form precipitation through the Bergeron process, but are

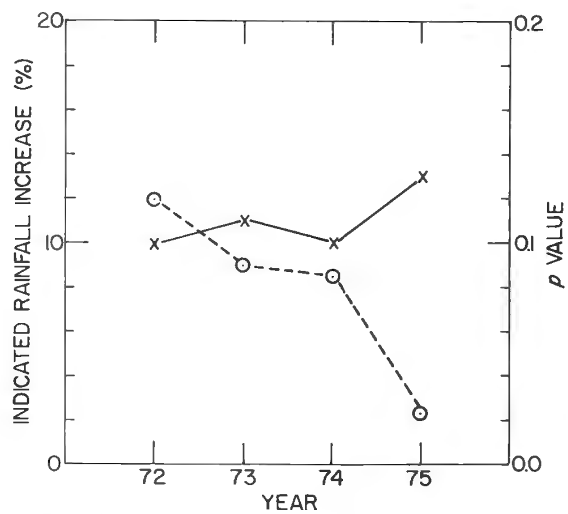


Fig. 7.5. Cumulative results of Israel II experiment at end of each of last four years. Indicated rainfall increase (——— × ———) in north experimental area determined by comparison with control area using a double ratio method. *p* values (– – –) are from a Monte Carlo (rerandomization) test.

often deficient in natural ice (Gagin and Neuman, 1974, 1976). The fact that the clouds form or intensify over orographic barriers may contribute to the stability of the rainfall statistics and therefore to the statistical significance of the results. The Israeli program has now been changed to an operational basis.

Arizona. The University of Arizona experiments (Arizona I and II) grew out of the observation that many supercooled convective clouds over the southwestern United States fail to precipitate [e.g., Battan and Braham, 1956]. This observation suggested a role for artificial ice nuclei, which were broadcast upwind from an airplane. However, rainfall per seed day between the hours of 1300 and 1800 MST was less than the rainfall per no-seed day. Battan (1966) and Battan and Kassander (1967) considered the results inconclusive.

Analyses by Neyman *et al.* (1972) provided additional evidence, which the present author finds convincing, that there was a net *negative* effect of seeding upon rainfall in the Arizona experiments. The ratio of seed day rainfall to no-seed day rainfall is consistently less than one for both Arizona I and II. It decreases for the combined data from 0.81 to 0.55 to 0.45 as one goes from low to medium to high cloud bases as estimated on the basis of 85 kPa temperature-dew point spreads. The decline in the ratio is consistent with the hypothesis that seeding produced smaller raindrops and increased evaporative losses below cloud base, that is, a hypothesis of a form of overseeding.

Missouri (*Whitetop*). The Whitetop Project also involved broadcast seeding of AgI from acetone generators on aircraft upwind of the target area (Table 7.4).

Whitetop is perhaps the most extensively analyzed cloud seeding project ever conducted in the United States. However, many of the papers published about it are simply reiterations of previously stated positions. The latest, by Braham (1979) with following commentaries by several authorities, can be used as a guide to the literature on Whitetop and to the main points of view concerning its disappointing results.

At the risk of oversimplification, we note that the principal investigator and his statistical advisers have produced evidence of rainfall increases on certain days, and of substantial rainfall decreases on other days. In particular, comparisons of conditions inside the seeded plumes with nonplume areas indicate that the seeding may have increased rainfall on days with westerly winds and cloud tops below 13 km, but caused substantial decreases on days with southerly winds and/or cloud tops above 13 km. There is evidence that radar echoes were more numerous immediately downwind of the seeding line (Panel, 1966, Vol. II, p. 19), but the total echoing area over a period of several hours after seeding was much less on seed days than on no-seed days (Braham *et al.*, 1971).

The other main school of thought about Whitetop is exemplified by the work of Neyman and his associates. At one time they viewed Whitetop as providing evidence of rainfall decreases over areas extending out 200 km or so in all directions. Later they developed serious reservations about the validity of the apparent results, and considered the possibility of type I errors (Lovasich *et al.*, 1971). Their reservations have not been removed so far (Neyman, 1979).

Discussion. It is clear that neither the Arizona or Whitetop experiments produced demonstrable rainfall increases, and may have produced net decreases. Explanations for the lack of rainfall increases on Arizona I and II and on Whitetop have ranged from complete lack of seeding effectiveness to overseeding. The former alternative is ruled out by the apparent changes in radar echoes (Battan, 1967) and the observed differences of either sign in rainfall on seed and no-seed days.

Cloud physics data collected on Whitetop showed rain forming by coalescence and considerable natural glaciation near $-10°C$ in clouds containing large droplets (Braham, 1964). This situation is cited sometimes as a reason for the lack of positive effects, but one is then at a loss to explain the apparent success of the Florida single cloud experiments (Table 7.3).

The present author sees possible reasons for the apparent rainfall decreases in Arizona I and II and Whitetop in the seeding generators and the seeding method. The AgI consumption rates were quite generous (Table 7.4). The AgI–NaI solution in the generators must have produced large particles with a hygroscopic component. Presumably, the particles were wetted during the ascent from cloud base and ineffective as ice nuclei at temperatures above $-12°C$, but quite effective at -20 to $-25°C$ (Chapter V). In this connection Battan's (1967) observation that radar echoes were more likely in seeded than in unseeded clouds in Arizona if the cloud top temperature was between $-18°C$ and $-42°C$, but not otherwise, is important. We suspect an overseeded condition in the cloud tops without any chance of compensating microphysical or dynamic effects at temperatures just below 0°C.

Most of the AgI on Arizona and Whitetop must have gone into the largest, mature storms as the seeded air parcels were entrained into them. This runs contrary to the results of single cloud experiments, which suggest best results from clouds of moderate size. The evidence of rainfall increases on certain days on Whitetop when *all* of the clouds were of small or medium size is tantalizing.

Colorado. Another broadcast seeding experiment with inconclusive results was started near Climax, Colorado, in 1966 (Grant *et al.*, 1972, 1974). Possible reasons for failure of the seeding to produce any significant effects upon rainfall (Table 7.4) include the use of AgI–NaI solutions in the generators, the small quantities used, and the weak statistical design of the project.

South Dakota. The Rapid Project in western South Dakota was started in 1966 with a randomized crossover design after some preliminary experiments in previous years in the same general area. Two pairs of target areas were set up to correspond to the prevailing directions of shower motions. The project utilized as the main seeding agent AgI delivered from airborne acetone generators into updrafts below cloud base. The generators were charged with AgI–NaI solution and the AgI consumption rate was 300 g hr^{-1}. This method was supplemented with dry ice at cloud tops and AgI from isopropylamine generators on the ground. A summary description of the project is given in Panel (1973, pp. 248–249).

Cloud physics investigations by aircraft on the Rapid Project showed cross-target contamination by AgI to be rare, the updrafts of unseeded cumulus clouds to be mostly ice free at -5 to -10°C, and more abundant ice crystals and snowflakes in the updrafts of seeded clouds near -10°C [e.g., Dennis and Miller (1977)]. It was the first randomized project in the United States to give indications of rainfall increases over fixed target areas by seeding convective clouds on a *prespecified* class of days. On shower days, as defined objectively in the work plans, rainfall in the north target areas was heavier on north seed days than on south seed days, and rainfall in the south target areas was heavier on south seed than on north seed days. On storm days results were mixed (Dennis and Koscielski, 1969; Simpson and Dennis, 1974, p. 250ff.).

Chang (1976) later computed $\overline{R}_s$ and $\overline{R}_{ns}$, the average rainfall by days in the seed and no-seed target areas, respectively, and used the permutation method to test the hypothesis that the number of days with $\overline{R}_s$ greater than $\overline{R}_{ns}$ was not affected by the random allocation of seeding treatments. Chang's analysis actually conforms better to the original statement of the seeding hypothesis than does the analysis by Dennis and Koscielski (1969), so his results rather than theirs are included in Table 7.4. [There is no conflict between the two sets of results.] Chang found the seed target area to be favored with rain vis-a-vis the no-seed target area on shower days and also showed a favorable result (p value = 0.02) for all southwest flow days as a class (Table 7.4).

North Dakota. Two projects specifically designed to test the impact of seeding for dynamic effects upon area rainfall were the Florida Area

Cumulus Experiment (FACE) and the North Dakota Pilot Project (NDPP) in its second phase (NDPP II).

The NDPP relied entirely on aircraft seeding from below cloud base in attempts to augment precipitation and suppress hail. During 1969–1970 seeding was done in new updraft areas beneath developing clouds with generators charged with AgI–NaI solution. There was no evidence of effects of seeding on average rainfall. The second phase of the project in 1971–1972 (NDPP II) involved seeding with acetone generators charged with $AgI{-}NH_4I$ solution. It yielded tentative evidence of large rainfall increases, but the results did not attain statistical significance at the 10% level (Table 7.4).

Statistical significance was obtained on a post hoc basis for days when a cloud model analysis of on-site rawinsonde data predicted increases in cloud height due to seeding. On those days precipitation per seed day exceeded that per no-seed day by a factor or more than three, with the apparent increase being attributable in part to increases in the frequency of rainfall events at individual gages (p value $= 0.04$) and in part to increases in the magnitude of rainfall events (p value $= 0.02$) (Dennis *et al.*, 1975b). A number of investigations into possible natural causes for the differences between seed and no-seed days (Fig. 7.6) have failed so far to come up with any explanation other than seeding to explain the differences.

The statistical task force of the Weather Modification Advisory Board (1978) pointed out that the relatively large number of stratifications and statistical tests used in the NDPP analysis involves the multiplicity pitfall (Chapter VI), so further experiments would be in order before the results are accepted as final.

Florida. The Florida Area Cumulus Experiment (FACE) was based on the success of single cloud experiments in the same region in 1968 and 1970. FACE began in 1971 and lasted intermittently through 1976. Operational days were limited to nondisturbed days when a one-dimensional cloud model predicted a dynamic response to AgI seeding. Silver iodide was delivered to growing clouds near the −10°C level on an individual basis by dropping 50 g pyrotechnics, often as many as 20 per cloud.

The FACE seeders consciously pursued the objective of stimulating cloud dynamics to promote inflow below cloud base and the artificial promotion of cloud mergers (Chapter IV).

Rainfall over the target area was estimated on the basis of radar data calibrated by a number of recording rain gages. The statistical methods used for analyzing the results varied somewhat from year to year, so FACE I should be viewed as an exploratory rather than a confirmatory

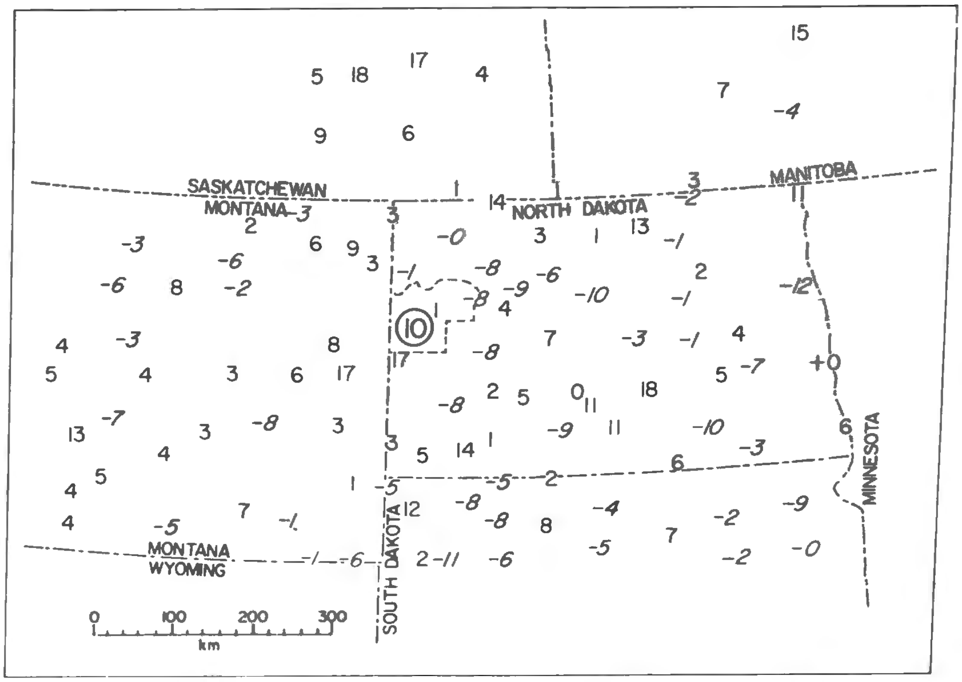


Fig. 7.6. Results of a "large-area" analysis in a search for uncontrolled background variations in rainfall on the North Dakota Pilot Project. Numbers are Wilcoxon test statistics (plotted in tenths) for individual rain gages comparing rainfall per seed day against that per no-seed day for the "dynamic seedability" stratification. The average regional value is 0.16 with a standard deviation of 0.70. The boldface number **10** in the target area (dashed boundary) is the average of the test statistics at 67 rain gages operated there. The positive rainfall anomaly on seed days in and just upwind of the target area does not appear to be part of any large-scale pattern. [After A. S. Dennis *et al.* (1975b). *J. Appl. Meteorol.* **14,** 959, by permission of American Meteorological Society.]

experiment. In any case, net results through 1975 for the total target area were inconclusive (Woodley *et al.,* 1976; Biondini *et al.,* 1977). There was a possible increase in rainfall over the target area on days with moving echoes (marching days), but no evidence of a net effect on days with stationary echoes.

In August 1975 the FACE seeders switched to pyrotechnics patterned after the TB-1 developed at the U.S. Naval Weapons Center, after wind tunnel tests showed the pyrotechnics previously used produced few ice nuclei active near $-5°C$ (Sax *et al.,* 1977a). Difficulties in monitoring the changes in ice concentrations due to seeding have been noted by Sax (1976), but Sax *et al.* (1977b) have presented evidence that glaciation of seeded clouds was more pronounced in 1976 than in previous years. The radar data for late 1975 and 1976 also showed more rain per seed day than per no-seed day (Woodley *et al.,* 1977). Unfortunately, there is strong evidence that this condition covered much of south Florida and not just the target area, so judgment on this second phase of FACE must be reserved (Nickerson and Brier, 1979).

Summary of Techniques Showing Apparent Effectiveness

Additional research may answer some of the questions still outstanding regarding the effects of seeding upon precipitation from convective clouds. The state of the art is summarized here for the convenience of the reader, even though modifications may well be required in future years. This summary is based principally upon the results of the randomized experiments noted above, but evaluations of operational programs [e.g., Pellett *et al.* (1977)], modeling results, and discussions with meteorologists and cloud seeding pilots in several countries have also played a part in shaping the conclusions presented.

Seeding very deep convective clouds for precipitation augmentation has not shown evidence of success, and is no longer generally practiced. The most favorable targets are clouds with top temperatures in the range of -10 to -30°C. The best formula that can be offered at this time is to *seed growing clouds* (Fig. 7.7), including new towers around existing showers, with glaciogenic agents while their tops are in this temperature range, *and to leave mature clouds alone*. Through some combination of microphysical and dynamic effects, as yet unspecified precisely, this approach leads in certain marginal situations to the conversion of cumulus congestus clouds into cumulonimbus yielding substantial showers.

Necaxa and Grossversuch III both showed that generators on the ground can affect convective rainfall over and near mountains. There can be little doubt that generators on flat terrain sometimes seed clouds effectively also, but the targeting problem is acute.

Broadcast seeding from aircraft is appropriate in dealing with numerous moderate cumulus or bands of cumulus, as in the Israeli winter situation, but has not been shown to be effective on large convective clouds. Optimizing results on more intense convection requires more specific targeting on the smaller but growing cloud towers, and a fairly rapid seeding response. Updraft seeding is effective in many situations. It may require guidance from controllers on the ground using weather radar sets to overcome visibility limitations.

Seeding from above with dry ice or AgI pyrotechnics may be preferable from an operational standpoint, but the data of Tables 7.3 and 7.4 do not show any better results for on-top seeding than for updraft seeding. The need for dispersed sources of ice particles is acute in the direct injection case. As noted in Chapter V, seeding with ten pyrotechnics is suit-

Fig. 7.7. Supercooled, growing, cumulus towers considered suitable for glaciogenic seeding by experienced cloud seeders. These towers grew on the flank of an older cloud out of view on the right [photo by Atmospherics Incorporated, Fresno, California].

able for a cumulus tower of moderate size, while up to 50 might be used in seeding a vigorously growing thunderstorm.

It appears prudent to avoid possible overseeding of the cloud tops. We have already alluded in Chapter V to the need for AgI generators which produce particles effective near −5°C, but which do not produce numerous small particles effective at −20°C and below.

The suggestions outlined should give maximum chances for demonstrable success. Nevertheless, decisions to proceed with operational programs to increase precipitation from convective clouds are calculated risks. The programs are not as dependable as those set up to increase precipitation from orographic cloud systems. Each one must be developed on the basis of local cloud climatology, and should be viewed in its beginning stages as an experiment. The more successful experiments can be expected to evolve into operational programs having significant impacts upon rainfall in the designated target areas, as has been the case in Israel.

7.6 LARGE AREA EFFECTS

We noted in Chapter IV a number of microphysical and dynamic mechanisms by which cloud seeding conceivably could lead to changes in clouds and precipitation at substantial distances from the site of release of the seeding agents.

The existence of conceptual models for the modification of weather at a substantial distance from the site of weather modification projects has led a number of investigators to study rainfall patterns outside target areas of weather modification projects in a search for effects at a distance. At one time, these studies concentrated on areas downwind of the target and persons spoke of downwind effects. With the increasing realization that dynamic effects could be as important as such microphysical effects as the transport of seeding agent or of ice cyrstals produced by seeding, the search for effects was extended in other directions. Today one commonly speaks of large area effects rather than downwind effects.

In 1948 Langmuir postulated that seeding with AgI generators might affect weather conditions thousands of kilometers away. Subsequently, he operated experiments in which AgI generators were operated every seventh day in New Mexico and effects were sought in rainfall patterns over the entire eastern United States (Langmuir, 1950, 1953). Although a seven day periodicity appeared for a time in rainfall patterns over the Ohio Valley, most meteorologists refused to accept the hypothesis that the periodicity was imposed by the seven day cycle in the operation of the generators in New Mexico.

A number of authors have published papers reporting evidence of seeding effects out to 100–200 km from primary target areas [e.g., Adderley (1968); Neyman *et al.* (1973)]. Most if not all of these analyses suffer from the defect that no hypothesis was stated in advance to predict where the effects would be observed. If one searches for effects over a large enough area, one is almost certain to find some rainfall or hailfall data suggestive of an effect from the seeding project under investigation. [This is another example of the multiplicity trap.] Therefore, the present author is hesitant to accept the statistical evidence that has been presented as anything other than suggestive and a guide to formulation of specific hypotheses for future experiments.

As a possible exception to the above rule, we note a randomized experiment in California set up for the specific purpose of studying the effects of cloud seeding at relatively long distances (Brown *et al.*, 1976). The data from that experiment suggest that rainfall increases in a target area can also be accompanied by increased rainfall downwind and to the

TABLE 7.5

Evidence of Large Area Effects Gleaned from Field Experiments

Project	Suggested effects on rainfall inside target	Suggested effects on rainfall outside target	Reference
Grossversuch III	Increase	Increase	Neyman and Scott (1974)
Victoria, Australia	Increase	Increase	Adderley (1968)
Arizona I and II	Decrease	Decrease	Neyman and Osborn (1971)
Israel	Increase	Increase	Brier *et al.* (1974)

right of the mean wind vector through the target. Brown *et al.* attribute the increases to dynamic stimulation of convective bands in winter storms and suggest that convective bands can retain their artificially stimulated intensity for a period of several hours following seeding. This suggested effect is directly contrary to the common perception that a rainfall increase in a target area should be accompanied by a rainfall suppression effect in downwind regions.

While keeping the cautions mentioned above in mind, it is still of some interest to tabulate the evidence for large area effects that has been produced so far. This is done in Table 7.5, which is drawn from a number of sources. The evidence in Table 7.5 suggests that effects of seeding do occur at large distances on occasion and that they can include both increases and decreases in rainfall.

There has been noted a general tendency for cloud seeding projects which have shown evidence of precipitation increases in the target area to show increases outside that target area, and similarly for projects with evidence of rainfall decreases (Table 7.5). For example, Brier *et al.* (1974) reported evidence of rainfall increases downwind of the Israel cloud seeding project, while Neyman and Osborn (1971) reported evidence of rainfall suppression downwind and to the right of the downwind vector in the case of the Arizona I and Arizona II experiments.

7.7 INCREASING PRECIPITATION FROM SYNOPTIC SCALE CLOUD SYSTEMS

Despite the evidence accumulated so far for large-area effects of seeding upon precipitation, it is not possible to state with any assurance whether the total precipitation from an extensive storm system such as a frontal cyclone of the eastern United States could be increased by seeding. Calculations of the precipitation efficiency of the extensive cloud decks associated with such storms suggest high precipitation efficiencies. If this is so, the only way to increase total precipitation from them would be through dynamic effects.

Two ambitious randomized experiments were conducted during the 1950s on synoptic scale systems. One of them was Project SCUD, an experiment to determine if cloud seeding could affect the rate of deepening of developing frontal cyclones near the east coast of the United States (Spar, 1957; Wells and Wells, 1967). The other was located in the Pacific northwest and was called the Weather Bureau Artificial Cloud Nucleation (ACN) Project (Hall, 1957; Neyman and Scott, 1967c).

Project SCUD provided no evidence of dynamic effects due to seeding nor of changes in precipitation. It only ran for two years, which is approximately the period estimated by Neyman and Scott as needed to detect a *40%* rainfall increase with the aid of predictors for that project (Table 6.1).

The ACN Project involved dry ice drops into extensive supercooled cloud systems. Three sets of variable but objectively defined target areas were established, called targets I, targets II, and targets III (Hall, 1957). The analysis for the U.S. Weather Bureau rated the ACN Project as inconclusive, and Hall's report emphasized the long period of time that would have been needed to establish definite results. (The project only ran for two winters.) However, Neyman and Scott (1967b, p. 297) calculated a two-tail significance level of 0.079 for the target III result (apparent rainfall increase of 109%) and listed the ACN Project as the only randomized project in the United States up to 1965 providing significant evidence of rainfall increases due to seeding.

Neither Project SCUD nor the ACN Project had any numerical model of how seeding treatments would lead to the postulated results. Recent theoretical studies on the modification of synoptic scale systems have discussed the possibility of dynamic effects, but so far no one has presented a convincing conceptual model as to how the large scale changes could be brought about. Therefore, we conclude that the present state of the art provides no means for modification of frontal wave cyclones or other synoptic scale systems to increase *total* precipitation.

The conclusion just stated does not mean that clouds associated with such systems cannot be treated to produce local increases in precipitation. Indeed, many of the types of orographic clouds and convective clouds discussed in the previous two subsections form in response to the passage of synoptic scale weather systems and some of these are, as we have seen, susceptible to cloud seeding to increase precipitation in target areas of a few thousand square kilometers.

7.8 INADVERTENT MODIFICATION OF PRECIPITATION

It is perhaps strange to speak of inadvertent weather modification as a part of a discussion of the state of the art, as there can scarcely be an art to the production of unintended or unwanted effects. Nevertheless, it is appropriate to summarize at this point the knowledge available concerning inadvertent weather modification effects.

The effects apparently produced by cloud seeders outside of intended target areas (Section 7.5) are a kind of inadvertent weather modification. There are many others. Inadvertent weather modification effects can be produced by any physical process which changes the condition of the atmosphere. All of the concepts described in Chapter IV for the modification of weather by intentional means, such as the modification of the CCN or the IN spectra, are applicable to inadvertent modification. A complete discussion of inadvertent weather modification would cover such possibilities as the modification of the ozone layer by high flying supersonic aircraft and the release of fluorocarbons from aerosol spray cans. A review article on inadvertent weather modification by Barrett (1975) lists 375 references! However, our attention in this section is restricted to changes in cloud and precipitation processes.

H. Dessens (1960) speculated on connections between extensive forest and grass fires in Africa and nearby cumulonimbus developments, and Warner (1968) attempted to relate a local reduction in rainfall in Queensland, Australia, to the burning of slash in sugarcane fields. Dessens considered the effect of additional heating, but Warner considered the addition of particulate matter to the atmosphere to be of primary importance. Warner later decided on the basis of an analysis of rainfall statistics that effects of burning sugarcane fields upon rainfall had not been established (Warner, 1971).

The idea that large cities modify local weather conditions has been accepted for a century or more in Europe, but was slow to win acceptance in North America, However, after a number of interesting anomalies related to industrial complexes and other urban areas had been pointed out [e.g., Changnon (1968)], a comprehensive study of urban effects on weather was started in the 1960s using St. Louis and its environs as the study area. This project, called METROMEX (Metropolitan Meteorological Experiment), involved studies of clouds, precipitation, and other weather elements in the St. Louis area by radar, instrumented aircraft and other means over a period of several years. The results have been described in numerous technical reports and in papers at scientific conferences. A useful review is that of Braham (1976).

Effects documented in METROMEX which are now generally accepted by atmospheric scientists include increases in rainfall downwind of the city [e.g., Huff and Vogel (1977)], increases in the frequency of thunderstorms and hail over and downwind of the city, and increases in nighttime temperatures within and close to the city.

In searching for possible explanations for the observed changes in clouds and precipitation, scientists have considered the role of smoke and other man made aerosol particles as artificial CCN or IN [e.g., Fitzgerald and Spyers-Duran (1973)], differences in evapotranspiration rates for city and countryside, changes in the mixing properties of the planetary boundary layer induced by the high surface roughness of the city with its buildings [e.g., Kropfli and Kohn (1977)], and differences in the radiative properties of pavement and buildings as compared to trees or grassland. Changnon *et al.* (1976) have hypothesized that a combination of these factors is responsible.

Obviously, every city would present its own set of conditions and the findings in one particular geographic area may not apply universally. The effects of the city on clouds and precipitation also vary with the season, air mass type, and time of day.

One problem studied in detail in St. Louis was the effect of the city on the formation of convective clouds and resultant afternoon and evening showers and thunderstorms. Probes by an instrumented aircraft showed regions of anomalously high temperature over the city as compared to surrounding open country. Cloud bases over St. Louis tended to be higher than those over the surrounding countryside, apparently because of the relatively small evaporation rates over the city as a whole compared to the evapotranspiration from forests and cultivated land and rainfall was if anything lighter.

Examination of the cloud droplet spectra showed a slight tendency for the clouds over the city to have higher cloud droplet concentrations and hence smaller droplets than clouds in a "normal" air mass. This change, which would tend to retard the formation of rain by coalescence, does not appear to be responsible for the slight deficiency in precipitation noted over the city itself. This appears rather to be due to the reduced evapotranspiration over the city and the slightly higher cloud bases. The effect on the cloud droplet populations, which can be ascribed to the particulates introduced into the air by the city and its industrial activity, does not extend very far downwind. However, intensified shower development was noted downwind of the city on summer afternoons, apparently because of increased convergence induced by the surface roughness factors and the extra heating associated with the metropolitan area.

The effects of St. Louis on rainfall, and especially on convective rainfall, hail, and thunderstorms, extend some 100–150 km downwind. Apparently the thunderstorms set off by the convergent wind fields and the additional heating in the metropolitan area tend to be self-propagating and persist for as much as 2 or 3 hours of mean air motion downwind from the metropolitan area.

7.9 CONCLUSIONS AND FUTURE OUTLOOK

Results of field experiments and operations confirm that at least some of the concepts of Chapter IV carry over into the real atmosphere.

Fog modification is an operational technology today. Although applied research may lead to some refinement of techniques, it is likely that operations will continue along the lines already developed.

Despite residual uncertainties about true increases versus redistributions, there is no doubt that cloud seeding has consistent effects of great economic impact on orographic precipitation.

Operational programs to increase snowpack from winter storms over the mountains of the western United States continue in several states, notably California and Utah. The U.S. Department of the Interior, Bureau of Reclamation, is making plans for an extensive randomized experiment called the Sierra Cooperative Pilot Project, which will involve seeding part of the basin of the American River, which drains part of the southwest side of the Sierra Nevada, in northeastern California (Silverman, 1976). Weather conditions will be monitored to 200–300 km downwind in an attempt to assess the large area effect of the seeding. The possibility of mesoscale dynamic effects will also be explored.

Another important experiment, which will deal with both stratiform and convective clouds in an orographic situation, is the Precipitation Enhancement Project (PEP) of the World Meteorological Organization (WMO). Planning for PEP through 1976 has been described by List (1976). The initial task was the selection of a suitable site for a randomized pilot project involving perhaps 50,000 km^2, of which 10,000 km^2 would be the target area proper. In 1975, 16 countries volunteered the use of parts of their territories as the site for PEP. Consideration of available facilities and suitability of available cloud types for a project using ice nucleants to enhance precipitation from supercooled clouds led to selection of six possible sites, namely, Algeria, Australia, India, Spain, Tunisia, and Turkey. Further studies, including computer simulations using available historical rain gage data, led to selection of the Spanish site, which occupies the upper part of the Duero Basin above the city of Valladolid. It is thought that PEP will involve one or two more years of observations in the experimental area by radar and instrumented aircraft, followed by five years of randomized seeding experiments and one or two additional years for evaluation and analysis. Some preliminary cloud physics observations were collected there during early 1979.

The modification of large convective clouds in nonorographic situations remains as an important challenge. Those clouds offer not only the potential of increased rainfall, but of a means to modify mesoscale and even synoptic scale systems through dynamic effects.

One method which has not been adequately explored yet in randomized programs on convective clouds is the injection of dry ice (Chapter V). Additional dry ice experiments are planned under the High Plains Cooperative Experiment (HIPLEX) of the U.S. Department of the Interior, Bureau of Reclamation (Silverman, 1976).

Convective clouds are highly variable over the earth, responding to different air mass situations and local topography. The complex interaction of sea breeze fronts from the east and west coasts of Florida over the Florida Penninsula on summer afternoons is an example of the peculiar factors controlling convective development in specific areas. Numerical models of such situations are now appearing.

Proper understanding of the diurnal cycle of convective activity in response to local conditions can lead to an ability to produce substantial modifications of the resultant rainfall patterns. As Howell (1960) has pointed out, the location of the first cloud in a given area that precipitates on a given day may be important to subsequent developments over several thousand square kilometers. There are undoubtedly many places in the world where even partial understanding of the factors controlling convective developments could lead to quite successful programs of local or mesoscale precipitation augmentation through cloud seeding.

CHAPTER

VIII

Suppression of Weather Hazards

Discussions of weather modification with lay persons nearly always turn to the possibility of relieving some of the misery experienced by human beings as the result of such weather hazards as floods, hailstorms, and tornados.

This chapter outlines some past and current attempts at suppressing losses due to hailstorms, lightning, and hurricane winds. Other hazards, such as thunderstorm winds, are mentioned briefly. The suppression of tornados is not discussed at all because we do not yet have a conceptual model to suggest how to go about suppressing them.

8.1 HAIL SUPPRESSION

Choice of Seeding Method

Two conceptual models for the suppression of hail by cloud seeding with glaciogenic agents were set forth in Chapter IV. Those two and a number of additional proposed concepts are listed in Table 8.1.

The trajectory lowering model holds that the size of a hailstone depends on its trajectory through the cold upper part of the cloud. The closely related coalescence model is based on the idea that stimulation of rain from below the 0°C level reduces the amount of water to be lifted

TABLE 8.1

List of Conceptual Models for Hail Suppression by Cloud Seeding

(1)	Complete glaciation of cloud water
(2)	Introduction of competing embryos
(3)	Trajectory lowering
(4)	Promotion of coalescence
(5)	Seeding for dynamic effects, e.g., premature weakening of updraft

above the 0°C level to feed hailstone growth. Not all of the proposed concepts in Table 8.1 are practical. Numerical modeling calculations [e.g., Young (1977)] have shown quite clearly that the complete conversion of cloud water droplets individually into cloud ice particles (the complete glaciation hypothesis) requires unrealistically large amounts of glaciogenic seeding agents, perhaps 1 kg of AgI per minute. The trajectory lowering and the promotion of coalescence hypotheses, which are closely related, would require large quantities of hygroscopic seeding agents. Young (1977) considers trajectory lowering by hygroscopic seeding to have a chance, judging from modeling results. The hypothesis that hail can be suppressed by dynamic cloud seeding admits of several variations, but none has yet been formulated in sufficiently precise form to permit a test of its validity.

Nearly all hail suppression programs have used glaciogenic seeding agents with the intent of increasing competition among the hailstone embryos. One project in the Caucasus Mountains in the southern U.S.S.R. utilized hygroscopic agents in combination with ice nuclei in a hail suppression program (Lominadze *et al.*, 1974), but all of the other major hail suppression programs have relied exclusively on artificial ice nuclei to promote additional freezing centers.

Artificial ice nuclei have been delivered to hailstorms by the three general approaches of broadcast seeding, updraft seeding, and direct injection. The relative merits of the three approaches have been discussed in general terms in Chapter V. Examples of each as applied to hail suppression are given in Table 8.2.

Because the competing embryo hypothesis holds that hail can be suppressed by providing freezing centers *in sufficiently high concentrations,* there has been a tendency for the amounts of seeding agent to increase from year to year on a given project. Despite Sulakvelidze's hopeful comment (Sulakvelidze *et al.*, 1967, p. 176) that "The 1963 experiments showed the possibility of reducing the doses somewhat in the future," seeding agent consumption per storm in operations in the U.S.S.R. has increased from a maximum of 900 g in one 1963 experiment to as much as

TABLE 8.2

Examples of AgI Consumption on Hail Suppression Projects[a]

Location	Area (km^2)	Year	Per season (kg)	Average per day (kg)
	Broadcast			
Alberta	6000	1968	266	17
France	70,000	1971	4500	55
	Updraft			
North Dakota	5000	1966	30	1.2
(Bowman-Slope)	5900	1967	55	2.9
	8200	1968	96	3.1
South Dakota	120,000	1974	230	3.2
Colorado (NHRE)	1500	1973	16	8
		1974	128	9.8
	Direct injection			
Caucasus Mts.[b]	300	1962	6.3	0.9
	500	1963	9.0	0.6
	—	1973	—	9.0
Alberta	14,000	1972	137	5.7
	19,000	1973	69	4.3

[a] After Dennis (1977), by permission of American Meteorological Society.

[b] It is believed that the reagent in these experiments was PbI_2. Cloud chamber measurements showed 1.7×10^{12} ice nuclei at $-10°C$ per gram of PbI_2 compared to 3.2×10^{12} per gram of AgI (Sulakvelidze *et al.*, 1967, pp. 174–176).

75 kg on a large storm, admittedly in a different project area, during the 1970s.

It should be noted that the figures given in Table 8.2 on reagent consumption are not an accurate guide to the numbers of ice nuclei introduced to the storms, because of the tremendous variations in generator efficiencies.

Results of Experiments and Operations

A comparison of the different methods of hail suppression is difficult due to the extreme variability of hailfalls which makes an accurate evaluation of any single project impossible. Simulated computer experiments have shown that more than 10 years could be required to detect a 50% suppression effect on an experimental project using the randomized single-area design [e.g., Long *et al.* (1976)], and target–control correlation coefficients are very low. Nevertheless, the sum total of all experiences to date may offer a clue to the relative effectiveness of the three seeding approaches.

Broadcast Seeding. Silver iodide has been broadcast from generators on the ground on randomized hail suppression experiments in Switzerland and Argentina and on many operational projects. The experiment in Switzerland (Grossversuch III) failed to demonstrate any evidence of a hail suppression effect (Schmid, 1967), but apparently increased rainfall in certain situations (Neyman and Scott, 1967b). In the Argentine experiment hail may have been suppressed during cold front situations but not otherwise (Iribarne and Grandoso, 1965).

Commericial projects to suppress hail using AgI ground generators have been operated intermittently over the years in Alberta. Operators and supporters [e.g., Krick and Stone (1975)] have been optimistic concerning the results but Petersen (1975) viewed the results as inconclusive.

A very extensive program of seeding for hail suppression with AgI generators on the ground was begun by H. Dessens in southwestern France in the 1950s and has continued each year since then under the auspices of the A.N.L.F.A. group (Association Nationale de Lutte contre les Fléaux Atmosphériques) of Toulouse. Ground based seeding is used because mountainous terrain and poor visibility in haze make aircraft operations dangerous. Consumption of AgI now amounts to several metric tons per season (Table 8.2). The annual report for 1978 indicates an attempt to reduce consumption by more restrictive rules on when to operate generators. The most recent English language report on the project is that of Dessens (1979).

The operators have reported some evidence of success on the basis of target–control comparisons [e.g., Dessens and Lacaux (1972)], but Boutin (1970) finds the results inconclusive due to instability of the variable tested, namely, the ratio of hail insurance claims paid to the total insured liability.

There was an eight year nonrandomized project in Germany in which AgI was broadcast by low altitude rockets. Mueller (1967) reported no evidence of hail suppression on that project.

Updraft Seeding. France has been the site of an operational project using a combination of direct injection and updraft seeding from aircraft (Picca, 1971). The project was operated by the Association Climatologique de la Moyenne Garonne (A.C.M.G.). A silver iodide–levilite mixture dispersed in powder form was the main seeding agent but some pyrotechnics were used. Boutin *et al.* (1970) have examined Picca's results, but rate the A.C.M.G. project as inconclusive.

There have been numerous operational hail suppression projects involving updraft seeding with AgI generators on aircraft in North America ever since 1948 (Frank, 1957). Most of these projects were never subjected to a scientific evaluation. However, Schleusener collected hailfall data on the northeast Colorado project in 1959. His results (Schleusener, 1962) and those of Butchbaker (1973) covering three years of observations on the Bowman–Slope Project in North Dakota presented an optimistic view of probable seeding effects, suggesting both suppression of large hailstorms and a stimulation of rainfall as compared to surrounding regions.

A comparison of hail damage to tea plants from seeded and unseeded thunderstorm cells in Kenya suggested a suppression effect (Henderson, 1970), but suffered from that fact that no-seed cases were not selected at random and that hail from a cell more than 15 min after the end of seeding was considered to be from an unseeded case. A subsequent brief randomized trial in Kenya provided no new significant evidence.

Hail research studies which began in northeast Colorado in association with an operational project in 1959 continued for almost 20 years, with participation by several university groups, the National Center for Atmospheric Research (NCAR), and agencies of the U.S. government. A randomized trial on 11 individual thunderstorms in 1964 showed no evidence of hail suppression through seeding in strong updrafts (Schleusener and Sand, 1964).

Following further study by national steering groups, plans were laid for a National Hail Research Experiment (NHRE) to conduct basic research on hailstorms and test the concept of hail suppression by direct injection of AgI. A "protected area" of 1600 km^2 was set up in northeast Colorado (Fig. 8.1). The actual experiment ran from 1972 to 1974 inclusive.[1] The method chosen for injecting the AgI was the firing of rockets upward from aircraft below cloud base. Unfortunately, technical problems occurred. The project started in 1972 with updraft seeding using AgI flares, and it was not until 1974 that the flares were replaced entirely by rockets. Therefore, we are treating NHRE under the heading of updraft seeding.

The research results of NHRE are treated in many places, notably in a monograph on "Hail: A Review of Hail Science and Hail Suppression," edited by Foote and Knight (1977). A preliminary analysis of the data showed no evidence of a hail suppression effect [e.g., Atlas (1977)]. Continued analysis efforts over the past few years have been generally inconclusive, which is not surprising for a three-year program using the randomized single-area design (Crow *et al.,* 1977).

[1] NHRE was planned as a five-year project, but was ended when it became apparent that no conclusive results could be expected from two additional years of seeding.

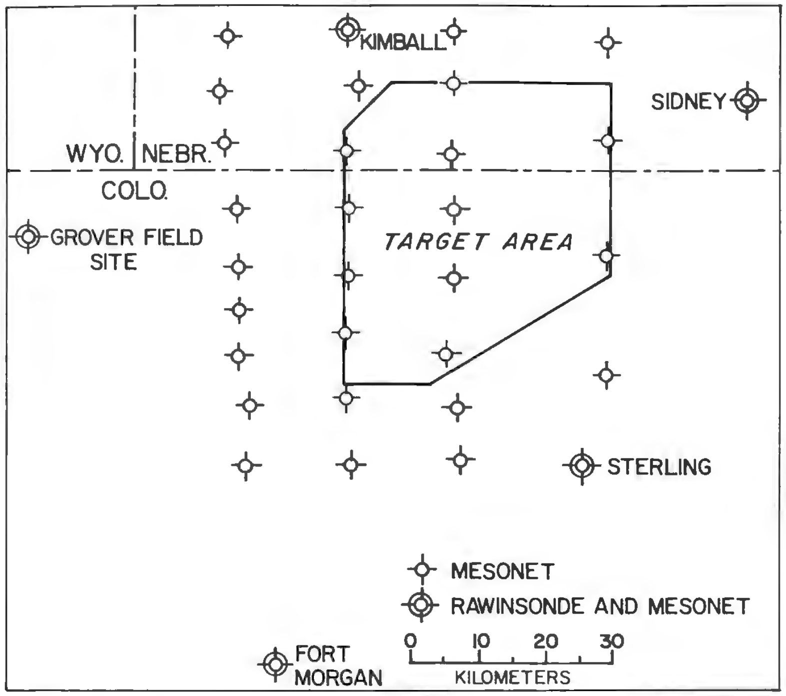


Fig. 8.1. **Map showing protected area and Grover operations center of National Hail Research Experiment (NHRE).**

Hail data collected on experimental programs of updraft seeding in the Dakotas suggested that hailfalls from seeded storms were generally less severe than those from unseeded storms (Schleusener *et al.*, 1972). Hail insurance data from the area of the randomized North Dakota Pilot Project (NDPP) of 1969–1972 show with marginal statistical significance that crop hail damage per seed day was less than crop hail damage per no-seed day (Miller *et al.*, 1975). The analysis of hail pad data from the same project suggest reductions of 50% or so in hail energy but do not meet standard tests for statistical significance.

The largest operational project to date using the updraft seeding technique is that operated by the State of South Dakota from 1972 to 1975. The project had the dual objectives of increasing rainfall and suppressing hail. The project expanded from 26 of the state's 67 counties in 1972 to cover approximately 45 counties in the three following years (Fig. 8.2). It was largely abandoned for 1976 because opponents of the program feared that it was suppressing rainfall as well as hail. Miller *et al.* (1976) analyzed crop hail insurance data and found that hail was less severe in the counties participating in the program than in nonparticipating counties. It was noted that this could have represented climatological differences. There-

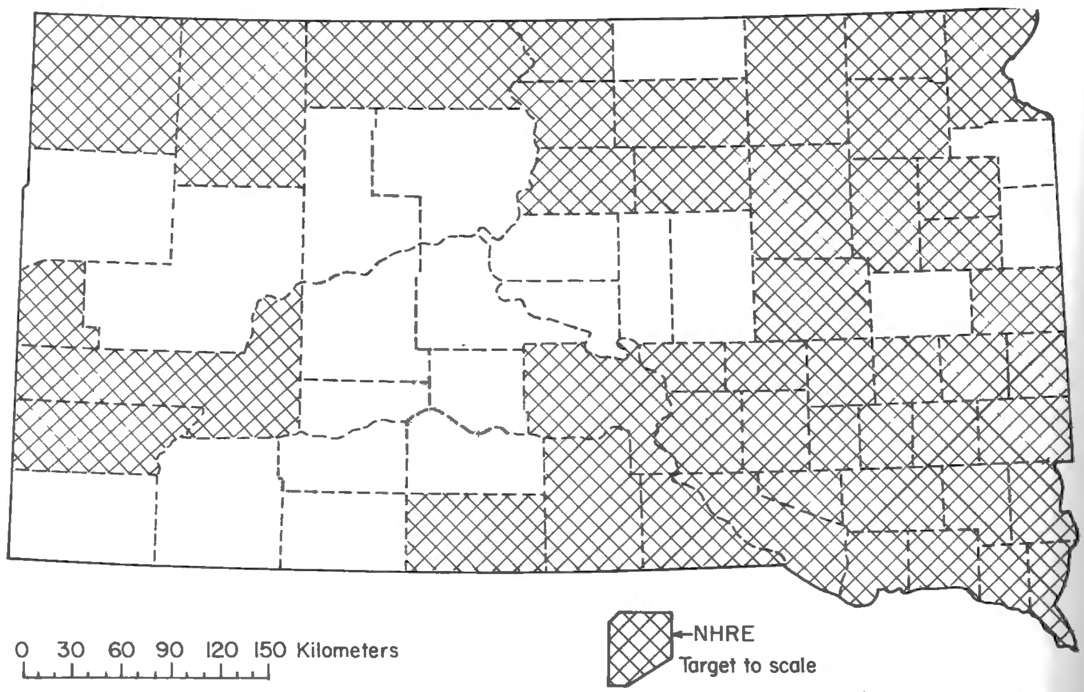


Fig. 8.2. **Area covered by cloud seeding program to stimulate rainfall and suppress hail in South Dakota in 1974. The NHRE protected area of Fig. 8.1 is reproduced to scale for comparison purposes.**

fore, simulated experiments were run in a computer to extend the actual combinations of seeded and unseeded counties backward in time to 1948 to establish the probability distributions of the differences observed in 1972–1975. It was concluded that the difference in hail experience in the seeded and nonparticipating counties from 1972 to 1975 was not due entirely to climatological effects. Apparently, the very large area involved, with over 100,000 km^2 seeded each year, tended to smooth out the natural variations in hailfall, and provide a stability to the results which was not possible in the small experimental area of NHRE (Fig. 8.2).

Henderson and Changnon (1972) and Changnon (1975) have presented some target–control analyses of operational projects in Texas which also suggest favorable results.

Despite the failure of any project to achieve a suppression effect meeting the usual standards for statistical significance, the total score card for updraft seeding seems more promising than that for broadcast seeding. Changnon (1977) reviewed the data available up to late 1976 and concluded that the preponderance of evidence supported the hypothesis that updraft seeding produces a modest suppression effect of 20 or 30% in hail damage.

Direct Injection. As noted in Chapter V, scientists in the U.S.S.R.

pioneered the development of the direct injection technique for hail suppression in the 1950s and 1960s after becoming convinced that the key to successful hail suppression was rapid response to threatening situations and that seeding agents introduced at cloud base would be ineffective as ice nucleants near 0°C. This latter condition would, of course, apply if their seeding agents contained significant amounts of soluble material, say lead oxide (Pb0), as impurities. We have already referred to the variety of rockets and artillery shells which they developed for hail suppression purposes.

Operational projects of hail suppression, which began in the Caucasus Mountains in the southern U.S.S.R., now extend from Kazakhstan in central Asia to Moldavia in the western U.S.S.R., and cover millions of hectares. However, the methods used are not identical. Some of the Soviet methods and equipment have been introduced on an operational basis into other countries of eastern Europe, including Hungary and Bulgaria, and one type of rocket and methods prescribed for its use are being tested on a randomized project in Switzerland (Federer, 1977).

The evaluation by Soviet scientists of their hail suppression projects are essentially target–control analyses utilizing crop damage data. Their reports of recent years are slightly less optimistic than the earlier reports by Sulakvelidze and others, which spoke of 90–100% suppression of hail. Tabulated data presented at the Second WMO International Conference on Weather Modification by Burtsev (1976) covered almost 50 project seasons and showed results generally in the range of 50–90% suppression for various projects. Despite some criticisms from foreign scientists over the lack of randomization and possible bias in the evaluations, the Soviet scientists remain convinced of the correctness of their approach. As Dr. A. Chernikov remarked at the WMO meeting of hail experts in 1977, the results will not change.

The direct injection technique has also been used in hail suppression experiments in Alberta. Droppable pyrotechnics were developed for that specific purpose by the Alberta Hail Studies Group [e.g., Summers *et al.* (1971, 1972)]. It is, of course, essential to use enough units to ensure widespread distribution of the seeding agent. In spite of an apparent reduction in hailstone size in some early single cell trials, the Alberta hail experiments have not yet generated any evidence for an overall suppression effect. As the Alberta group is now supplementing its direct injection seeding with AgI releases below cloud base, and have largely abandoned the use of randomization due to pressures from farm organizations for operational seeding, it is hard to see how they can obtain any conclusive results about direct injection of pyrotechnics.

The direct injection method has been introduced to the Nelspruit

Project in South Africa by a private American firm. Operators of the Nelspruit Project claim favorable results [e.g., Mather *et al.* (1976)] and independent evaluations of that project are in progress.

Possible Engineering Deficiencies

It should be emphasized again that hailfalls are extremely variable. The results for certain individual projects may actually be opposite to those indicated by the evaluations. The present author considers that the results for updraft seeding and direct injection, taken in toto, indicate a net suppression effect. Only one thing is certain: The goal of 100% hail suppression has eluded everyone.

The failure of hail suppression projects based on the competing embryo concept to suppress hail completely could be due to engineering deficiencies, for example, failure to produce high enough concentrations of ice nuclei in the hail forming regions or inadequate diffusion of the seeding agent within the hail producing storm cells.

With regard to the first possible problem (inadequate concentrations), it is fair to note that there are few data showing any correlation between hail severity and natural ice nucleus concentrations. This suggests that production of a detectable effect must require ice nucleus concentrations well above the natural background counts. Such concentrations have been measured on several projects. To the suggestion that higher concentrations would be better, one can only say that increases in AgI consumption from year to year on various projects (Table 8.2) have not yielded consistent evidence of improved performance.

Seeding with aircraft in strong updrafts below cloud base results in various distorted cylinders of seeded air moving upward into the hail cells, as discussed in Chapter V. As noted there, a single aircraft can seed a moderate convective cell fairly well, but it cannot seed a fast updraft effectively because the time available before the seeding agent passes the 0°C level is too short to spread it throughout the updraft. Further confirmation of this view is provided by Linkletter and Warburton (1977), who analyzed precipitation from 18 supposedly seeded hailstorms on NHRE for silver content by the atomic absorption technique. They estimated that less than 10% of the samples contained enough silver to represent a significant seeding effect.

Some practitioners of updraft seeding have been aware of the problem of inadequate diffusion for years. They have concentrated seeding under newly developing cells before the updrafts become intense or in generalized inflow regions several kilometers from the most intense updrafts.

There are not sufficient data in hand to say whether this difference in procedure accounts for the apparent success (at *very* weak significance levels) on certain updraft seeding projects as compared to NHRE (Changnon, 1977). It is interesting to note, however, that the plans for a resumption of NHRE field experiments in 1976, which never materialized, called for a similar seeding approach on that project. Some Soviet hail fighters (e.g., Abshaev and Kartsivadze, 1974) have also modified their approach in recent years to emphasize seeding in incipient hail cells as well as in mature cells with strong radar echoes.

Further Comments on Competing Embryo Hypothesis

Evidence of rainfall increases associated with some hail suppression projects (to be discussed below) argues that the seeding has not been totally ineffective, and that failure to suppress hail dramatically may be due to basic flaws in the competing embryo hypothesis rather than engineering deficiencies.

Dennis (1977) has noted three objections to the competing embryo concept ranging from technical to fundamental, as follows: (1) The production of competing embryos by AgI crystals acting on cloud droplets in a strong updraft is not feasible, as the particles produced in the time available are not large enough to function as hail embryos (see footnote 2 in Chapter III). (2) If artificial hail embryos are introduced directly, one must transport unrealistically large masses of material. Young and Atlas (1974) find that 1-mm embryos would have to be introduced in concentrations around 10^3 m^{-3} to reduce the mass of individual hailstones by a factor of 5. With a large hailstorm drawing in 10^9 m^3 of air per second, a requirement for tons of embryos per storm is found. (3) The assumption that the amount of hail to fall from a cloud is constant and that the only effect of extra embryos would be to reduce hailstone diameters appears to be a *dangerous oversimplification*. Many hailstorms are inefficient producers of precipitation, and it is quite conceivable that extra embryos could increase the total mass of hail from a storm (Browning and Foote, 1976).

Examination of these problems led Dennis (1977) to suggest the following hypotheses:

1. Enough competing embryos to suppress hail can be produced quite readily by seeding with less than 1 kg of AgI per hour if the hailstorm contains significant amounts of supercooled rainwater. The raindrops are not frozen by direct action of the AgI crystals, but by collision with ice crystals induced by the artificial ice nuclei. The ice crystals in

turn are most efficiently produced by condensation-freezing, rather than by collision-freezing of cloud droplets (Chapter V).

Storms which produce hail from cloud water without any intervening rain stage pose two possibilities.

2. In storms which consist of a number of cells which grow, produce hail, and dissipate, in turn, additional hail embryos can be produced by seeding early in the lifetime of each new cell.

3. In quasi-steady-state storms (supercells), where hailstones grow by direct accretion of supercooled cloud droplets blowing upward past them, for example, above a weak echo region, there is no obvious way in which cloud seeding can retard the growth of the largest hailstones.

A complete study of these suggestions, which are based on kinematic cloud models and hailstone growth models, and of others which may be offered, will only be possible with the aid of dynamic cloud models with interacting microphysics. Timing of treatments may prove critical, as the proponents of direct injection methods have long asserted. Preliminary results by Farley *et al.* (1976) point to a wide range of possible outcomes depending on initial conditions (Fig. 8.3).

In view of the wide range of physical conditions which can exist within storms yielding damaging hail at the ground, it appears that any field experiment that fails to allow for these variations can never provide convincing evidence for the effectiveness of any proposed hail suppression technique. The distinction between storms with frozen raindrop embryos and storms with rimed ice crystals as embryos may well be crucial, as Atlas (1977) and others have suggested.

The consensus of a panel of hail experts convened by the WMO in 1977 was that, while the concept of an international hail suppression experiment was interesting, more understanding of hail growth mechanisms should be achieved prior to the launching of a formal, international field experiment. They did, however, endorse such endeavors as the Swiss hail experiment (Grossversuch IV), and noted with approval the participation in it of French and Italian scientists.

Effects of Hail Suppression Seeding on Rainfall

The effects of hail suppression seeding on rainfall are important from an economic point of view and because they provide some insight into the ways that seeding can affect cloud processes.

The concept of hail suppression by provision of additional competing embryos bears some resemblance to the concept of suppression of rainfall

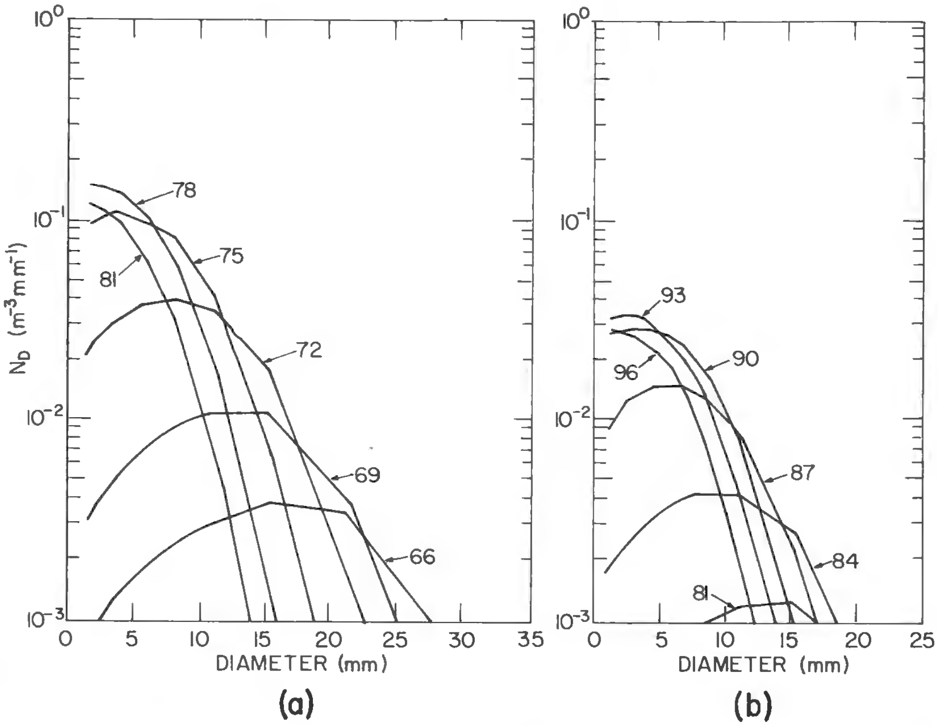


Fig. 8.3. Model generated hail spectra for a storm observed in Switzerland. Spectra are labeled in minutes from time of initialization of model. (a) Unseeded cases, (b) seeded to provide one ice nucleus per liter active at $-5°C$ and 150 liter^{-1} active at $-20°C$. For the unseeded case the model predicted total hail impact energy at the ground of 1290 J m^{-2}, while the actual storm produced 1810 J m^{-2}. Model indicates seeding would have reduced total impact energy in this case to around 275 J m^{-2}. The one-dimensional model used here reflects effects of the timing of seeding of individual hail cells. [After R. D. Farley *et al.* (1976). Papers presented at *WMO Scientific Conf. Weather Modification, Boulder, 2nd,* p. 349, by permission of World Meteorological Organization and senior author.]

by overseeding. Some practitioners of weather modification projects have even spoken of overseeding for hail suppression. It is perhaps for that reason that many persons have expressed concern that hail suppression projects would inevitably involve suppression of rainfall in the treated areas.

Closer examination of the concepts in Table 8.1 indicate that it is the concept of hail suppression by complete glaciation that corresponds to overseeding in the case of rainfall projects. The objective in the competing embryo concept is to produce additional freezing centers so that the same quantity of precipitation reaches the ground with only the size distribution of the particles and hence their probability of melting to rain being affected.

Such studies of rainfall in and around hail suppression projects as have been conducted generally support the concept that the provision of the extra freezing centers has a tendency to increase rainfall from the treated storms rather than suppressing it. We have already mentioned the experience of Grossversuch III, where the operation of generators on the ground failed to produce any significant effect on hailfalls but was associated with significant rainfall increases when the seeding was conducted under inversions in advance of convective storm outbreaks.

Several evaluators of hail suppression projects have used hail/rainfall ratios as a guide to whether or not a hail suppression effect was being achieved. The cautions sounded in Chapter II about the use of ratios in evaluations apply with additional force here, because the rainfall and hailfall probability distributions are quite different. Nevertheless, it may be significant that every comparison which has come to the present author's attention shows that hail/rainfall ratios tend to be smaller for seeded than for unseeded storms. A number of such studies are covered in a review paper by Schleusener (1968). This result could obviously reflect a wide variety of responses to seeding in terms of both rainfall and hail, but it is encouraging. It suggests that the percentage suppression of hailfall damage exceeds the suppression of rainfall, if any exists, by a wide margin. However, a more penetrating examination is required.

Schleusener (1962) concluded on the basis of target–control studies that the operational project in northeast Colorado in 1959 had a tendency to increase rather than decrease rainfall. The evidence from NHRE, while not conclusive with regard to either hail suppression or rain stimulation, pointed in the same direction (Atlas, 1977). The North Dakota Pilot Project of 1969–1972 pursued the objectives of rain stimulation and hail suppression in the same area during the same season (but not on the same clouds) and found tentative evidence of success in both objectives (Dennis *et al.*, 1975b; Miller *et al.*, 1975). The same results were found on the large scale operational program operated with the dual objectives of rain stimulation and hail suppression in South Dakota (Miller *et al.*, 1976; Pellett *et al.*, 1977). Operators of the operational projects for hail suppression in Alberta have also presented target–control analyses indicative of a stimulation rather than a suppression of rainfall from those activities (Krick and Stone, 1975).

The most extensive hail suppression operations in the world remain those in the U.S.S.R. Soviet scientists have not published extensive or detailed analyses of the effects of their seeding upon rainfall. However, the question has obviously been examined to some extent. Sulakvelidze *et al.* (1974) report that target–control studies suggest rainfall increases of 10–15% associated with some of their hail suppression projects.

In summary, we note the somewhat ironic result that programs conducted for the purpose of hail suppression have apparently had greater success in stimulating rainfall than programs designed specifically for the latter purpose. This might reflect the differences in the operational procedures employed, but the present author suspects that it reflects differences in the characteristics of the cloud systems treated. Hailstorms are often inefficient producers of precipitation, as noted in Chapter II, and present far more attractive targets for rain stimulation efforts than the tropical cumulonimbus clouds treated on certain days on Whitetop, for example.

8.2 LIGHTNING SUPPRESSION

The principal motivation for lightning suppression experiments is found in the economic losses produced by forest fires ignited by lightning.

Use of Ice Nucleants

The first cloud seeding flights to suppress lightning in the United States were made over the northern Rocky Mountains on 16 June 1949 (Schaefer and Gisborne, 1977). The California Division of Forestry conducted a randomized experiment using AgI ground generators in 1958 and 1959 following some earlier trials with aircraft seeding (Court, 1967, p. 247). No significant changes in rainfall or lightning caused fires were observed in California. Experiments in the northern Rockies continued through the 1950s and 1960s with support from the U.S. Forest Service under the code name Project Skyfire.

Project Skyfire was intended to test the hypothesis that glaciogenic seeding would reduce the number of cloud-to-ground lightning strokes. Support for this hypothesis is found in suggested thundercloud charging mechanisms which depend on the coexistence of graupel and supercooled water (Mason, 1971). The working hypothesis was eventually refined to concentrate upon those hybird lightning flashes with continuing currents of 40 ms or greater duration following one or more of the return strokes. Such flashes are thought to be responsible for most lightning fires (Dawson *et al.*, 1974).

A two-year randomized trial in 1960–1961 was followed by a second experiment of three-year duration from 1965 to 1967. The latter experiment used both ground based and airborne generators burning AgI–NaI solution in acetone. Consumption of AgI averaged about 2 kg hr^{-1}. Artificial cloud glaciation was observed with some growing cumulus glaciated at temperatures as high as $-7°C$ (MacCready and Baughman, 1968).

Various analyses of the Skyfire lightning counter data have been published. The most authoritative is probably that of Baughman *et al.* (1976). They noted an apparent reduction in the frequency of the cloud to ground flashes and in the duration of lightning activity in a given storm. A decrease in the duration of the long continuing currents was found with a p value of 0.04 (two-tailed test). However, the analysis by Baughman *et al.* departed from the original design, which called for analysis by paired days, so one cannot accept the calculated significance levels strictly at face value.

Gaivoronsky *et al.* (1976) have reported on randomized experiments carried out in Moldavia from 1973 to 1975. They concluded that rocket seeding with either PbI_2 or cupric sulfide (CuS) was followed by short term increases in the frequency of intracloud flashes and then by substantial decreases in the total electrical activity of the storms as measured by lightning counters and electric field meters. However, there is some possibility of bias in their data. Observations of seeded clouds were geared to periods before, during, and after *seeding,* while observations of control clouds were geared to periods before, during, and after the *maximum lightning activity*. A disturbing feature of their results is the evidence that rainfall from seeded clouds lasted on average 36 minutes compared with 64 minutes in the control cases. This could indicate a rainfall suppression effect or the selection (by chance) of the more vigorous storms as control cases. If the latter situation were true, any conclusion that seeding suppressed the lightning would constitute a type I error.

Chaff Seeding

An alternative approach to lightning suppression is the release of chaff (finely divided metal foil) to create corona discharge, thereby ionizing the air and increasing its conductivity sufficiently to discharge clouds without lightning flashes. Field tests of the concept were carried out at Flagstaff, Arizona, in 1965 and 1966. Kasemir presented evidence of vertical electric fields decaying after the release of chaff (Dawson *et al.*, 1974, pp. 612–622).

Battan (1977) reported that chaff seeding with a carbon substance was being investigated by the Central Aerological Observatory of the U.S.S.R. in 1976 with actual field experiments planned in Moldavia. No information is available yet on the results.

Present lightning research in this country is organized around the Thunderstorm Research International Program (TRIP), which emphasizes observations and development of theoretical and numerical models of thunderstorms with no attempt at modification [e.g., Pierce (1976)].

8.3 MODIFICATION OF HURRICANES

Project Stormfury

The only large scale research program in the world dedicated to the modification of tropical hurricanes is Project Stormfury. Project Stormfury was organized as a joint effort of the U.S. Weather Bureau and the U.S. Navy about 1962. At the present time responsibility for Stormfury rests with the National Oceanic and Atmospheric Administration (NOAA).

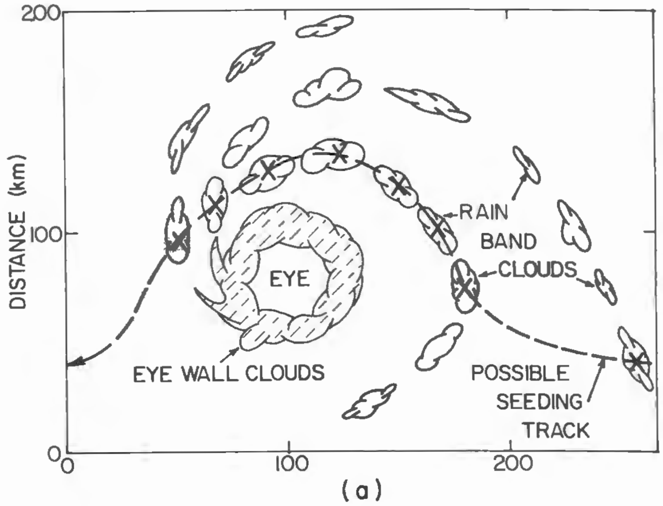
Project Stormfury has gone through several major changes, including changes in seeding devices and other equipment, and even in the conceptual model on which the program is based. Reviews of the project history are available [e.g., Gentry (1974)]. The evidence of success in the past experiments is fragmentary and unconvincing. As of this date (May 1979), the project is in a state of suspense pending completion of international agreements for renewal. Discussion here will be limited to the present conceptual model and the plans for its implementation in renewed experiments.

Conceptual Model

Hurricanes draw their energy mainly from the release of latent heat in the convective cloud towers making up the eye wall cloud and the rain bands spiraling inward toward the eye wall. A number of modeling studies as reviewed by Rosenthal (1974) support the concept that the rainfall rates in mature hurricanes balance the moisture convergence in the lowest kilometer of the atmosphere, and that the latent heat released thereby drives the hurricane. It is therefore reasonable to think that seeding to change cloud dynamics could affect the entire storm.

The basic seeding concept embodied in the present Stormfury hypothesis is the generation of a new eye wall from rain band clouds by dynamic seeding (Fig. 8.4). The seeded rain band clouds are stimulated to grow up to the divergent region in the upper troposphere while increased inflow into them near the sea surface deflects air away from the existing eye wall. The total energy adjustment involves additional release of latent heat of condensation and is several times larger than the direct adjustment due to freezing of supercooled water.

Success in Project Stormfury depends on the development of numerical models capable of assessing probable seeding effects. The models potentially available for testing hurricane modification hypotheses include two-and three-dimensional models, but the three-dimensional model is



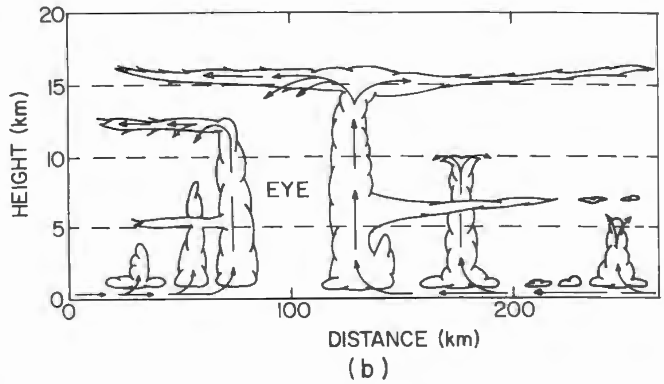


Fig. 8.4. **The current seeding hypothesis for Project Stormfury. Latent heat releases in pyrotechnic drop zones (×) in rainband clouds (see plan view) stimulate their growth toward the tropopause, reducing the inflow to the eye wall clouds. Eventually the original eye wall dissipates, and a broader eye is formed with a reduction in maximum winds.**

still under development and not ready for testing of seeding hypotheses. Various difficult problems remain to be solved, including the handling of the convective clouds within the storms. The release of latent heat in the convective updrafts is recognized as the hurricane's main source of energy and so must be represented in the models. However, the updrafts are very small compared to the spacing between grid points in the models. Parametrization is being explored as one possible solution to the dilemma.

Equipment

The equipment assembled so far for renewed Stormfury experiments is formidable. A total of five long range aircraft equipped with seeding devices or meteorological sensors or both will be deployed for continuous operations far from land over periods of several hours.

New seeding devices have been developed for Stormfury by the U.S. Naval Weapons Center. They are small droppable pyrotechnics measuring 250 mm in length by 15 mm in diameter and containing 17 g silver iodide in a formulation containing 78% $AgIO_3$ by weight. Up to 120 can be launched per minute and it is expected that each seeding aircraft will carry up to 4000 units. The units are ignited as they leave the firing tube and burn out in a drop of 3 km.

Experimental Design

The present thinking with regard to the experimental design is that no storms will be reserved as no-seed cases, but that randomization will be introduced in determining the time of seeding. It is proposed to determine how well the total sequence of events following a seeding mission conforms to specific hypotheses about seeding effects. Methods are being devised to set significance levels for tests of the null hypothesis that events following seeding are not different from those during other periods of observation (Knight and Brier, 1978).

One problem with Stormfury has been the small number of storms available per year in the North Atlantic for experimentation. At the present time negotiations are underway with Mexico and Australia regarding the possibility of joint experiments on hurricanes in the eastern and southwestern Pacific, respectively.

8.4 SUPPRESSION OF OTHER HAZARDS

The discussion in Chapter VII and Sections 8.1–8.3 summarizes the principal applications of cloud seeding technology to date. Additional concepts for suppressing threatening weather developments have been proposed, and a few have been tested on a limited scale in field experiments.

Excessive Rainfall

Rainfall suppression by a secret chemical in the State of Washington in 1952 has already been mentioned. The suppression of excessive rainfall and local floods by overseeding with AgI has often been suggested as a possible application of cloud seeding. A few such attempts were made in the 1950s in California, mostly directed against thunderstorms during wet periods. No serious evaluation of those attempts is possible on the basis of the published data.

Thunderstorm Winds

As early as 1948 the United Fruit Company was seeding thunderclouds over Honduras with dry ice in an attempt to reduce the number of banana trees blown down by the associated winds (Byers, 1974, p. 17). Wind suppression operations near Santa Marta, Colombia, have been described by Lopez and Howell (1961). Lopez and Howell considered the release of convection early in the day and the shading of the ground by artificial cirrus to be important in preventing severe storms later in the day. Evaluation of the Santa Marta operation was difficult because of the absence of suitable controls.

Dynamic Destruction of Convective Storms

Vulfson *et al.* (1976) have developed a theoretical basis for experiments in which convective clouds are supposedly weakened or destroyed by imparting downward momentum to parcels of air near the cloud top. Methods used to initate the downdrafts have included explosions and the dropping of powdered cement or other hydrophilic material into the cloud tops.

Cumulus dissipation experiments have been conducted over the Gulf of Mexico by the U.S. Naval Weapons Center. Lewis and Hawkins (1974) dropped cement from a C-130 aircraft in some trials in 1972. They concluded that visual observations alone could not determine the effects of seeding cumulus with hydrophilic agents.

Frost Prevention

The possibility of preventing frost by artificially generating cirrus clouds has been mentioned in Section 4.3. However, most applied work in frost prevention has followed more prosaic approaches.

There are numerous techniques for modifying microclimate and micrometeorological conditions, including reduction of risk of frost, by construction of protecting walls, planting of wind breaks and hedges, artificial production of smoke clouds to reduce radiative heat losses, and using radiant heaters directly on trees threatened by frost. All of these techniques are appropriate in certain situations and are widely used. However, they are not discussed in detail in this monograph as they fall outside of the topic of weather modification by cloud seeding. Their effects are on such a small scale that they normally are of no concern to persons other than their practitioners.

Drought

Drought is not spectacular. It works so insidiously that only the most calamitous examples of it attract much attention. Nevertheless, it is the greatest weather hazard of all.

The possibilities that exist for stimulating precipitation during moderate to severe droughts in Illinois have been explored by Huff and Semonin (1975). They conclude that favorable conditions for cloud seeding occur frequently enough to provide occasional relief in most droughts. Regardless of scientific studies, the urge to relieve human misery during droughts is very strong. Drought relief programs have been operated in many countries, for example, in some of the Sahel countries during 1973 and 1974.

Stimulation of precipitation is not the only possible response to drought. As we shall see in Chapter IX, crop-yield models are being used increasingly to predict what the effects of weather modification programs might be on agricultural production in specific instances. Modeling studies show rather clearly that heat stress is a key important factor during droughts. Decreases in maximum temperature of even 2 or 3°C during droughts would alleviate the damaging effects considerably. Indeed, if such changes could be produced systematically, the relief produced would be comparable to that associated with the potential increases in rainfall.

As noted previously, attempts to produce clouds by provisions of CCN have no hope of success in any practical sense, but provision of IN may produce artificial cirrus clouds in parts of the atmosphere where the vapor pressure is intermediate between that required for ice and water saturation. This possibility has been discussed at numerous workshops and meetings. It deserves closer study, particularly to determine how frequently the required vapor pressure for its implementation might exist during drought periods.

Another possibility for generating artificial cirrus clouds to suppress daytime temperatures would be the deliberate overseeding of supercooled convective clouds to convert them into artificial anvil clouds. The temperature contrasts set up at the boundary between regions shielded by thunderstorm anvils and nearby areas without cloud cover often cause mesoscale convergence lines, which favor the organization of further shower and thunderstorm development. It requires little imagination to see a number of possible ways to manipulate a cloud system, once one adopts as the objective the promotion of crop yields rather than the modification of a specific weather variable such as precipitation or temperature.

CHAPTER

IX

Impacts of Weather Modification on Society

9.1 INTRODUCTION

The material presented in the above chapters summarizes presently accepted ideas regarding the physical mechanisms through which cloud seeding modifies weather conditions, and the results attainable in pursuing specific operational objectives.

The fact that a certain operational objective may be attainable through weather modification is not in itself sufficient justification for launching an operational program. The implementation of a weather modification program can be a very complex issue.

Most operational weather modification programs are launched with the hope of securing economic and social benefits. For example, fog modification programs at airports are undertaken because of economic losses to airline companies and other parties if flight schedules are interrupted, and because the inconvenience of air travelers is seen as an undesirable thing. As the effects of the fog modification treatments are generally localized, there has not been much debate about the wisdom of employing them, provided there is a net economic benefit.

Projects designed to influence weather over substantial areas present far more complex questions. There may be persons involved who stand to lose from such operations as well as persons who stand to gain. Attempts to reconcile the conflicting interests of various groups have led to new social and political arrangements, and weather modification is likely to have even greater impacts in the future.

The varied impacts of weather modification on human society have been the subject of considerable research over the past two decades [e.g., Sewell (1966)]. This chapter summarizes some of the more important of the research efforts, which are organized according to whether they fall primarily into the fields of economics, ecology, sociology, or law and political science.

9.2 ECONOMIC CONSIDERATIONS

In view of the uncertainties which attend the application of weather modification by cloud seeding, one might inquire why it has been so widely adopted around the world. The answer, of course, lies in the large economic benefits that it offers. Although difficult to measure exactly, the perceived benefits far outweigh the cost of implementing programs.

The suppression of hail in the United States alone offers a potential saving of roughly $500,000,000 per year. Elimination of a single tropical hurricane could save as much as $1,000,000,000 in property and several thousand lives. In this connection it is sufficient to recall that the human death toll in a tropical cyclone in East Pakistan (now Bangladesh) in the fall of 1970 was variously estimated at from 100,000 to 300,000.

However, the economic studies which have been made to date show beyond any doubt that it is the augmentation of precipitation which has the greatest potential economic impact. In this connection one ordinarily thinks first of agriculture, but additional water has great economic value for hydroelectric power generation, the production of timber and pulp wood, and for many industrial applications.

We shall consider briefly two types of situations, first those in which additional water produced by cloud seeding is impounded and dispensed in a controlled manner for later use and, second, those in which additional rainfall is induced artificially for its direct impact upon crops.

Use of Weather Modification to Augment Managed Water Supplies

In many of the arid and semiarid parts of the world elaborate engineering facilities exist for the collection, storage, transportation, and distribution of water, which is usually collected as runoff from mountain watersheds.

Irrigation and other water management programs represent the ideal place for application of weather modification technology. As noted in Chapter VII, the seeding of winter storms over mountains to produce additional runoff is the type of precipitation augmentation whose effects can be most accurately predicted. The fact that water is produced during a wet season (usually winter) for use in a subsequent dry season, perhaps several years later, avoids the awkward problem of having no clouds to seed when the need is greatest. Although some of the additional snowpack or rainfall is lost by evaporation, evaporative losses are substantially reduced once the runoff is collected into reservoirs. In the case of the Santa Clara Valley Water Conservation District, which has sponsored cloud seeding since 1951, the water collected in reservoirs is passed through percolation beds to underground aquifers, where it is virtually immune to evaporative losses, but can be subsequently recovered by pumping.

In managed-water situations the economic value of the additional water produced by cloud seeding can be fairly readily calculated. Managed water is normally assigned a value equal to the cost of obtaining it elsewhere and transporting it to the region of ultimate use. For example, irrigation water in southern California is commonly assigned values in the range 5–10 cents per cubic meter, that being the approximate cost of bringing water from water surplus areas in northern California and elsewhere.[1] Those are conservative figures. Critics of large irrigation projects in the western United States have complained that the charges to the water users often do not reflect the project costs in full. Unsubsidized municipal water often costs one dollar or more per cubic meter.

As target–control regression analyses over many years have indicated that runoff from water sheds on the southwest side of the Sierra Nevada and similarly situated ones can be increased by 5–10% by AgI seeding, the size of watershed required for an economically viable program is only 1000 km^2 or so. Most of the operational programs in that region involve a single major stream and its tributaries. The Kings River above Pine Flat Dam, for example, encompasses some 4000 km^2 and has a mean annual flow exceeding 10^9 m^3. Statistical studies suggest increases in annual runoff of approximately $5–8 \times 10^7$ m^3 (Henderson, 1966), indicating that the seeding program which has operated in that watershed since 1954 has produced, on the average, several million dollars worth of additional water per year.

Several power companies have found it feasible to operate orographic projects solely for the sake of the additional hydropower. Projects in which additional water is used to generate hydroelectric power and then

[1] The English unit for irrigation water is the acre ft. 1 acre ft = 1233 m^3.

for irrigation are especially attractive. It should be emphasized that additional hydropower can be generated by snowfall redistributions as well as by absolute increases in precipitation. Reducing the terminal speeds of snowflakes by seeding or mild overseeding has already been shown to be feasible (Chap. IV and VII). Seeding on a windward slope might result in less precipitation at low elevations but increase the snowpack at higher elevations, above hydroelectric installations for example.

The other examples of snowfall redistribution mentioned in Chapter VII also have potential economic impacts. Shifting snowpack eastward over the Cascade Mountains is one possible way to move water into the dry, eastern parts of Washington and Oregon with a minimal expenditure of energy. The Great Lakes Project was motivated by the fact that large cities in the lee of the Great Lakes spend millions of dollars each year on snow removal, to say nothing of the personal inconvenience and losses of industrial production associated with snowstorms off the lakes.

Effects of Rainfall Increases on Crop Yields in Dry Farming Operations

There are many places in the world where irrigation is not available but where crop yields are limited by available rainfall. Large areas of the Great Plains of North America and of the grain producing areas of Africa, Australia, and the U.S.S.R. fall into this category. Cloud seeding has been widely used in an attempt to stimulate crop production in these regions. Much of the summer rainfall in these regions falls from convective clouds, and as we have seen, the technology for increasing precipitation from convective clouds is not as well established as that for increasing orographic precipitation. Nevertheless, the importance of the agricultural production of such areas has led to extensive studies of the potential benefit of increased precipitation.

Studies of the crop increases that might be obtained by precipitation augmentation in dry farming areas have generally made use of previously developed crop-yield models. Crop-yield models have been developed by agricultural economists in order to study such questions as the value of agricultural land as a function of rainfall, and to predict the potential impact of irrigation developments.

The simplest crop-yield models make use of seasonal values of a few meteorological variables, usually beginning with rainfall and temperature, to predict yields. Much more sophisticated models have been developed in recent years using multilinear regression analyses and other statistical techniques to take account of the distribution of the meteorological elements and the ways in which they interact with one another to affect crops. Some research groups now have crop yield models which can utilize as input daily observations of soil moisture, temperature, solar radiation, etc.

Some of the earliest predictions of the economic effects of weather modification in dry farming areas simply took note of the fact that cloud seeding might increase growing season rainfall by 10–20%, converted that to depth of precipitation, and applied the simplest crop-yield models based on annual data. The answers were useful, if only because they forced attention to the tremendous potential impact of weather modification.

Ramirez (1974) reported that every millimeter of additional rainfall during the growing season in western North Dakota is associated with an additional yield of 9 kg of hay per hectare, 7 kg of wheat per hectare, and so on.[2] By applying such multipliers to the total area farmed in North Dakota, one can estimate that gross farm income in North Dakota would be increased by $400,000,000 per year by a 10% increase in growing season rainfall.

The analysis by Ramirez (1974) is admittedly oversimplified. A more detailed study by Huff and Changnon (1972) attempted to determine how the economic benefits of weather modification would vary from year to year in Illinois. They concluded that corn and soybean yields would be improved by additional rainfall in a majority of the situations studied.

The relationships observed between crop yields and rainfall in past years might not continue to exist in the presence of a cloud seeding program. A year with natural rainfall equivalent to 90% of normal and the deficiency made up by cloud seeding is not exactly the same as a year with natural rainfall equal to 100% of normal. It is well known too that crop yields are influenced by distribution of rainfall throughout the growing season and its relationship with other meteorological variables.

Difficulties of the type just mentioned can be attacked with more precise models in which monthly or even daily rainfall amounts are entered along with other weather elements to predict the crop yields. For example a recent study indicates that Illinois corn yields usually would not be increased by additional rainfall during the month of June, but that additional rainfall during a typical July or August would increase the yields. These more complex programs also permit the modeling of the responses to seeding in more realistic ways. As noted in Chapter VII, seeding convective clouds does not increase rainfall from every cloud system contribut-

[2] Corresponding English units: 200 lb of hay and 2.7 bushels of wheat per acre per inch of additional rainfall.

ing rainfall at a given point. Questions to be posed for the computer simulations should therefore resemble the following: Would the occurrence of three additional showers of 20 mm each timed at random between June 1 and July 31 have a positive impact on corn yields in Iowa? What if the timing could be controlled to within one or two days?

The possibility of simulating crop response to augmented precipitation in such detail also offers the possibility of programs in which operations would be limited to situations where additional precipitation would actually be of benefit to the affected crops. Obviously, such fine tuning of a program would provide for the suspense of operations on the basis of existing soil moisture conditions or of forecasts of sufficient rainfall without cloud seeding, provisions which even today are features of well managed programs in dry farming areas.

Taking as many as possible of the above factors into account, the Agricultural Experiment Station (1978) of Kansas State University has projected yield increases for a number of crops in different parts of Kansas due to an assumed "precipitation alteration scheme." As an example, they estimated that wheat production would be increased by over 100 kg per hectare (2 bushels per acre) per year in western Kansas. The impact on gross agricultural receipts in western Kansas would approach $100,000,000 per year.

Economic Impact of Hail Suppression

The principal motivation for the adoption of hail suppression programs around the world has been the damage which hail causes to agricultural crops. It has been estimated that hail damage to crops in the United States averages close to $700,000,000 per year [e.g., Borland (1977, p. 160)]. Much of the crop damage is concentrated in "hail alley," a strip of the High Plains stretching from Texas to Montana and North Dakota at elevations between 1300 and 2000 m. In parts of this region, crop hail insurance rates run as high as $25 per $100 of insured value.

There are many places in the world with hail problems comparable to or exceeding those in hail alley. Regions downwind of the Andes Mountains in Argentina, downwind of the Alps in northern Italy, and near the Caucasus Mountains in the southern U.S.S.R. are examples of regions with severe hail. Economic losses in some of these regions are even worse than in hail alley because of the more valuable crops affected. Orchards and vineyards are involved; hail damage to fruit trees and vines can extend the loss from one severe storm to several seasons.

The value of crops destroyed is not an accurate measure of total loss to hail, because the *threat* of hail undoubtedly causes farmers to choose less than optimum production strategies. For example, a farmer may harvest a wheat crop a few days before what would otherwise be regarded as the ideal time, simply because every day the crop is left standing increases the chance of damage by hail. Furthermore, hailstorms introduce additional variance into farm incomes, which is itself an undesirable effect.

The need to spread hail losses accounts for the acceptance of hail insurance programs. Even though hail insurance programs must cost as much in the long run as the hail damage in an area, along with an appropriate markup for administrative and overhead expenses, hail insurance is nevertheless seen as desirable by most farmers in hailprone regions. Exceptions are found in areas of extreme hail loss, however, where premiums sometimes run to 30–40% of insured crop value. In such areas, some farmers insure only enough of their crop to ensure that they will have enough money to plant another year, and leave the remainder of their crop at risk.

It is obvious from the above that successful hail suppression programs could produce large economic benefits. Operators of the hail suppression projects of the U.S.S.R. estimate on the basis of their target–control analyses that annual economic benefits amount to tens of millions of rubles[3] (Sulkavelidze *et al.*, 1974, p. 410). However, the effects of changes in rainfall on crop yields are so important that a hail suppression program rapidly loses its economic attractiveness if it simultaneously reduces rainfall. With the exception of seeding to suppress hail just before or during harvest, a hail suppression program is likely to produce economic benefit only if rainfall can be left unchanged or slightly increased. This question of combined effects has been studied at length in a technological assessment of hail suppression (TASH) carried out for the National Science Foundation [e.g., Changnon *et al.* (1977); Farhar *et al.* (1977)]. The study considered sociological and legal factors as well as economic ones. The economic studies are of a particular interest because the investigators attempted to assign quantitative values to the benefits that could be obtained by hail suppression assuming that the program had certain specified effects upon rainfall.

The results indicated that the regions of the United States where hail suppression is likely to be adopted are quite limited. Figure 9.1a shows as an example the estimated area in the western United States in which hail suppression would likely be adopted by 1985 if a 40% hail suppression effect could be accomplished simultaneously with an 8% increase in grow-

[3] 1 ruble = $1.50 U.S.

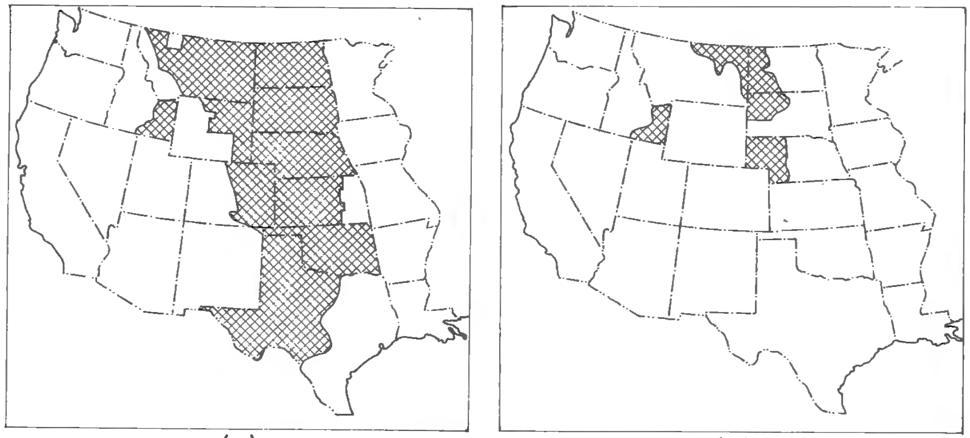


Fig. 9.1. Areas of the western United States in which cloud seeding for hail suppression would likely be adopted under two different sets of assumptions. (a) By 1985, assuming a 40% hail suppression effect and an associated 8% increase in rainfall. (b) By 1995, assuming a 30% hail suppression effect and no effect upon rainfall. Adoption analysis considered scientific, economic, legal, and social factors. [After B. C. Farhar *et al.* (1977). Hail Suppression and Society. 25 pp., Rep., Illinois State Water Survey, Urbana, Illinois, by permission of Illinois State Water Survey.]

ing season rainfall. Figure 9.1b shows the areas where hail suppression would likely be adopted by 1995 if a 30% suppression effect were accompanied by no change in rainfall. In the latter case, only regions with extremely severe hail problems would be likely to adopt hail suppression.

Distinction between Gross and Net Benefits

The discussion to this point has focused on the gross economic benefits that might be obtained from weather modification.

In calculating a potential net benefit from weather modification, it is obvious that the direct cost of the operations must be deducted. The indications are that this would not be a serious obstacle to the implementation of programs to increase precipitation or suppress hail, provided of course that the desired effects could be obtained. State wide programs in both South Dakota and North Dakota have been funded at roughly $1,000,000 per year, and evidently could be extended to cover all counties in either state for less than $2,000,000 per year. This is not a formidable obstacle to a program with suggested gross benefits of over $100,000,000 per year per state. However, the cost of operations is a serious factor in some other types of programs, like fog modification, where the potential economic return is much smaller.

There are substantial indirect costs associated with weather modification programs. One of them is the cost of processing the additional crops. An increase in the wheat crop in one state of, say, 500,000,000 kg imposes additional costs for harvesting, storage, and transportation. Over many seasons, the farmers might find it necessary to provide more fertilizer to maintain the higher level of production.

The most important factor limiting the producer's benefits from weather modification is likely to be its impact on prices. Prices of farm products are elastic and even volatile, with changes of only a few percent in supply driving prices up or down by much larger proportions. There is little doubt that the potential increase of 5% in the total U.S. soybean crop foreseen by the Council of Renewable Resources in 1976 as a possible consequence of cloud seeding would depress prices. Price drops due to weather modification would let the benefits of the applied technology slip from the farmers' pockets to those of food processors or consumers (Borland, 1977, pp. 166-167).

The ideal situation, from the point of view of an individual producer, is for him to use a new technology while other producers do not, so that there is no perceptible impact on prices. While it is not feasible for a single producer to have a weather modification program, groups of producers might do so, so it is appropriate to consider local and regional impacts. For example, if it turned out that weather modification could increase rainfall in the northern plains but not in the southeast U.S., the economy of the northern plains would obtain some advantage at the expense of the southeast. The Kansas study referred to earlier (Agricultural Experiment Station, 1978) found that the western part of Kansas might gain at the expense of eastern Kansas if a successful rain stimulation program were instituted statewide.

Regional competition on the U.S. farm scene is not new. The development of irrigated agriculture in the west around the turn of the century certainly hurt the agricultural economies of many states in the east, even while consumers in eastern cities were benefiting from lower prices for vegetables. In those days there were no computers and economic models which permitted any estimate in advance of what the total impact of such changes would be, but they have been shown in retrospect to have been very large.

Work on the impact of weather modification on regional economies and on the country as a whole with sophisticated econometric models is just getting under way.[4] The models themselves are still in the development stage and can give conflicting output, but their use is likely preferable to widespread adoption of the new technology with no guidance whatever concerning its ultimate economic impact.

[4] An early and noteworthy study at Stanford Research Institute (now SRI International) on the impact of a project to increase streamflow on the Colorado River is described by Weisbecker (1972).

9.3 ECOLOGICAL CONSIDERATIONS

The environmental aspects of technological change have been coming under increasingly close scrutiny in the United States during the past 20 years. It is not surprising that weather modification, which deliberately sets out to modify the environment over considerable areas, should be examined very closely in this regard.

In considering the environmental impact of weather modification, it is necessary to distinguish between the incidental changes due to side effects of the seeding agents dispersed into the atmosphere, and the changes in weather or climate actually produced.

Side Effects of Seeding Agents

Most of the concern about seeding agents has centered about the release of AgI. The other seeding agents have been used less widely. In addition, most of them are soluble compounds like NaCl or urea, or biodegradable organic compounds such as 1,5 dihydroxynapthalene that do not persist in the atmosphere. Dry ice quickly vaporizes into CO_2 gas, although some concern has been expressed over the fact that commercial dry ice is sometimes contaminated with other gases. Lead iodide, widely used in the U.S.S.R. for hail suppression, has never been used much in the United States. The concerns about damage to the ecology and to human health therefore come down, in this country, to a question about the effects of AgI on the environment.

Investigations into the possible side effects of AgI have covered such points as the total amounts released per year, the AgI concentrations in rain or snow falling from seeded storms, the fate of the AgI in streams and in the soil, and its degree of toxicity.

Environmental research sponsored under Project Skywater has established the general outlines of silver transport in the atmosphere, in water, and in the soil as a result of cloud seeding projects [e.g., Howell (1977)]. The total amount of AgI released to the atmosphere by cloud seeders in the United States in 1977 was 1512 kg, according to reports filed with the Department of Commerce (Charak, 1978). To put this number in perspective, one can note that industry apparently discharges somewhere between 135 and 360 Mg of silver into the atmosphere over the United States each year (Environmental Protection Agency, 1973).

The concentrations of AgI in rainwater and snow from seeded storms are of the order of 10^{-12}, that is, well under 1 part per billion (1 ppb). As AgI is not soluble in water, it is not readily washed into the sea, but tends to be deposited in the soil and on the bottom of stream beds.

Early studies of the toxicity of AgI were mainly literature searches. It was noted that the substance is not toxic in the ordinary sense.

A number of laboratory investigations of the effects of AgI on organisms have been made in the last few years. These experiments are in principle like the now famous experiment in Canada in which rats were fed saccharin with their food in concentrations much larger than would ever occur in an ordinary diet. The analogies in the AgI investigations include aquaria filled with saturated solutions of AgI which serve as homes to lowly forms of plant life, and grasshoppers fed on vegetation grown in soil saturated with AgI or other silver compounds. The results up to 1977 were generally reassuring, with no evidence that AgI was interfering with biological processes [e.g., Klein (1977)].

Ecological Impacts of Weather Changes

The effects of changes in the weather upon plant and animal species are likely to be more serious than the side effects of seeding agents.

It is sometimes argued that because the effects produced by weather modification are within the range of natural weather variations, there will be no long-term effects on plants and animals. This argument is not correct. Plant and animal species react in subtle ways to the weather. Sometimes the mean annual rainfall is a controlling factor; sometimes the driest year out of a decade or more is the factor limiting the spread of a species; sometimes weather factors act in unison to produce their impacts. A working weather modification program *can* alter the weather statistics and, therefore, the local ecology (Cooper, 1975).

As an example of studies of these questions, we may note those conducted under sponsorship of the U.S. Department of the Interior on possible impacts of weather modification on plant and animal life in the Rocky Mountains (Howell, 1977). Some of the work indicates that the principal impact of a program to increase snowpack would be the delay in the melting of snow cover in the spring. An increase of 10% in snowpack could easily delay the exposure of the ground by one or two weeks in the spring, which certainly would have an impact on the appearance of vegetation and on the animals dependent on that vegetation. Whether the native species, such as the elk, could successfully adjust to the change by staying at lower altitudes until the high mountain meadows became snow-free is a subject for more study.

While stressing the importance of these questions, we note that they all involve some value judgments. We may prefer eagles and elk to magpies and porcupines, but there is no objective way to determine an intrinsic superiority of any species over another. Therefore, it is not possible to say unequivocally that changes in the flora and fauna that might accompany cloud seeding are good or bad.

Most ecologists now agree that a static ecosystem is not necessarily superior to a dynamic one, and that many natural ecosystems are evolving. Uninformed people sometimes speak of the balance of nature as if (1) it were established long ago, but (2) is in a precarious state, and any disturbance would have catastrophic consequences. As a matter of fact, the composition of species in an area is usually in a state of flux due to natural fluctuations in climate, the erosion or deposition of soils, and unexpected changes arising out of the extremely complex ways species interact with one aother. Epidemics due to overcrowding or to the arising of new strains of viruses, forest fires set by lightning, and even volcanic eruptions are among the factors which ensure that change will remain the order of the day, with or without intervention by human beings and their technology.

In considering ecological effects, a systems approach is definitely in order. Consider, for example, the case of seeding orographic clouds to obtain runoff for power generation. This has some ecological effects. In judging them, one is helped to retain perspective by noting that the alternatives might include additional mining of coal or the construction of power lines to import power from another region. In agriculture, the ecological disturbances due to cloud seeding should be compared to those involved in production and application of fertilizer, aerial application of insecticides, and the other methods which might be used to increase food production.

9.4 SOCIOLOGICAL CONSIDERATIONS

Even in situations where weather modification offers potential economic benefits and where there are no overriding ecological difficulties, the attitudes of individual persons toward the adoption of the new technology vary greatly.

Perhaps 5 or 10% of the population (the innovators) are highly receptive to almost any proposed change. At the opposite extreme are persons who would prefer to see no technical changes whatever. The great majority of the population occupies an intermediate position, neither actively promoting introduction of new technology nor actively opposing it.

Sociologists speak of persons introducing new technology as "change agents." In the case of weather modification the change agents may be representatives of cloud seeding firms, university research workers, or government officials charged with implementing or regulating weather modification programs. The acceptance of a program may depend as much on the extent to which the members of the user groups and the general public identify with the change agents as on the objective data on which the acceptance is supposedly based (Johnson and Farhar, 1977).

Farhar (1975) noted, following earlier writers in sociology, that the opinions of many people about cloud seeding are shaped by "influentials," who might also be called "opinion makers." The opinion makers tend to be highly successful farmers and ranchers, local businessmen, county commissioners, and others who are perceived by their neighbors as possessing superior knowledge about new technical developments. These people act as an interface between the general mass of users and the change agents referred to above.

Social scientists have conducted numerous surveys of attitudes toward weather modification programs and attempted to correlate them with the sex, age, education level, job status, and religious beliefs of the respondents. Among the most extensive surveys are those done in connection with the National Hail Research Experiment in Colorado (e.g., Borland, 1977), and the operational weather modification program in South Dakota [e.g., Farhar (1974, 1975)]. A sample of about 300 South Dakota respondents was drawn up in 1972 prior to the adoption of extensive operational cloud seeding in the state and the opinions of those persons were sampled several times over the next few years (Table 9.1). Events which transpired during the total sampling period included the in-

TABLE 9.1

Changes in Opinions about Efficacy of Cloud Seeding during Early Years of South Dakota Operational Program[a]

	Spring 1972	Fall 1972	1973
Do you think cloud seeding can actually increase moisture?			
No	13 (%)	11 (%)	14 (%)
Don't know	39	15	17
Yes	48	74	69
Number of respondents:	435	368	326
Do you think cloud seeding can actually suppress hail?			
No	14 (%)	18 (%)	19 (%)
Don't know	67	43	29
Yes	19	39	52
Number of respondents:	435	367	326

[a] After Farhar (1975), by permission of American Meteorological Society and the author.

auguration of the state's operational cloud seeding program, the Rapid City flood of 1972, severe drought in 1973 and 1974, and the termination of state support for operational cloud seeding in 1976.

The South Dakota program was initially accepted on the basis that the program had the endorsement of the scientific community and that it would lead to demonstrable economic benefits. When the failure of the cloud seeding program to avert drought in the late summers of 1973 and 1974 was combined with a growing perception that the scientific community was divided concerning the effectiveness of such programs, opposition to the program increased. Although polls suggest that a majority of the people of South Dakota would have voted for continuation of the program through 1976, the pressure exerted on the state legislature by the members of the population violently opposed to its continuation overwhelmed the efforts of its supporters (Mewes, 1977a).

Among the interesting points uncovered by the studies of Farhar and others were the finding that religious beliefs were effective in shaping opinions about weather modification for only a very small minority of the population, and that the attitude of persons in South Dakota toward weather modification was not changed appreciably by the occurrence of the Rapid City flood in 1972, despite widespread publicity given by the news media to the experimental cloud seeding flights that took place on the day of the flood (Farhar, 1975).

One situation that inevitably leads to difficulty in the application of weather modification technology is the division of the population of a region along economic lines such that certain groups can be identified (rightly or wrongly) as losers in a weather modification program (Haas, 1973). For example, the Blue Ridge Weather Modification Program, which was initiated in 1957 by fruit growers in the Shenandoah Valley and adjacent parts of Virginia, West Virginia, Maryland, and Pennsylvania to suppress hail damage to orchards, was ended about 1964 as the result of fears on the part of farmers in the same area that the program was reducing rainfall (Howell, 1965; Mewes, 1977b). A program in the San Luis Valley of southern Colorado foundered on differences between the sponsors, who were farmers under contract to grow barley for a brewery, and neighboring ranchers, and on a loss of confidence in the scientific merits of the program (Mewes and Farhar, 1977).

Another example of divided interests concerns the projects to stimulate snowfall in the upper reaches of the Colorado River Basin for the benefit of irrigation interests downstream. Fear of avalanches on the part of residents in the vicinity of Ouray, Colorado, led to protests about the program. The protestors were mollified by the institution of rules calling for suspension of operations during periods with over normal snowpacks in the target regions or during heavy snows with avalanches predicted as likely.

9.5 LEGAL CONSIDERATIONS

It is inevitable that a subject which impinges upon so many persons as does weather modification would give rise to legal problems. These exist on all scales from local to international.

The legal aspects of weather modification were the subject of a conference sponsored by the American Bar Association at Duke University in 1976. The proceedings volume of that conference provides a useful, albeit disjointed, introduction to the subject (Thomas, 1977). Useful papers have also been contributed [e.g., Davis (1974)].

Legal action with regard to weather modification activities has taken the forms of (1) court cases and (2) legislation or regulation.

Court Cases

Most of the court cases have been filed by opponents of weather modification programs seeking to enjoin further activities or to recover for alleged damages. The bases on which suits have been brought against weather modification operators include the concepts of (1) trespass, (2) public nuisance, (3) negligence, (4) the concept that a land owner has a riparian right to the rainwater that nature would bring to his land, and (5) strict liability [e.g., Davis (1974)].

The concept of strict liability assumes that weather modification is intrinsically an ultrahazardous activity. In law it is commonly held that the practitioners of an ultrahazardous activity are strictly liable for all damage resulting from that activity, even in cases where negligence has not been proven. It is small wonder, then, that weather modification operators have sought to have included in state statutes dealing with weather modification the explicit statement that weather modification is *not* an ultrahazardous activity. They have succeeded in a few cases (Davis, 1974).

In order to recover damages under any of the concepts mentioned above, a plaintiff must show that he has suffered financial losses. Suits against weather modifiers have been almost uniformly unsuccessful. It is sometimes stated that this is because no one can prove that weather modification has *any* effect. While this statement may have an element of truth, the defendants have generally not taken refuge behind such a defense. Court records show that in most cases the defendants have insisted that their cloud seeding activities had effects, but have argued in detail that the effects of their seeding were other than claimed by the plaintiffs [e.g., Mann (1968); Davis and St.-Amand (1975)].

Legislative and Regulatory Actions

Most of the legal activity with regard to weather modification programs in recent years has had to do with the passage of legislation, rather than court cases. About 30 of the 50 states now have some form of regulation of weather modification activities. Many of the state codes incorporate parts of a model law prepared by the Weather Modification Association.[5] Common elements in the state laws include requirements for licensing of weather modification operators, permits for the conduct of specific projects, and the conduct of public hearings prior to the issuance of permits. A tabulation of the key elements contained in the state laws as of late 1974 has been made by Farhar and Mewes (1975) and is shown as Table 9.2.

The state laws range from laws that favor weather modification by providing taxing authority for weather modification districts to laws which are so restrictive as to constitute an effective ban on weather modification.

A common feature of many state laws or regulations is the requirement that a public hearing be held in the affected area before a permit is issued for the conduct of a project. This requirement gives a voice to local residents other than the sponsors of the project. In 1972 it resulted in denial of a permit for an operation in the San Luis Valley of Colorado, where cloud seeding has been an issue of contention for several years (Davis, 1974).

Up to this time there have been no federal regulations to control weather modification research or operations. However, federal reporting requirements have existed intermittently since the Advisory Committee on Weather Control was authorized to collect reports from operators in the mid-1950s. At the present time the agency empowered to collect and compile reports is the National Oceanic and Atmospheric Administration [e.g., Charak (1978)].

[5] The model law appears in *J. Weather Modification* **2**, 221–224 (1970).

Weather Modification and International Law

A number of legal scholars have written scenarios of possible interactions between nations arising from the application of weather modification technology [e.g., Weiss (1972)]. While dramatic confrontations have not actually arisen yet, it is possible that weather modification progress in one country will become the subject of controversy with neighbors. This would seem particularly likely in the case of programs operated by very small countries in water deficient regions. Nevertheless, the fact that the tentative indications now available point to rainfall increases rather than decreases downwind of such projects may reduce the probability of serious confrontations.

Most suggestions of geophysical warfare, such as the steering of hurricanes into unfriendly nations to disrupt their economies, have no realistic prospects for implementation for the time being. Nevertheless, possible changes in the intensity or tracks of tropical hurricanes are of such consequence that several countries have expressed concern over Project Stormfury. A plan developed about 1972 to move Stormfury to the western Pacific Ocean was abandoned because of opposition from China and Japan.

Attempts by the U.S. military to use weather modification to disrupt transportation systems of the Viet Cong forces in Laos and Vietnam during the Vietnam war are now well known. Public revulsion led to the passage of legislation by the United States Congress in 1974 prohibiting the use of weather modification and other forms of geophysical warfare. An international resolution initiated by the United States and the U.S.S.R. in 1974 prohibiting the hostile use of environmental modification techniques was approved by the United Nations General Assembly in 1976. It will take effect if and when 20 countries ratify it. The Weather Modification Advisory Board (1978) recommended that the United States ratify the resolution.

9.6 CONCLUDING REMARKS

It is symbolic that a story which began with a scientist dropping dry ice from a light aircraft in 1946 should end with a mention of a resolution passed by the U.N. General Assembly in 1976.

Articles have appeared in the *Journal of Weather Modification* and elsewhere expressing impatience with the slow progress of weather modi-

TABLE 9.2

Selected Characteristics of State Weather Modification Statutes[a]

State	Legislative Characteristics 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
Ariz.	f	no	—	no	no	yes	no	—	no	yes	yes	abc	yes	no	no	no	no	no	no	no
Calif.	b	no	—	no	no	yes	no	—	no	yes	yes	b	yes	yes	no	yes	no	no	no	no
Colo.	a	yes	yes	no	no	yes	yes	no	yes	yes	yes	ac	yes	no	no	no	a	no	yes	yes
Conn.	d	yes	no	no	no	no	no	—	no	no	no	—	yes	no	no	no	no	yes	no	no
Fla.	a	no	—	no	no	yes	yes	yes	no	yes	yes	abc	no	no	yes	yes	no	no	no	no
Hawaii	a	no	—	no	no	no	no	—	no	no	no	—	no	no	no	no	no	no	no	no
Idaho	c	no	—	no	no	no	no	—	no	yes	no	—	no	no	no	no	no	no	no	no
Ill.	f	yes	yes	no	no	yes	yes	no	yes	yes	yes	b	yes	no	yes	no	b	yes	no	yes
Iowa	d	yes	no	no	yes	no	no	—	no	no	no	—	no	no	no	no	no	no	no	no
Kan.	b	yes	yes	no	no	yes	yes	yes	yes	yes	yes	b	yes	yes	yes	yes	a	no	no	no
La.	c	no	—	no	no	yes	no	—	no	no	yes	a	no	no	no	no	no	no	yes	no
Mass.	d	no	—	no	no	yes	no	—	no	no	yes	a	no	no	yes	no	a	no	no	no
Minn.	f	no	—	no	yes	no	no	—	no	no	no	—	no	no	no	no	no	no	no	no
Mont.	a	yes	no	no	no	yes	yes	no	yes	yes	yes	b	yes	yes	yes	yes	b	no	no	no
Neb.	c	no	—	yes	yes	yes	no	—	no	no	yes	a	yes	no	no	no	no	yes	yes	no
Nev.	a	yes	no	no	no	yes	yes	yes	yes	yes	yes	b	yes	no	yes	yes	no	no	no	no
N. H.	f	no	—	no	yes	no	no	—	no	no	no	—	no	no	no	no	no	yes	no	no
N. M.	d	no	—	no	no	yes	yes	no	no	yes	yes	b	yes	no	no	no	no	no	yes	no
N. Y.[b]	f	no	—	no	yes	no	no	—	no	no	no	—	no	no	no	no	yes	no	no	no
N. Dak.	e	no	—	yes	yes	yes	yes	no	no	yes	yes	no	no	no	no	no	no	yes	yes	no
Okla.	b	yes	no	yes	yes	yes	yes	no	yes	yes	yes	abc	yes	no	no	yes	no	yes	no	yes
Ore.	c	no	—	yes	yes	yes	yes	yes	no	yes	yes	b	yes	no	yes	yes	no	no	no	no
Penna.	c	yes	yes	no	no	yes	yes	yes	no	yes	yes	abc	yes	yes	yes	yes	no	yes	no	no
S. Dak.	a	yes	yes	yes	yes	yes	yes	no	yes	no	yes	abc	yes	no	yes	yes	no	yes	yes	no
Texas	b	yes	no	no	no	yes	yes	no	yes	yes	yes	abc	yes	no	yes	yes	no	yes	no	no
Utah	b	yes	no	no	no	yes	yes	no	yes	yes	yes	b	yes	no	yes	yes	b	yes	yes	no
Wash.	a	yes	no	no	no	yes	yes	yes	yes	yes	yes	b	yes	yes	yes	yes	a	yes	no	no
W. Va.	e	no	—	no	no	yes	yes	yes	no	yes	yes	abc	yes	no	yes	yes	no	no	no	no
Wis.	f	no	—	no	no	no	no	—	no	yes	yes	a	no	no	no	no	no	no	no	no
Wyo.	d	no	—	no	no	no	no	—	yes	yes	yes	ac	yes	no	no	no	no	no	yes	no

1. Administering agency
   a. Dept. of Natural Resources/Conservation
   b. Dept. of Water Resources
   c. Dept. of Agriculture
   d. Special Weather Control Boards
   e. Aeronautics Commission
   f. Other
2. Advisory committee provision?
3. Advisory committee membership specified?
4. Weather modification district?
5. Tax provision?
6. License required?
7. Proof of financial responsibility required?
8. Amount of financial responsibility specified?
9. Permit required?
10. Reporting required of operators?
11. Sanctions for violations specified?
12. Punishment specified?
    a. fine
    b. revocation of license or permit
    c. jail sentence
13. Safeguards for public safety, etc.?
14. EIS required? (not full scale)
15. Provision for emergency operations?
16. Publication of notice of intent required?
17. Public hearing provision?
    a. required
    b. optional
18. Intergovernmental cooperation?
19. State sovereignty over atmospheric water?
20. Prevention of interproject contamination?

[a] After Farhar and Mewes (1975) by permission of American Meteorological Society and senior author.

[b] In our attempt to establish the existence of weather modification legislation for the State of New York, we called the State Attorney General's Office and a state legislator who had proposed weather modification legislation which failed to pass. Neither of these sources knew of a state law regulating weather modification. However, as this paper was going to press, a copy of the State's General Municipal Law, Sec. 119-p, regarding "Projects Relating to the Use of Atmospheric Water Resources" arrived in the mail from the Cornell Law Library. This statute, in essence, authorizes municipal cooperations to "conduct or engage in projects, experiments, and other activities designed to develop the use of atmospheric water resources, and to make scientific evaluations of such projects, experiments, and other activities, or to contract therefore, and to appropriate and expend moneys therefore." According to the criteria outlined in this paper, this law is noncomprehensive in nature.

fication during that 30 year period. However, the effects of the original cloud seeding experiments have been felt in almost every country of the world during those 30 years. The fact that the United States and other countries are still groping for development of national policies on weather modification simply attests to the complexity of the issues involved. Not only must scientific uncertainty regarding the effects of seeding be reduced, but equitable mechanisms must be established to balance the interests of groups within one country and in different countries.

Despite the hopeful stance of the Weather Modification Advisory Board's 1978 report, the present author is not convinced that an expansion of research support by the United States government would solve the outstanding problems in 20 years. It is unlikely that the research results would reduce the uncertainties sufficiently so that decisions on whether or not to proceed with weather modification programs would be risk free. One must anticipate that weather modification technology will evolve, if at all, in a setting of imperfect knowledge and clumsy social and legal arrangements.

Over the long run, there can be little doubt that human beings will exploit their increasing knowledge in all branches of the physical and biological sciences to increase their well being as they perceive it. It would be very surprising if they made an exception in the case of weather modification.

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