CHAPTER I

FORMATION OF CLOUDS

1.1 Introduction

Clouds form wherever air is cooled below its dew point. The discussion in this report is restricted to warm clouds whose tops do not extend to temperatures colder than 0°C. The microphysical processes of these clouds exclude ice phase. Only condensation and coalescence processes provide means for growth of droplets in these clouds.

1.2 Microphysics of a Cloud Droplet

The amount of water vapour which can exist in a given volume in equilibrium with a plane surface of pure water is a function of temperature. Air containing this amount of water vapour is said to be saturated. As water vapour closely obeys the ideal gas laws, it is convenient to express the amount present in terms of the pressure exerted by it (e) and of the saturation equilibrium vapour pressure, $e_s(T)$.

Any water vapour in excess of the amount required for saturation is theoretically available for formation of a water cloud. The rate at which water vapour 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 vapour pressure to T, the temperature in degrees Kelvin, as follows (Dennis, 1980).

$$e_s(T) = e_s(T_o)\exp\left[\frac{L_v}{R_w T_o} - \frac{L_v}{R_w T}\right] \tag{1.1}$$

Where $e_s(T)$ is the saturation or equilibrium vapour pressure at the temperature T. $e_s(T_o)$ is the saturation vapour pressure at some reference temperature T_o. L_w is the latent heat of vapourisation, and R_w is the specific gas constant for water vapour. The above equation 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.

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

$$S = \frac{e}{e_s(T)} \tag{1.2}$$

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

1.3 Surface Tension (Kelvin) Effect

Energy is stored in water surfaces because of the effects of surface tension. The energy stored in the surface of water droplet of diameter d is

$$E_{\nu} = \pi d^2 \nu \tag{1.3}$$

Where ν, 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 vapour 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) vapour pressure $e_{s,d}(T)$ above a pure water droplet of diameter d is given as

$$e_{s,d}(T) = e_s \exp\left[\frac{4\nu}{\rho_l R_w T d}\right] \tag{1.4}$$

where e_s is the saturation vapour pressure above a plane surface of pure water, ν is the surface tension coefficient, ρ_l is the density of liquid water, R_w is the specific gas constant for water vapour and the other symbols used are as previously defined.

1.4 Role of Dissolved Solute

Consideration of (1.4) 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 vapour state requires very large supersaturation ratios. Only for values of S around 4.5 (150% supersaturation) do the droplet embryos resulting from chance assemblages of water molecules and large enough to grow further by capture of additional moisture have an appreciable probability of appearing in a reasonable period of time, say 1 sec (Pruppacher and Klett, 1978).That is why clouds normally form by heterogeneous nucleation upon cloud condensation nuclei (CCN). The role of CCN is to overcome the energy barrier that must be surmounted to form new cloud droplets.

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

$$\frac{(e_s - e_s')}{e_s} = \frac{\eta}{(\eta + \eta')} \tag{1.5}$$

where e_s' is the saturation vapour pressure over the solution, η is the number of molecules of dissolved solute, and η' 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 vapour pressure over a solution droplet of diameter d as

$$e'_{s,d} = e_s\left[1 + \frac{4\nu}{\rho_l R_w T d} - \frac{6i M_w m_s}{\pi \rho_l M_s d^3}\right] \tag{1.6}$$

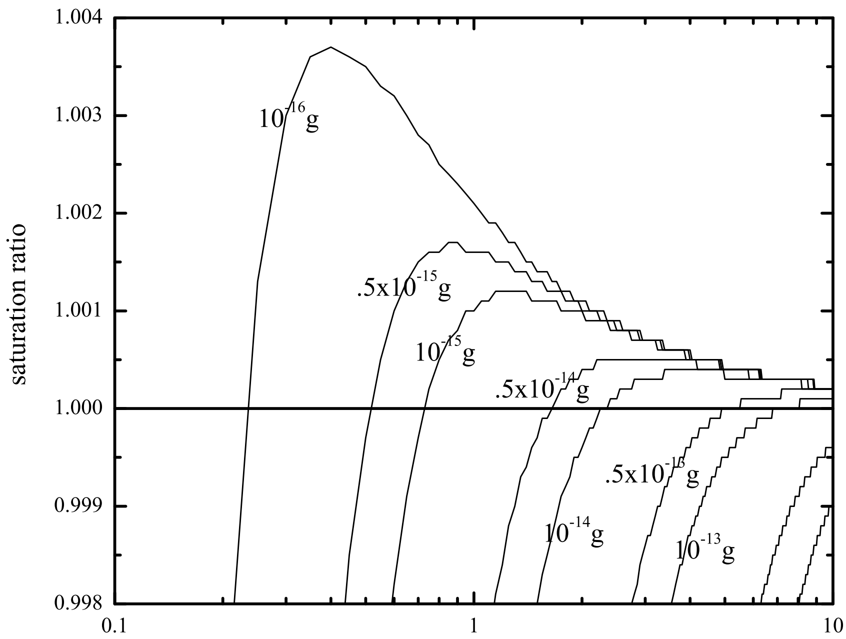
Where M_w is the molecular weight of water, M_s is the molecular weight of solute, m_s is the mass of solute, i is the dissociation factor, and the other symbols are as previously defined.

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

$$S_{eq} = \frac{e'_{s,d}}{e_s} = \left[1 + \frac{4\nu}{\rho_l R_w T d} - \frac{6i M_w m_s}{\pi \rho_l M_s d^3}\right] \tag{1.7}$$

Solutions to (1.7) for various sized NaCl particles are shown in Fig. 1.1. Each curve shows the equilibrium saturation ratio as a function of d for one size of NaCl particles. Such curves are called Kohler curves.

Kohler curves



drop diameter um

solute(NaCl) mass in grams is given for each curve

(Figure 1.1)

Figure 1.1 shows that for each nucleus there exists a critical diameter for which the supersaturation required to prevent the droplet from evaporation 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.

1.5 Droplet Growth Rate Equation

In considering the growth or evaporation of cloud droplets, it is necessary to consider not only the saturation or equilibrium vapour 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 vapour pressure in the 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 vapour diffuses to the droplet surface and the rate at which the latent heat of vapourisation released at the droplet surface is conducted away from the droplet.

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

$$\frac{\mathrm{d}d}{\mathrm{d}t} = \frac{4}{d}\frac{\left(S - S_{eq}\right)\mathrm{F}_v}{\dfrac{\mathrm{L}_v^2\rho_l}{\mathrm{KR}_w T^2} + \dfrac{\rho_l \mathrm{R}_w T}{\mathrm{D}e_s(T)}} \tag{1.8}$$

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

Equation (1.8) has important implications for cloud formation. It points out that, for the 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 (1.8) 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 droplets 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 (1.8) giving the rate of increase in droplet mass is

$$\frac{\mathrm{d}m}{\mathrm{d}t} = \frac{2\pi d\left(S - S_{eq}\right)\mathrm{F}_v}{\dfrac{\mathrm{L}_v^2}{\mathrm{KR}_w T^2} + \dfrac{\mathrm{R}_w T}{\mathrm{D}e_s(T)}} \tag{1.9}$$

Equations (1.8) and (1.9) predict that the growth rate will vary with the temperature for a given value of S. The thermal conductivity of air R and the diffusivity D are themselves functions of the temperature and density of the air (Table 1.1). The most rapid growth, therefore, occurs under conditions of low density at relatively high temperatures, such as in the upper portions of tropical clouds.

Table 1.1			
Dynamic Viscosity and Thermal Conductivity of Air and Diffusivity of Water Vapour in Air[a,b] (after Dennis, 1980)			
Temperature (°C)	Dynamic Viscosity (μ) (10^{-6} kg m^{-1})	Thermal Conductivity (K) (10^{-3} J m^{-1} s^{-1} k^{-1})	Diffusivity of Water Vapour (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			

A commonly quoted formula for the ventilation factor is

$$F_v = 1 + 0.22(\mathrm{Re})^{1/2} \tag{1.10}$$

where Re, the Reynold's number of a droplet in air is given by

$$\mathrm{Re} = \rho_a d \frac{u}{\mu} \tag{1.11}$$

where ρ_a is the density of air, *d* the diameter of the droplet, *u* the fall speed of the droplet and μ the dynamic viscosity.

Equations (1.8) and (1.9) can be modified further to take account of molecular processes at the droplet surface (Prupaccher and Klett, 1978). They overestimate growth rates especially for droplets in the 1 μm size range, but they are satisfactory for many applications (Dennis, 1980).

Equation (1.9) 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 the terminal fall velocity, 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.

1.6 Simulation of Cloud Formation

Use of equation (1.9) 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 (Dennis, 1980). If the cooling is caused by the ascent and adiabatic expansion of the air parcel, a number of adjustments should be introduced at each time step. They cover changes in the number concentrations of CCN vis-à-vis N_c, the number concentration of cloud droplets, the coefficients of thermal conductivity and diffusivity, and in the concentration of the excess water vapour.

Howell (1949) was the first to calculate the growth of the droplets by condensation. He described the growth of the individual droplet with an ordinary differential equation that included the droplet radius, super saturation and nuclei mass. His model consisted of a parcel of air containing discrete size distribution of Sodium Chloride particles, rising adiabatically within the steady updraft. He noticed that the radial growth rate of an individual droplet increases with increase in droplet size, and the droplet size distribution narrows with age in the case of simple steady updraft. He suggested that the condensation might be the dominating process creating homogeneous distributions found in some young clouds. He found that updraft speed is most important in determining the shape of the droplet size distributions whereas the Sodium Chloride particle size distribution has a minor effect. He concluded that evaporation or turbulent mixing might create a broader size distribution.

From his calculations, Mordy (1959) concluded that the particle size distribution is as important as the updraft speed in determining final droplet population. Mordy's numerical computations (1959), shows the history of nuclei of different sizes as a parcel of air is cooled by an upward motion of 0.15 ms^{-1}. As 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 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 vapour is being condensed exceeds the rate at which excess water vapour 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 50m thick. Further condensation as air parcels continue their ascent increases the sizes of the existing droplets rather than the droplet concentration.

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. 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 (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 the 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 the size distribution of CCN present. An abundance of large hygroscopic particles to act as CCN generally results in a high value of N, the number concentration of cloud droplets.

Twomey and Wojciechowski (1969) have derived empirical formulae which quantify the three way relationship among updraft speed, the maximum super saturation during cloud formation, and the concentration of CCN activated.

1.7 Comparison with Observations

Cloud droplet size distributions have been measured by several investigators (Diem, 1942; Weickmann and Aufm Kampe, 1953; Warner, 1969; Spyers-Duran, 1972) noticed that the measured distributions were considerably broader than the computed microphysical models of Howell (1949). Mason and Ghosh (1957) showed that the large droplets may be formed on giant (r>1µm) salt nuclei. This may be acceptable for maritime clouds but does not explain the observed droplet size distribution in continental clouds where giant size salt nuclei are extremely rare or non-existing.

A number of workers have attempted to explain the broadening of droplet spectra beyond what can be accounted for by condensation alone by taking into consideration factors omitted by Howell (1949) in his original formulation. Two of these factors are turbulence and the mixing of environmental air into the cloud (Srivastava, 1974).

The effects of turbulence were extensively investigated by Russian and Chinese researchers (E.G.Beliaev, 1961; Sedunov, 1965); Mazin, 1965; Levin and Sedunov, 1966; Jaw Jeou-Jang, 1966; Wen Ching-Sung, 1966; Stepanov, 1975; 1976) who claim that the spectrum is considerably broadened by turbulence. In these studies, the super saturation is usually taken as a function of the mean vertical air velocity and the size distribution of the cloud drops is computed at a given time rather than at a given height. This approach has been criticized (Srivastava, 1974). Warner (1969) and Bartlett and Jonas (1972) showed that turbulence does not sensibly broaden the size drop spectra. The effects of mixing of environmental air on cloud drop spectra were investigated by Mason and Chien (1062) and Warner (1973). Warner finds that mixing broadens the spectra but in a way not observed in natural clouds. He concluded that droplet spectra observed in the lower levels of the cloud could be fairly closely reproduced by computations if an accomodation coefficient (fraction of the molecules hitting the liquid surface that are actually captured) between 0.02 and 0.05 was assumed. Using an accomodation coefficient of 0.036, and simultaneous measurements of cloud drop size spectra and cloud condensation nuclei, Fitzgerald (1972) found good agreement between observed and computed spectra. Fitzgerald calculated the droplet spectrum at a height of a few hundred meters above the cloud base and compared the results with observed spectra. Comparisons were made for five continental and two maritime clouds. Both the observed and computed spectra were quite monodisperse. Close agreement was found between observed and calculated spectra, as indicated by comparing the droplet concentrations, mean droplet diameters and the standard deviation of diameters. The average values of the dispersion coefficient (the ratio of standard deviation to mean diameter) was 0.17 for the observed distributions and 0.12 for the computed distributions.

The reason for the discrepancy, thought to exist earlier, was not a shortcoming in theory but inadequate observations. Fitzgerald measured the droplet distributions just above cloud base and found them to be as narrow as predicted by the diffusional growth theory. In earlier work the observations had not been confined to the lowest cloud level, and broader distributions were found (Rogers, 1976).

From these studies, it may be concluded that the droplet spectra within a few hundred meters of cloud base can be satisfactorily accounted for by the existing theory. The same has probably not been demonstrated for spectra observed higher up in the cloud. In some of these cases coalescence may have been an appreciable effect on the size distribution.

1.8 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 (Squires and Twomey, 1958). In general, droplet concentrations 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. Wide variations occur in each case.

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. Study of the cloud droplet spectra and aerosol concentrations vis-à-vis the rain activity in modified maritime and continental regions in India was carried out (Kapoor et al., 1976c). Results of the study indicated marked differences in the cloud drop spectra, drop concentration, median volume diameter of cloud drops, liquid water content and the giant size hygroscopic aerosols in the maritime and continental regions. Twomey and Wojciechowski (1969) compared the number of CCN activated as a function of supersaturation for the continental and maritime air masses sampled by the US Naval Research Laboratory. The result agrees in general with the cloud droplet counts.

CHAPTER II

FORMATION OF RAIN BY COLLISION-COALESCENCE PROCESS

2.1 Introduction

Most of the world's precipitation falls to the ground as rain, much of which is produced by clouds whose tops do not extend to temperatures colder than 0ºC. The mechanism responsible for precipitation in these "warm" clouds is the collision-coalescence process. This is the dominant precipitation forming process in tropics. It also plays a role in id-latitude cumulus clouds whose tops may extend to sub-freezing temperature.

Growth of rain by a condensation was proposed in the seventeenth century by A.Le. Grand in 1680 and E.Mariotte in 1686 (Middleton, 1965; List, 1977). The idea of coalescence of droplets was first put forth by E.Baslow in 1715. After Thomson derived the famous law on the dependence of water vapour pressure on the curvature of water surfaces, Osborne Reynolds (1877) was able to conclude that growth by condensation alone in warm clouds could never account for the production of raindrops with radii of several millimeters. Yet rain does form in warm clouds, especially in the tropics. The enormous increases in size required to transform cloud droplets into raindrops is illustrated by the size difference between cloud and raindrop as follows. A cloud droplet 10 μm in radius requires an increase in volume of one million fold to grow to a raindrop of 1mm in radius. However, only about one droplet in a million (say, about 1 per litre) in a cloud has to grow by this amount for the cloud to rain.

Current theories of rain development in warm clouds are adequate to account for both the basic characteristics of the cloud base spectra and rapid production of showers by coalescence once sufficient number of drops with radii greater than 20-25 μm are produced. The physical mechanisms for initial production of the first large drops however remain unclear. In maritime clouds drops of this size may be produced by condensation on giant sea salt particles.

In general, two basic approaches have been used to explain the origin of first large drops which are capable of initiating the coalescence (Johnson, 1979). The oldest and probably the widest accepted source for these large drops is condensation from giant sea salt particles which are generally present in sub-cloud air. Measurements of these particles in maritime air, (e.g., Woodcock and Gifford, 1949, Woodcock, 1950; 1952; 1953; Lodge, 1955) have generally confirmed their presence in significant concentrations. Although similar concentrations of salt nuclei have occasionally been reported in continental regions (e.g., Byers et al., 1957), it is almost universally agreed that these particles do not play major role in continental clouds (Mason, 1971).

This section contains a review of the microphysical properties of warm clouds, basic theory relating to droplet growth by collision and coalescence, and current status of numerical modeling of coalescence growth of droplets.

2.2 Microphysical Properties

2.2.1 Cloud Droplet Size Distribution

Cloud droplet spectra may be characterized by the function $n(r)$, with the property that $n(r)\mathrm{d}r$ is the number of droplets per unit volume with radii in the interval (r, r+dr). In general the distribution will vary with position in a cloud and with time at any one location.

The droplet concentration are number density N is the total number of droplets per unit volume and equals the integral of $n(r)$ over all droplet sizes present. In continental cumulus clouds a typical value of N is 200 cm^{-3}. In Hawaiian orographic clouds, an extreme case, it is about 10 cm^{-3}.

Squires (1956; 1958) reported measurements of cloud droplet size distributions in maritime and continental warm clouds. The droplet size spectrum for continental cumulus clouds is much narrower than that for the maritime cumulus clouds. Also the average droplet radius is significantly smaller. The measurements point out that, in the maritime clouds droplets with radius of about 30 μm exist in concentrations of about 1 $litre^{-1}$ whereas in continental clouds droplets of about 20 μm exist in concentrations of 1 $litre^{-1}$ (Wallace and Hobbs, 1977).

Measurements of cloud droplet spectra in different types of clouds have been reported by Aufm Kampe and Weickmann (1957) and they showed the average distributions. These differences in microstructure have important effects on the formation of rain in warm maritime and continental cumulus clouds.

Measurements of cloud drop size distributions made in warm monsoon clouds in different regions in India have indicated that the total concentration of cloud droplets varies between 50 cm^{-3} and 500 cm^{-3} (Paul et al., 1980; 1982; Kapoor et al., 1976b; Mary Selvam et al., 1980a). The width of the cloud droplet spectra in maritime clouds is broader than that in continental clouds. Aircraft observations made in more than 2000 warm monsoon clouds (Mary Selvam et al., 1980a) have shown that the horizontal extent of individual cumulus clouds at cloud base levels is in the range of 3 to 14 km. The horizontal cross-section of cloud liquid water content (LWC) shows a number of peaks and crests. The location of maximum LWC in the horizontall cross-section is more or less at the center of the cloud. The vertical profile of LWC shows an increase with height from the base of the cloud. The cloud drop spectra broadens with height from the base of the cloud.

2.2.2 Cloud Liquid Water Content

The cloud liquid water content q is the mass of condensed water per unit volume of air, defined by

$$q = \frac{4}{3}\pi\rho_l \int r^3 n(r)\,\mathrm{d}\,r$$

In the above equation n(r) is the number concentration of droplets in the radius size range r to r+dr

In the non-precipitating cumulus clouds a typical value of q is 0.5 gm^{-3}, with peak values of about 1 gm^{-3}. In stratus clouds the values tend to be smaller. In cumulonimbus clouds q can exceed 5 gm^{-3}. The upper limit of q is approximately q_a, the adiabatic liquid water content from parcel theory. The liquid water content varies very rapidly from place to place within a cloud. The maximum values of water content are almost always found to be considerably less than the adiabatic value, and of course mean values are lower still. A value of about 0.5 gm^{-3} is probably typical for liquid water content averaged throughout a normal cumulus cloud, the peak value within the cloud being perhaps twice this quantity.

Average dropsize distributions, droplet concentration and cloud liquid water content measured at different heights above the cloud base show that the average droplet diameter and the width of the distribution showed increases with height above the cloud base, whereas the droplet concentration decreased with height (Zaitsev, 1950; Squires, 1958).

2.3 Droplet Growth by Collision-Coalescence

Collisions may occur through differential response of the droplets to gravitational, electrical, or aerodynamic forces (Rogers, 1979). Gravitational effects predominate in clouds, large droplets fall faster than smaller ones, overtaking and capturing a fraction of those lying in their paths. The electrical and turbulent fields required to produce a comparable number of collisions are much stronger than those thought usually to exist in natural clouds. As a drop falls it will collide with only a fraction of the droplets in its path because some will be swept aside in the air stream around the drop. The ratio of the actual number of collisions to the number for complete geometric sweep out is called the collision efficiency, and depends primarily on the size of the collector drop and the sizes of the collected droplets.

Collision does not guarantee coalescence. When a pair of drops collide several types of interactions are possible: (1) they may bounce apart (2) they may coalesce and remain permanently united (3) they may coalesce temporarily and separate apparently retaining their initial identities (iv) they may coalesce temporarily and then break into a number of small drops. The type of interaction depends upon the drop sizes and collision trajectories, and is also influenced by the existing electrical forces and other factors. For sizes smaller than 100 μm radius, the important interactions are those given at (1) and (2).

The ratio of the number of coalescences to the number of collisions is called the coalescence efficiency. The growth of a drop by the collision-coalescence process is governed by the collection efficiency, which is the product of collision efficiency and coalescence efficiency. Laboratory studies of small colliding droplets indicate that the coalescence efficiency is close to unity if the droplets are charged or an electrical field is present. Because weak fields and charges exist in natural clouds, theoretical studies of droplet growth by collision-coalescence usually make the assumption that the collection efficiency equals the collision efficiency. The problem of explaining the initial development of rain then reduces to one of determining collision rates among a population of droplets.

(a) Droplet Terminal Fall Speed

The drag forces exerted on a sphere of radius *r* by a fluid is given by

$$F_R = \frac{\pi}{2} r^2 u^2 \rho C_D \tag{2.1}$$

where u is the velocity of the sphere relative to the fluid, ρ is the fluid density, and C_D is the drag coefficient characterizing the flow. In terms of the Reynold's number $R_e = \rho du/\mu$ where μ is the dynamic viscosity of the fluid and d equal to $2r$ is the drop diameter, (2.1) may be written in the form

$$F_R = 6\pi\mu r u\left(\frac{C_D R_e}{24}\right) \tag{2.2}$$

The gravitational force on the sphere is given by

$$F_G = \frac{4}{3}\pi r^3 g(\rho_l - \rho)$$

where ρ_l is the density of the sphere. For the case of a water drop falling through air $\rho_l >> \rho$ and

$$F_G = \frac{4}{3}\pi r^3 g\rho_l \tag{2.3}$$

to a good approximation. When $F_G = F_R$ the drop falls relative to the air at its terminal fall speed. For this equilibrium situation

$$u = \frac{2}{9}\frac{r^2 g\rho_l}{\left(\frac{C_D R_e}{24}\right)\mu} \tag{2.4}$$

For very small Reynolds numbers the Stokes solution to the flow field around a sphere shows that $(C_D R_e/24)=1$. In this case (2.4) reduces to

$$u = \frac{2}{9}\frac{r^2 g\rho_l}{\mu} = k_1 r^2 \tag{2.5}$$

with $k_1 \approx 1.19 \times 10^6 cm^{-1}$. This quadratic dependence of fall speed on size is called Stoke's Law and applies to cloud droplets up to 40 μm radius.

Experiments with spheres indicate that for sufficiently high Reynold's numbers C_D becomes independent of R_e and has a value of about 0.45. Using this value in (2.4) leads to

$$u = k_2 r^{1/2} \tag{2.6}$$

$$\text{where } k_2 = 2.2\times 10^3\left(\frac{\rho_l}{\rho}\right)^{1/2} cm^{1/2}\,sec^{-1} \tag{2.7}$$

In (2.7) ρ is the air density and ρ_l is a reference density of $1.2 \times 10^{-1} g\ cm^{-3}$, corresponding to dry air at 1013 mb and 20°C. Raindrops have high Reynolds numbers but are not perfectly spherical. Consequently, (2.6) describes the fall speed of raindrops reasonably well only over a limited range of size.

Observational data on raindrop fall speed have been reported (Gunn and Kinzer, 1949) and these values have been used to verify the empirical formulas developed by several investigators (Beard, 1976; Rogers, 1979). It can be determined from the data that (2.6) provides a reasonable approximation to the fall speed in the radius interval $0.6 < r < 2$ mm but with $k_2 \sim 2.01 x 10^3$ cm$^{1/2}$ sec^{-1}. In the intermediate size range, between the Stoke's Law region and the square-root law, an approximate formula for fall speed is

$$u = k_3 r \qquad 40\mu m < r < 0.6mm$$
$$\text{with } k_3 = 8x10^3 \sec^{-1} \tag{2.8}$$

(b) Collision Efficiency

For a drop of radius *R* overtaking a droplet of radius *r*, if the droplet had zero inertia it would be swept aside by the stream flow around the larger drop and a collision would not occur. Whether a collision does in fact occur depends on the relative importance of the inertial force and the aerodynamic force, and the separation *x* between drop centers called the impact parameter. For given values of *r* and *R* there is a critical value x_c of the impact parameter within which a collision is certain to occur and outside of which the droplet will be deflected out of the path of the drop. Mason (1971) has summarized calculations of collision efficiency and established the values of x_c over a wide range of spherical drop sizes. Results are presented in the form of collision efficiencies defined by

$$E(R, r) = \frac{x_c^2}{(R + r)^2} \tag{2.9}$$

The collision efficiency is equal to the fraction of those droplets with radius *r* in the path swept out by the collector drop that actually collide with it. Alternatively, E may be interpreted as the probability that a collision will occur with a droplet located at random in the swept volume. Clearly $E << 1$.

Collision efficiencies for small collector drops as obtained from three sets of theoretical calculations were compared. The computed results of Hocking (1959) were accepted as accurate for about a decade and figure prominently in the cloud physics texts of Fletcher (1960) and Byers (1965). These results have been superseded by two more recent calculations which are seen to agree fairly well with each other. One of the most significant differences between Hocking's original results and the new calculation, concerns the '18 μm cutoff'. The early work indicated that collision was impossible unless the larger droplet exceeded 18 μm in radius. The new results show that collisions can still occur with collector drops at least as small as 10 μm, though with small efficiency. The new calculations also indicate small but non-zero collision efficiency for droplets of the same size, unlike the original findings of Hocking. Mason (1971) gave the field of collision efficiency (E) as a function of *R* and *r*. E is generally an increasing function of *R* and *r*, but for *R* greater than about 80 μm E depends largely on *r*.

Some of the papers on collision efficiency refer to a quantity called the linear collision efficiency, defined by

$$y_c = \frac{x_c}{R} \tag{2.10}$$

From (2.9) and (2.10)

$$\mathrm{E} = \frac{y_c^2}{(1+p)^2} \tag{2.11}$$

Where $p = r/R$. Alternatively the efficiency is sometimes defined by

$$\mathrm{E}' = \frac{x_c^2}{R^2}$$

Since $E' = E(1 + p^2)$, it is clear that E can take on values greater than unity.

(c) Growth Equation

Suppose a drop of radius R is falling at terminal speed through a population of smaller droplets. During unit time it sweeps out droplets of radius r from a volume given by

$$\pi(R+r)^2[u(R)-u(r)]$$

where u denotes terminal fall speed. Thus the average number of droplets with radii between r and r+dr collected in unit time is given by

$$\pi(R+r)^2[u(R)-u(r)]n(r)E(R,r)dr$$

where E(R,r) denotes the collection efficiency, which equals the product of collision efficiency times coalescence efficiency. When the drops are all smaller than about 100 µm it is usually assumed that the coalescence efficiency is unity, so that the collection efficiency is identical to the collision efficiency. The total rate of increase in volume of the collector drop is obtained by integrating over all droplet sizes:

$$\frac{\mathrm{d}v}{\mathrm{d}t} = \int_0^R \pi(R+r)^2 \frac{4}{3}\pi r^3 \mathrm{E}(R,r)n(r)[u(R)-u(r)]dr \tag{12.12}$$

In terms of drop radius

$$\frac{\mathrm{d}R}{\mathrm{d}t} = \frac{\pi}{3}\int_0^R \left(\frac{R+r}{r}\right)^2 [u(R)-u(r)]n(r)r^3 \mathrm{E}(R,r)dr \tag{12.13}$$

The change of drop size with altitude may be obtained from

$$\frac{\mathrm{d}R}{\mathrm{d}z} = \frac{\mathrm{d}R}{\mathrm{d}t}\frac{\mathrm{d}t}{\mathrm{d}z} = \frac{\mathrm{d}R}{\mathrm{d}t}\frac{1}{(w-u(R))} \tag{12.14}$$

Where w is the updraft speed

Equation (2.13) is general in that it allows for the sizes and fall speeds of the collected droplets. If these droplets are much smaller than the collector drop, then an approximation to (2.13) follows by setting $u(r) \sim 0$ and $(R+r) \sim r$.

Thus
$$\frac{\mathrm{d}R}{\mathrm{d}t} = \frac{\mathrm{E}q}{4\rho_l}u(R) \tag{2.15}$$

Where E is the effective average value of collection efficiency for the droplet population and q is the cloud liquid water content. From (2.14), the change of radius with altitude is then given approximately by

$$\frac{\mathrm{d}R}{\mathrm{d}z} = \frac{\mathrm{E}q}{4\rho_l}\frac{u(R)}{w-u(R)} \tag{2.16}$$

2.4 Models for Coalescence Growth of Droplets

(a) Continuous Collision Model

In this model it is assumed that larger drops grow by continuously capturing smaller drops uniformly distributed in space. This model predicts that drops of a given size grow at the same rate (Wallace and Hobbs, 1977).

Let us now consider a collector drop of radius, which has a terminal fall speed v_t, when it falls through still air. Let us suppose that this drop is falling in still air through a cloud of otherwise equal sized droplets of radius r_2, and terminal fall speed v_2 in still air. We will assume that the droplets are uniformly distributed in space and that they are collected uniformly at the same rate by all collector drops of a given size.

The rate of increase in the mass M of the collector drop due to collisions is then given by

$$\frac{\mathrm{d}M}{\mathrm{d}t} = \pi r_1^2(v_1 - v_2)q\mathrm{E}_c \tag{2.17}$$

where q is the liquid water content (in kg m^{-3}) of the cloud droplets of radius r_2. Substituting $M = \frac{4}{3}\pi r_1^3\rho_l$ into (2.17), where ρ_l is the density of liquid water we obtain

$$\frac{\mathrm{d}r_1}{\mathrm{d}t} = \frac{(v_1 - v_2)}{4\rho_l}q\mathrm{E}_c \tag{2.18}$$

If $v_1 >> v_2$ and we assume that the coalescence efficiency is unity, so that $\mathrm{E}_c = \mathrm{E}$, and (2.18) becomes

$$\frac{\mathrm{d}r_1}{\mathrm{d}t} = \frac{v_1 q\mathrm{E}}{4\rho_l} \tag{2.19}$$

Since v_1 increases as r_1 increases and E also increases with r_1 it follows from (2.19) that $\mathrm{d}r_1/\mathrm{d}t$ increases with increasing r_1: that is, the growth of a drop by collisions is an accelerating process. For small cloud droplets growth by condensation is initially dominant but beyond a certain radius growth by collisions dominates.

If there is a steady updraft velocity w in the cloud, the velocity of the collector drop will be $(w\text{-}v_1)$ and the velocity of the cloud droplets $(w\text{-}v_2)$. Hence $\mathrm{d}r_1/\mathrm{d}t$ will still be given by (2.18) but the motion of the collector drops is given by

$$\frac{\mathrm{d}h}{\mathrm{d}t} = w - v_1 \tag{2.20}$$

Where h is the height above a fixed level (say, cloud base) at time t. Eliminating $\mathrm{d}t$ between (2.19) and (2.20) and assuming $v_1 >> v_2$ and $\mathrm{E_c} = \mathrm{E}$ we obtain

$$\frac{\mathrm{d}r_1}{\mathrm{d}h} = \frac{v_1 q \mathrm{E}}{4\rho_l (w - v_1)}$$

or, if the radius of the collector drop at height H above cloud base is r_H and at cloud base is r_0

$$\int_0^H q\,\mathrm{d}h = 4\rho_l \int_{r_0}^{r_H} \frac{(w - v_1)}{v_1 \mathrm{E}} \mathrm{d}r_1$$

Hence, if we assume that q is independent of h,

$$H = \frac{4\rho_l}{q} \left[\int_{r_0}^{r_H} \frac{w}{v_1 \mathrm{E}} \mathrm{d}r_1 - \int_{r_0}^{r_H} \frac{\mathrm{d}r_1}{\mathrm{E}} \right] \tag{2.21}$$

If values of E and v_1 as a function of r_1 are known, (2.21) can be used to describe the value of H corresponding to any value of r_H and vice-versa. We can also deduce from (2.21) the general behavior of cloud drops growing by collisions by following qualitative argument. When the drop is still quite small $w > v_1$, and the first integral dominates over the second, H then increases as r_H increases: that is, a drop growing by collisions is carried upward in the cloud. Eventually, as the drop grows, v_2 becomes greater than w and the value of the second integral becomes larger than that of the first. H now decreases with increasing r_H : that is, the drop begins to fall through the updraft and, if it holds together, it will eventually pass through the cloud base and reach the ground as a raindrop. Some of the larger drops (with radius greater than 1 mm) may break up as they fall through the air, particularly those involved in collisions. The resulting fragments may then grow and break up again. Such a chain reaction may enhance the rain.

In the early 1950s several investigators used (2.21) and similar equations to investigate the feasibility of rain formation by the collision-coalescence mechanism (Wallace and Hobbs, 1977). A few drops, large enough to grow by collisions, were generally arbitrarily assumed to be present near the base of the cloud. Some of the earliest estimates of collision efficiencies were used and the coalescence efficiency was assumed to be unity. These calculations indicated that rain should be able to form by the collision-coalescence mechanism within reasonable time periods in cumulus clouds. For example, in a cloud containing 1 g m^{-3} of water (consisting of 10 μm radius droplets) and having an updraft velocity of 1 m s^{-1}, an initially small collector drop was predicted to grow to a radius of about 0.15 mm in about 45 min during its travel upward in the cloud to a height of 2.2 km. During the next 15 min the radius of the drop was predicted to increase to 0.75 mm (which corresponds to a fairly large raindrop) as it fell back to cloud base. With the same updraft velocity but a liquid water content of about 0.5 g $m^{=3}$, the calculations indicated that the collector drop would be carried up to a height of about 3.2 km and finally emerge from cloud base with a radius of about 0.65 mm. With a liquid water content of 1 g m^{-3} and an updraft velocity of 0.1 m s^{-1} these simple model calculations indicated that a collector drop would only be carried 0.5 km above cloud base, from which height it would take nearly 2 h for it to fall back to cloud base where its radius would be only about 0.1 mm (which corresponds to a small drizzle drop). Hence, these model calculations indicated that warm clouds with strong updrafts should produce rain in a shorter time than clouds with weak updrafts, but clouds with strong updrafts must be quite deep in order for raindrops to be produced. Also, raindrops which form in deep, warm clouds with strong updrafts should be considerably larger than raindrops from shallower clouds with weak updrafts.

In the continuous collision model it is assumed that the collector drop collides in a continuous and uniform fashion with smaller cloud droplets which are uniformly distributed in space. Consequently, the continuous collision model predicts that all collector drops of the same size will grow at the same rate when they fall through the same cloud of droplets.

(b) Stochastic collision Model

An important advance in our understanding of the growth of drops in warm clouds has been made by replacing the continuous collision model by a stochastic (or statistical) collision model. The stochastic model allows for the fact that collisions are individual events, statistically distributed in time and space.

Consider for example 100 droplets, initially the same size. After a certain interval of time some of these droplets (let us say 10) will have collided with other droplets so that the distribution will now be as follows. Because of their larger size these 10 larger droplets are now in a more favourable condition for making further collisions. The second collisions are similarly statistically distributed, giving a further broadening of the droplet size spectrum. Hence by allowing for a statistical distribution of collisions, three size categories of droplets have developed after two time steps. This concept is extremely important, since it not only provides a mechanism for developing broad droplet size spectra from the fairly uniform droplet sizes predicted by condensation, but it also reveals how a small fraction of the droplets in a cloud can grow much faster than average by statistically distributed collisions.

(b) Results of Numerical Models

The basic problem in explaining the formation of rain by coalescence of cloud droplets is to understand how drops of different sizes are first introduced into the cloud. It has long been recognized that the rate of growth of a cloud droplet by condensation is inversely proportional to the radius. If, as was formerly believed, all the nuclei involved in cloud formation are much smaller than typical cloud droplets, then continued condensational growth results in an increasingly monodisperse drop distribution. Since these drops will all fall at approximately with the same velocity, collisions between these drops would be exceedingly rare. It was this difficulty which led Bergeron to underestimate the potential of drop coalescence for producing the rain showers. Houghton (1938) recognized the potential importance of coalescence mechanism, and suggested several possible explanations for the initial production of different sized drops.

Langmuir (1948) introduced the concept of the "chain reaction" to explain the development rain in warm cumulus clouds. According to this concept, drops grow by coalescence until they reach a size at which they become unstable and break up into smaller fragments. Each of these fragments in turn can grow by coalescence until an unstable stage is reached and it too disintegrates. This process may continue through several cycles.

Das (1950) computed the droplet trajectories taking the size of the droplets (collected) into account and obtained new values of collision efficiency for different combinations of collector and collected drops. Subsequently he computed drop size spectra and liquid water content for different initial conditions and compared his values with observed spectra. It was shown that the number of large drops in a cloud cannot increase by coalescence, unless the drop spectrum has a very large range or a sharp peak (Das, 1956).

Bowen (1950) and Ludlam (1951), using the "continuous coalescence" concept, investigated the warm rain process in clouds with constant updrafts. These studies indicated that clouds with strong updrafts require greater depth to produce rain than the weak updraft clouds, but strong updraft clouds would produce rain in a shorter period of time. The initial size of the collector drop at the cloud base had this similar effects, large drops require less depth than smaller drops, but the small initial drops developed into larger raindrops. The results of these calculations suggest, the possible role of sea spray drops in the development of warm rain.

Improvements on the calculations of Bowen and Ludlam were made by Telford (1955) with coalescence being treated as a "Stochastic" rather than a "continuous" process. Telford showed that the stochastic process could produce significant number of large drops in much shorter time than the coalescence concept. The stochastic process also led to the formation of the complete raindrop spectra. He concluded that the stochastic process would also produce a few drops large enough to rupture within much shorter time than were indicated by previous studies and thus the possibility of chain reaction processes. East (1957) performed detailed calculations of coalescence and showed that condensation is the driving force behind the rapid coalescence growth.

Increasingly more sophisticated stochastic models have been developed since Telford's work. They have shown that Telford over-estimated the role of production of larger drops but that his conclusions were correct quantitatively.

Twomey (1964) verified Telford's conclusions for a continuous distribution of cloud droplets. Barlett (1966) studied growth by coalescence for different initial cloud droplet spectra. He concluded that there must be a significant number of drops of radii greater than 20 μm (more than 1 per litre) and also adequate supply of drops with radii between 10 – 20 μm for rain to form. He found that the minimum collector size was the most important single parameter governing the onset of the coalescence mechanism. Twomey (1966) performed similar calculations and reached essentially the same conclusions. Twomey (1966) found that the cloud droplet number concentration to be much more important than the dispersion of the distribution. The lower the number concentration, the more rapid the development (maritime-type size distributions much more rapidly than continental-type size distribution).

Berry (1967) studied cloud droplet growth by stochastic coalescence by a variety of Kernels (hydrodynamic capture, electric field capture and geometric sweepout). He found that the droplet growth rate is proportional to the magnitude of the collection kernel and the growth pattern dependent on the derivative of the kernel with respect to the size of the droplet. Warshaw (1967) included vertical sedimentation for a non-updraft situation for sizes less than 15 μm.

Kovetz and Olund (1969) included the effects of growth by condensation, coalescence and drop sedimentation in a somewhat more realistic model.

Nelson (1971) investigated the initiation of warm rain. He modeled the physical processes of advection, coalescence and drop breakup but ignored condensation. A steady-state updraft profile was assumed and the initial cloud droplet distribution was based on observations of tropical warm cumulus. He noted the importance of a localized accumulation of liquid water (above the updraft maximum) in leading to an acceleration of the coalescence process. Drops growing rapidly in these accumulation zones soon become large enough to fall against the updraft and move downward through the cloud. He also simulated the effects of water-spray seeding in his model. The water-spray "seeds" were introduced at the same level where the accumulation zone later developed. He noted earlier rainout for the seeded cloud but a lower peak rain intensity and less total rainout. He suggested that this decrease was caused by earlier removal of much of the water from the accumulation zone. Haman (1968) and Iribarne (1968) also studied accumulation zones in models with more realistic dynamics but very crude microphysics. Both stressed the critical role of dynamics in subsequent development of the accumulation zone.

Danielsen et al. (1972) noted some effects of drop break up in their study of hail stone growth in a one-dimensional, time-dependent model. Drop breakup had a significant effect on drop distribution between 150 μm and 1 mm. The effects of breakup were only noticeable in a situation where the updraft was strong enough to suspend large drops in a region of rapid growth by coalescence (in the lower, warmer regions of the cloud).

Ogura and Takahashi (1973) studied the development of warm rain in a one-and-a-half dimensional, time dependent model. The processes of vertical advection, condensation/evaporation, stochastic coalescence and drop breakup were modeled in detail in their study. They also noted the accumulation zone, and attendant rapid growth by coalescence, in the upper portions of the model cloud. Their model indicated no significant difference in the overall dynamic behavior of cloud with respect to the breakup process.

The apparent inability in explaining the rapid development of large drops in continental clouds has led a large number of investigators to consider more elaborate mechanisms. Woods et al. (1972) and Almeida (1975) have examined the role of small scale turbulence in increasing the collision efficiency between the small drops. Alternately, Mason and Jonas (1974) looked at drop growth in successive small thermals, where a residue of an individual thermal is allowed to interact with the later ones. These studies have not produced convincing explanations for the initial production of the large drops.

An alternate hypothesis has been proposed by Johnson (1979) which assumes that ultra giant aerosol particles of size as large as 100 μm or more could act as embryos for rapid development of precipitation sized particles, even in colloidally stable, continental clouds. Measurements of aerosol particles in mixing layer below the cloud base have shown that particles as large as 100 μm or more are a regular part of the atmospheric aerosol (Nelson and Gokhale, 1968; Hindmann, 1975; Johnson, 1976; Hobbs et al. 1977) which support the hypothesis of Johnson. He argued that these ultra giant aerosol particles may initiate coalescence growth without the necessity of prior condensational growth and termed these particles as coalescence nuclei.

Manton (1979) developed a more general theory of the interaction of the cloud droplet distribution with turbulent cloud. He also showed that the vertical velocity fluctuations cannot of themselves induce broadening of droplet distribution. This is because of variations in time for an air parcel to reach a given level above cloud base are compensated by variations in supersaturation. By assuming that convective cloud turbulence has a high vertical coherence and that fluctuations in the integral radius of the droplet distribution are negatively correlated with vertical velocity, Manton (1979) showed that an initial unimodal distribution is transformed into a bimodal distribution at heights above cloud base.

Baker et al. (1980) have also calculated the effects of turbulence on the initial broadening of a droplet spectrum growing by condensation. In their model, entrainment and mixing are viewed as taking place inhomogeniously. Calculations with the model demonstrate a bimodal droplet spectrum similar to observed spectra and that the dispersions of the spectra increase with height above the cloud base.

Blyth et al. (1980) provided circumstantial evidence from observations of mountain clouds that the hypothesized inhomogeneous mixing process and associated droplet spectral broadening is occurring in real clouds. While one may question the details of the inhomogeneous mixing models of Manton (1979) or Baker et al. (1980), they have clearly strengthened the case for turbulent mixing as being a major factor in the formation of embryonic precipitation droplets. Telford and Chai (1980) examined the effect of mixing of dry air into a cloud top and its effect on droplet spectra. They have shown that the resulting cycling of the air up and down in the cloud can double the largest drop radius and generate cloud parcels containing drops of all sizes upto this maximum. These changes in the droplet distribution are not greatly influenced by the cloud condensation nuclei since "maritime" like spectra can develop in "continental" type cumuli. The large number of cloud condensation nuclei should not have much effect in inhibiting the rain-forming process by reducing coalescence growth.

Johnson (1980) has pointed out that cloud base temperature also influenced the activation of cloud droplets. Other things being the same, clouds with colder cloud bases will activate more cloud droplets. The aerosol distribution and updraft velocity at cloud base, are the most important factors controlling the concentration of activated droplets, and, hence, the colloidal stability of the cloud.

In addition to the physical and dynamical processes responsible for the initial broadening of the cloud droplet spectrum, the regions of highest liquid water contents (LWC) in warm clouds are most favourable for initiation of precipitation. Twomey (1976) showed that if locally enhanced regions of LWC comprise only one percent of the cloud volume and exist for periods of a few minutes, such regions can produce significant concentrations of large drops averaged over the entire volume of the cloud. Thus, the presence of protected updrafts having nearly moist adiabatic liquid water contents (Heymsfield et al. 1978) can have significant bearing upon the initiation of precipitation in warm clouds. However, the ultimate amount of rainfall from a given cloud is controlled by the overall time space character of its updrafts and its liquid water contents (Cotton, 1982).

2.5 Effects of Electric Field and Drop Charge on Collision and Coalescence

The effects of drop charge and electric field on collision and coalescence have been examined both theoretically and experimentally by numerous investigators over the years. Lord Rayleigh (1879) has first suggested a possible connection between rain formation and electrical effects. Cochet (1952) concluded that charged water drops could accelerate the collision and coalescence process. Numerous laboratory studies (Sartor, 1954; Goyer et al. 1960; Cataneo et al. 1970; Abbott, 1975; Dayan and Gallily, 1975; Smith, 1972; Paul et al. 1979) and modeling studies (Sartor, 1960; Lindblad and Semonin, 1963; Davis, 1965; Krasnogorskaya, 1965; Plumlee and Semonin, 1965; Semonin and Plumlee, 1966; ;atham, 1969; Schlamp et al. 1976; Cohen and Gallily, 1977) have demonstrated that charged droplets and electric fields can significantly enhance collection efficiencies. In some of these studies it is noticed that the effect is most pronounced for the smallest collector droplets and droplets of comparable size that exhibit little difference in relative momentum. The electric field strengthens producing the greatest effect have magnitudes of the order of those found in thunderstorms and pre-thunderstorm conditions. Aircraft observations of microphysical and electrical parameters in warm monsoon clouds (Mary Selvam et al., 1976a; Murty et al., 1976; Kapoor et al., 1976b) and the results of the laboratory investigations (Paul et al., 1979) indicate the drop charges and electric fields observed in some of the warm monsoon clouds are large enough to enhance the collision and coalescence and promote the initiation of precipitation embryos. However, Pruppaccher and Klett (1978) concluded that weak charges and weak fields that would be expected to be present in developing warm cumuli will probably not significantly enhance collection efficiencies.

2.6 Evolution of Raindrop Spectra

Raindrops which form by the coalescence process in a cloud of liquid water droplets do not continue to grow indefinitely. The distributions of raindrops usually indicate a rapid decrease in drop concentration with increasing size, at least for diameters exceeding about 1 mm (Rogers, 1979). Also they generally show a systematic variation with rainfall rate.

Measurements, in general suggest that the distributions are approximately negative – exponential in form. Marshall and Palmer (1948) first suggested this approximation on the basis of a summer's observations in Ottawa, Canada. Comparison of drop spectra at three values of rainfall rate with the best fit exponential approximations show straight lines on the semi-logarithmic coordinates. Thus the drop-size distributions, except for very small sizes may be approximated as

$$N(D) = N_0 e^{-\Delta D} \tag{2.22}$$

Where N(D) dD is the number of drops per unit volume with diameters between D and D+dD.

Marshall and Palmer found that the slope factor Δ depends only on rainfall rate and is given by

$$\Delta(R) = 41R^{-0.21} \tag{2.23}$$

Where Δ has units of cm^{-1} and R is measured in mm hr^{-1}. They also found that the intercept parameter N_0 is a constant given by

$$N_0 = 0.08cm^{-4} \tag{2.24}$$

It may be noted that not all drop size distributions have this simple exponential form. Measurements from different regions have shown that an exponential tends to be the limiting form as individual samples are averaged.

Theoretical investigations of raindrop size distributions in rain – showers from warm cumulus clouds have been reported (Roy and Srivastava, 1958; Srivastava, 1960; Srivastava and Roy, 1962). The results suggest that the computed raindrop size distributions as obtained from certain assumptions are in agreement with the accepted theory of coalescence growth of raindrops. The anomalies, regarding the time interval between the cumulus cloud development and precipitation release and width of raindrop spectrum were explained by Ramana Murty and Roy (1962) on the basis of existing theories at that time (East, 1957; Scorer and Ludlam, 1953).

Extensive measurements have been made on dropsize distributions in orographic monsoon showers at Khandala in the Western Ghats regions during August 1956 and non-orographic monsoon showers at New Delhi during July –October 1956 (Ramana Murty and Gupta, 1959). The observed distributions follow the Marshall and Palmer distribution for very low rain intensities (up to 5 mm hr^{-1}).

Sivaramakrishnan (1960; 1961; 1965) and Sivaramakrishnan and Mary Selvam (1967) made extensive observations of raindrop size distributions and the associated parameters like liquid water content, rate of rainfall and the electric charge carried by rain. The study suggested that the measurement of rain current (i.e., the total charge brought down by rain per cm^2 per second) would help in understanding the type of process involved in the rain.

(a) Raindrop Break up

An explanation of the tendency for raindrop size distributions to approach a negative – exponential form is provided, at least in part, by the phenomenon of break up. Once a diameter of about 3 mm is attained it is no longer sure that surface tension can hold the drop together. A drop as large as 6 mm in diameter is unstable and can exist only briefly before breaking apart. Langmuir (1948) suggested that once raindrops grow to a critical size of ~ 6 mm in diameter, they will break-up due to hydrodynamic instability. He hypothesized that each break-up fragment will act as a new precipitation embryo, which can grow to break-up size and create more raindrop embryos. He referred to the process as the 'chain reaction theory of warm rain formation". Other observations (Blanchard, 1948; Magarvey and Geldart, 1962; McTaggart-Cowan and List, 1975) have suggested that collisions amongst droplets on the order of 2-3 mm in diameter and smaller can initiate break-up.

Using a numerical cloud model (Farley and Chen, 1975) have concluded that a necessary condition for a Langmuir chain reaction to develop is that a cloud must develop sustained updrafts 10 m s^{-1}.

Komabayasi et al. (1964) have given data on the probability of spontaneous break-up as a function of drop size and the size spectrum of small drops thereby produced.

Another cause of break-up is collisions between drops. Collisions at grazing incidence produce a spinning elongated drop which may quickly fly apart resulting in the formation of satellite drops. Disruption occurs when the rotational kinetic energy of the coupled drops exceeds the surface energy required to produce separate drops (Brazier-Smith et al., 1972).

Srivastava (1971) attempted to determine theoretically whether raindrop coalescence and break-up would lead to a Marshall- Palmer distribution. The drop-size distributions that resulted were not found to fit the simple negative exponential law, being rather too flat in the size range from about 2 to 5 mm diameter. He suggested that the failure to approach exponential form for his distributions might have arisen from the neglect of condensation or collision-induced break-up.

Other studies of coalescence and spontaneous and collisional break –up (Young, 1975; Gillespie and List, 1978; Srivastava, 1978; 1982) showed the development of an equilibrium size distribution of raindrops.

Computer simulations of the evolution of a population of falling raindrops growing by coalescence with cloud droplets and with one another, but subject to break-up collisions have shown that these processes lead to an exponential distribution, but the indicated values of Δ (slope factor) do not always agree with observations (Dennis, 1980).

CHAPTER III

PHYSICAL CONCEPTS AND MODELS FOR WARM CLOUD MODIFICATION

3.1 Introduction

Attempts at warm cloud modification have been under way for more than 50 years. Several techniques have been adopted for modification of warm clouds in the various experiments and operations carried out over different regions of the globe. The physical concepts and models advanced for warm cloud modification are reviewed in this chapter.

3.2 Physical Concepts

The growth of precipitation in warm clouds begins with condensation of water vapour into droplets in a rising air current. The number and size of droplets near the base of the cloud depends on the available supply of condensation nuclei in the rising air and the velocity of the updraft. Thus, the initial number of droplets in a particular volume is determined soon after that volume enters the cloud. Thereafter, unless the droplet number is altered by evaporation or by additional nuclei, as air is mixed in from outside, the droplet concentration in the volume either remains the same or is decreased by the coalescence of droplets or the removal of droplets by sweepout.

The droplets in a rising volume of air undergo growth within the cloud by condensation of water vapour directly on to the droplets. Additional drop growth within the cloud results from collision and coalescence of one or more droplets to form a larger drop. These collisions take place more readily when the colliding drops are of different size and have different fall velocities. The collision efficiencies of small droplets (< 19μm radius) are very small and the rate of droplet growth by coalescence, among drops of these sizes, is very slow. Under proper conditions of moisture and initial drop sizes, however, growth by coalescence can be rapid enough to form precipitation elements within the cloud which subsequently fall through the base.

The processes described above take place within the interior of the cloud. For the production of rain, it is vital that the cloud lifetime exceed the time required for the droplet growth process to develop precipitation-size elements. The lifetime and nature of the updrafts in the cloud are largely controlled by the characteristics of the environment surrounding the clouds. Temperature lapse rate, air moisture content and wind shear contribute substantially to the cloud behaviour. Organised motions in the atmosphere on a scale of a few kilometers to several hundred kilometers may ultimately be the dominant factor in determining the nature of the cloud systems and the ability of clouds to give greater precipitation.

The warm cloud processes which might be subject to modification are examined in the following.

The rate of coalescence growth within the cloud is critically dependent on the cloud droplet size distribution. A predominance of small drops slows the growth rate considerably while large drops collide more effectively and stimulate growth. This suggests a modification technique based on control of initial cloud drop size. The method for altering the drop size is through the addition of condensation nuclei. Large number of small condensation nuclei should result in small cloud drop sizes. On the other hand, a limited number of large ($0.1 < r < 1 \mu m$) or giant ($r > 1 \mu m$) size condensation nuclei will create large cloud drops necessary for obtaining an increase in the coalescence growth rate.

At a large scale it may be possible to alter the dynamics of the cloud system itself by massive seeding of clouds to help generate the buoyancy within the cumulus updrafts for invigorating the sub-cloud moisture flux. Any increase in the updraft speed would lead to an increase in the water content available for the drop growth process and would prolong life-time of the cloud. An increase in the updraft speed and available cloud water would result in an increase in effective buoyancy in the natural updraft. A small increase in the upward moisture can cause an appreciable increase in cloud water content. This could be accomplished by massive seeding of warm clouds with hygroscopic particles which would result in the release of latent heat due to condensation and or with other chemicals such as Calcium Carbide (CaC_2) and Calcium Oxide (CaO) which hydrate prior to deliquescence and in the process release substantial quantities of heat. Such concept (dynamic seeding) is more advanced and experimented in the case of cold cloud modification than in the case of warm clouds.

3.3 Models for warm cloud modification

Once the basic discoveries of Langmuir (1948a,b; 1951, 1955) and Schaefer (1941, 1948, 1949) were announced many persons developed ideas on how clouds might be modified to obtain specific operational objectives. The important ones were summarized by Howell (1966) under the "Conceptual models that guide applied cloud seeding".

The principal conceptual models advanced for modifying the microphysics of clouds by artificially altering the CCN spectrum have been discussed by Dennis, 1980.

3.3.1 Artificial formation of clouds

The question is to be examined whether or not it is possible to produce a cloud artificially by supplying CCN. This would require a situation where air is super saturated, but no clouds form because of absence of CCN. This situation occurs, but only rarely. For e.g., above the hot pools of Yellow Stone National Park, supersaturation ratios sometimes reach three or four where the water temperature in a thermal pond is about 85^{o}C and the ambient air temperature above the pond is approximately -15 ^{o}C. The atmosphere there contains very large amounts of water vapour drawn from the hot pools, while the constant diffusion of water vapour upward and away from the surfaces of the hot pools drives away most of the CCN. It has been demonstrated that small clouds (supercooled) could be produced over the hot pools simply by burning a match or otherwise introducing hygroscopic particles capable of acting as CCN. Occurrences of high supersaturation 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 40mg per kilogram of air. They reasoned that the heat of vapourisation released might provide sufficient added buoyancy to produce a convective cloud.

Although visible clouds were produced in the experiments of Woodcock and Spencer over the Pacific Ocean off Hawaii, the clouds consisted of solution drolets in equilibrium with the ambient vapour pressure and none of the droplets was driven past their critical diameters. Aircraft observations indicated temperature excess of about 0.4 ^{o}C. As updrafts in vigorous cumulus are often 1-3 ^{o}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 layers are ordinarily capped by a subsidence inversion.

3.3.2 Changing the microphysics of a cloud

Maritime clouds produce rain more readily than do continental clouds 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 dropsize distribution characteristics of maritime clouds and thereby increase their chances of producing precipitation. The "problem" with continental clouds is a cloud droplet concentration (N) higher than optimum for rain formation by coalescence. It might appear that a lower N could be obtained by nucleus poisoning, i.e., the addition of some chemical that would deactivate many of the CCN. However, no chemical is readily available which has been demonstrated to produce the desired nuclei poisoning effect. Also, no matter how many CCN are poisoned, there would still be numerous particles which would be capable of serving as CCN.

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 1-3μm in concentrations of 25-100cm^{-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 stand point. Assuming that 2μm 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 10^6m^3 of air s^{-1} to be 30kgmin^{-1}. The logistics involved in applying the seeding concept to an entire cloud are obviously formidable.

3.3.3 Artificial raindrop embryos

A second compromise approach to the promotion of coalescence in water clouds is the introduction of artificial raindrop embryos. A direct method of introducing raindrop embryos into a cloud is to spray it with fine water droplets.

However, this method of seeding 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. 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 provide the giant CCN that play a role in the formation of rain showers over the sea. For the artificial raindrop embryos to be effective the giant particles introduced must yield embryos well beyond the Hocking limit (diameter = 38μm).

Calculations by Biswas and Dennis (1972), for e.g., 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 to 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.

Klazura and Todd (1978) consider condensational as well accretional growth of the embryos in their cloud model. They have run their growth model in a one-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. They conclude that hygroscopic particle 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 within updraft speed. For a cumulus of moderate depth (5km) and with a moderate updraft ($12ms^{-1}$) NaCl particles introduced at the base must be more than 40 μm in diameter for the resultant drops 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 that the embryo concentration required to speed the coalescence process is $1000m^{-3}$, 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 break up and thereby create more raindrops. 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 break up. The process is unlikely in clouds where updrafts do not exceed $5ms^{-1}$.

Farley and Chen (1975) performed a numerical model calculations in which they concluded that for seeding to be effective in stimulating the warm rain process, there must be a mechanism for drop break up, that for a chain reaction to develop, vertical velocities greater than $10m\ s^{-1}$ are required, and that salt seeding acts primarily to initiate a Langmuir-type chain reaction.

A reasonable compromise appears to be the use of 25-50kg 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 (Dennis, 1980, Krishna et al., 1974).

3.4.4. Overseeding with CCN

If one wishes to delay onset of rainfall in maritime clouds they should be converted to continental clouds with regard to their microphysics. It would be necessary in such cases to introduce additional CCN into the clouds. Such an action would be an example of overseeding, which is loosely defined as suppression of precipitation by interference with coalescence process. The artificial CCN required to introduce would have to be of size comparable to the natural CCN present and be provided in concentrations of hundreds per cubic centimeter. The logistical problems involved would, therefore be greater than those involved in hygroscopic seeding due to the larger CCN concentrations required. Overseeding with CCN is, therefore, not a practical concept.

CHAPTER IV

MODIFICATION OF THE MICROSTRUCTURE

4.1 Introduction

Different techniques of seeding for modification of the microstructure have been experimented by different investigators. These can be broadly classified into the following two categories.

i. Water drop seeding

ii. Hygroscopic particle seeding

The experiments and operations carried out in different countries are reviewed in this Chapter.

4.2 Water drop seeding

In the 1950s water drop and hygroscopic particle seeding experiments on warm clouds were carried out in several countries. In some cases rain appeared to be initiated by seeding, but since neither extensive physical nor statistical evaluations were carried out the results were inconclusive (Wallace and Hobbs, 1977).

Langmuir experimented the drop breaking "chain reaction" theory proposed by him through an experiment conducted on a cumulus cloud in the Hondura Region in the dry season of 1948. The technique adopted consisted of seeding the top of the cloud with coarse water spray. The cloud, following seeding, grew rapidly and gave a heavy rainfall most unusual for that season of the year, no other rain occurring nearby (Langmuir, 1955).

Coons et al., (1949) and Bowen (1952) carried out limited exploratory experiments in water spray seeding. Bowen (1950) and Ludlam (1951) carried out model calculations of water spray seeding. They suggested seeding at cloud base height with a spray of 20-30 µm droplets would produce optimum effects.

Water drop seeding experiments are conducted by the University of Chicago, USA, in the warm cumulus clouds over the Carribean during 1953-1954. All cloud were seeded with coarse water spray, at first at the rate of 130 gal/mile (small value), later at the rate of 400 gal in 18 seconds (large value). The seeding was done at 3000-5000 ft above the cloud base. The clouds were selected for seeding on a randomized basis. It was concluded that the small value treatment was ineffective in initiating rain, but the large value treatment appeared to increase the average probability of rain from 4 to 44 percent and to be significant at level of 4 percent when the results from the two periods of experimentation were combined. In the treated clouds, rain was detected in 2 mins to 16 mins following seeding, averaging 8½ minutes for the small value and 6½ minutes for the large value treatments, compared with the echoes observed in 2 to 26 minutes (average time 12 minutes) in the unseeded clouds (Braham et al., 1957).

Warm cloud modification experiments were carried out over the level plains of the Southern Camaquey Province of Cuba during the winter dry season of 1954. Cumulus clouds with bases at 4000-5000 ft above sea level and of vertical thickness of a few thousand feet were seeded with fine droplets of water spray. The seeding was carried out within the lowest few hundred feet of the cloud (cloud base level) in updrafts whenever possible. These clouds of vertical thickness 800-2300 ft dissipated in about 15 minutes following seeding. One cloud of vertical thickness 3500 ft darkened and bulged at the base immediately after seeding but did not rain. A cloud of vertical thickness 4800 ft gave a light shower that began 11 minutes after seeding and lasted for 30 minutes. Six clouds of vertical thickness greater than 7000 ft gave heavy rain, beginning from 11 minutes to 25 minutes after seeding (average 16 minutes that lasted 49 minutes or more). Isohyetal maps for the period of the experiments showed an unusually heavy concentration of rain in the seeded zone (Howell, 1962). Also cloud seeding experiments using solid CO_2 and NaCl were conducted using aircraft during April-September months of the period 1968-1978. Two target areas of 2500 and 500 km^2 were chosen for the experiments. Results are not available (WMO, 1978).

Cloud seeding experiments were conducted over selected agricultural areas in the French Equatorial Africa by the Meteorological Services of the Afrique Equatoriale Francaise (Raoul du Chaxel, 1955; Howell (1962). Stratocumulus and cumulus clouds were seeded with water spray and finely powdered salt. The water spray seeding was carried out at the rate of 200 litres per run. 20 gms ampoules of finely powdered salt were used in salt seeding experiments for 11 cases of stratocumulus seeding, rain was noticed in 3 cases, 17 minutes after seeding. In the above cases the cloud vertical thickness was 1200-2000 ft. When the thickness of the seeded clouds was 500-2000 ft no rain was noticed. The thinnest layer clouds dissipated following seeding. Out of the 16 cases of seeded cumulus clouds rain was observed in 9 cases, 3 to 17 min (average time 8½ min) following seeding. The vertical thickness of the seeded clouds which developed rain was in the range 2000-4000 ft 9average 2700 ft). All seeded cumulus clouds thicker than 4000 ft yielded rain.

The Hong Kong Royal Observatory conducted warm cloud modification experiments by spraying water from nozzles placed on hill tops. The records from a network of rain gauges failed to indicate any effect of the seeding (Ramage and Bell, 1955; Howell, 1962).

Rokicki and Young (1978) performed numerical simulation of water drop seeding in a Lagrangian parcel model. They introduced a spray of seed droplets of radii 600-200μm. The effect of seeding was evaluated on the basis of time required to initiate precipitation. They concluded that effect of water drop seeding is generally greater than Silver Iodide seeding with no danger of over seeding and suggested that this technique should be considered for precipitation enhancement in mid-latitudes as well as tropics.

The technique of water drop seeding has not been viewed as an economically attractive proposition because large quantities of water must be carried aloft by aircraft (Cotton, 1982).

4.3 Hygroscopic particle seeding

In general, there have been major approaches to hygroscopic particle seeding, first, attempting to increase the number of raindrops in clouds that are already capable of producing rain and, second, trying to form rain earlier or faster than would have occurred naturally. The second approach seems to offer the possibility of making clouds rain that would have not rained by themselves (Johnson, 1980). Computer modeling studies, (e. g., Rokicki and Young, 1978; Klazura and todd, 1978) seem to show a promising future for this type of seeding by predicting major changes in precipitation formation through the addition of modest amounts of seeding material.

Hygroscopic particle seeding experiments on warm clouds have been carried out in several countries. In some cases rain appeared to be initiated by the seeding. However, for obtaining more conclusive results, and for a better understanding of the warm cloud responses to salt seeding, extensive cloud physical measurements are required particularly in seeded and not-seeded clouds. A summary of hygroscopic particle seeding experiments conducted in different countries is given below.

East African Meteorological Department carried out a series of cloud seeding experiments in Tanganyika, Kenya and Uganda. Silver Iodide and salt were used for seeding. The experiments conducted during 1952 suggested no increase in rainfall on days of Siver Iodide seeding as compared to not-seeded days, but on days of salt seeding the rainfall 6-12 miles downwind from the release point was substantially greater. Rainfall at the release point upwind from the target was apparently lighter on seeded than on not-seeded days. In 1953, salt seeding experiments were conducted on 64 cumulus clouds. Out of these rain was noticed on 37 occasions and light rain or virga was noticed on 10 occasions. The time between seeding and appearance of rain ranged from 7 minutes to 35 minutes averaging 22 minutes. In 1954 clouds were seeded on a random basis. The results indicate that 24 out of 34 seeded clouds produced rain. In 1966 rockets were used for dispersing salt into clouds. Thirty rocket firings were carried out on 7 seeded days. All clouds seeded produced rain. The times between seeding and rain ranged from 3 minutes to 37 minutes. On 70 percent of occasions rain appeared within 8 minutes. The rainfall pattern of the 7 seeded days as compared with that of six seedable unseeded days showed a shift of the maximum from the upwind rim of the water shed to a point 2 to 5 miles down-wind from the firing site, along with a considerable intensification of the maximum (Davies et al., 1951, 1952, 1955; Sansom et al., 1955; Brazel and Taylor, 1959).

Salt seeding experiments were carried out in Madagascar during 1953-54. Finely powdered salt was dispersed into the clouds at a rate of 10 to 100 ml km^{-1} by blowing heated air up through a can of ground salt. Clouds were seeded on 101 days scattered over 22 targets. The results were analysed principally by comparison of rainfall in target gauges with their climatic averages. Eight positive, 8 indifferent, and 4 negative outcomes were reported. It was also reported that no effect of seeding was seen unless the clouds were more than 3000 ft thick (Augustin and Angola, 1956; Howell, 1962).

Pakisthan Meteorological Department carried out cloud seeding experiments in Punjab region during the monsoon season of 1954 (Fournier de'Albe et al., 1955; Howell, 1962). Powdered salt was dispersed during southeasterly winds at a rate of about 10 gm sec^{-1} into the air near the ground by a heater-bellows device so that some of it would be carried by convective currents into clouds forming down-wind. Seeding was done in two areas, one mountainous and one more plain. Results were analysed by establishing from a previous 40-years' period the normal relationship between rainfall in the 50 degrees sector down-wind from the seeding sites and neighbouring sectors on either side and comparing this ratio for the seeded year with this normal. In one area the ratio in the seeded year was the highest on record and 2½ times the normal. In other area, it was the second highest on record and about 80 per cent above normal.

In Mexico randomized salt seeding experiment was carried out in 1958 using ground-based seeding generators (Fournier de'Albe and Aleman, 1976). Salt was dispersed at the rate of 50 Kg hr^{-1} during hours of maximum convective activity during the summer. Salt particle masses were estimated to range from 10^{-10} gm to 10^{-8} gm. In all, 54 days were seeded while 22 days were used as control days. The analysis showed that there was significantly less rainfall on seeded days than on control days. It was suggested that these results may have been due to exceptionally heavy natural rainfall on the limited number of control days, rather than a decrease due to seeding.

A well designed randomized salt seeding experiment was conducted in three areas (Delhi, Agra and Jaipur) in North India during the summer monsoon season (July-September) of 1957-1966 (Ramanamurty and Biswas, 1968). Seeding was carried out by spraying from the ground a dilute salt solution or by dusting of finely powdered salt and soapstone mixture using power sprayers and air compressors. The estimated dispersal rate at the source is approximately $2x10^{10}$ salt particles (radius 5 μm) per second. Control and target areas were defined upwind and down-wind of the central seeding locations and comparisons were made between rainfall in these two areas for seeded and unseeded days. Days on which heavy rain occurred frequently or continuously were considered as unseedable. Seedable days were selected on the basis of certain meteorological criteria, particularly in respect of low cloud amount, wind shear and humidity in the lower levels. As days with heavy or continuous rain have been considered as unseedable it is unlikely that the rainfall on seeded days was from clouds penetrated well above the freezing level which will involve ice phase. Radar climatology for the Delhi region suggests that the frequency of occurrence of warm, cold and combined warm and cold rain processes are respectively 41 % , 16 % and 43 % (Ramanamurty et al., 1960; Devara and Ramanamurty, 1984). The results of the statistical analysis of the experiment suggested that the rainfall in the target area was on the average 20 % greater than that in the control area which is significant at better than 0.5 % level (Biswas et al., 1967). Radar observations were made in the Delhi region during the five summer monsoon seasons (1961, 1963-1966) of the salt seeding experiment. The evaluation of these observations indicated positive result in four seasons, negative result in one season (Chatterjee et al., 1969). Also results of computer numerical simulation of the cloud seeding experiment carried out for North India using historical rainfall data suggested that a 30 per cent increase in rainfall due to seeding could be detected over a period of 12 summer monsoon seasons with more than 80 % probability (Mary Selvam et al., 1980b). However, the ground based salt seeding experiment carried out in North India during the summer monsoon seasons of 1957-1966 (Ramanamurty and Biswas, 1968) has been criticized for lack of cloud physical observations in support of the significant increase in rainfall noticed in the target area on seeded days (Warner, 1973; Cotton, 1982).

Randomised salt seeding experiment using aircraft was conducted in the Delhi region during the summer monsoon of 1962 (Roy et al., 1964). This was the first attempt to carry out aircraft-based cloud seeding experiment in India. Seeding was carried out on 18 days. Assessment of the results has been made on the basis of the data collected both by raingauge and radar. Raingauge data indicated positive trend on 7 occasions, negative trend on 2 occasions and no specific trend on the remaining 9 occasions. Measurements by radar on 14 out of the 18 cases considered suggested positive trend in 8 cases, negative trend in 4 cases and no specific trend in the remaining two cases. The results of the limited number of trials conducted during the season have been encouraging, though by no means conclusive.

Ground-based salt seeding experiments for modifying orographic warm clouds were carried out at Munnar, South India during the summer monsoon seasons of 1964 and 1965. The results of these experiments suggested rainfall in target area was on the average 38.8 per cent more than that in the control area which is not statistically significant (Ramanamurty and Biswas, 1968).

Randomised salt seeding experiments, using ground-based generators, with fixed control-target design, was carried out at Thiruvallur in the Madras region in South India during the southwest and northeast monsoon seasons of 1973, 1975-1977. The results of the experiments suggested an increase of 32 per cent in rainfall on seeded days during the southwest monsoon and 17 per cent decrease during northeast monsoon. The results are not statistically significant (Pillai et al., 1981).

Salt seeding experiments using aircraft were carried out on operational basis in the Rihand catchment in the state of Uttar Pradesh, North India during the summer monsoon seasons of 1973 and 1974 (Kapoor et al., 1976a). The rainfall analysis indicated increases in rainfall on seeded days by 16 to 28 per cent. The result is not statistically significant. Another operational salt seeding experiment using aircraft was carried out in the Linganamakki catchment in the state of Karnataka, South India during the summer monsoon season of 1975. The reservoir water level during the period of seeding operation was estimated to be 25.7 percent greater than the largest increase reported during the preceding 10 years (Murty et al., 1981).

A warm cloud modification experiment was carried out in an area of 4800 Km^2 in the Pune region (18°32'N, 73° 51'E, 559 m asl) during the 11-summer monsoon (June-September) seasons (1973-74, 1976, 1979-86). A double-area cross-over design with area randomization was adopted and an instrumented aircraft was used for seeding and cloud physical measurements. Finely pulverised salt (sodium chloride) particles were released into the monsoon clouds (cumulus and stratocumulus) during aircraft penetrations into the clouds at a height of 200-300 m above the cloud-base. The results of the Indian Experiment have clearly emphasized the need for the physical understanding, sequential development (stepwise programmes to test the applicability of the warm cloud modification hypothesis), preditor variables, model simulations for obtaining conclusive results. The warm cloud responses to salt seeding are found to be critically dependent on the cloud physical characteristics e.g., vertical thickness and liquid water content. Clouds with vertical thickness > 1 km, LWC > 0.5 gm m^{-3} when seeded with salt particles (modal diameter 10 μm, concentration 1 per litre of cloud air) produced increase in rainfall of 24 per cent significant at 4 per cent level. Shallow clouds (vertical thickness < 1 km, LWC < 0.5 gm m^{-3}) when seeded showed tendency for dissipation. The cloud physical observations made in not-seeded (control) and seeded (target) clouds have provided some useful evidence to test the applicability of the warm cloud modification hypothesis. The results of the cloud model computations suggested that moderate convergence at the cloud-base is essential for the cloud growth and development of precipitation in the real world. Hygroscopic particle seeding of warm clouds under favourable dynamical conditions (convergence at the cloud-base level) may result in the acceleration of the collision-coalescence process resulting in the enhancement of rainfall (Murty et al., 2000)

Warm cloud modification experiments were conducted in Phillippines during 1948. Mostly warm clouds were seeded with dry ice. One of the seeded clouds yielded precipitation shortly following seeding and also appeared to take an increased rate of development (Webber, 1948; Howell, 1962). The seeding effect was attributed to the mechanical influence of dry-ice pellets in sweeping out cloud droplets and releasing them as larger particles. Salt seeding experiments were carried out in an area of 6216 km^2 in the Bohl area during April-June 1975. Results are not available (WMO, 1975).

In Thailand the first Royal Rainmaking Research and Development Team was formed in 1969, and the first experimental operation was conducted. The result of the operation was observed to be promising and encouraging. The rainmaking operation conducted in 1971 was reported to be successful. The Royal Rainmaking Research and Development Institute was established in 1975. The chemicals used for operation are Calcium Carbide, Calcium Chloride, Sodium Chloride, dry ice and urea. These were used separately or mixed in certain proportions and used either in powder or liquid form depending upon the prevailing meteorological conditions. Seeding was accomplished, sometimes, both at the cloud-top and cloud-base. The government allocates a yearly budget enough for three rainmaking teams. These teams are responsible for about 10 million acres of cultivated land in 30 out of the total 72 provinces. The rainmaking operation takes place for 150-200 days from February to October every year. Eight single-engine or five twin-engine aircraft are used by each one of the rainmaking teams for dispersal of chemicals. The total net chemical load carried by the aircraft per team is over 2000kg. There is a proposal to collect rainfall data from all the raingauges functioning in the target area for preparing a report which may include some kind of evaluation (RRDI, 1981).

In Indonesia the first preliminary experiment was conducted during 19-28 July 1977 in Bogor. The second experiment took place during 9-11 August 1977 in the Solo-Pacitan area. The chemicals used in the experiments were Sodium Chloride powder, Calcium Chloride powder, urea powder, urea solution and dry ice. Aerosil was used for prevention of coagulation. Six to seven aircrafts used for dispersing the chemicals and for surveying purpose. Precipitation was reported to have been achieved as a result of the experiments (Hadijaya, 1977). The next series of experiments commenced on 3 November 1979 in Jatilunhur area and continued until 20 November. After the experiments there was rain for 15 days and the water level in the dam rose by 3.55m. As the water level in the dam normally shows a rise of 1.00m in 15 days, the observed rise of 3.55m during the period of the experiment was reported to be considerable as it has seldom occurred. Some useful visual observations were made about rain developments in clouds following seeding (BPPT, 1980). Experiments took place over a period of 16 days between 25 March – 14 April 1980 in Lombok Island. Rain was reported to have fallen in the Island on each of the experimented days (BPPT, 1981). The above experiments were in the nature of operations. A five-year warm cloud modification non-randomisation program was set up in 1979 in the Jatilunhur Hydroelectric dam area to evaluate the effectiveness of seeding as it is possible to measure accurately the water flow into the reservoir.

Pineapple Research Institute and Hawaiian Sugar Planter's Association carried out cloud seeding experiments during 1948-1949 in Hawaii (Leopold and Halstead, 1948; Howell, 1962). Mostly warm clouds were seeded with dry ice and the seeding effect was attributed to the mechanical influence of dry ice pellets in sweeping out cloud droplets and releasing them as larger particles. Out of the 54 tests conducted on 15 days, rain or virga was noticed in 20 tests, generally in less than 15 minutes after seeding. Both heavier rainfall and higher frequency of rainfall was observed on seeded days, and the heaviest rains seem to be associated with sub-freezing cloud-tops. Cloud thickness was distinguished as the most significant factor in determining whether seeding would produce rain.

In USA, salt seeding experiments were conducted by the Pennsylvania State University in the St. Corix, VI. region during 1968 and 1969. The results were inconclusive (Simpson and Dennis, 1974). Naval Weapons Center at China Lake, California carried out warm cloud modification experiments during 1969-1971 over the Gulf of Mexico in the vicinity of Brownsville, Texas. The first of the three phases of the experiments consisted of a life cycle study of untreated warm clouds. The other two phases covered measurements of relevant cloud physics parameters before and after treatment of clouds with seeding agents. The seeding material included hygroscopic solutions of Ammonium Nitrate and urea, and hydrophilic powders such as Portland cement. The effect of treatment ranged from no change to complete collapse and dissipation depending upon the dosage rate, state of cloud development, and area of application (Clark et al, 1972). Randomised hygroscopic seeding experiments were carried out in the state of South Dakota in the USA during 1969-1970 as a part of the Project Cloud Catcher (Biswas and Dennis, 1971; Dennis and Koscielski, 1972). Results were evaluated based on the radar data relating to 41 seeded and 38 not-seeded cloud cases. The preliminary results suggested that the first radar echoes in the new cloud towers appear closer to cloud base in the salt-seeded than in not-seeded clouds. The average height of first echoes above cloud-base was roughly 1.5 and 3 km for salt-seeded and not-seeded clouds respectively. Also, it was concluded that rainfall from clouds of moderate depth (~ 5km) can be increased by several hundred-acre ft. From an example of the salt-seeding case of 23 July 1970, Biswas and Dennis (1971) concluded that the salt seeding had initiated a Langmuir chain reaction. Blanchard (1972) disputed this claim saying that the seeded rain embryos could not have grown to break-up dimensions in a cloud of only 3 km depth. Biswas and Dennis (1972) performed further calculations to show that the embryos of 100μm diameter are greater, when inserted at a cloud base would have a good chance of growing to break-up size (they chose diameter 5mm as break-up threshold) and descending through the cloud depth. From these calculations they concluded that Langmuir chain reaction had been initiated in seeded clouds. Mayers and Orville (1972) have shown that changing a parameter in the Marshall-Palmer distribution of the rain equation to simulate salt seeding would result in rainfall from the seeded model cloud and not from the non-seeded model cloud. A redistribution of the rain water loading may have been the determining factor in the rainfall in the model although a modified Marshall-Palmer distribution implies increased break-up as well. Farley and Chen (1975) performed numerical model calculations in which they concluded that for seeding to be effective in stimulating the warm rain process, there must be a mechanism for drop-break up, that for a chain reaction to develop vertical velocities greater than 10 m s^{-1} are required and that salt seeding acts primarily to initiate a Langmuir-type chain reaction. Klazura and Todd (1978) also performed numerical simulations of the salt seeding experiments in South Dakota. They used a one-dimensional, steady-state, adiabatic, condensation-coalescence model to simulate hygroscopic seeding. They concluded that: (i) for weak updrafts, larger hygroscopic seeded particles travel through a lower trajectory and sweep out less water than small seeded particles. The smaller seeded particles are more likely to grow large enough to break-up and create additional rain drop embryos (ii) for strong updrafts, the large hygroscopic seeded particles grow into precipitation and have a better chance of breaking up and initiating the Langmuir chain reaction, while smaller particles are not able to initiate precipitation before evaporating at cloud-top. The Meteorology Laboratory Air Force Cambridge Research Laboratory (AFCRL) performed measurements and seeding experiments in warm cumulus clouds off the East coast of Florida in September 1971. The goal of the AFCRL program was to collect data characterizing the development of these clouds and test the use of encapsulated urea as a seeding agent to stimulate rainfall (Cunningham and Glass, 1972). In the course of the program 2 to 6 clouds of similar dimensions were selected in each daily flight. Half the clouds were seeded while the other half served as controls. The order of selection was randomized. Some 26 clouds were selected on 7 days over a period of 2½ weeks. All clouds penetrated were analysed to determine the initial cloud conditions. The subsequent history of each cloud in terms of its growth, liquid water content and buoyancy changes were also determined. Statistical results of these experiments were not reported (NOAA, 1972).

In Malaysia the first experiments on cloud modification were carried out in 1973 with the objective to precipitate prematurely the clouds in the South China Sea. Seeding was done with large droplets of Calcium Chloride solution dispersed from aircraft in convective clouds. Reduction in rainfall was observed in the target area. Experiments in the Kedah State under "MADA Cloud Seeding Project" have been undertaken since 1977 for Agricultural purposes (The and Ho, 1974). Solutions/powders consisting of Sodium Chloride, Calcium Chloride and urea were released into warm cumulus clouds either at cloud-base or within the cloud depending upon the cloud cover. Seeding was undertaken during January – April, May-June and September- October months. Also, experiments in Perak State under "TEMENGOR Cloud Seeding Project" were undertaken since 1979 for Hydro-Electric purposes. The seeding technique consisted of releasing saturated solution of Sodium Chloride and urea at the rate 40 litres per minute. Dry particles of Sodium Chloride, Ammonium Nitrate, urea or Calcium Chloride and in some cases mixed with crushed dry ice were released at the rate of 30 kg min^{-1} at the cloud-top, cloud-base or into the cloud. The experiments presently deploy 2 to 4 aircraft of Skyvan – type, each of 1000kg payload. Rainfall in the target area could not be measured due to inaccessibility of the areas. Attempts to gauge the effects of seeding by determining the reservoir inflow were inconclusive because of the poor correlation of the reservoir inflow with rainfall. There is a proposal to install additional raingauges in the Temengor Catchment and to randomize the seeding operation for the purpose of evaluation.

Precipitation enhancement operations in Upper Volta (West Africa) began in 1974 and stopped in 1977. The seedings were carried out at the beginning and at the end of the rainy season to increase its duration. Urea pyrotechnique flares were used for warm clouds or Silver Iodide pyrotechnique flares for cold clouds. The pyrotechniques were launched from a Piper Navajo aircraft. Randomisation was not used in the operations. The effectiveness of seeding has been evaluated by comparing the precipitation observed in the seeded period with past records over the same area. The average increase noticed in precipitation is found not significant (Boutin, 1980).

In Spain urea seeding experiments were carried out in a area of 3000 km^2 in the Gran-Canarias- Tenerife regions during October-December 1975. Warm clouds with base temperature of $+8^oC$ and top temperature of $-5\ ^oC$ were seeded. The seeding was carried out by dispersing urea into the clouds of vertical thickness 800-1000 m. Results are not available (WMO, 1975).

Salt seeding experiments were conducted in an area of 10480 km^2 in the Xinanjinang and Zhejinang regions of China during the April-June months of 1968-1979. Salt was released from aircraft at the rate of 10 to 40 kg/min. Results are not available (WMO, 1979).

The Indian salt seeding experiment referred to earlier was carried out in North India during 1957-1966 using ground-based salt seeding generators. Though the experiment was statistically well designed and the result obtained was significant at less than 0.5 per cent level, physical evaluation could not be carried out for lack of cloud physical observations. In view of this limitation as well as for precise targeting of seeding material, a salt seeding experiment using aircraft with a randomized cross-over design has been undertaken in 1973 in Maharashtra State. The experimental area consisted of North and South sectors of each 1600 km^2 area and a buffer sector of 1600 km^2 area. The raingauge density in the target-control areas in 1 per 40 km^2. A DC-3 instrumented aircraft is used for seeding. The seeding material used is the same as in the ground-based experiment, namely, pulverized mixture of salt (Sodium Chloride) and Soapstone in the ratio 10:1 with the modal size of the particle being 10µm. A specially designed seeding gadget fitted to the aircraft is used for spraying salt at cloud-base levels. The seeding is carried out inside clouds at a height of a few hundred metres above the base of the cloud. The seeder aircraft has instruments for making the following observation. (1) Cloud droplet size distribution (2) Liquid water content (3) Vertical electric field (4) Electric charges carried by cloud drops and rain drops (5) Electrical conductivity (6) Corona discharge current (7) Vertical air velocity (8) Temperature of cloud and cloud-free air (9) Dew point temperature (10) Pressure altitude (11) Cloud condensation nuclei (12) Giant size condensation nuclei (13) Aitken nuclei. In addition to the above cloud physical observations a specially designed gadget for collection of cloud and rain water samples has been fitted to the aircraft and cloud/rain water samples from seeded and not-seeded clouds were collected and analysed for their chemical composition. Also rain water samples at the surface were collected from the surface at the target and control areas during the period of the experiment and their chemical composition was determined. The microphysical, dynamical and electrical observations and the chemical composition of the cloud and rain water were used for the physical evaluation of the warm cloud responses to salt seeding.

A warm cloud modification experiment was carried out in an area of 4800 Km^2 in the Pune region (18°32'N, 73° 51'E, 559 m asl) during the 11-summer monsoon (June-September) seasons (1973-74, 1976, 1979-86). A double-area cross-over design with area randomization was adopted and an instrumented aircraft was used for seeding and cloud physical measurements. Finely pulverised salt (sodium chloride) particles were released into the monsoon clouds (cumulus and stratocumulus) during aircraft penetrations into the clouds at a height of 200-300 m above the cloud-base. The results of the Indian Experiment have clearly emphasized the need for the physical understanding, sequential development (stepwise programmes to test the applicability of the warm cloud modification hypothesis), predictor variables, model simulations for obtaining conclusive results. The warm cloud responses to salt seeding are found to be critically dependent on the cloud physical characteristics e.g., vertical thickness and liquid water content. Clouds with vertical thickness > 1 km, LWC > 0.5 gm m^{-3} when seeded with salt particles (modal diameter 10 µm, concentration 1 per litre of cloud air) produced increase in rainfall of 24 per cent significant at 4 per cent level. Shallow clouds (vertical thickness < 1 km, LWC < 0.5 gm m^{-3}) when seeded showed tendency for dissipation. The cloud physical observations made in not-seeded (control) and seeded (target) clouds have provided some useful evidence to test the applicability of the warm cloud modification hypothesis. The results of the cloud model computations suggested that moderate convergence at the cloud-base is essential for the cloud growth and development of precipitation in the real world. Hygroscopic particle seeding of warm clouds under favourable dynamical conditions (convergence at the cloud-base level) may result in the acceleration of the collision-coalescence process resulting in the enhancement of rainfall (Murty et al., 2000)

4.4 Warm cloud electrical and microphysical responses to salt seeding

The responses of warm clouds to salt seeding have been studied based on the results of the analyses of the cloud physical observation referred to in the earlier section. The values of the median volume diameter, Liquid water content, electric field, coronal discharge current in clouds and rain water samples showed significant changes in seeded versus not-seeded clouds.

Increases have been noted in the mean volume diameter upto 478% and in the computed liquid water content upward of 60% (Kapoor et al., 1976b). The electric field in maritime warm cumulus clouds, which developed rain following seeding showed a sign reversal from the initial negative to positive occasionally preceded by intensification (Murty et al., 1976). The field reversal noticed is due to the transport of large positive charges from the upper levels of the clouds to the base through rain drops in the vigorous up draft regions. The intensification of the electric field is due to the invigoration of the updraft by salt seeding promoting increased charge separation. The corona discharge current has shown an increase up to 400 per cent following seeding in maritime cumulus clouds (Murty et al., 1981).

4.5 Chloride and Sodium ion concentrations in cloud and rain water samples

Chloride and Sodium ion concentrations in samples collected from seeded clouds have been found to be upward of 100 % higher than those in samples collected from not-seeded clouds (Khemani et al., 1982). The conductivity of the samples collected from seeded clouds has been higher by about 60 % than that observed in samples collected from not-seeded clouds.

In addition to the experiments and operations involving water drop and hygroscopic particle seeding techniques described in Sections 4.2 and 4.3 numerous laboratory and modeling studies have been carried out for stimulation of calescence by electric fields or drop charging. Recent investigations involving this technique have been aimed at increasing fog visibility by enhancing coalescence (Davis, 1983). The logistical problems of performing substantial modification of drop charges or electric fields in cumulus clouds are so complex that it may not be a feasible approach to the modification of warm rain (Cotton, 1982).

CHAPTER V

MODIFICATION OF THE DYNAMICS

5.1 Introduction

The concept of precipitation modification by ordering the dynamics of warm clouds is not new. Various techniques have been suggested and also experimented. These are described briefly below.

5.2 Initiation of thermal plumes (Météotron)

Fuel-oils were burnt for initiating the thermal plumes from the ground by an artificial source called "Météotron" (Dessens, 1960; Dessens and Dessens, 1964; Benech, 1976). In the French Météotron experiment an array of 105 fuel-oil burners is deployed in a three-armed spiral pattern within a 140x140 m^2. Temperatures in the plume (Benech et al., 1980; Radke et al., 1980) have been measured to be 60°C at the 30 m level, and 25°C at the 60 m level with vertical velocities of 7-10 m sec^{-1} at both levels. Under stratocumulus conditions, the plume was observed to produce a cloud-free ring around the plume. There is no documented evidence that the plume is associated with any precipitation anomaly (Cotton, 1982).

5.3 Asphalt coating for Albedo modification

Black and Tarmy (1963) have proposed to coat large areas of coastal deserts with asphalt to change the Albedo and induce sea breeze-like, meso-scale circulations and vertical motions, which, in turn, could initiate precipitating cumuli. Asphalt coated surfaces frequently showed temperature excesses over surrounding vegetated land, as large as 11°C, and sustain a temperature anomaly through a considerable portion of the diurnal cycle. The technique is considered to be worthy of consideration in oil-rich semi-arid coastal regions where Asphalt is by product of oil processing industry (Cotton, 1982).

5.4 Carbon black seeding

Another hypothesis for cloud modification that utilizes the sun's radiation as an energy source is carbon black seeding. In the late 1950's the US NAVAL Research Laboratory (Van Straten et al., 1958) seeded eight cumulus clouds with carbon dust, each of which was presumed to have dissipated prematurely. Five seeding runs in clear air were followed by the formation of small clouds. Downie (1960) reported on clear air seeding with carbon black which produced no obvious results. Gray et al., (1976) argued that dispersal of carbon black on the meso-scale (~ 100-300 km) with a time-scale of application of 1-2 days, would significantly modify meso-scale circulations. They estimate that carbon dust cloud of 10 % horizontal area coverage could absorb on the order of 100 joules per day which is enough to warm a I km layer of air by about 4°C per day. The validity of this concept should be examined by more extensive numerical simulations (Cotton, 1982).

Chen and Orville (1977) performed two-dimensional model simulation of carbon dust seeding on cloud-scale. They found that a 10 % area coverage can produce a vertical velocity of ~ 30 cm s^{-1} in 10 min but that the vertical velocity thereafter decreases rapidly with time.

The results were not encouraging for direct formation of cloud lines by dispersion of carbon dust in the tropical atmosphere.

In Brazil experiments on climate modification through carbon dust seeding were undertaken during 1975-1979 in two areas of size 1.6x10^6 km^2 and 0.9x10^6 km^2. Ground-based generators and airborne generators were used for carbon dust particles (dia ~ 0.1 μm) by complete combustion of hydrocarbons. It was hypothesized that when a significant portion of the incoming solar energy over the Atlantic Ocean could be absorbed by carbon dust in the atmospheric boundary layer during day-light hours, a stimulation of cumulus convection would occur. Results are not available (WMO, 1979).

5.5 Utilisation of heats of hydration for dynamic effects

Seeding of warm cumulus cloud regions with chemicals such as Calcium Carbide (CaC_2) has been carried out in parts of SouthEast Asia (DeMott, 1982). This substance hydrates prior to deliquescing and in the process releases substantial quantities of heat. This may well be a more efficient method of generating buoyancy within cumulus updrafts with the hope of invigorating the sub-cloud moisture flux. If the heat of hydration alone, or a combination of the heats of hydration and condensation are sufficient to drive the relative humidity to 100 % or below, heating produced could be strong and sustained. Calculations of the heats of hydrations for different compounds were reported by DeMott (1982). The most energetic f the reactions are the hydrations of Aluminium Chloride ($AlCl_3$) and Calcium Carbide (CaC_2). For any degree of success the material must be dispersed quickly and in very large quantities. Assuming reasonable seeding rates, it was calculated that the substances Aluminium Chloride and Calcium Carbide will likely heat seeded cloud volumes by 0.3-0.4°C and maintain this buoyancy through deep (perhaps hundreds of metres) cloud layers. Seeding should be performed in the early cloud growth stages with the incentive of accelerating the growth process. Due to the low solubility of the hydrated complex produced from CaC_2 and the slightly higher heat of hydration of $AlCl_3$, the latter compound appears more favourable. However, an enormous practical problem is the dispersion of these materials. The effects of heats of hydration generated through seeding of warm clouds with chemicals are to be investigated through modeling studies. Some of them are:

a) Effect of seeding on cloud microphysical processes

b) Effect of heat released on dynamics of the cloud particularly in the turbulent scale

c) Effects of buoyant production of turbulence on droplet coalescence process

Some of the above effects have been investigated in the warm cloud model (Mary Selvam et al., 1983b).

5.6 Utilisation of heats of condensation for dynamic effects

The concept of dynamic seeding of cumulus clouds with ice nucleating substances, in order to promote rapid glaciations for massive latent heat release, has received much attention in the literature. The concept of dynamic enhancement of cumuli by inducing heat release in the warm cloud region is, however, an area of scientific investigation in its infancy. Salt seeding of warm cumuli has been conducted primarily to change microphysical properties although it is possible to release some heat due to some condensation. This can promote buoyancy in local cloud regions. Results of investigations relating to modification of dynamics of warm clouds through utilization of heats of condensation are briefly given below.

Woodcock and Spencer (1967) hypothesized that the latent heat of condensation liberated to the atmosphere by the dispersion of NaCl particles in a nearly water-saturated atmosphere would be sufficient to initiate a cumulus cloud. They estimated that the release of 40 mg of NaCl per kilogram of air would raise the air temperature a few tenths of a degree Celsius. Experiments in releasing salt from aircraft in the warm, moist marine boundary layer near Hawaii where the relative humidity was 80-90 % created small but visible cumulus clouds. Aircraft measured temperature excesses were of the order of 0.4°C, less than expected temperature anomalies in cumulus clouds having a precipitation potential. It has been reported that the clouds created were actually "haze" clouds, and the supersaturations generated were not sufficient to force the solution droplets beyond their critical radii (Cotton, 1982).

Calculations of the warming at a convective cloud-base by seeding with NaCl to form artificial rain embryos in concentrations around 10^3 m^{-3} show that it would amount to perhaps 0.02-0.5 ^{o} C (Dennis, 1980). However, reduction of transient thermals by repeated seeding in the updrafts above the cloud base might be a method of attempting to enhance sub-cloud base moisture flux (DeMott, 1982). One-dimensional modeling studies of salt seeding (Hirsch, 1971; Fields and Robinson, 1971) have failed to predict dynamic effects due to seeding. However, in both cases, salt concentrations were too small to produce thermodynamic effects. It has been pointed out that the realization of dynamic effects by using heats of condensation alone is a very difficult task (DeMott, 1982). The local heating is the net difference on the heat of condensation on the hygroscopic particles and the heat lost to evaporating the natural cloud droplets. The buoyancy produced will eventually be lost to evaporation of additional cloud droplets and expansional cooling upon rising. The only net cloud heating possible is the heat of solution of the seeding material. NaCl has a negative heat of solution and the net effect of any seeding with the chemical could be to cool a cloud as a whole. Nevertheless, it has been argued that the critical parameters in seeding for dynamic effects due to condensation heat release are cloud liquid water content, seeding rate and initial dispersion rate. Under certain favourable conditions of these parameters, namely cloud liquid water content 0.5 gm m^{-3}, seeding rate, 500 gm s^{-1} and initial dispersion rate, 10 m s^{-1} in horizontal and vertical, seeding through updraft regions with NaCl particles of diameter 10 μm, could produce non-trivial amounts of dynamic effects. Local liquid water contents would be higher by as much as 100 % when 10 % of the updraft would become buoyant temporarily by as much as 1-2°C. Knowledge and control of the above critical parameters is essential to ensure that massive seeding might indeed create local buoyant plumes in cumulus cloud updrafts. Salt seeding in this manner will affect microphysical processes as well.

Observations of possible dynamic effects by hygroscopic seeding have been recorded by two research groups in the past twenty years, one at the Naval Weapons Center (NWC) in USA and the other at the Indian Institute of Tropical Meteorology (IITM). In hygroscopic field tests conducted by NWC (Clark et al., 1972) clouds seeded in the updraft at or below mid-cloud level (before the mature stage) with solutions of Ammonium Nitrate, urea, and water displayed accelerated vertical growth, increased updrafts and liquid water content and extended life times. The Indian Institute of Tropical Meteorology has been conducting salt seeding experiments in Maharashtra State since 1973 using instrumented aircraft. The cloud physical observations before and after seeding the clouds and in seeded and not-seeded clouds showed the evidence of possible dynamic effect of salt seeding (Murty et al., 1975; Parasnis et al., 1980a, 1982). These observations suggested rise in cloud temperature (1-2ºC) increase in cloud liquid water content (upto 200 %) and increase in cloud depth upto (60 %) in all the seeded clouds whose initial thickness was more than 500 ft at the time of seeding. The observations relating to liquid water content (LWC), temperature, cloud dimensions together with seeding details are given. Similar dynamic effects of salt seeding were noticed in the course of experiments conducted over the ocean region ((Murty et al., 1981). The observational evidence of the dynamic effect of salt seeding described above is corroborated from the results of cloud-base horizontal temperature spectra and the cloud liquid water content (LWC) variations in seeded and not-seeded clouds (Parasnis et al., 1982).

Study of the turbulence in seeded and not-seeded clouds is important for the understanding of dynamic responses of warm clouds to salt seeding as it enables to identify the sources and sinks of energy (Mary Selvam et al., 1983a). The temperature spectra in clear air and cloud air tend to follow the -5/3 power law (de Almeida, 1979). The longer wavelengths of temperature spectra inside the cloud are generated as a result of latent heat of condensation and shorter wavelengths primarily represent the small scale turbulence (Warner, 1970). Hence variations noticed in the in-cloud temperature spectra can beneficially be used for detecting the dynamical responses of warm clouds to salt seeding. Results of the high resolution observations of temperature and cloud liquid water content obtained during air craft penetrations at a single level in warm cumulus clouds before and after massive salt seeding indicated the following. The slope of the spectra relating to seeded traverses increased when LWC increased and rain formed. The temperature of the seeded traverses showed a net energy loss in the shorter wavelengths. The net energy gain is due to condensation of water vapour on the salt particles, the net energy loss to the decrease in the small-scale turbulence resulting from the invigoration of the updraft. These features are considered to be manifestations of the alteration of the dynamics of the cloud through salt seeding (Parasnis et al., 1982).

As a part of the warm cloud modification experiment which is being conducted in Maharashtra State by the Indian Institute of Tropical Meteorology, extensive aircraft measurements of cloud dynamical, microphysical and electrical observations have been made in more than 2000 monsoon warm cumulus clouds (Mary Selvam et al., 1980a). Also aircraft measurements of giant size condensation nuclei, cloud condensation nuclei, Aitken nuclei at cloud base levels and Sodium and Chloride ion concentrations in cloud and rain water samples collected from seeded and not-seeded clouds have been made during the above warm cloud modification experiments (Khemani et al., 1982). Based on these observations and certain physical processes taking place in the atmospheric planetary boundary layer, a simple cloud model has been developed, incorporating the dynamic effects due to salt seeding (Mary Selvam et al., 1983b). These results indicate that massive salt seeding could enhance the sub-cloud-base moisture flux which results in the dynamic modification of warm clouds.

CHAPTER VI

INADVERTENT MODIFICATION

6.1 Introduction

It has long been known, from Radar measurement and other observations, that continental and maritime clouds differ in the ease with which they precipitate. By making extensive measurements of cloud droplet size spectra in continental and maritime type clouds, Squires (1956, 1958a) and Battan and Reiten (1957) were able to show that a relationship existed between cloud microstructure and efficiency of warm rain mechanism. An increase in colloidal stability was associated with an increase in droplet concentration. Further studies by Squires (1958b), Squires and Twomey (1960), Twomey and Warner (1967) and Fitzerald (1972) have shown that cloud microstructure is in turn controlled to a considerable extent by the updraft speed and by the sub-cloud supersaturation spectrum of cloud condensation nuclei (CCN). These studies suggest that the concentration of CCN in an air mass can exert a significant influence on precipitation processes in warm clouds. It may be reasoned therefore, that one of the possible ways in which urban pollution may induce changes in precipitation is by modifying the cloud nucleus content of air passing over a city (Fitzerald and Spyers-Duran, 1973). A number of interesting anomalies of local weather conditions in urban areas have been reported by several workers (Chagnon, 1968; Landsberg, 1974;.Braham, 1976; Khemani and Ramanamurty, 1973; Mary Selvam et al., 1976b). Hence, it is appropriate to summarise the results of investigations relating to inadvertent weather modification effects in this report, particularly those concerning modification of warm clouds.

6.2 Inadvertent modification over urban areas

6.2.1 Positive rainfall anomalies

A comprehensive study of urban effects of weather was started in the 1960's over St. Louis in USA. 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.

Increases in summer rainfall amounting to 25 % in the downwind of St. Louis, and even greater percentage increases in thunderstorm and hail, and increases in night time temperatures within and close to the metropolitan area have been reported (Chagnon et al., 1976; Huff, 1975). Results of METROMEX have been reviewed by Braham (1976). The observed changes in clouds and precipitation have been attributed to several causes like smoke and other man-made cloud condensation nuclei and ice nuclei (Fitzerald and Spyers-Duran, 1973), differences in evapotranspiration rates for urban city and countryside, changes in the mixing properties of the atmospheric planetary boundary layer induced by the high surface roughness of the city with its buildings (Dennis, 1980) and differences in the radiative properties of pavements and buildings as compared to trees or grassland. In another study it has been hypothesized that a combination of the factors described above is responsible for the observed changes in clouds and precipitation (Changnon et al. 1976).

Observations of greater droplet concentrations and narrower droplet distributions in urban and downwind fair weather cumulus rather than in up wind clouds led Braham (1974) and Semonin and Changnon (1974) to hypothesize the instance of giant CCN to explain lower first-echo heights in urban areas, rather than rural areas. Support for the giant CCN hypothesis was provided by Johnson (1976), who found that the average down wind total ultra giant (particles > 5 μm) particle volume was 1.8 times that of the average up wind volume. Ochs and Semonin (1979) carried out a number of numerical experiments using as input aerosol distributions up-wind and down-wind of St. Louis. They concluded that variations in CCN concentrations in the small, large and giant-size ranges did not explain the METROMEX observations of urban-rural first echo differences. Further examination of urban-rural first echo differences reported by Ochs and Johnson (1980) support the conclusion that urban induced dynamic effects were responsible for the observations.

As already mentioned the results of the subsequent studies (Ochs and Semonin, 1979; Ochs and Johnson 1980) have indicated that the aerosol effects on the urban precipitation anomaly area are of secondary importance. However, Changnon et al (1976) conclude that the major impact of the urban area is that it results in a greater frequency of cell merger. Since cell merger has been shown to result in enhanced convective activity and greater rain volumes (Simpson et al., 1971; 1980), this seems to be a plausible connection to the observed rainfall anomaly. Thus it seems likely that the enhanced cell merger activity would be associated with the thermodynamic effects of the urban environment (Cotton, 1982). These concepts, however, require further investigation for confirmation.

Large increases in mean annual precipitation (over 30 % in some cases) adjacent to or down-wind of some large industrial sources of CCN in the State of Washington, USA have been reported (Hobbs et al., 1970). Precipitation anomalies were inferred from stream flow records. It has been postulated that the plume from paper mills contain both higher than normal concentrations of CCN and large particles ≥ 0.2 μm. Subsequent studies (Eagen et al., 1974; Hindman et al., 1977a, b) revealed that the paper mill plumes contained higher concentrations of large (0.2-2 μm diameter) and giant (> 2 μm diameter) CCN, but no significant change in small CCN (<0.2 μm diameter). Moreover, cumulus clouds in the plume contained higher concentrations of large droplets (≥ 30 μm), but smaller concentrations of droplets ≥ 5 μm diameter in clouds outside the plume. This study is interesting because in a region of an inferred significant precipitation anomaly a well defined anomaly has been observed in the aerosol distribution and in the cloud droplet distribution, the aerosol/droplet anomalies being in favour of an enhanced warm rain process (Cotton, 1982). However, Hindman et al. (1977c), using the observed aerosol/cloud droplet distributions as input into an one-dimensional cumulus model and a stratus cloud model, showed that the large and giant CCN emitted by the paper mills are unlikely by themselves to be responsible for the observed 30 % increase in precipitation near the paper mills. Hindman (1976) suggested that the dynamical processes in warm cumulus clouds may be invigorated by heat and moisture released from a paper mill. This invigoration, in conjunction with the acceleration of warm cloud precipitation processes by additional large and giant CCN from the mill, provides a potential mechanism for the observed rainfall anomaly. Many of the clouds that contributed to the observed rainfall anomaly may not wholly be warm clouds and some of them extend deep into supercooled cloud layers. The possibility for a modified warm cloud micro structure to interact with the evolution of precipitation in the ice-phase with the additional potential for dynamic effects, was pointed by Cotton (1982).

A study of the rainfall records for the period 1901 -1969 in the urban industrial city of Bombay, India suggested significant increases in rainfall in the downwind region by about 15

% during the period of increased industrialization from 1941 to 1969 (Khemani and Ramanamurty, 1973). Similarly analysis of rainfall data for the period 1958-1974 indicated rainfall anomalies in the downwind region of the thermal power station at Neyveli, South India. The rainfall in the immediate vicinity of the power plant showed an increase by over 25 % (Mary Selvam et al., 1976b). Aircraft observations relating to atmospheric thermal, microphysical and chemical conditions in the downwind regions of the urban industrial complexes at Bombay indicated the following. The air temperatures, concentrations of gaseous pollutants Sulphate, Ammonium and Nitrate ion concentrations in rain water and concentrations of CCN, giant size condensation nuclei, concentrations of cloud droplets and large cloud drops are significantly higher in the downwind regions compared to those in upwind regions of the urban industrial complexes (Khemani et al., 1980). Similarly aircraft observations of electrical, microphysical and thermal conditions in the upwind and downwind regions of the 1100 MW thermal power station, Obra, North India, suggested that the parameters in the downwind region are systematically higher than those in the upwind region (Mary Selvam et al., 1976a). Under cloud-free conditions, the downwind concentrations of the cloud condensation nuclei and the computed cloud droplet concentrations were higher up to two times, the atmospheric electric field was higher up to seven times and the air temperature up to 2.3°C. Under in-cloud conditions the downwind values of the electric field and the cloud droplet charge were higher up to two times, the measured cloud droplet concentration was higher by one and half times, the cloud air temperature up to 1.5°C and the cloud liquid water content up to four times. The results of aircraft observations (Mary Selvam et al., 1976a; Khemani et al., 1980) indicate that the observed increases in rainfall in the downwind regions of the thermal power plant/urban complexes could be due to enhanced coalescence process in warm clouds, forming in the down-wind region.

6.2.2 Negative rainfall anomalies

Decreases in rainfall observed in some areas have also been attributed to the anomalies in the cloud condensation nuclei concentrations. Warner and Twomey (1967) found that the smoke from sugar cane fires was a prolific source of CCN, and the clouds in the down-wind plume were observed to have higher than normal cloud droplet concentrations. Warner (1968) found that a reduction in rainfall at inland stations over Eastern Australia coincided with increasing sugar production. He could not eliminate the possibility that other climatic factors contributed to the trend. However, Woodcock and Jones (1970) attempted to obtain independent confirmation of Warner's hypothesis that enhanced CCN caused a reduction in rainfall. They analysed precipitation records over Hawaii in two locations. One location was down-wind from a major cane-growing region, while the other was not. Downward trends in rainfall over a 30-60 year period were detected in both areas, but the trends were not statistically significant. They concluded that factors other than sugar cane burning are probably involved in the rainfall trends. Subsequently, Warner (1971) attempted to find independent confirmation of his hypothesis by examining rainfall records in another sugar cane producing region in Eastern Australia. The study did not lend further support to his hypothesis.

6.3 Inadvertent modification over rural areas

Evidence for possible inadvertent modification of precipitation in rural areas has also been reported. Schickendand (1974) has found rainfall increases ranging from 14 to 91 % associated with irrigated regions in the US High Plains. Since surface moisture is a major factor in the surface energy budget, one would expect irrigated areas to be cooler than non-irrigated regions at mid-day and warmer than non-irrigated regions in the early morning hours (Cotton, 1982). This should set up off-irrigated-area flow (divergence) during the day and an on-irrigated-area flow (convergence) during the early morning hours. Since meso-scale convergence has been shown to be a major control on convective precipitation (Ulanski and Garstang, 1978; Chen and Orville, 1980; Tropoli and cotton, 1980), the associated anomalies in meso-scale convergence would be likely contributors to the observed rainfall anomalies.

The dominance of soil moisture content variations (hence latent heat transport changes) in surface albedo on predicted meso-scale circulations also has been demonstrated by Mahrer and Pielke (1978). They found that surface moisture completely masked the effect of surface albedo changes in the simulation of land-use changes over the Negev and Sinai desert areas.

CHAPTER VII

NUMERICAL SIMULATION OF MODIFICATION EXPERIMENTS

One important requirement in a cloud seeding experiment is availability of means to detect, at an acceptable level of significance, any increase in rainfall resulting from seeding. The probability of detecting prescribed increase in rainfall due to seeding in a given area with a specified degree of confidence can be determined by undertaking computer simulation experiments. A numerical technique for the purpose has been developed by Twomey and Robertson (1973). Using this technique numerical simulation experiments were carried out for selected areas in Australia (Twomey and Robertson 1973; Smith and Shaw, 1976). These experiments require a great deal of computer time even on high speed modern computers. A different simulation technique has been developed and tested (Mary Selvam et al., 1979). This technique not only reduces the computational time by an order of magnitude, but also defines the exact lower limit for the double ratio value which can be detected at 5% level of significance. Using this technique and that of Twomey and Robertson (1973) numerical simulation experiments were conducted for a few selected regions in India (Mary Selvam et al., 1978, 1979, 1980b, 1980c, 1980d).

Recently, a new simulation technique has been developed (Mary Selvam et al., 1984). This technique is simple and involves application of ratio estimators and requires coefficients of rainfall variation and correlation of target and control areas as input data. Using the coefficients of rainfall variability computed from the weekly total rainfall of the 35 meteorological sub-divisions in India for the 5-year period 1976-1980, numerical simulation experiments were carried out. Nomograms for detection of prescribed increases in rainfall and a map showing probabilities of their detection have been prepared for different regions in India (Mary Selvam et al., 1983a). The nomograms can be used for any region in the world provided the rainfall variability in that region is known. This technique will also be helpful for identifying suitable regions in the world for undertaking weather modification experiments.

CHAPTER VIII

CLIMATIC ZONES FAVOURABLE FOR WARM CLOUD MODIFICATION

The climatic zones in the world favourable for cloud seeding with hygroscopic nuclei have been identified (Fournier d'Albe, 1976). The criteria considered in this regard are:

a) Convective cloud base temperature should exceed 10^{o}C.

b) Frequency of occurrence of convective clouds should be sufficient to justify seeding operations, and

c) Natural concentration of giant hygroscopic particles in the atmosphere should be sufficiently low to allow significant increase to be achieved by seeding.

Cloud base temperatures have been estimated by using the aerological data (i.e. by evaluating height of the lifting condensation level). The estimated cloud base temperatures have been plotted at intervals of 5^{o}C for January, April, July and October. The 10^{o}C isotherm which delimits the area in the world which are suitable for cloud seeding with hygroscopic particles are shown in figures 1-2 (Fournier d'Albe, 1976). The monthly rainfall has been taken as an index for the occurrence of suitable clouds for seeding. Rainfall of 25mm in the reference month has been taken as an index for lack of suitable clouds and a monthly rainfall of more than 100mm is considered to be the limit for not undertaking cloud seeding. Monthly rainfall between 25 and 100mm is, therefore, taken as the index of favourable conditions for seeding. The vertically hatched areas in the map are those with rainfall between 25-100mm in the reference month and with cloud base temperature higher than 10^{o}C. However, some of the areas thus delimited have abundant rain at other seasons of the year. Cross-hatching areas are those with (i) cloud base temperature higher than 10^{o}C in the reference month. (ii) rainfall in the month between 25 and 100mm and (iii) annual rainfall less than 1000mm. If the annual rainfall exceeds 1000mm attempts to increase rainfall artificially may not be judged economically rewarding.

The concentration of giant hygroscopic nuclei of natural origin in the atmosphere is found to vary from less than 10 to more than 1000m^{-3} (Fournier d'Albe, 1976). The higher value is commonly observed in maritime air particularly in areas with frequent, moderate or strong winds. The lower value is characteristic of inland areas more than 300km from the nearest coast or separated from the coast by mountain ranges. A concentration of 1 particle per m^{3} of mass 10^{-9} gm corresponds to 1 kg of salt per 1000 km^{-3} (the volume of air likely to be involved in a seeding experiment with a target area of 1000 km^{-2}). With appropriate dispersion techniques, it is thus feasible to achieve a significant increase in the concentration of giant hygroscopic nuclei over an area of this extent provided that the natural background concentration does not exceed about 100 particles per m^{3}.

CHAPTER IX

CONCLUSIONS AND RECOMMENDATIONS

The number of warm cloud experiments carried out to test the feasibility of increasing rainfall by artificially enhancing the efficiency of the collision-coalescence process is limited. Only a few experiments have indicated change in the radar echo characteristics and increase in rainfall. However, none of them has the requisite combination of successful rainfall increases based on physical and statistical evidence.

The fundamental obstacle to developing a reliable warm cloud modification technology is lack of adequate understanding of the precipitation processes and the dynamics of warm clouds. The high natural variability of clouds and rainfall makes it difficult to detect the seeding effects. The above obstacles can be overcome to a certain extent by undertaking extensive aircraft cloud physical observations in the areas of the experiments and evaluation of the suitability of the sites for undertaking warm cloud modification through numerical simulation experiments. A numerical simulation experiment for the simulation of weather modification experiments and a cumulus cloud model for understanding the precipitation processes have recently been developed at the Indian Institute of Tropical Meteorology, Pune, India.

At present, a randomized warm cloud modification experiment with cross-over design is underway only in one country (India). In this experiment, hygroscopic particles are released from aircraft inside the clouds, a few hundred meters above the cloud base and extensive cloud physical observations are being collected besides rainfall data from rain gauges located in the test areas. Statistical analysis of the rainfall data, relating to the experiment conducted on 93 pairs of days spread over a period of seven monsoon seasons has indicated increases in rainfall by 6.6 per cent, significant at 15% level. Results of the numerical simulation experiment for the test areas have suggested that the minimum period for detection of 5% and 10% increases in rainfall due to seeding is respectively 28 and 14 years. Aircraft observations of cloud microphysical, dynamical and electrical parameters in seeded and not-seeded clouds, chemical composition (chloride and sodium ion concentrations) of cloud and rain water samples collected from aircraft in seeded and not-seeded clouds and the concentration of giant size condensation nuclei at cloud-base levels before and after seeding showed significant differences.

The need for increasing rainfall is felt by many tropical countries due to fluctuations in the annual rainfall and weather systems particularly in the recent past. In many of these areas there are clouds that precipitate inefficiently and therefore opportunities to increase rainfall through warm cloud modification techniques may exist. There is, therefore, need to undertake statistically well-designed experiments and organize cloud physical observations in suitable areas in many countries. Such effort will help to obtain scientific proof of the efficiency of warm cloud modification techniques.

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