Fundamentals of Weather Modification

The purpose of these lectures can be enounced in a simple sentence: **to help pilots to become good cloud physics observers**.

Cloud Physics and Chemistry Primary
Historical Brief

On July 12th, 1946 Vincent Schaefer discovered by accident that supercooled water can be transformed into ice using dry ice (solid CO2);

Four months later, on November 13th, Schaefer dropped about 3.3 pounds of dry ice pellets from a light aircraft into a supercooled lenticular stratocumulus over Western Massachusetts. After about five minutes the cloud turned into snowflakes, which started to fall into a 2000 foot-dry layer below before subliming completely;

On November 14th, whereas collaborated with Schaefer and Irving Langmuir, **Bernard Vonnegut** found that silver iodide aerosols were excellent ice-nuclei;

On November 15th Schaefer's landmark paper appeared published in Science (Schaefer, 1946);

After these pristine issues a plethora of experimental and operational projects were developed in different countries around the world, which have made the history of the discipline rich although winding. There has been periods of hectic activities followed by other of oblivion (the hydro-illogic cycle). In general we can talk about four main stages:

First stage (1946-1980), a period of intense research and optimistic applications;

Second stage (1981-1990), a period of skeptical applied research and operations;

Third stage (1991-2000), a period of scarce funding and the development of proper techniques and technologies; these new skills and tools have permitted the birth of the called "scientific management".

I hope we are entering in a new stage (**the fourth one**) in which operational programs will approach the structure of applied scientific research without abandoning their focus in human needs.

2.2 Microphysical and chemical bases

2.2.1 Atmospheric Moisture

Clouds are formed by the lifting of moist air which cools by expansion as it reaches falling pressures at higher levels; therefore, the first factor to consider in our lecture is the atmospheric moisture. Gases in the atmosphere are within the range of pressures and temperatures in which their state is well represented by the equation for an ideal gas:

$$pV = R\,T \qquad \text{(equation 2.2.1.}\mathbf{1}\text{)}$$

If V is the volume occupied by a mole of the gas, R is called the universal constant for all the gases (= 8.3144 x 10ergs K mole).

In the meteorological applications we usually use the specific volume, which is the volume of a weight unit (the inverse of density), v = V / m, being m the molecular weight. The equation becomes then:

$pv = (R / m) T = RT$ or $p = d RT$ (equation 2.2.1.**2**)

where R refers only to the gas having molecular weight **m** and density **d**.

What is a mole? It is a counting unit. A mole of anything is the amount of this thing that contains

6.022 X 100 000 000 000 000 000 000 000

simplest entities of this thing.

Problem: What is the mass, in grams, of one atom of carbon?

The atomic mass of carbon is 12.011 amu (see the Periodic Table of the Elements).

One mole of carbon has a mass of 12.011 g and contains
6.022 x 100 000 000 000 000 000 000 000 atoms of carbon. Therefore:

1 atom of carbon has a mass of

1.995 x 0.000 000 000 000 000 000 000 01 grams.

Dry air consists of 78.08 % of molecular nitrogen (N), 20.95 % of molecular oxygen (O), 0.93 % of argon (Ar), and 0.0365 % of carbon dioxide (CO) by volume (it is an excellent approximation), the “molecular weight” of dry air becomes

$$m = (0.7808 \times 28.0134 \text{ amu}) + (0.2095 \times 31.9988 \text{ amu}) + (0.0093 \times 39.948 \text{ amu}) + (0.0004 \times 44.0098 \text{ amu})$$

$$= 28.9657 \text{ amu} \sim 28.97 \text{ amu}$$

then a mole of dry air has a mass of 28.97 grams (and a volume of 22.4 liters at **s**tandard **t**emperature and **p**ressure (**stp**) which means 0 °C and 1 atmosphere).

In the case of moist air which is composed by dry air and water vapor (it is the fifth major component, but its concentration is variable ranging from 0.5 to 3.5 %) a typical calculation would give us a small value because the molecular mass of water vapor (HOH) is

Molecular mass of water = 2x1.0079 + 15.994 = 18.0098 amu

which is smaller than the corresponding values for the other gases in air.

Therefore, we can write our first important conclusion:

Moist air is lighter than dry air. Additionally, we know that the concentration of water vapor in moist air is variable and ranging from 0.5 to 3.5 %.

Above 80 km (~262 320 feet), the concentrations of these major species begin to change significantly, due mainly to photochemical processes that cause the dissociation of dinitrogen (NN) and dioxygen (OO). There is where cosmos begins.

There are more expressions for the water-vapor content. We already used the concentration of mixing ratio in our previous considerations, which is defined as the quotient between the density of water vapor and the density of dry air. This quantity is of the order of

0.01 (remember its range from 0.005 to 0.035).

Water vapor is said to be at saturation at a given temperature when it is in equilibrium with a flat surface of pure water at that temperature.

The state of equilibrium means that there is no net movement of molecules between the two phases when they are in contact with each other. The saturation values of water vapor pressure, density, and mixing ratio describe the water vapor content under these conditions. Supersaturation exists when these values for a given temperature are exceeded, and subsaturation exists when the vapor content is lower than that represented by these values.

$pv = (R / m)\, T = RT$ or $p = d\, RT$

Table 1: Saturated water vapor pressure and density values at some temperatures

Temperature (°C)	saturated vapor pressure (mb)	saturated vapor density (g/m)
- 10	2.86	2.36
0	6.11	4.85
10	12.25	9.40
20	23.33	17.30
30	42.29	30.40
40	73.55	51.10

The approximated mathematical expression for saturated vapor pressure associated to this table is:

p_{sat} = 6.11x 10 ^ {7.5 T / (237.3 + T)} (equation 2.2.1.**6**)

whereas the mathematical expression for the actual water vapor pressure uses the dew point temperature:

p = = 6.11x 10 ^ {7.5 T_d / (237.3 +T_d} (equation 2.2.1.**7**)

The saturation of water vapor pressure over water and over ice are different, as the molecular forces bind much more in an ice crystal than in a water bubble. Therefore, the saturation pressure over ice is smaller than over water. The expression for the former is:

$$e = 6.11 \times 10^{\{9.5\, T/ (265.5 + T)\}} \qquad \text{(equation 2.2.1.}\mathbf{9}\text{)}$$

The saturation water vapor pressure over ice at – 10 °C is:

$$= 6.11 \times 10^{\{9.5\, X(-10)/ (265.5-10)\}} = 2.60 \text{ mb} < 2.86 \text{ mb}$$

Precisely, this difference is the basis of the Bergeron-Findeisen mechanism for the formation of precipitation which states that

“ice particles will grow at the expenses of vapor and liquid water through deposition of vapor over them”.

2.2.2 Atmospheric aerosols

In addition to its gaseous constituents, the low atmosphere contains quantities of suspended material both liquid and solid, and although their concentrations are relatively small, these aerosols play a disproportionately important role in the atmosphere. These particles can be charged or uncharged. The smaller particles affect the electrical and optical properties of the atmosphere, where the large and giant particles serve as nuclei for the condensation of atmospheric water.

Size: Atmospheric aerosols cover a size range from below 0.01 μm to over 10 μm in diameter.

Particles with diameters of **less than 0.2 μm** are called **Aitken particles**;

those with diameters between **0.2 and 2 μm** are called **large particles**;

and the others with diameters **in excess of 2 μm** are called **giant particles**.

Recently, particles with diameters **in excess of 20 μm** have been called **ultra-giant particles.**

Concentration: The total aerosol concentration varies widely, from as low as 1 000 000 000 per cubic meter in clean country air to over 100 000 000 000 per cubic meter in heavily polluted areas.

However, not only is the total aerosol concentration higher in the polluted air, but the particle size distribution is different. The pollution consists mainly of large particles and particles near the upper limit of the Aitken range in concentration that ranged from about 100 000 000 to over 100 000 000 000 per cubic meter in polluted continental air.

Pollutants are defined as chemicals that are present in the air in sufficiently high concentrations to be harmful to humans, other species, or ecosystems as a whole.

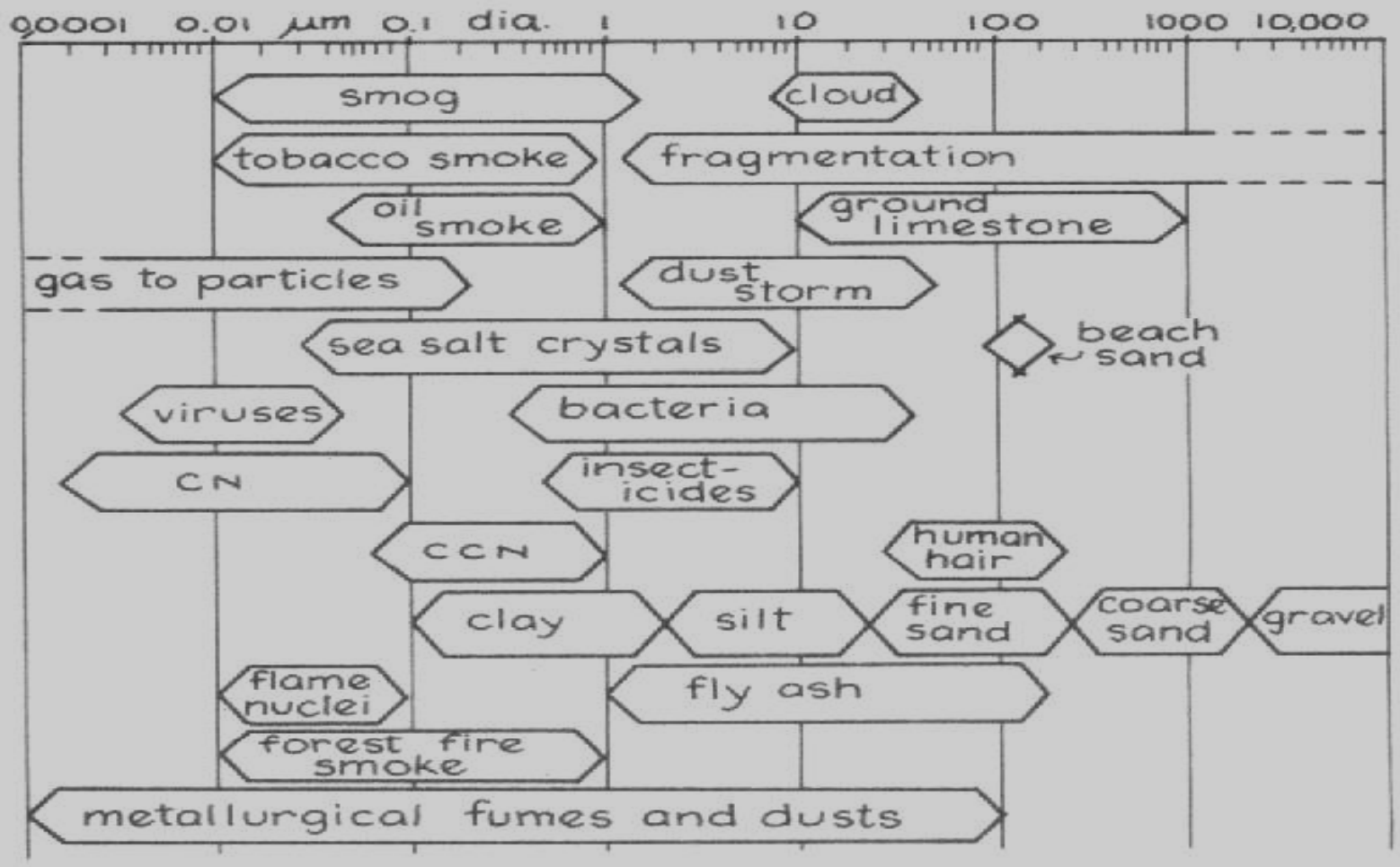


Fig. 1 Size range of airborne particles.

From "The Peterson Field Guide Series", page 4

Vincent J. Schaefer and John A. Day

Houghton Mifflin Company Boston, 1981

Mechanics: The concentration and size distribution of the atmospheric aerosols are controlled by the initial size distribution itself, and by the processes of collision, coagulation, and removing. These processes are strongly size dependent.

- Fall speed: An aerosol particle in free fall quickly reaches its terminal speed, U, at which point the downward force of gravity corrected for the buoyancy of the medium, is exactly balanced by the drag force.
- Brownian Motion, Diffusion, and Coagulation: In addition, aerosol particles suffer bombardment by air molecules and undergo in erratic motions know as Brownian motion. Large particles are struck on all sides and because of their relative great masses they suffer less perturbation than the sufficient small particles, which are diffused better. Precisely, the diffusion leads to eventual collisions among the particles and to coagulation. The most likely collisions of all are those between small and large particles, because the former move more whereas the latter have appreciable collision cross sections, for example, the probability that a 10 μm-diameter particles collides with a 0.1 μm-diameter particle is over 70 times the probability of collision between two 10 μm-diameter particles.

- Removal: The tendency for aerosol particles to coagulate into larger ones provides a mechanism for cleansing the atmosphere because the larger particles have appreciable fall speed. However, the turbulence of the atmosphere, wind, and convection are factors that distribute the particles widely and block to some extension the removal. Furthermore, aerosol particles, particularly hygroscopic ones, are removed from the atmosphere so effectively by clouds and precipitation that direct deposition on the earth's surface is of secondary importance. This type of removal (wet removal) has been coined as **washout** when it is done by clouds and as **fallout** when it is done by precipitation (rain, hail, snow). **Washout** is very efficient, mainly because of the earth's cloud coverage (20- 45 %; 1- 5 % of the troposphere volume), the very large surface area provided by cloud droplets, and the effects of condensation and coalescence in transferring aerosol particles from a small size range abruptly into a much larger size range, and eventually become fallout. Three main factors have been identified in the dynamics of fallout:
  - Fallout always leaves behind a large amount of heat that appeared when water was condensed and/or frozen;
  - Fallout causes downdraft by applying its weight to the air through which it falls, and by cooling that air by evaporation into it (downburst, microburst);

- Fallout alters the appearance of nearby clouds by its falling motion which cools the air below cloud (or beside cloud). The downdraft of cold air mimics a cold front at microscale which can create an outflow boundary easy to detect with Doppler radars, and sometimes with conventional radars because the temperature discontinuity, birds soaring in the up-currents ahead, and also insects. Such cold outflows can be filled with dust (typical situation in West Texas). On the other hand, the outflow boundary can promote the formation of new clouds with cellular appearance since it may prevent convection in some places and trigger it in other places.
- On the other hand, washout has been well analyzed. It has been noted that every cloud droplet contains a CCN. As a typical raindrop consists of approximately one million collected cloud droplets, each raindrop brings to the earth's surface at least about one million aerosol particles. Cloud droplets that exist for any appreciable time collect additional aerosols. This process is called **scavenging** and explains how the clouds and precipitation can clean the air. Observations during fog and drizzle have indicated that the concentration of aerosols can drop from 1000 000 000 to 10 000 000 per cubic meter over 1 hour period.

- Vertical distribution: In the normal atmosphere the concentration of aerosol particles of a given size decreases with height. Observational data fit exponential functions.

- Sedimentation: The final stage in the removal of an aerosol particle from the atmosphere is its impaction upon a collector surface. Some surfaces are more effective than other in removing the particles. For example, forests are very effective scavengers of aerosol particles and produce a great cleansing action called “green area effect”.

- **Chemistry**: Solar radiation capable of inducing chemical reactions among the gaseous components of the earth's atmosphere is strongly absorbed before reaching the troposphere and therefore, there is relatively little chemistry involving the major constituents of the lower atmosphere. The exceptions are the reactions between oxygen and nitrogen produced by electrical discharges during thunderstorms. However, aerosol particles can act as sites for chemical reaction to take place. The most significant of these reactions are those that lead to the destruction of the stratospheric ozone (O). In general there are at least two types of chemical processes of great impact: the chemistry of air pollution, and the phase changes of water, which are the most obvious features of Weather.

- Aerosol particles exist in a rich variety that includes aqueous solutions, smokes, spores, pollen, asbestos fibers, continental dust, talc, volcanic emissions, and so on. Their chemical composition is highly variable whereas, due to their sizes, interfacial processes appear to be of prime importance with aerosols acting as catalysts for much of the chemistry that occurs in the lower levels.

Origins of aerosol particles:

Two types of aerosols can be easy isolated according to the major two sources: continental and marine. These classification has been later split in detail as the following table shows:

Aerosol	Natural (N) or anthropogenic (A)	Annual flux (10 Tg/year)
Sea spray	N	1000-1500
Dust	N, A	100-750
Forest Fires	N, A	35-100
Volcanic variable)	N	50 (highly
Meteors	N	1
Combustion	A	50
Condensation	N, A	1500

The total annual global production of aerosol particles appears to be between

2500 and 4000 x 10Tg/year.

From measurements at the ground, Junge in 1958 found a mathematical expression that relates the concentration of aerosols and their sizes: the amount of aerosols per a given volume of air is about the same in all size classes from about 0.2 to 20 μm diameter.

Ions form a special class of aerosols. There is an inverse relation between the number of small ions and the number of larger aerosols at any one time and place.

The common aerosols capture and immobilize the ions. About a half of the aerosols in the 0.02 to 0.2 μm diameter range carry a net charge as a result of ion capture and therefore, they are classed as large ions. Small ions become important for condensation only when the air is cleaned of other aerosols.

Not all the aerosol particles become CCN. In general the amount of CCN depends on the local supersaturation in a specific region. The vast majority of CCN are the large nuclei (diameters between 0.2 and 2 μm).

It is necessary to distinguish between the **hygroscopic particles,** which readily take on water, and the **hydrophobic ones**, which do not. The chemical analysis of large nuclei shows that ammonium sulfate $(NH_3)_2\ SO_4$

is the most common constituent, which indicates that the source is land rather than sea. This constituent is produced by gas-to particle conversions. Its concentrations over the ocean are appreciable too.

The insoluble particles play an important role in the nucleation of ice. Most natural ice nuclei (**IN**) are insoluble clay particles picked up from the ground by the wind. Considering activity and abundance together, kaolin minerals (complex silicates), with a threshold temperature around – 9 °C, are among the most the most important Ice nuclei. Data from West Texas suggest that wind speed of 12 -15 m/s (~23-29 knots), depending on direction, are sufficient to raise dust clouds. The size spectrum of particles raised by the wind also depends on the wind speed. Winds of 25 m/s (~48 knots) can raise particles as large as 100-200 μm into the atmosphere although these particles are too heavy to stay longer. The most numerous particles are in the range of 2 to 20 μm. The finer dust particles remain suspended for several days in some cases and travel long distances. Dust particles from the Sahara Desert have been identified in the atmosphere over the Caribbean islands.

There are four modes of IN activation:

Deposition (sometimes called sublimation);

Condensation-freezing (sometimes called soption)

Contact nucleation

Bulk freezing (immersion freezing)

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

$$N(D) = N \exp [A(D)] \quad ; D \text{ between } 10\text{-}30°C \quad \text{(equation 2.2.2.}\mathbf{2}\text{)}$$

where D is the supercooling in degrees Celsius D = -T, N(D) is the concentration of active nuclei, and Nand A are adjustable parameters. A typical value for Nis 100 per m(10 per liter) while A varies from 0.4 to 0.8, and is usually near 0.6. The table below shows values of N(D) using N= 100 and A = 0.6:

Temperature	**N(D)**
- 10 °C	0.004 per liter
- 20 °C	1.628 per liter
- 25 °C	32.69 per liter
- 30 °C	656.6 per liter

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

2.2.3 The Formation of Clouds and Precipitation

In the previous paragraphs we described the factors that control the formation of water clouds in the earth's atmosphere, now we will study the whole process of formation of clouds and precipitation.

Clouds form wherever air is cooled below its dew point, whether by radiation, by mixing with cooler air, or by ascent in the atmosphere with the resultant decompression. The amount of water vapor which can exist in equilibrium with a plane surface of pure water is a function of the temperature only (remember equation 2.2.1.5). Any water vapor in excess of saturation is in principle available for formation of a water cloud. We define the saturation ratio as

$S = p/ p_{sat}$ for water (equation 2.2.3.1)

$S = e/ e_{sat}$ for ice (equation 2.2.3.2)

Supersaturation is reached when there is an excess of water vapor available for the formation of cloud, and values of S are then greater than 1.

Because of the surface tension effects, there is energy stored in all water surfaces. The vapor pressure required to maintain a small water droplet in equilibrium with its environment is greater than that required to maintain equilibrium above a plane surface of pure water at the same temperature. The formula associated is:

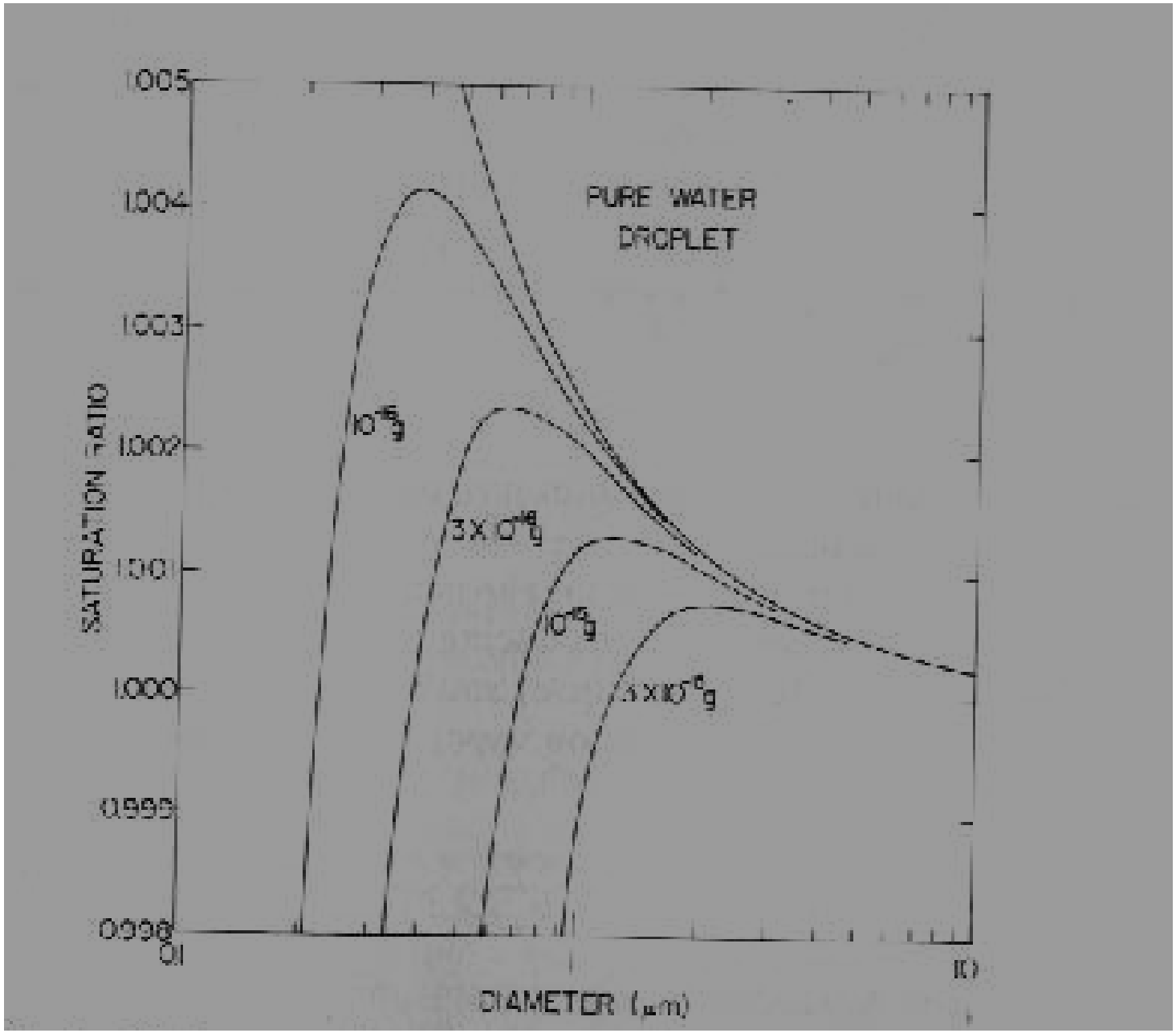
$$p_{sat\ droplet} = p_{sat} \exp [4\gamma/ (dRT\ D)] \sim p_{sat} \{1 + 4\gamma/ (dRT\ D)\}$$

(equation 2.2.3.**3**)

Really, clouds form by heterogeneous nucleation upon cloud condensation nuclei. The reason of this phenomenon is that the presence of dissolved solute in water reduces the saturation pressure (Raoult's law) and helps in the formation of droplets. The expression for the droplet saturation pressure becomes:

$$p_{sat\ droplet} = p_{sat} [1+\{4\gamma/ (d_w RT D)\} – \{6iM_w m_s / (\pi d_L M_s D^3\}]$$

(equation 2.2.3.**4**)


1.005
1.004
1.003
1.002
1.001
1.000
0.999
0.998
SATURATION RATIO
PURE WATER
DROPLET
10^{-16}g
3×10^{-16}g
10^{-15}g
3×10^{-15}g
0.1
1.0
DIAMETER (μm)

In general we can say that the process of warm rain formation has the following steps:

Air cooling followed by Heterogeneous Nucleation of water vapor;

Growth by condensation;

Definite growth by collision-coalescence (coalescence means that cloud droplets are captured by larger droplets).

The process of coalescence is very interesting depends on the size of the ‘catcher’ (or collector). Observational studies, experiments and models have showed that the process becomes efficient only if the collectors have diameters greater than 38 μm (the Hocking limit). The coalescence process is stochastic (random), and a collector (a lucky large droplet) increases its mass dramatically, whereas others (the unlucky ones) almost do not change, but later the race will favor other collectors and drops of intermediate sizes are able to capture small droplets while they can be captured by larger drops. The computer simulations indicate that the coalescence can lead to the appearance of raindrops in 20-30 minutes, which agrees with the observations. These simulations confirm also the importance of the initial cloud droplet spectrum.

Droplet concentrations are lower in maritime than in continental clouds, with typical values of 50 and 500 cm respectively (but with very high variability in both cases). For example, cumulus with strong updrafts in heavily polluted air have exhibited droplet concentrations as high as 1500 cm. Clouds over the oceans do not exhibit the strong updrafts characteristic of continental clouds, although the main reason for the difference in cloud droplet concentration appears to lie in the differences between aerosol distributions in maritime and continental air-masses. The aerosol distribution over the ocean generally has fewer of the large hygroscopic nuclei than the continental one contains.

A similar theory has been development for the growth of ice crystal from water vapor. It is important to say however, that the ice particles may appear in the clouds either as new particles or through freezing of supercooled cloud droplets and drops. Ice crystals take on a wide variety of forms, all of them basically hexagonal structures, that we call **habits**. Ice crystal habits depend on temperature and ice saturation ratio, as shown in the following graphic:

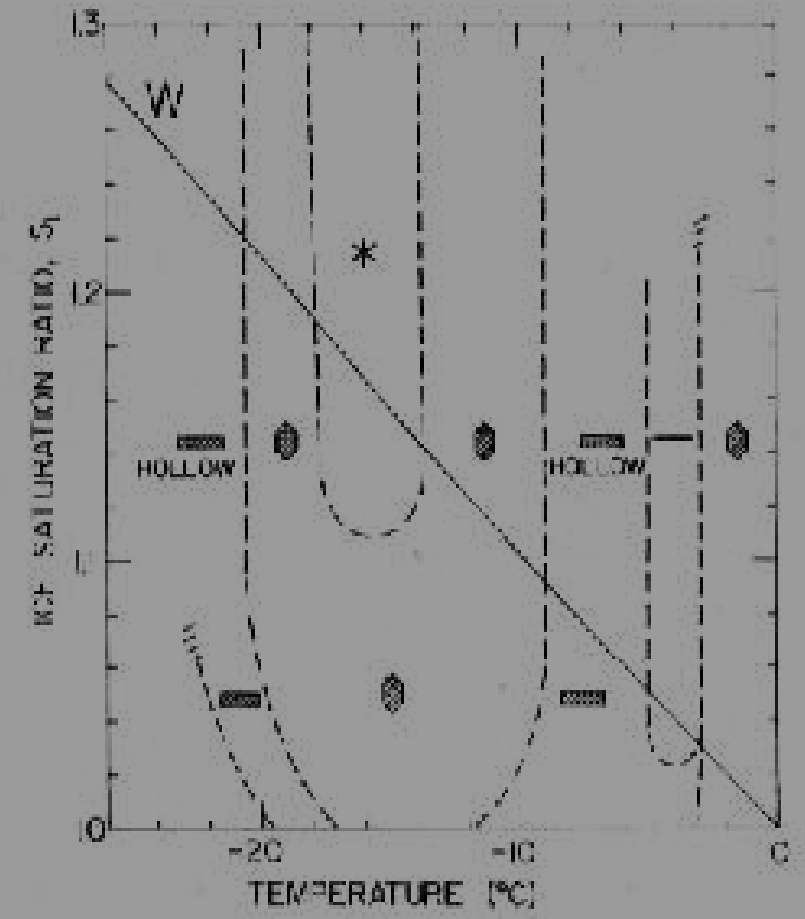
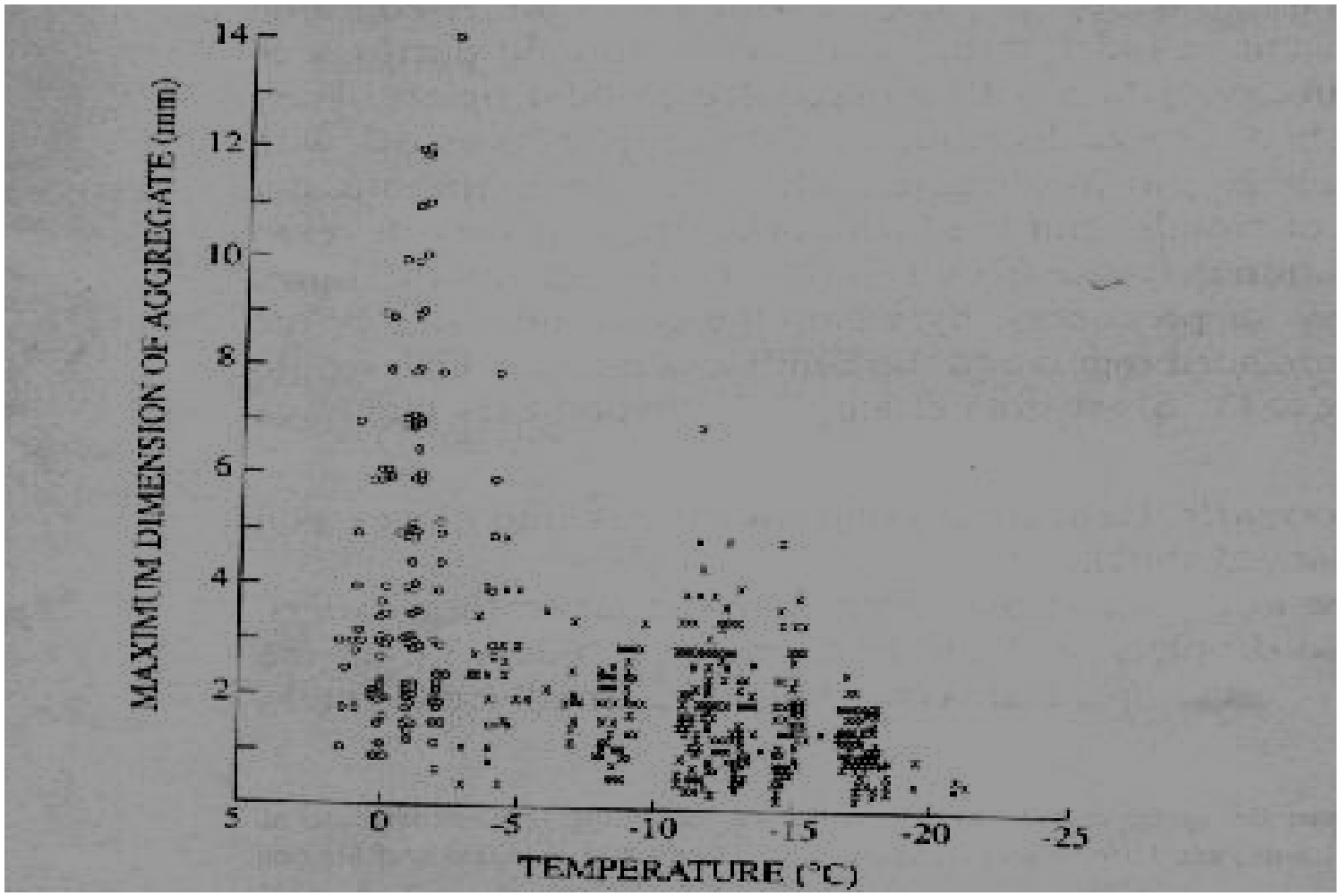


Fig. 3.7. Predominant ice crystal habits as a function of temperature and ice saturation ratio. Line W indicates water saturation. ——— indicates needles, ▬ indicates columns, ⬮ indicates plates, * indicates dendrites and stellars.

Ice crystals can grow by **deposition of water vapor** (some scientists do not believe deposition is a real process in the atmosphere), which reaches its maximum near – 12 °C. Furthermore, two processes of ice multiplication have been identified: **aggregation** and **riming** (some authors use the term accretion for both processes).

If ice particles collect other ice particles, the process is called **aggregation**. This process depends strongly on temperature. The probability of adhesion of colliding ice particles becomes much greater when the temperature increases to above – 5°C, at which the surfaces of ice crystals become sticky. Below – 20 °C aggregation does not appear to exist. A secondary maximum occurs between – 10 and – 16 °C, where dendritic crystals become entangled. See graphic below:


MAXIMUM DIMENSION OF AGGREGATE (mm)
14
12
10
8
6
4
2
0
0
-5
-10
-15
-20
-25
TEMPERATURE (°C)

If ice particles collect liquid drops the process is called riming. Extreme riming produces hailstones. These particles are commonly 1 cm in diameter but have been observed to be as large as 10-15 cm. When the temperature remains below zero during the process of growth, the hailstone remains dry, whereas if the temperature rises to zero, the melting process produces a wet growth of spongy hail. Moreover, the time required for an ice crystal to reach its riming threshold is minimized if it can grow below – 15 °C.

In general natural ice enhancement can occur in many different ways enumerated below:

Fragmentation of ice crystal due to collisions and/or thermal shock;

Ice splinter in riming (- 3 to – 8 °C);

Contact nucleation;

Deposition-nucleation;

Aggregation;

Riming

2.3.4 Concepts of Cloud Modification

2.3.4.1 Modification of CCN Spectrum

It has been suggested that seeding operations could cause continental clouds to assume the drop size distribution characteristic of maritime clouds, and therefore, increase their chances of producing precipitation. One alternative is called **nucleus poisoning**, which consists in the addition of some chemical that would deactivate many of the CCN. However, the idea has been rejected (impractical). A more promising approach would be to add artificial CCN of sufficient size and in sufficient quantity to prevent the activation of the natural CCN. By introducing particles with diameters of the order of 1 – 3 μm in concentration of 25 – 100 per cubic cm one could, in certain cases, ensure the formation of a maritime cloud even in the most polluted continental air-mass. The large artificial CCN would capture the available moisture and prevent the more numerous, but smaller, CCN from participating in the cloud formation process. However, practical limitations are in place. Calculations in the case of a vigorous cumulus cloud which ingest about 1million cubic meter of air per second indicate that we would need 30 kg of sodium chloride per minute to obtain the desire particle concentration of 50 per cubic cm. The logistics involved in applying this seeding concept to an entire cloud are formidable. Nevertheless, one could think that the formation of precipitation particles in a cloud can act as an infection, which may contaminate the cloud after a while once the precipitation gets underway anywhere in a cloud. This idea is more feasible.

A second approach might be the introduction of artificial raindrop embryos. The water spray seeding method has the disadvantage that very large quantities of water must be transported to a cloud by aircraft to produce detectable effect.

One way to reduce the logistical problem is to treat the cloud with hygroscopic agents, either dry particles or spray droplets with the sizes of giant or ultra-giant CCN. Actual recommendations suggest the use of ultra-giant hygroscopic particles 50 μm-diameter in concentrations of a few per liter. The basic intent of this hygroscopic seeding is to accelerate the coalescence process through the modification of the cloud droplet size distribution. However, that modification can also affect ice phase processes in the clouds if they reach temperatures below 0°C.

The effect of seeding clouds with different background CCN concentrations shows that those growing in an environment with less than about 350 CCN per cubic cm, having more maritime characteristics with enough large size CCN, are not responding favorably to hygroscopic seeding. Apparently, the presence of the large natural nuclei produced large drops on their own, preventing the seeded particles from growing. One important feature of the results from modeling is that for hygroscopic seeding to be effective it has to take place within a very narrow time window, corresponding to the time when rapid growth of the drops is just beginning.

The previous considerations are very encouraging, but it is important to point out a number of cautionary points:

The hygroscopic seeding technique has not yet been demonstrated to be cost beneficial for an area-wide program;

The hygroscopic seeding results can not be automatically transferred to a new geographic area, since the background aerosol and other related variables must be studied;

The process in general is not well-known;

Without this physical understanding we cannot have full confidence about the statistical results.

2.3.4.2 Glaciogenic Seeding

Glaciogenic seeding is defined as seeding designed to add ice particles to clouds or to portions of the clear atmosphere. This objective can be accomplish in two ways, (1) by chilling the air to temperatures below –40°C, where homogeneous ice nucleation takes place, or sufficiently to activate the natural ice nuclei present, and (2) by adding artificial ice nuclei capable of producing ice particles.

The first way is almost impractical because of economical reasons. The second way is in use in Texas and around the world.

Two modes of glaciogenic seeding have been defined, the static mode, and the dynamic mode. Early studies demonstrated that concentration about 50- 100 ice crystal per liter can suppress riming and favor aggregation. This concentration range defined the border between the two modes of glaciogenic seeding. Glaciogenic seeding releases latent heat of fusion, and may thereby alter the dynamics of clouds. Nevertheless, there are many situations in which the atmosphere is so stable that the heat released by glaciogenic seeding would not be able to have any detectable impact upon the air motion in the clouds. Even in unstable situations, seeding might be conducted in such a way that no major effect upon cloud dynamics is anticipated. To obtain dynamic effect the operation must be conducted with high precision, in the right place at the right time, to promote effects not only in a specific isolated cloud but in its neighbors.

One attractive feature of the dynamic seeding concept is that it provides a way to modify precipitation from convective clouds that already contain ice particles near – 5 to – 10 °C. It is because dynamic seeding looks for the release of latent heat in enough amount to impact the whole cloud scale.

The dynamic mode of seeding may be applied to obtain precipitation enhancement effects and hail suppression effects as well.

Over-seeding is a potential problem, especially in top seeding operations. The result of over-seeding may be the production of larger than usual cirrus anvil, the separation of upper parts of the clouds, the dissipation of turrets, and also the opening of vents to cold upper winds.

Seeding effect at large distances has been identified. The possibility of effects not only downwind, but crosswind or even upwind has been noticed. Silver iodide particles may remain active for several hours, especially if released at night, when they would not suffer photodeactivation.

The chain of events associated to dynamic seeding may be summarized as follow:

Stage 1 (~ 20 minutes or more): **Initial vertical tower growth**: rapid glaciation of the updraft regions of supercooled water by seeding agent; invigoration of the updraft through buoyancy increase produced by the release of latent heat; pressure falls beneath the actively growing tower.

Stage 2 (~ 40 minutes): **Horizontal cloud expansion, secondary growth**: Enhanced downdraft, convergence at the interface between the downdraft and the ambient low-level flow, growth of secondary towers; horizontal enlargement.

Stage 3: **Interaction with neighboring clouds**: Seeding of secondary towers, additional growth and merger of clouds on the mesoscale.

Stage 4: **Increased area rainfall**: more rainfall is obtained from the available moisture than would have obtained naturally; there is an enhancement of moisture supply to the area.

Cumulus Dynamics

Diurnal heating at the ground level by solar radiation results in the development of a convective layer at a low level above the ground that thickens during the day time. If the conditions are favorable for subsequent development of thicker thermal convection, thermals appear. Convective clouds will appear if thermals can reach condensation levels.

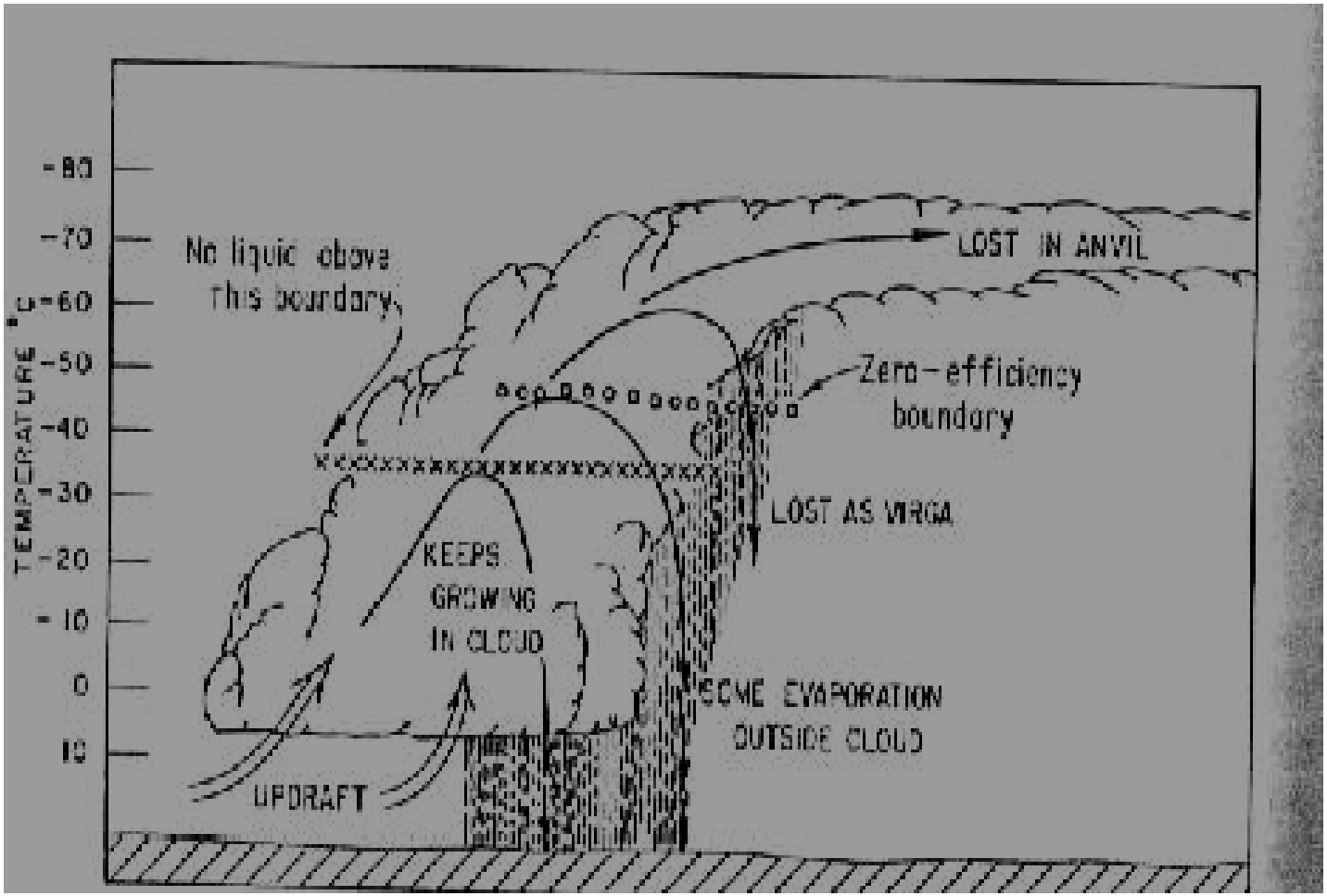
An atmospheric thermal is a body of buoyant air. Initially a thermal is a compact body of warm air that quickly develops a cauliflower appearance due to the surrounding instability. Vorticity in the form of ring around is also present. Thermals were thought to have the shape of bubbles with wakes in rear, but detailed observations showed that there is turbulence on the top of the thermals, and there is no wake below. Thermals can be **puffs** (turbulent bodies ejected into a relative undisturbed environment), **jets** (conical thermal with top hat profile at the beginning and bell profile later), or **plumes** (like jets but with a steady source of buoyancy). The latter is the predominant type in the formation of clouds.

When buoyant convection reaches the condensation level convective clouds appear.

The desert provides a very hot surface but one in which very little heat is stored. Infrared radiation from the surface is the agent of a large fraction of the heat loss, and this is absorbed and reemitted over a depth of a few meter mainly by the water vapor in place. As soon as a small thermal rises, cooler air descends to the surface. The air spaces between the sand particles make the sand layer a very effective insulator, so a depth of a few centimeters there is no perceptible diurnal variation of temperature in spite of variations of the order of 50 °C at the surface. Consequently, one thermal is not followed by others, and large thermals are inhibited. In this situation the only system able to sweep up heat from the ground as it travels over it is **the dust devil**. They vary from a foot or two to several hundred meters in height.

In order to a cloud to rise into dry air it must be large enough so that parcels may rise to its top without being subject to evaporation. This situation is most likely if parcels ascend along a path already moistened by previous thermals. The wind shear plays an important role in the actual shapes of convective clouds.

If the described process is successful a large cloud with a long enough life is formed. The general pattern of this cloud usually starts as a single-cell cloud (an original radar concept), which may present different turrets during its evolution. The following figure is very explicit:


TEMPERATURE °C
-80
-70
-60
-50
-40
-30
-20
-10
0
10
No liquid above
This boundary
LOST IN ANVIL
Zero-efficiency
boundary
LOST AS VIRGA
KEEPS
GROWING
IN CLOUD
SOME EVAPORATION
OUTSIDE CLOUD
UPDRAFT

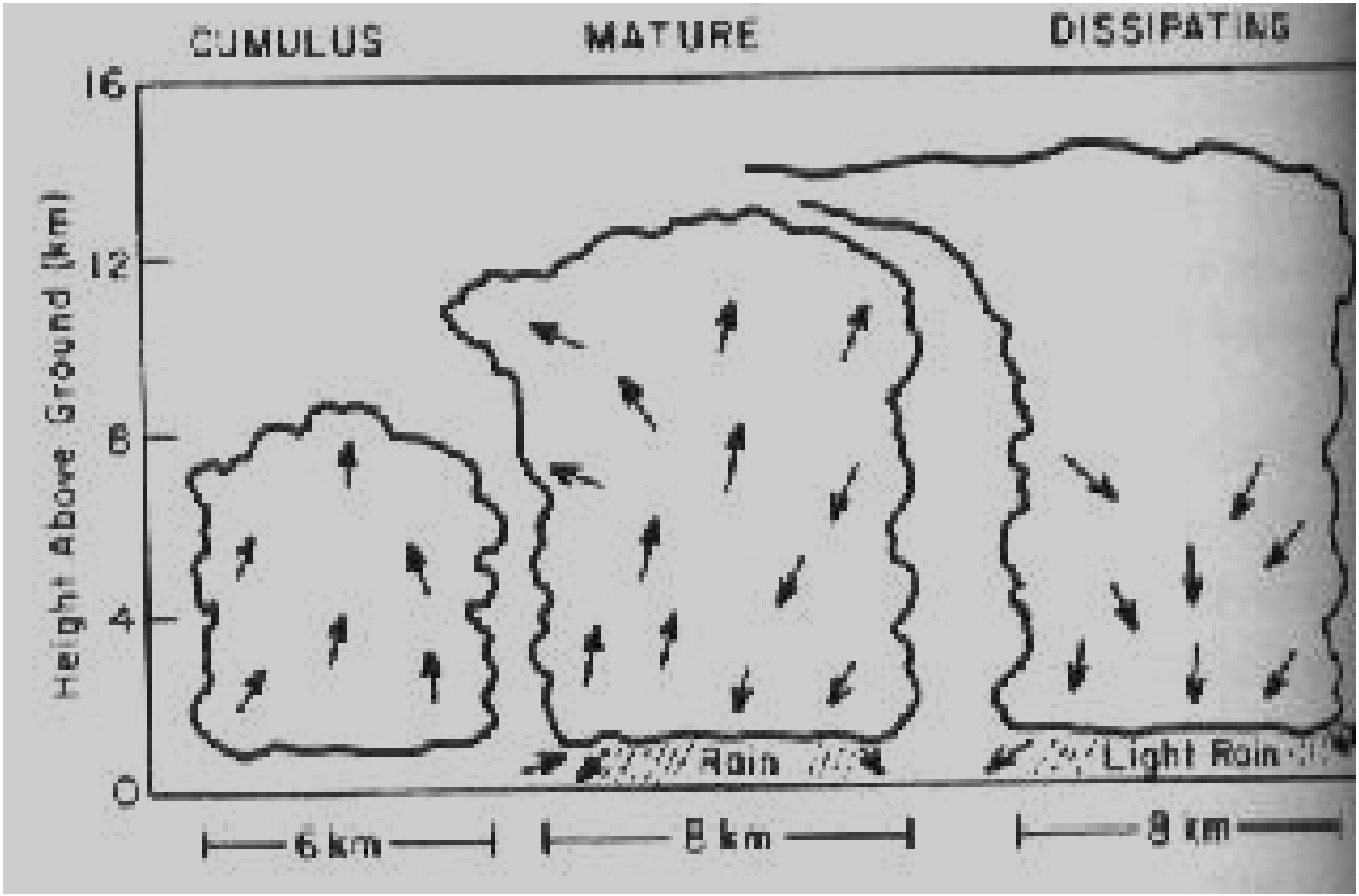
However, the cloud might evolve to a multi-cell stage, in which new cells can coexist with old cells in a family. Every ordinary cell suffers a process (its life process) consisting of three stages:

A first stage of cumuliform growth marked by the establishment and intensification of the updraft accompanied by the rapid increase in height of the cloud top and echo intensity;

A maturity stage with the coexistence within the same cloud of a specific updraft and a specific downdraft in the presence of strong and localized precipitation;

A dissipation stage during which only the downdraft exists with precipitation tending to generalize to the entire cell while its intensity diminishes.

Several causes combine to create the downdraft: entrainment, evaporation, fluctuations in pressure of dynamic origin around the cloud.

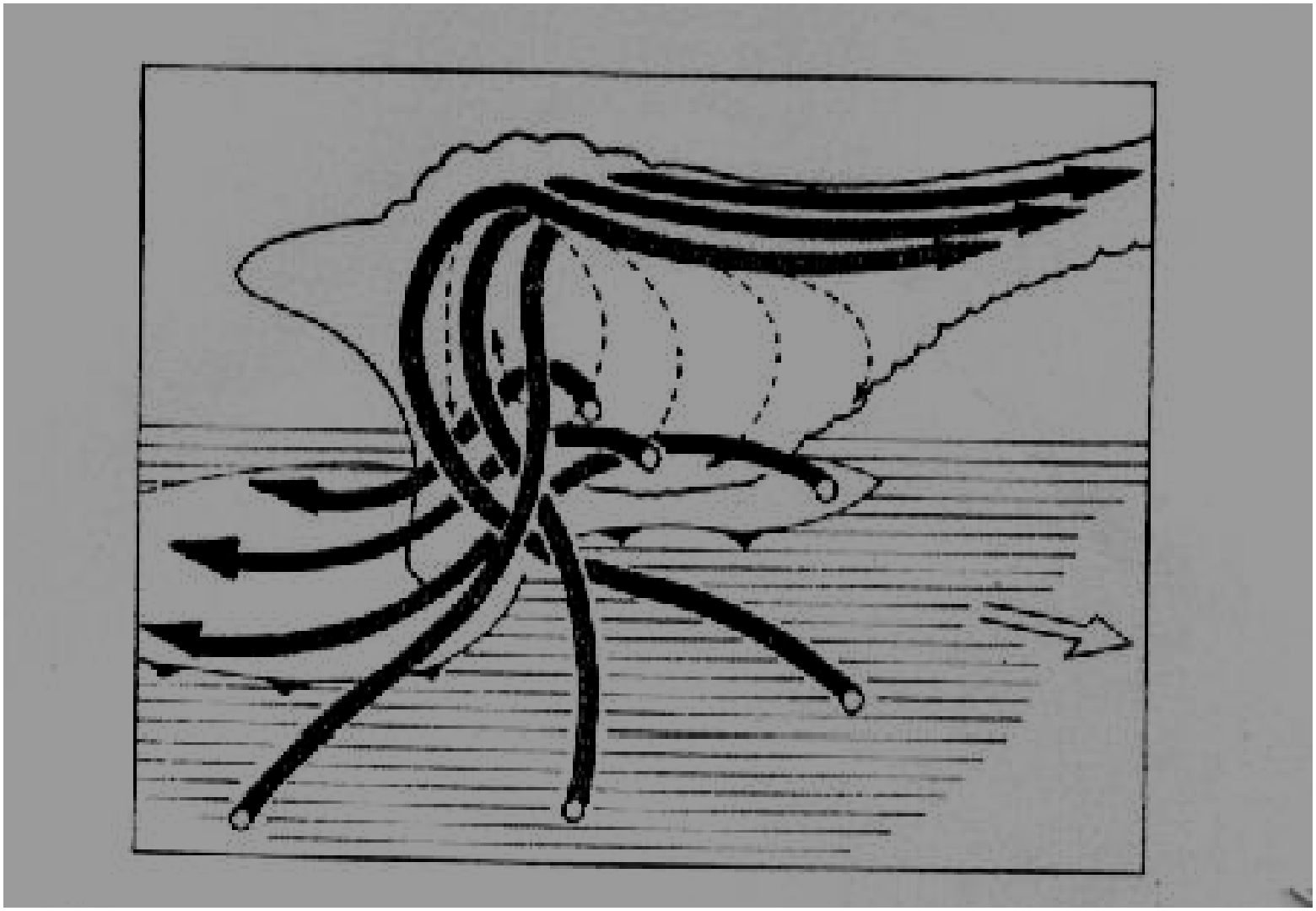

CUMULUS
MATURE
DISSIPATING
16
12
8
4
0
Height Above Ground (km)
Rain
Light Rain
6 km
8 km
8 km

R_i = (CAPE) / (0.5 <u>^2) (equation 3.1)

where CAPE is the Convective Available Potential Energy (value obtained from sounding analysis)

and <u> represents the cell relative inflow vector, which can be measured by the pilots or calculated by taking the difference between the mean wind in the lowest 6 km (cell motion) and a representative surface layer wind (500m mean wind). For moderate values of CAPE numerical studies indicate that supercells tend to form for values of R_i smaller than, whereas multi-cells tends to form for values of R_i greater than 50.

Super-cells may be considered the most perfected organization in convective clouds. They have an intense three-dimensional circulation, which roughly consists of an updraft and a downdraft. This organization is maintained for more than one hour (sometimes several hours), during which the storm covers considerable distances, producing intense precipitation, possible hail, strong gusts of wind, and tornadoes. The structure is shown in the figure below:



In summary:

Single-cell clouds move with the mean environmental wind over 5- 7 km (3- 4 miles), last 30- 60 minutes, produce moderate to heavy precipitation, with possible small hail. They develop in an environment with small and disorganized wind shear.

Multi-cell clouds consist of evolving cells, forming on right or right rear flank about 15 minutes, with motion deviate from the mean wind, produce moderate to heavy precipitation in which hail is usually present, and possible short-lived tornadoes along the gust front or in rapidly developing updraft. In general there are two types of multi-cell storms: clusters and lines.

Mesoscale Convective Systems are cluster of storms that persist and evolve like multicells.

Super-cell clouds are far rarer are more violent than the other two patterns. Their pervasive circulation is connected to mesociclons. They are very severe.

The electrical activity of the storms

Ever since Benjamin Franklin proved, in 1752, the presence of electricity in thunderclouds, the subject has provided challenging scientific problems which are still waiting for solutions. In essence, a thundercloud can be considered as an electrostatic generator which produces electrical charges, both positive and negative, both separated in different regions. Before we enter in details it is better to study the fair weather atmospheric electrical field.

Fine weather electrical field

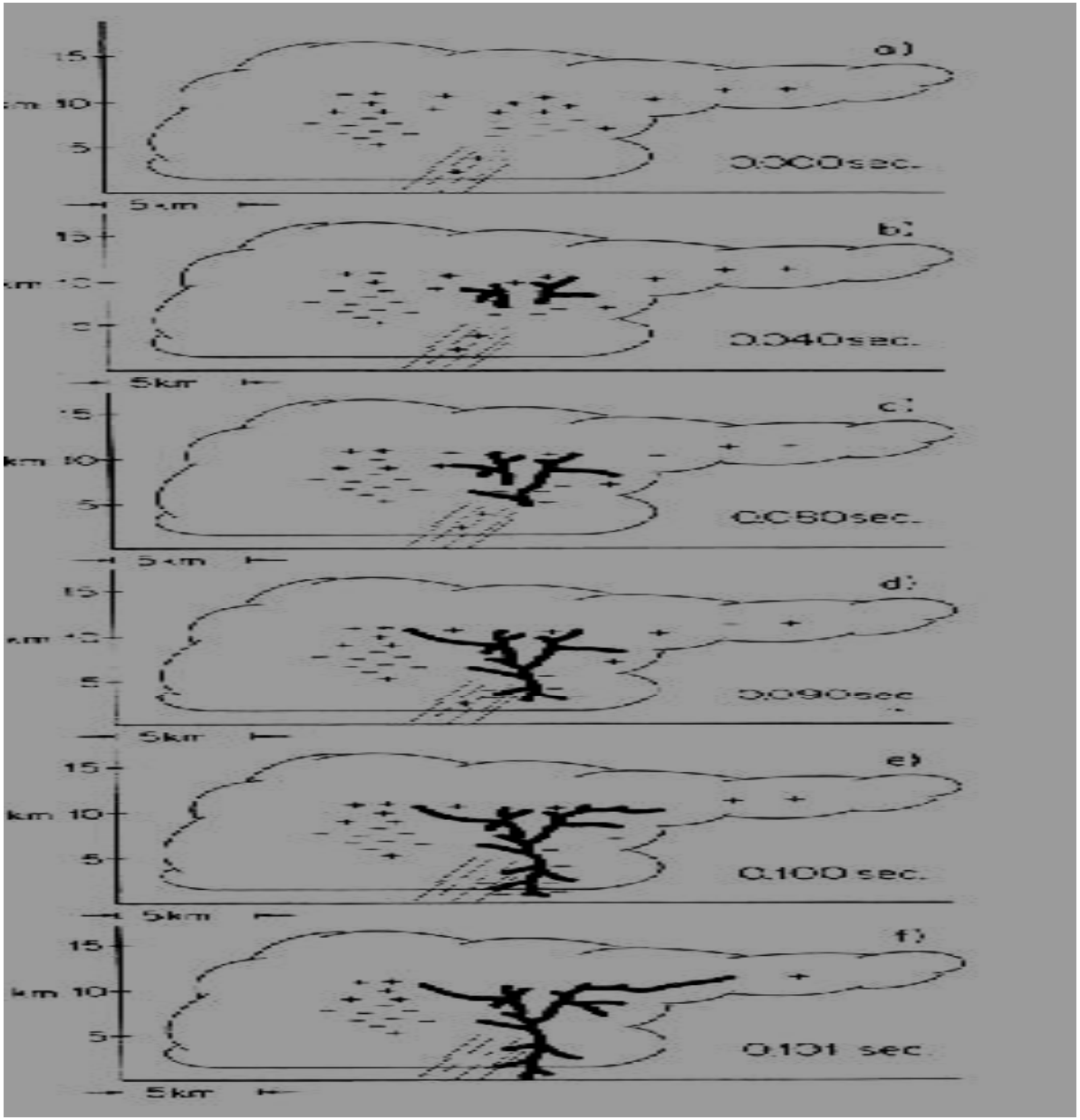
In undisturbed weather the atmosphere exhibits a fairly uniform, steady, downwardly directed electrical filed due to the existence of a negative charge on the earth's surface and a net positive space charge in the atmosphere. The intensity of the vertical electrical field has a maximum value at the ground where its magnitude is 120 V/m when averaged over the whole earth, and 130 V/m over the ocean; in heavily polluted areas the field may be considerably enhanced (~ 360 V/m). The field intensity decreases at greater heights, at 10 km (~ 32 800 feet) falling to only 3 % of its surface value. The potential of the atmosphere with respect to the ground increases with altitude up to 20 km (65 600 feet), above which it remains nearly constant at about 400 000 V. The air at these levels is highly conducting.

This so-called 'fine-weather' electric field is violent disturbed in the presence of storms and particularly by the occurrence of lightning flashes which may cause short-period changes, of order of 100 000 V/m, as they transfer charge from one part of a cloud to another.

Physics of Lightning

Lightning is a transient, high-current electric discharge that occurs in the atmosphere and has a path length on the order of kilometers. Most lightning is produced by thunderclouds, and well over a half of all discharges occur within the cloud. Cloud-to ground flashes (CG) are not so as frequent as intra-cloud flashes but more dangerous. The continental USA receives an estimated 40 million CG strikes each year.

CG lightning almost always starts within the cloud with a process that is called **the preliminary breakdown.** The location of the preliminary breakdown is not well understood, but it may begin in the high-filed region between the positive and the negative charge region. After several tens of millisecond, the preliminary breakdown initiates an intermittent, highly branched discharge that propagates horizontally and downward and that is called the stepped-leader. The individual steps of the stepped-leader have lengths of 30 to 90 m and occur at intervals of 20 to 100 μsec. When the tip of the stepped-leader gets close to the ground, the electric field just above the surface becomes very large and causes one or more upward discharges to begin at the ground and initiate **the attachment process.** The upward propagating discharges rise until one or more attach to the leader channel at a junction point that is usually a few tens of meters above the ground. When contact occurs, the first stroke begins. The peak currents in return strokes have typical values of 40 kA. The peak power dissipated by the return stroke is on the order of 10watts per meter of channel, and the peak channel temperature is at least 30 000 K. After a pause of 40 to 80 milliseconds, most CG flashes produce a new leader, the dart leader, which propagates without stepping down the previous return-stroke channel and initiates a subsequent return strokes. Most flashes contain two to four strokes.


a)
15
km 10
5
5km
0.000sec.
b)
15
km 10
5
5km
0.040sec.
c)
15
km 10
5
5km
0.050sec.
d)
15
km 10
5
5km
0.090sec
e)
15
km 10
5
5km
0.100 sec.
f)
15
km 10
5
5km
0.101 sec.

Important results of recent research indicate, (1) remote measurements of thunderstorm electric and magnetic fields can be used to infer properties of lightning currents, (2) sources of radio-frequency noise can be used to trace the geometrical development of lightning channel, and (3) small rockets can be used to trigger lightning artificially.

While most CG flashes transfer negative charge from the cloud to the ground, early documentation of + CG flashes has been found. For example, Benjamin Franklin wrote on 12 April 1753: **"The clouds of a thundergust are most commonly in a negative state of electricity, but sometimes in a positive state- the latter, I believe, is rare."**

Today we know that the Franklin's statement is perfect if we replaced the word 'clouds' for 'bases of the clouds'.

4.3 Electrical structure of thunderstorms

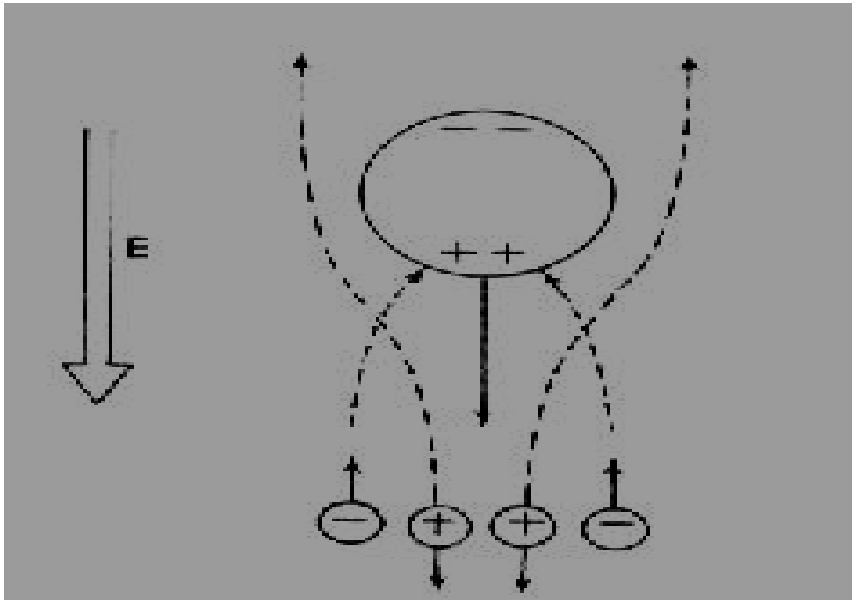
Initial electrification: It is found that a storm does not become strongly electrified until its radar echo extends above a certain altitude threshold and is growing vertically. The threshold altitude depends somewhat on the radar sensitivity but is about 8 km MSL (26 240 feet MSL) in summer months, corresponding to an air temperature of about – 20 °C. Electric field values of 1000 V/m are not detected until the radar echo top grows above about – 5 °C.

It is generally accepted that convective growth is very important in the process of electrification. Moderately strong precipitation can develop in a storm before to its electrification. Different results indicate that the electrification process operates at temperature of less than 0 or – 10 °C. One of the biggest questions has been whether the precipitation particles cause the electrification, or the convective motions themselves directly electrify the storm without involving or requiring precipitation. The former mechanism is favored in the scientific community.

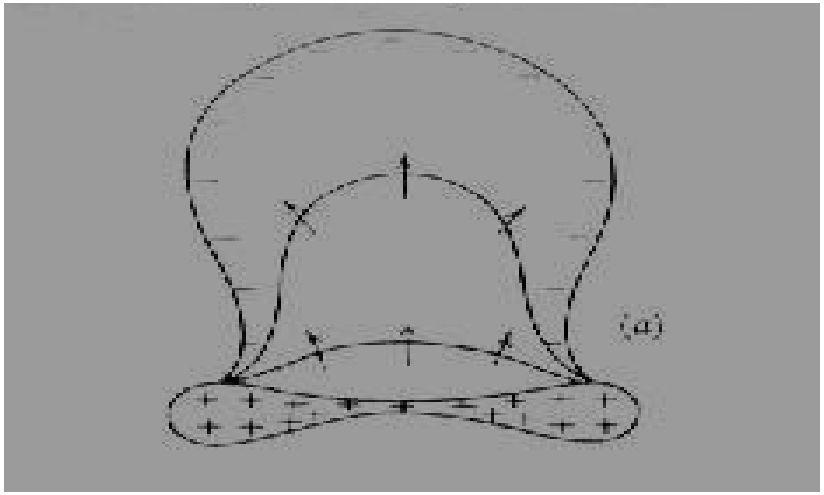
Electrical structure: The interior of a storm contains a dipolar charge distribution consisting of positive charge in the upper part and negative charge below the positive. These are the dominant accumulations of charge in the storm and are called the “upper positive” and “main negative” charges, respectively. The upper positive charge attracts negative ions to the top of the cloud, which form a negative screening layer at the edges. The main negative charge causes point discharge or corona from trees, vegetation, and other pointed or exposed objects on the ground below the storm. Positive charge is also found beneath and inside the base of the cloud below the main negative charge. This is called the “lower positive charge” and is caused by positive ions carried upward, and by precipitation particles descending.

In response to the dipolar structure of the storm, the initial lightning discharges are usually intracloud flashes (IC) that transport charge vertically between the two dominant charges. The first CG flash usually follows an initial sequence of IC flashes, but sometimes simple CG flashes begin the lightning activity. The initial lightning is associated with the cell having the greatest vertical development in the storm. Other cells do not generate lightning until they develop vertically above 7- 8 km altitude MSL (in summertime).

Different mechanisms have been proposed to explain the cloud charge separation. Here we illustrate one, the selective ion capture from droplet polarization in a downward-directed field (Wilson effect):



Additionally, charge separation can appear during the breakup of a large raindrop



Correlation between lightning and precipitation:

Observations and measurements of the electric field have reported that intensification of the electric field and the radar echo goes hand in hand. However the correlation, in both space and time, is very complicated.

Although thunder and lightning make a great show, and are very helpful especially during night flights, the amount of energy involved in such displays is small compared with that involved in the mechanical mechanism of the storm. Electrical activity must be considered as a byproduct of the buoyant forces, but a product that can help us to detect the formation of new developments.

5. Terra Incognita

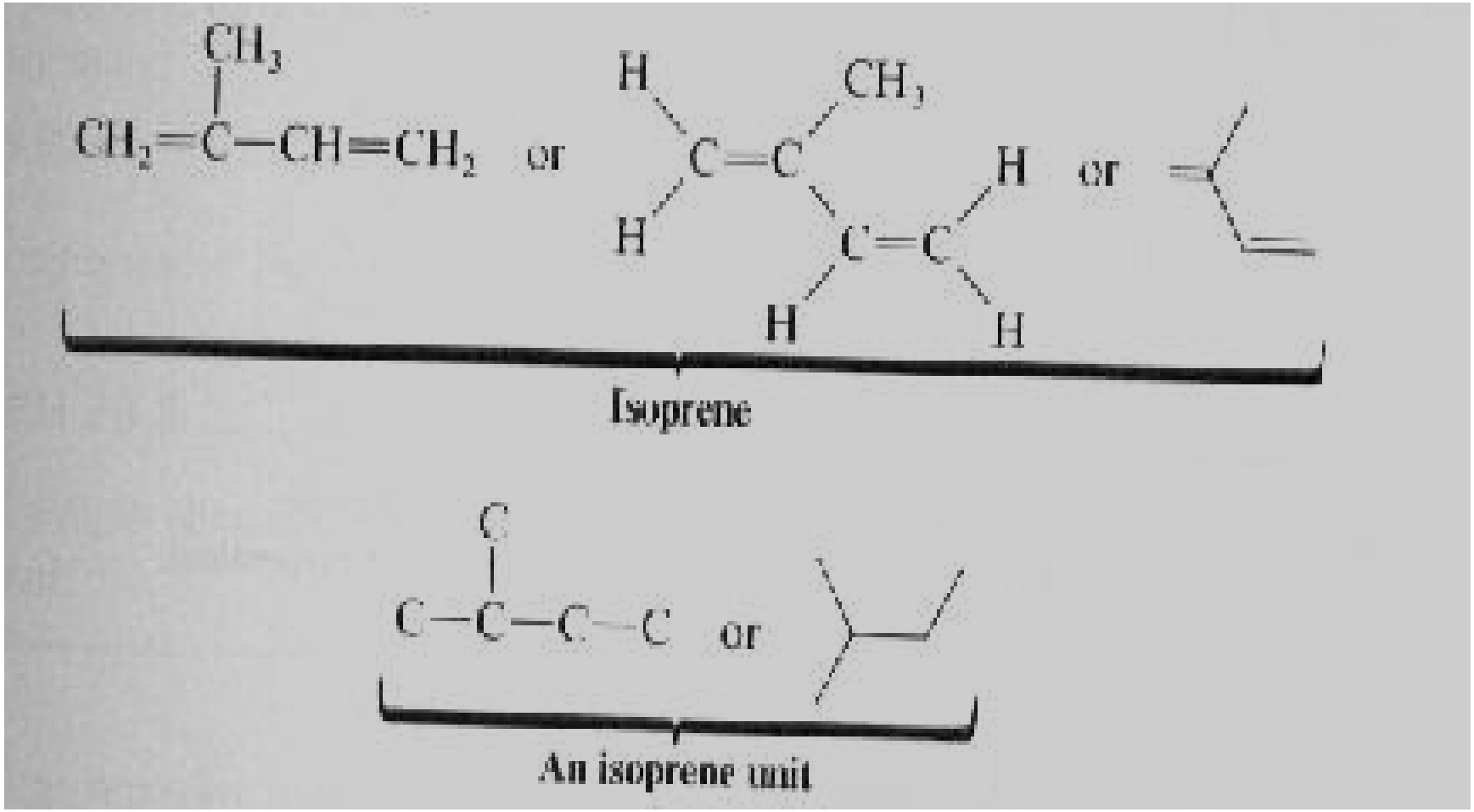
Our travel through the fundamentals of Weather Modification is ending. I hope we can understand better the complexity of the problems we are facing in every cloud seeding mission, and the necessity of a deeper understanding. However, to understand better the best way is to develop research, and here are some of the main tasks we will need to do:

Improving our radar tools and skills. The future use of Doppler radar data may improve the TITAN-based analyses.

Cloud water samples. The study of cloud water samples in different Texas areas may teach us better ways to improve our work.

CCN and IN counting. The atmosphere is in constant change. New research must be done now to know the CCN and IN background.

Other alternatives. There are different alternatives to improve the seeding operations, and also to extend the horizon of these operations. From cell-based cloud seeding toward environmental improving, the field of Weather Modification has a long road in front. The existence of isoprene in the emanations of forest might suggest ways to improve the atmosphere before the precipitation events.


CH3
CH2=C–CH=CH2
or
H
CH3
C=C
H
H
C=C
H
H
or
Isoprene
C
C–C–C–C
or
An isoprene unit

In 1966 it was reported that various terpenes and other tree oils can combine with iodine to form very active freezing nuclei. For example, oils of eucalyptus formed iodine compounds with the nucleation temperature of – 4 °C.

In 1972 another amazing result was reported: In the presence of strong electrical fields, created when an electrified cloud passes overhead, the tips of pine needles and other wax-covered plant surfaces release terpenes directly in the form of aerosol particles. These particles subsequently serve as IN.

The travel has ended