PHYSICS OF CLOUD AND PRECIPITATION

By

NIGATU KUCHI SHEGIRO

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Dated: April 11, 2022

Advisor: _____
Dr. Yitagesu Elfagd

Examiners: _____
Dr. Lemi Demeyu

_____
Dr. Feraol Fana

Chairman: _____
TBA

ADISS ABABA UNIVERSITY

Date: **April 11, 2022**

Author: **NIGATU KUCHI SHEGIRO**

Title: **PHYSICS OF CLOUD AND PRECIPITATION**

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Table of Contents

List of Figures

Abstract

In this project we present a review of the physics of clouds and precipitation in the atmosphere.

Thunderstorms are a global phenomenon in the mid-latitudes and tropics. They form where and whenever the ingredients for their formation come together: instability, moisture and lift. Especially upon interaction with vertical wind shear, they may develop in to well-organized systems that produce hazards such as large hail, severe winds, heavy precipitation, and tornadoes.

We must look in to the processes by which the clouds are formed and precipitation is produced in order to the meaning of clouds they related to weather. We will see how clouds are classified and, named and what kinds of precipitation certain types of clouds produce. And also in this project we study how to clouds form and precipitation develop in the atmosphere must be saturated with moisture. As a cold air passes over warm water, rapid evaporation takes places and the saturation is quickly reached.

Lifting of air, and the resulting adiabatic expansion, is the most important cooling method. The lifting may be accomplished by thermal orographic or frontal action. It produces most of the clouds and precipitation.

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Chapter 1

Introduction

In this project we introduce the thermodynamics of droplets in the atmosphere. Droplet nucleation is described by modifications of the Clausius-Clapeyron equation due to surface tension [1].

These modifications also provide the framework for understanding droplet nucleation due to electric charge. We also introduce some of the processes involved in droplet growth [2].

Thermodynamics plays an important role in our quantitative understanding of atmospheric phenomena ranging from the smallest cloud microphysical processes to the general circulation of the atmosphere. Thunderstorms are among the most fascinating objects of study in atmospheric physics, thunderstorms are very complex clouds, driven by buoyancy effects they develop strong vertical motions in both upward and downward directions [3].

Cloud droplets will be carried by air currents within the cloud, and if they bump into each other, it is called a collision. However, if they collide then stick together, that is called coalescence. Although this process is important, especially in the tropics, it falls short of being the primary mechanism for the formation of raindrops [4].

The smaller droplets are also falling at their terminal velocity, $u(r)$ We have to correct for this, because if the small droplets were falling at the same speed of the larger droplet there would be no increase in mass since there would be no collisions. The greater the difference in the terminal Lectures in Clouds Physics velocities, the fewer collisions and the less mass will be collected [5].

The drop size distributions of actual clouds can be quite complex, and varies from cloud to cloud. Some distributions have more than one peak, while others have a single peak. In general, maritime cumulus clouds have larger drops and broader distributions than do continental clouds [7].

Many physical characteristics of clouds, in particular the cloud phase makeup and water content, are temperature dependent, and it is precisely this dependence which frequently determines the seasonal and latitude variations of the physical parameters. Consequently, it is important to have an idea of the mean atmospheric temperatures. Mean air temperatures for the hemispheres and mean global temperatures are given and mean temperatures at various latitudes in the Northern Hemisphere are given [11].

The concentration of ice nuclei, which determines the probability of the appearance of crystals in the cloud, increases sharply with a drop in temperature. The droplet size is the next most important factor: The smaller the droplets, the more likely it is that unfrozen droplets will be present in the cloud at the given temperature [12].

The storms mostly extend throughout the whole troposphere from the boundary layer to the upper troposphere and even penetrate into the stratosphere with their overshooting tops. Water vapor condenses to form cloud droplets or ice crystals. Collisions among these particles lead to larger precipitation particles like raindrops, snow flakes, graupel, or hailstones [14].

The atmospheric boundary layer the layer affected by the surface moisture content and temperature typically increase as a result of solar heating. If the temperature higher up in the troposphere is low enough, convective available potential energy (CAPE) is created [18].

Clouds play an eminent role in solar energy production and are a fundamental aspect of the hydrological cycle. However, the most striking feature about clouds is their impact on Earths energy budget and so on climate [19].

Most clouds formed in the Tropics, and many in the middle latitudes, are warm clouds. Those clouds have temperatures greater than 0^0C throughout. The Collision-coalescence process generates precipitation. This process depends on the differing fall speeds of different-sized droplets. It begins with large collector drops which have high terminal velocities [22].

If a cloud containing supercooled water droplets is saturated with respect to water, then it is supersaturated with respect to ice, and the relative humidity with respect to ice is greater than 100 percent. The force resulting from the difference between vapor pressure over water and over ice causes vapor molecules to be attracted to ice crystals, and the ice crystals will grow rapidly [29].

Precipitation is any product of the condensation of atmospheric water vapor that falls under gravity.Precipitation occurs when a portion of the atmosphere becomes saturated with

water vapor, so that the water condenses and "precipitates". Precipitation is a major component of the water cycle, and is responsible for depositing most of the fresh water on the planet Mechanisms of producing precipitation include convective, stratiform and orographic rainfall [30].

Precipitation is measured on the basis of the vertical depth of the water or melted snow. Snow, sleet, hail, and other solid forms are also measured on the basis of the depth of the unmelted form [31].

In chapter two, we present a review of atmospheric thermodynamics. In chapter three, we discuss in detail about atmospheric clouds. In chapter four, atmospheric precipitation is discussed in detail. Finally, chapter five is conclusion and summary.

Chapter 2

Atmospheric Thermodynamics

Thermodynamics is one of the cornerstones and crowning glories of classical physics. It has applications not only in physics, chemistry, and the Earth sciences, but in subjects as diverse as biology and economics.

Thermodynamics plays an important role in our quantitative understanding of atmospheric phenomena ranging from the smallest cloud microphysical processes to the general circulation of the atmosphere. The purpose of this chapter is to introduce some fundamental ideas and relationships in thermodynamics and to apply them to a number of simple, but important, atmospheric situations. Further applications of the concepts developed in this chapter occur throughout this project.

Thunderstorms are weather systems that feature strong vertical motions on the scale of a few kilometers in horizontal diameter and are primarily driven by buoyancy, i.e, free convection. Upward and downward motions in convective storms are among the most intense observed in Earths atmosphere, reaching wind speed values of many tens ofm/s, and they may develop within much less than an hour.

There, weather systems of hundreds to thousands of kilometers evolve over many hours to days, exhibiting vertical velocities of less than $0.1m/s$. There are a few ingredients required to initiate convective storms in the atmosphere. To start, two ingredients are needed that create the so-called convective available potential energy (CAPE) that storms transform into motion: a rapidly decreasing temperature with height and an abundance of moisture in the lower troposphere.

In the atmospheric boundary layer the layer affected by the surface moisture content and temperature typically increase as a result of solar heating. If the temperature higher up in the troposphere is low enough, CAPE is created. Such a situation is called one of latent

instability. The only additional ingredient needed to create a storm is a source of lift, also called a trigger mechanism.

The vertical profiles of moisture,temperature,and the horizontal wind components measured at a given geographic position and height are displayed. These measurements are typically obtained by releasing a weather balloon that carries a thermometer, hydro-meter and barometer-aradiosonde. During the afternoon, strong thunderstorms developed in this atmospheric environment[1].

2.1 Water in the atmosphere

Water vapour is the most important variable constituent of the atmosphere. For this reason, water vapour is normally treated as a separate constituent alongside the well-mixed gases (nitrogen, oxygen, and argon) that make up the rest of the bulk of the atmosphere. When water vapour cools down enough it will condense to a liquid or even freeze to as olid; it will experience a phase transition. In this chapter we examine the thermodynamics of phase transitions and derive its key result: the Clausius-Clapeyron equation, which describes the equilibrium pressure of the vapour at coexistence with the liquid. Although we use these results to understand the properties of water in the Earths atmosphere, we can use them equally to understand freezing of CO_2 in the Martian atmosphere or the formation of methane clouds on Titan.

One result of phase transitions in the atmosphere is the formation of clouds and precipitation. However, the formation of drops from vapour depends critically on additional factors such as surface tension and solute concentration. Phase transitions are also responsible for a large amount of heat transfer in the atmosphere. To evaporate water requires energy, the latent heat of evaporation. This latent heat is given back to the atmosphere on condensation. So vapour is a storage and transport medium of energy.

Latent heat is one of the most important contributions to the global energy cycle. For example, the atmosphere is dominantly heated in the tropics and the main source of heat for the tropical atmosphere is latent heat in convective clouds. Water vapour is some times confused with moist air or steam.Water vapour is a gas made up of H_2O molecules. Moist air is a mixture of dry air and water vapour. Steam is technically a vapour at high temperature (above boiling point) and pressure. In informal language, steam is a mixture of air and vapour

where the water vapour is condensed into small haze droplets. The partial pressure of water vapor in the atmosphere. The value of density of the water vapor is 4.77×10^{-3}kg/m^3 at ^{o}C.

2.2 Clausius-Clapeyron (C-C) equation

Consider a closed cylinder with a liquid and its vapour, but nothing else, coexisting. All the time some of the liquid molecules will evaporate to vapour, while some of the vapour molecules will condense to liquid. This situation will equilibrate: when excess molecules are leaving the fluid to go into the vapour, the pressure of the vapour will increase and consequently then umber of molecules moving back into the liquid will increase, reducing the excess pressure until equilibrium is restored.

For systems in equilibrium, where the liquid and the vapour phases coexist, the specific Gibbs functions g_v and g_v for the liquid and vapour phases have to be the same. The equality of the specific Gibbs functions in turn determines how the vapour pressure changes with temperature, as described by the Clausius - Clapeyron equation. Now we give a more complete argument why the specific Gibbs functions for the liquid and the vapour have to be the same when they coexist.

Consider a cylinder with a piston, which is flled with a liquid and its vapour and held under a constant temperature T and pressure (the pressure of a vapour is denoted by the symbol e to distinguish it from the total pressure P of air). Because the system is held at constant pressure and temperature, the Gibbs function for the whole system

$$G = U - TS + eV \tag{2.1}$$

where G is Gibbs free energy, U is internal energy, T is temperature, S is entropy, e vapor pressure and V volume. TS express energy you get from the system environment by heating. Where eV is the work to give system final.

Has a constant value as it is a function of temperature and pressure only, but the total Gibbs function has contributions from both the liquid and the vapour,

$$G = M_l g_l + M_v g_v \tag{2.2}$$

with M_l and M_v the masses of the liquid and vapour and g_l and g_v the specific Gibbs functions for the liquid and the vapour . If now, by some fluctuation, some amount δM of liquid

evaporates to the vapour, the total Gibbs function changes by

$$\delta G = M_l \delta g_l + M_v \delta g_v + (g_v - g_l)\delta M$$

$$\delta G = M_l \left(-s_l \delta T + v_l \delta e\right) + M_v \left(-s_v \delta T + v_v \delta_e\right) + \left(g_v - g_l\right) \delta M \tag{2.3}$$

Because the total Gibbs function is kept constant, its first order variation has to vanish, δG=0. The pressure and the temperature are prescribed, so we have δe=0 and $\delta\, T = 0$. This requires the cylinder to be below the critical temperature, the temperature above which the gas and liquid phases become indistinguishable and no phase separation can occur no matter how high the pressure is, the critical temperature for water is $647K$.

$$g_v = g_l \tag{2.4}$$

The specific Gibbs functions for the liquid and vapour phases are the same if they coexist in the cylinder. At first sight it appears that thee quality of g_v and g_l depends on the fact that the system was in a cylinder at constant temperature and pressure. This is not the case. Suppose we had fixed the piston and the system was at constant temperature T and constant volume V. In that case the total free energy

$$F = U - TS \tag{2.5}$$

would have to stay constant. Using the same arguments as above(with $F = M_l f_l + M_v f_v$) it then follows that:

$$\delta F = M_l \delta f_l + M_v \delta f_v + (f_v - f_l)\delta M$$

$$\delta F = M_l \left(-s_l \delta T - e\delta V_l\right) + M_v \left(-S_v \delta T - e\delta V_v\right) + \left(f_v - f_l\right) \delta M \tag{2.6}$$

Now we have to take in to account that, although the total volume V is kept constant, the specific volumes v_l and v_v can change when there is a mass transfer δM from the liquid to the vapour phase. We have

$$M_l \delta v_l = \delta V_l - v_l \delta M_l, M_v \delta v_v = \delta V_v - v_v \delta M_v. \tag{2.7}$$

Substituting this in Eq. 2.6 we find

$$\delta F = -\left(M_l s_l + M_l s_v\right) \delta T - e\left(\delta V_l + \delta V_v\right) + \left(f_v + e v_v - f_l - e v_l\right) \delta M$$

$$\delta F = -(M_l s_l + M_v s_v)\delta T - e\delta V + (g_v - g_l)\delta M, \tag{2.8}$$

where we have used the fact that $g = f + ev$. Because the free energy is kept constant, its first order variation has to vanish, $\delta F = 0$ the temperature and volume are prescribed so we have $\delta V = 0$ and $\delta T = 0$. It again follows that $g_v = g_l$, that is, the specific Gibbs functions for the two phases have to be the same if they coexist. It turns out we can choose any set of constraints and still find that the specific Gibbs functions for both phases have to be the same.The Gibbs function is the only thermodynamic potential that depends on two intensive variables: pressure and temperature. This is the ultimate reason why for coexisting phases their specific Gibbs functions, rather than any of the other thermodynamic potentials, are the same. Having established that when the phases coexist their specific Gibbs functions have to be the same, we will now change the temperature T of the cylinder by dT. This temperature change will lead to a change in vapour pressure e byde, see Figure 2.1. On changing e and T we have for the specific Gibbs functions the coexistence curve $g_v = g_l$ determines how much the vapour pressure needs to change (d_e) when the temperature is changed (dT).

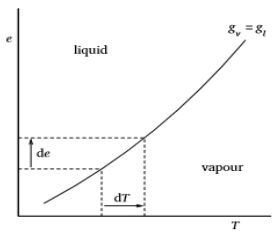

Figure 2.1: Derivation of the ClausiusClapeyron equation.

$$dg_l = v_l d_e - s_l dT, dg_v = v_v d_e - s_v dT \tag{2.9}$$

Because for the coexisting phases the specific Gibbs functions g_l and g_v are equal, we must have $dg_l = dg_v$ From this follows

$$\frac{d_e}{dT} = \frac{s_v - s_l}{v_v - v_l} \tag{2.10}$$

The entropy difference between the liquid and the vapour is related to the heat required to evaporate a certain amount δ M of liquid into the vapour phase. This is most easily formalized

by noting that at equilibrium. by Eq. 2.4,

$$g_v = g_l$$

From Gibbs free energy:

$$h_v - Ts_v = h_l - Ts_l$$

where h_v enthalpy of unit mass of vapor and h_l enthalpy of unit mass of liquid. Enthalpy difference between the vapour and the liquid is,

$$L = T(s_v - s_l) \tag{2.11}$$

Substituting this in Eq. 2.10, we find the ClausiusClapeyron equation,

$$\frac{de_s}{dT} = \frac{L}{T(v_v - v_l)} \tag{2.12}$$

where , e_s is saturation vapor pressure and T is temperature, L is the specific latent heat of evaporation (2400 kJ/kg at 25^0C to 2600 kJ/kg at -4^0C), v_v and v_l are specific volumes for water vapor and liquid water respectively. This equation describes how the pressure of the vapour changes when the temperature changes; the subscript s has been added to emphasize that the vapour is saturated, that is, the vapour is in equilibrium with the coexisting liquid(If no liquid is available, the vapour pressure is not determined by the Clausius-Clapeyron equation but by the equation of state for the vapour). Because $v_v \gg v_l$ and $L > 0$, we find that the saturated vapour pressure is an increasing function of temperature; at higher temperatures more molecules can escape the attractive potential in the liquid. For $v_v \gg v_l$ the above equation no.(2.12) be come

$$\frac{de_s}{dT} = \frac{L}{T(v_v)} \tag{2.13}$$

2.3 Calculation of saturated vapour pressure (e_s)

2.3.1 How the air becomes saturated?

Cooling air to its dew point

The dew point is the temperature to which a parcel of air must be cooled in order to become saturated, and (unless super-saturation occurs) condenses to water. Water vapor normally begins to condense on condensation nuclei such as dust, ice, and salt in order to form clouds.

The cloud condensation nuclei concentration will determine the cloud microphysics. An elevated portion of a frontal zone forces broad areas of lift, which form cloud decks such as altostratus or cirrostratus.

Stratus is a stable cloud deck which tends to form when a cool, stable air mass is trapped underneath a warm air mass. It can also form due to the lifting of advection fog during breezy conditions. There are four main mechanisms for cooling the air to its dew point: adiabatic cooling, conductive cooling, radiational cooling, and evaporative cooling. Adiabatic cooling occurs when air rises and expands. The air can rise due to convection, large-scale atmospheric motions, or a physical barrier such as a mountain (orographic lift). Conductive cooling occurs when the air comes into contact with a colder surface usually by being blown from one surface to another, for example from a liquid water surface to colder land.

Radiations cooling occurs due to the emission of infrared radiation, either by the air or by the surface underneath. Evaporative cooling occurs when moisture is added to the air through evaporation, which forces the air temperature to cool to its wet-bulb temperature, or until it reaches saturation. Adding moisture to the air The main ways water vapor is added to the air are: wind convergence into areas of upward motion precipitation or virga falling from above, daytime heating evaporating water from the surface of oceans, water bodies or wet land transpiration from plants cool or dry air moving over warmer water and lifting air over mountains.

The ClausiusClapeyron equation needs to be integrated to find the vapour pressure at a certain temperature. To do this, assume that the vapour is an ideal gas, so we have $e_s v_v = R_v T$, with R_v the specific gas constant for the vapour (in the case of water vapour R_v =461.5$Jkg^{-1}K^{-1}$). Further assume that $v_v \gg v_l$, which is true away from the critical temperature. The Clausius Clapeyron equation then becomes

$$\frac{de_s}{dT} = \frac{Le_s}{R_v T_S} \tag{2.14}$$

Further assuming that L is constant (this is not a very good assumption, see below) this integrates. From ideal gas equation we have that:

$$e_s = \rho_v R_V T \tag{2.15}$$

where, ρ_v density of water vapor R_V gas constant of vapor and T is the temperature of the

gas particles. From ideal gas equation we have that

$$PV = RT$$

$$e_s v_v = R_V T \tag{2.16}$$

$$v_v = \frac{R_V T}{e_s} \tag{2.17}$$

Substitute Eq.(2.17) in Eq.(2.13) it become,

$$\frac{de_s}{dT} = \frac{e_s L}{R_V T^2} \tag{2.18}$$

$$\int_{e_s(T_o)}^{e_s(T)} \frac{de_s}{e_s} = \int_{T_0}^{T} \frac{L}{RT^2} dT = -\frac{L}{R}\left[\frac{1}{T} - \frac{1}{T_O}\right] \tag{2.19}$$

$$ln\frac{e_s(T)}{e_s(T_O)} = -\frac{L}{RT} + C \tag{2.20}$$

Where C is some constant.

$$e_s(T) = e_{s_o} exp\left(\frac{L}{R}\left[\frac{1}{T_0} - \frac{1}{T}\right]\right) \tag{2.21}$$

with, es_0 the saturated vapour pressure at temperature T_0.Because the specific latent heat,L reduces with temperature this integration is only valid for small temperature differences. It can be seen that the saturated vapour pressure is a strongly increasing function of temperature. This aspect is independent of the presence of any other gases alongside the vapour. By Daltons law, the total pressure of a mixture is the sum of the partial pressures of the constituents; the partial pressure of vapour above a liquid surface is determined by the ClausiusClapeyron equation, independent of any other gases present. This clarifies the common misconception that warm air holds more water; the air does not hold the water, and its presence does not change the vapour pressure.

If the vapour pressure is the same as the atmospheric pressure, the liquid has reached its boiling point. Above the boiling point, the vapour cannot coexist with the liquid as its equilibrium vapour pressure is higher than the prescribed pressure in the system. From the ClausiusClapeyron-equation it then follows that the boiling point of aliquid reduces if the environmental pressure reduces, the boiling point of water is set at 100^0C because at that

temperature the vapour pressure is equal to the standard atmospheric pressure of $1013.25hPa$. we derived that the specific latent heat of evaporation can be approximated as

$$L = L_0 - (c_{pl} - c_{pv})(T - T_0) \tag{2.22}$$

with L_0 the latent heat of evaporation at some reference temperature T_0. With this approximation for the specific latent heat of evaporation as a function of temperature we can now rewrite the ClausiusClapeyron equation, Eq. 2.14, as

$$\frac{de_s}{e_s} = \frac{L_0 + (c_{pl} - c_{pv})\,T_0}{R_v}\frac{dT}{T^2} - \frac{c_{pl} - c_{pv}}{R_v}\frac{dT}{T} \tag{2.23}$$

This integrates to;

$$e_s = e_{s_0} exp\left[\frac{L_0}{R_v}\left(\frac{1}{T_0} - \frac{1}{T}\right)\right]\left(\frac{T_0}{T}\right)^{\alpha} exp\left[\alpha\left(1 - \frac{T_0}{T}\right)\right] \tag{2.24}$$

with $\alpha = \frac{(c_{pl} - c_{pv})}{Rv}$. At $T_0 = 25^oC$ we have $L_0 = 2.444106Jkg^{-1}$, $c_{pv} = 1865.1Jkg^{-1}K_{-1}$, and $c_{pl} = 4179.9Jkg^{-1}K^{-1}$. Using these values, the accuracy of Eq. 2.24 is better than 1 part in two thousand when $0^0C < T < 40^0C$ and better than 1.5 parts in a hundred when $0^0C < T < 100^0C$; an impressive result as the saturated vapour pressure varies by more than two orders of magnitude over this temperature range. For realistically varying L and without ignoring v_l there are several empirical formulae which give values of the saturated vapour pressure as a function of temperature. A particularly simple empirical formula is Tetens formula.

$$e_s(hp_a) = 6.112exp\left[\frac{17.67T(^0c)}{T(^0c) + 243.5}\right] \tag{2.25}$$

with $T(^0C)$ the temperature in degrees Celsius. This formula has an accuracy level similar

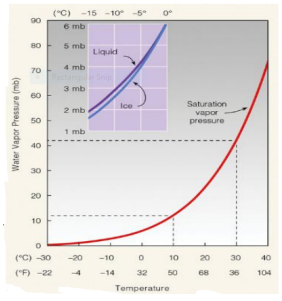

Figure 2.2: Water vapor pressure (mb) directly proportional to Temperature[2].

to the analytical expression, Eq. 2.24 Such formulae can also be produced for the saturated

vapour pressure over ice. The argument leading to the Clausius-Clapeyron equation can also be applied to the balance between sublimation and deposition of vapour over ice. Equation 2.21 with $L_{ice} = 2.826 \times 106 Jkg^{-1}$ gives the vapour pressure over ice accurate to two parts in a hundred down to temperatures of $50^0 C$. Again, accurate empirical formulae exist to calculate the saturated vapour pressure over ice.

As the temperature of the liquid increases, the kinetic energy of the molecules also increases and the kinetic of the molecule increases, the number of molecules transition in to a vapor also increasing the vapor pressure.

The higher the temperature is the more molecules have enough energy to escape from the liquid or solids to higher vapor pressure values.

Generally, a substance's vapor pressure increases as temperature increases and decreases as temperature decreases.(i.e; vapor pressure is directly proportional to temperature).

In these equations, the enthalpy of vaporization L always occurs in the fraction $\frac{L}{Rv}$ with R_v the specific gas constant for the vapour. It is often more convenient to rewrite this fraction as $\frac{L}{R_v} = \frac{\wedge}{R^*}$ with$\wedge$ the molar enthalpy of vaporization and R^* the universal gas constant [3].

Chapter 3

Atmospheric Clouds

3.1 Definition of cloud and cloud composition

A cloud is aggregate of tiny water droplets or ice crystals suspended in the air. Clouds consist of an agglomeration of suspended liquid water droplets or ice crystals and are formed when air parcels become saturated. Formation mechanisms include cooling of the air, for instance by adiabatic, convective, or orographic lifting, and adding water vapor to the air, e.g, by evaporation of the oceans.

High spatial variability down to the centimeter and millimeter scale characterizes the microscopic cloud. Some are found only at high elevations whereas others nearly touch the ground. They can be thick or thin, big or little and exist in a seemingly endless variety of forms. Clouds consist of minute water droplets, ice crystals, or a mixture of the two in sufficient quantities to make the mass discernible. Some clouds are pretty, others are dull, and some are foreboding. But we need to look beyond these aesthetic qualities.

Clouds are visible evidence of atmospheric moisture and atmospheric motion. Those that indicate instability may serve as a warning to the fire-control man. Some produce precipitation and become an ally to the fire fighter. We must look into the processes by which clouds are formed and precipitation is produced in order to understand the meaning and portent of clouds as they relate to fire weather. We will see how clouds are classified and named, and what kinds of precipitation certain types of clouds produce.

In order for clouds to form and precipitation to develop, the atmosphere must be saturated with moisture. In this chapter we learned that at saturation the atmospheric vapor pressure is equal to the saturation vapor pressure at the existing temperature and pressure. There are two principal ways in which the atmospheric vapor pressure and saturation vapor pressure

attain the same value to produce 100 % relative humidity, or saturation. These are through the addition of moisture to the air, or, more importantly, through the lowering of air temperature.

Moist air may be cooled to its dew point and become saturated as it passes over cool and or water surfaces. Nighttime cooling of the ground surface by radiation, and the subsequent cooling of adjacent moist air, may produce saturation and fog. Air can become saturated by the addition of moisture. This may occur by evaporation as cold, dry air passes over warm water, or as warm rain from above a front foils through cold air beneath the front. The most important method of cooling air to saturation is adiabatic cooling because of lifting [1]. Lifting may be thermal, orographic and frontal as shown in figure 3.1 and 3.2

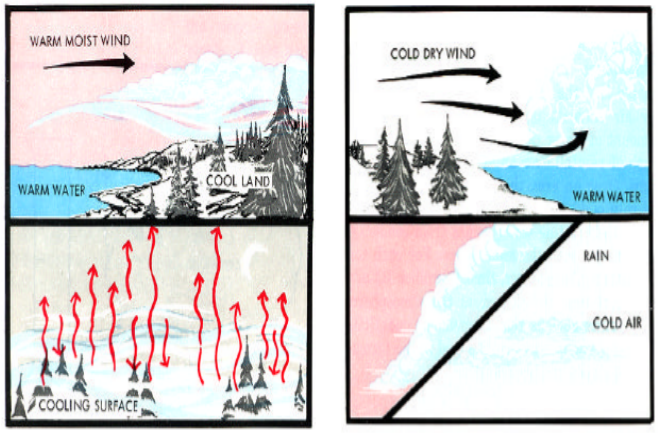

Figure 3.1: Saturated as it passes over cool and or water surfaces [1].

As cold air passes over warm water, rapid evaporation takes place, and saturation is quickly reached. Cold continental polar air crossing the warmer Great Lakes in the fall and early winter gathers large amounts of moisture and produces cloudiness and frequently causes rain or snow to the lee of the lakes. Saturation may also be reached when warm rain falls through cold air; for example, beneath a warm front. Rain falling from the warm clouds above the front evaporates in the cold air beneath and forms scud clouds. Contrails made by high-flying aircraft are a type of cloud formed by the addition of moisture from the planes exhaust.

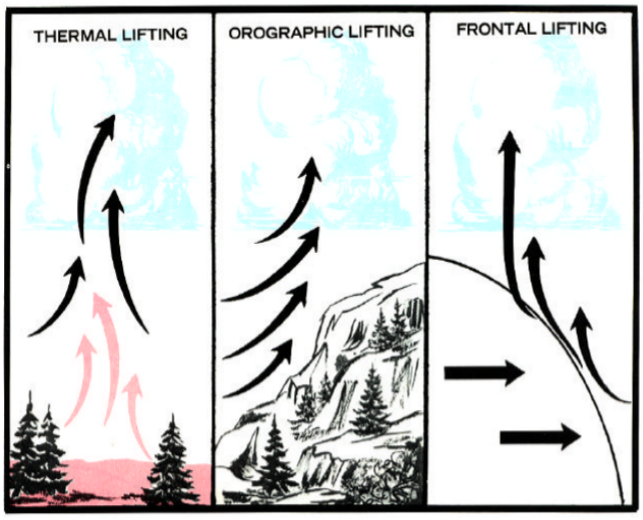

Figure 3.2: Thermal, orographic and frontal lifting [1].

The more important method of reaching saturation, by lowering air temperature, is accomplished in several ways. Warm, moist air may be cooled to its dew point by passing over a cold surface. The cooling takes place near the surface so that, with light wind conditions, fog is formed. If the winds are strong, however, they will cause mixing of the cooled air, and clouds will form several hundred or even a thousand or more feet above the surface.

Lifting of air, and the resultant adiabatic expansion, is the most important cooling method. The lifting may be accomplished by thermal, orographic, or frontal action. It produces most of the clouds and precipitation. Local heating will result in thermal lifting. As heated surface air becomes buoyant, it is forced aloft and cools. The air cools at the dry-adiabatic rate of 5.5^{o}F per thousand feet, while the dew point lowers only about 1^{o}F per thousand feet, reflecting the decreasing absolute humidity with expansion. Thus the temperature and dew point approach each other at the rate of 4.5^{o}F per thousand feet. If the locally heated air contains enough moisture and rises far enough, saturation will be reached and cumulus clouds will form.

In fact, a common method of estimating the base of cumulus clouds formed, usually in the summer months, by thermal convection in the lower layers is to divide the difference between the surface air temperature and dew point by 4.5. This gives the approximate height of the cloud base in thousands of feet. Thermal lifting is most pronounced in the warm seasons.

It may turn morning stratus clouds into stratocumulus with the possibility of light showers. More frequently, depending on stability, continued heating develops cumuli-form clouds that result in heavier showers and thunderstorms.

Rainfall associated with thermal lifting is likely to be scattered in geographic extent. In flat country, the greatest convective activity is over the hottest surfaces. In mountain country, it is greatest over the highest peaks and ridges. Orographic lifting, in which air is forced up the windward side of slopes, hills, and mountain ranges, is an important process in producing clouds and precipitation.

As in thermal lifting, the air is cooled by the adiabatic process. In the West, maritime polar air flowing in from the Pacific Ocean produces winter clouds and precipitation as it is lifted over the mountain ranges. If the temperature at the surface of a thermally lifted parcel of air was $84F$, the wet-bulb $71F$, and the dew point $66F$, saturation would be reached at 4,000 feet above the surface.

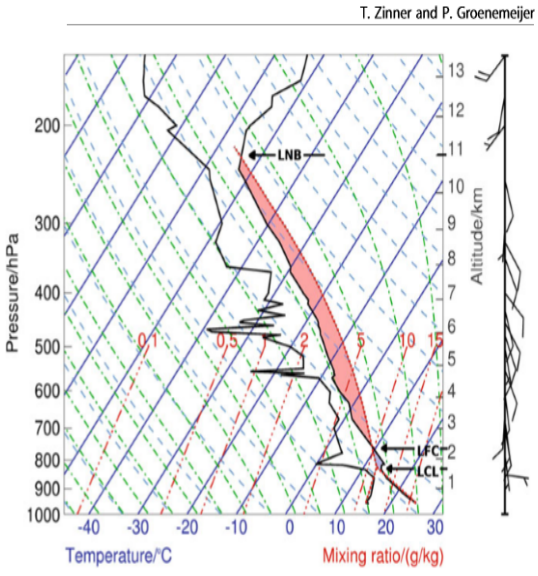

Figure 3.3: The temperature at the surface of a thermally lifted parcel of air [1].

Lifting of moist air over mountain ranges is an important process in producing clouds and precipitation. In the West the winter precipitation is heaviest on the western slopes of the Coast Ranges, Sierra- Cascades, and Rocky Mountains. Low lands to east of the ranges are comparatively dry.

Figure 3.4: Lifting of moist air over mountain ranges [1].

The Coast Ranges, Sierra-Cascades, and Rocky Mountains are the principal mountain systems involved. Lifting in each case occurs on the western slopes, and it is these that receive the heaviest precipitation. The lee slopes and adjacent valleys and plains receive progressively less as the air moves eastward. Similarly in the East, maritime tropical air that has moved into the central portion of the United States and Southern Canada is lifted and produces precipitation in the Appalachian Mountains as it progresses eastward. Thermal lifting usually produces cumulus clouds. Continued heating in moist air will result in showers and possibly thunderstorms.

Other air masses, such as continental polar and maritime polar, will also cause precipitation in these mountains if they have acquired sufficient moisture before being lifted. Frontal lifting, as air is forced up the slope of warm or cold fronts, accounts for much cloudiness and precipitation in all regions in the winter and in many regions during all seasons of the year. East of the Rockies and along the west coast, warm fronts, because of the gradual slope of their frontal surfaces, typically produce steady rains over extensive areas.

Cold fronts, with characteristically steeper and faster moving leading surfaces, frequently produce more intense rainfall from cumulonimbus clouds along the front or along a squall line ahead of the front. This rainfall, however, is usually more scattered and of a shorter duration

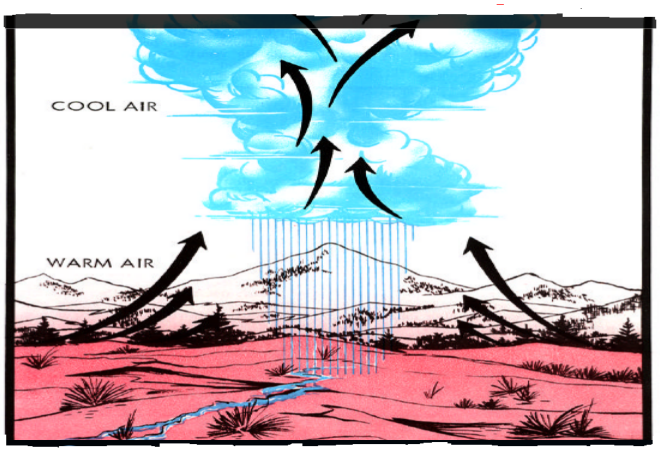

Figure 3.5: Thermal lifting [1].

than that produced by a warm front.

3.2 Cloud composition

Clouds consist of water droplets. When clouds become very cold, the water vapor condenses to form tiny six-sided crystals of ice, many of them like the snowflakes. Some clouds contain both water droplets and ice crystals. Water droplets do not freeze instantly when the temperature drops below freezing but supercool to varying extents first. Super cooled liquid water is water at a temperature below freezing but still in the liquid state. As the temperature continues to drop, the fraction of frozen droplets increases until they all freeze at -40^o C. The exact distribution of frozen to liquid droplets is not constant for all clouds but depends on the prevalence of trace atmospheric constituents called ice nuclei [3].

3.3 Atmospheric Aerosols

An aerosol is by definition a collection of solid or liquid particles in suspension in a gas. Strictly speaking the term aerosol includes both the particles and the suspending gas. In atmospheric sciences, it is usual however to use the term aerosol in its plural form to refer to theaerosol particles with out the atmosphere as the suspending gas. We follow that convention in this chapter, but occasionally refer to the aerosol when we refer to atmospheric aerosols in a generic way.

Atmospheric scientists also like to differentiate cloud particles from other types of particles in the atmosphere. For this reason we define aerosols as solid or liquid particles in suspension in the atmosphere to the exception of all hydro meteors (cloud droplets, ice crystals, raindrops, snowflakes, and graupel).

Aerosols are always present in the atmosphere but in extremely variable concentrations. This is due to the very large heterogeneity in aerosol sources and their relatively short residence time in the atmosphere(of the order of hours to weeks). The vast majority of aerosols are not visible to the naked eye because of their microscopic size but one can easily see the collective effect of aerosols in the atmosphere as soon as their concentrations are large enough. For example, a haze that reduces the atmospheric visibility and whitens the sky is nothing else than a collection of aerosol particles that interact with solar radiation.

A smoke plume is composed of microscopic particles that stem from incomplete combustion of carbonated fuels;these particles collectively darken the sky(see Fig.3.1). Aerosols canal so be visible when they get deposited in great quantity on the Earthsurface, as it is occasionally the case for Saharan dust particles (Fig. 3.2). The large variety of sizes and shapes among atmospheric aerosols is already striking from these photographs.Vegetation fire in the Alps on a cloudy day. One can notice the blueish color of the smoke plume which contrasts with the white color of clouds. The blueish color is due to the small aerosol size in the smoke plume and their larger effectiveness at scattering radiation in the shorter wavelengths of the visible spectrum [4].

The amount and properties of aerosols are extremely variable in space and time. This is why one is usually interested in characterizing a population of aerosols rather than individual particles. The most important characteristics of an aerosol population are the size distribution, chemical composition, and. It is useful to classify aerosols in different categories according to their properties. There are several possible classifications:

1. Primary aerosols have been emitted into the atmosphere as particles. This is the case of aerosols produced by the effect of the wind friction on an oceanic or terrestrial surface and aerosols produced during an incomplete combustion. Secondary aerosols designate those particles that have not been emitted directly in the particulate phase but come instead from the condensation of atmospheric gas-phase species. These gas-phase species, which can undergo a number of chemical transformations before they condense, are called aerosol precursors.

Figure 3.6: Vegetation fire in the Alps on a cloudy day [4].

The primary or secondary origin of aerosols offers a first way to categorize the atmospheric aerosol. The chemical composition of the aerosol usually provides a first idea as to whether the aerosol is primary or secondary.

2. Aerosol properties vary spatially and some of these properties can vary more or less systematically with the type of environment. One can thus speak of urban aerosols, semi-urban aerosols, continental aerosols, desertic aerosols, marine aerosols, volcanic aerosols or stratospheric aerosols. This is an imperfect categorization in that aerosols can be transported a long way and are therefore not necessarily representative of the location where they can be found. It is for example possible to observe marine aerosols above the continents and continental aerosols above the ocean. When local effects dominate, it may be useful though to refer to such aerosol types in a broad sense.

3. Aerosols can also be classified according to their origin. One can distinguish natural from anthropogenic sources. Natural sources consist of emissions from the ocean, soils, vegetation, fires, and volcanoes. Anthropogenic sources are largely dominated by emissions from the combustion of fossil fuels (i.e. coal and oil), bio fuels (plant biomass including wood, vegetable oils, animal waste), other fuels (e.g. peat), or vegetation fires caused by humans. Industrial activities, transportation, heating, or even domestic activities related to cooking in developing countries, are important sources of aerosols. Some industrial and agricultural

The snowpack has taken an orange color due to the deposition of dust.

Figure 3.7: Saharan dust event over the Alps [4].

activities can also emit primary aerosols referred to as industrial dust and arable dust, respectively. None of these classifications can categorize fully and systematically the aerosol.The different aerosol populations mix and interact with each other in the atmosphere so that some of the terms and aerosol classes that we have just introduced are some how misuses of language.

3.4 Nucleation

Nucleation is the process when water-vapor molecules accumulate to build an initial drop or ice crystal. In the air, water molecules collide randomly and build embryo droplets.A small fraction of collision between water vapor molecules in the atmosphere are inelastic, leading to formation of molecules aggregates. The size of aggregate can be increased by inelastic collision between the molecular aggregates, or by accretion of individual molecules. Most aggregates have a short life time, since they disintegrate under continual bombardment. When an aggregate attains a size sufficient for survival, then nucleation of a water drop has occurred. As shown in figure below

Because of surface tension effects, the escaping tendency of the water molecules in a spherical drop is reduced, and equilibrium vapor pressure over a spherical drops can be substantially

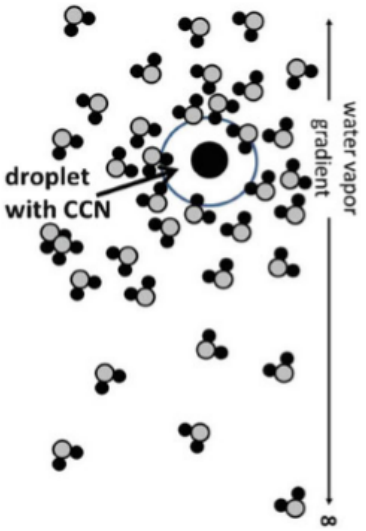

Figure 3.8: Schematic view of diffusion [13].

greater than values derived for bulk water without significant surface tension effects. This higher value of saturation vapor pressure over a curved surface related to a bulk flat surface at the same temperature is a consequence of the work that must be performed on the system to increase the drop's surface area. For a drop with a surface area A=$4\pi r^2$ the surface tension work is determined from,

$$dW_{st} = \sigma_{lv} 8\pi r d_r \tag{3.1}$$

Where σ_{lv} is the surface tension between the liquid and vapor phases. Since the work of expansion against a difference in pressure must equal the work done in changing the surface drop area,

$$\sigma_{lv} d_A = \Delta p d_v \tag{3.2}$$

where, Δp is the pressure differential between the external ambient pressure and the internal pressure of the drop, For spherical drop, we can write.

$$\sigma_{lv} 8\pi r d_r = \Delta p 4\pi r^2 d_r \tag{3.3}$$

$$\delta p = \frac{2\sigma_{lv}}{r} \tag{3.4}$$

Note that the smaller drop, the larger the pressure differential. Under standard atmospheric pressure condition, the internal pressure of a drop with μm radius is 1.5_{atm}. The existence

of this excess internal pressure in a small drop is a fundamental consequence of the surface tension. The surface tension for the vapor-liquid interface of water σ_{lv} is a function of temperature and is given by.

$$\sigma_{lv} = 0.0761 - 1.55 \times 10^4 T \tag{3.5}$$

To determine the condition for nucleation of a water drops, we use Gibbs function

$$dG = -\eta dT + \upsilon d_p + \sigma_{lv} dA + \mu_l d_{nl} + \mu v d_{nv} \tag{3.6}$$

If we assume that nucleation occurs at constant temperature and pressure, and that$d_{nl} = -d_{nv}$,we have

$$dG = \sigma_{lv} 8\pi r d_r + (\mu_l - \mu_v)\, d_{nl} \tag{3.7}$$

Since dG is an exact differential, we can consider nucleation to occur in two stages first, the bulk condensation of supersaturated water vapor on to a plane surface; and second, the formation of drops from the bulk water. For the first stage, we can write as

$$dG = (\mu_l - \mu_v)\, d_{nl} \tag{3.8}$$

The term$(\mu_l - \mu_v)$ can be evaluated as follows. For an isothermal process involving one mole of water vapor, we can

$$d\mu_v = dG = R^* T d\,(\ln_e) \tag{3.9}$$

or

$$mu_v = \mu_{vo} + R^* T \ln \frac{e}{e_o} \tag{3.10}$$

Where μ_{vo} is a reference chemical potential that varies only with temperature and ϵ_0 is the corresponding reference vapor pressure. At saturation, we can write

$$d\mu_{vs} = dG = R^* T d\,(\ln_{e_s}) \tag{3.11}$$

and,

$$\mu_{vs} = \mu_{vo} + R^* T \ln \frac{e_s}{e_o} \tag{3.12}$$

Since $\mu_l = \mu_{vs}$ when the two phases in equilibrium over a plane surface,we can writ

$$\mu_l - \mu_v = R^* T \ln \left(\frac{e_s}{e}\right) \tag{3.13}$$

The number of moles of water vapor, dn_v that condense on to spherical drops can be written

$$dn_v = \frac{1}{M_v} dm_v = \frac{\rho l}{M_v} 4\pi r^2 dr \tag{3.14}$$

Substitution of Eq.(3.12 and 3.13) in to Eq.(14) yields

$$dG = \left(-R_v T \ln \frac{e}{e_s} \rho_l 4\pi r^2 + \sigma_{lv} 8\pi r \right) dr \tag{3.15}$$

Integration of (3.14) for the nucleation process gives

$$\Delta G = 4\pi r^2 \sigma_{lv} - \frac{4}{3} \pi r^3 \rho_l R_v T \ln \varsigma \tag{3.16}$$

where $\varsigma = \frac{e_s(r)}{e_s}$ is the saturation ratio and e_s is the saturated vapor pressure over a plane surface.

3.5 Thermodynamics of charge enhanced nucleation formation and growth

Droplets are often charged.There can be many sources of this charge, typically charge separation by the differential motion of drops of different sizes, or they always-present electric current between ionosphere and the Earths surface, which can be charge up individual droplets at cloud boundaries. A net charge corresponds to an electrostatics energy on the drop which modifies the Gibbs free energy budget and therefore the saturation vapour around drop.

A spherically symmetric charge distribution has an electrostatics V_e at radius r of

$$V_e = \frac{Q}{4\pi\epsilon r} \tag{3.17}$$

Where Q the charge and ϵ the electric permittivity; for all practical purposes we can ϵ_0, the permittivity of the vacuum . Increasing the charge by an amount dQ requires work dW against the electrostatic potential of

$$dW = V_e d_Q \tag{3.18}$$

So to charge up a sphere from no charge to a total charge Q requires a total electrostatics energy W of

$$W = \frac{Q^2}{8\pi\epsilon r} \tag{3.19}$$

For a given charge, a charge in a radius of the drop would charge the electrostatics energy. That mean that on evaporating the mass δm from a charged drop, its electrostatics energy will charge. Analogous to the Gibbs function budget is modified to

$$\delta G = (g_v - g_l)\,\delta M - \frac{Q^2}{8\pi\epsilon r^2}\delta_r \tag{3.20}$$

We have ignored any effects of solute or surface tension here in order to isolate the specific effects of the charge and we have omitted the term proportional to δ_e and δ_T as these are zero by construction. As before,δ_r is related to δ_M. The relevant inverse permittivity is

$$\frac{1}{\epsilon} = \frac{1}{\epsilon_{air}} - \frac{1}{\epsilon_{water}} \tag{3.21}$$

But the permittivity of water is 80 times larger than that of air, which in turn is very similar to that of the vacuum, ϵ_0.

$$\delta_r = \frac{\delta_r}{A} = -\frac{v_l}{4\pi r^2}\delta M \tag{3.22}$$

With V the volume of the sphere, A its area and v_l the specific volume of the liquid phase(remember that positive δM corresponds to a reduction in droplet volume). For the variations in the total Gibbs function to vanish, therefore we have

$$g_v\,(e,T) = g_l\,(e,T) - \frac{Q^2 v_l}{32\pi^2\epsilon r^4} \tag{3.23}$$

At infinite radius, the charge effect vanishes and the vapour pressure is the same at the flat surface value from the Clausius Clapeyron equation. This result is a new charge-induced saturation ratio S_Q

$$S_R = \frac{e_s(Q)}{e_s(0)} = exp\left(\frac{-Q^2 v_l}{32\pi^2\epsilon R_v T r^4}\right) \tag{3.24}$$

where have returned to the usual notation of e_s for saturation vapor pressure and with $e_s(0)$ the saturation vapor pressure for an uncharged drop. We see that the charge of drop decrease the vapor pressure around the drop by an exponential factor, $s_r < 1$. The effect is called the Rayleigh effect. The physical picture is that on a charged drop it is more difficult to evaporate water because the associated reduction of droplet radius requires extra electrostatic energy. Analogous to the electrostatic term can be absorbed in the pressure dependency of the Gibbs function,

$$g_l\,(e,T) - \frac{Q^2 v_l}{32\pi^2\epsilon r^4} = g_l\,(e',T) \tag{3.25}$$

with,

$$e' = e - \frac{Q^2}{32\pi^2\epsilon r^4} \tag{3.26}$$

In other word, the effective pressure in side in drop decreased by the charge. This decrease of pressure in side in a drop should, again, come as no surprise: the repulsive electrostatic charges try to expand the drop. The pressure drop due to the charge is p_Q, with

$$p_Q = \frac{Q^2}{32\pi^2\epsilon r^4} \tag{3.27}$$

The charge effect can be interpreted as a Boltzmann factor,

$$exp\left(-\frac{Q^2 v_l}{32\pi^2\epsilon R_v T r^4}\right) = exp\left(\frac{-\Delta E}{K_B T}\right) \tag{3.28}$$

With the excess energy per molecule written in terms of the pressure drop p_Q due to the charge.

$$\Delta E = \frac{p_Q V}{N} \tag{3.29}$$

Where N the number of molecules in the drop and V its volume. The charge helps to droplets by counteracting the Kelvin effect. It is of interest to see when the two effect compensate. Using Eq.(2.18 and 2.14). We find compensation when $S_K S_R$=1 or equivalently

$$p_r = p_Q \tag{3.30}$$

For a given charge Q this compensation will occur at a critical radius r_R, the so-called Rayleigh radius, with

$$r_R^3 = \frac{Q^2}{64\pi^2\epsilon r} \tag{3.31}$$

Rayleigh showed that for droplets with a radius smaller than r_R the repulsive electrostatic force become so strong that the droplets disintegrate explosively; thermodynamically this can be interpreted as the destabilizing charge pressure overcoming the stabilizing capillary pressure. With the definition of the Rayleigh radius r_R and the Kelvin radius r_k we can rewrite the combined Rayleigh and Kelvin effect to the saturation ratio S as

$$S = S_K S_R = exp\left[\frac{r_k}{r}\left(1 - \frac{r^3 R}{r^3}\right)\right] \tag{3.32}$$

For this equation it becomes clear that charge effect is only important when the droplet radius is close to the Rayleigh radius. Only for droplets smaller than the Rayleigh radius, the charge effect is dominates the Kelvin effect and saturation can occur at relative humidities below 100

percent. So what is atypical values for the Rayleigh radius r_R. For a unit charge,e(an electron has charge-e), we can put in standard values for a water droplet. To find $r_R \simeq 0.4nm$,for Very tiny indeed; in fact this is so small that quantum effects become important. For droplets with such high saturation ratios to grow we need relative humidities in excess of $RH = 400$ percent; this does not occur in nature. We are forced to conclude that in the atmosphere charge ions cannot serve as condensation nuclei. This is of importance in the present discussion on influence of cosmic rays on climate.

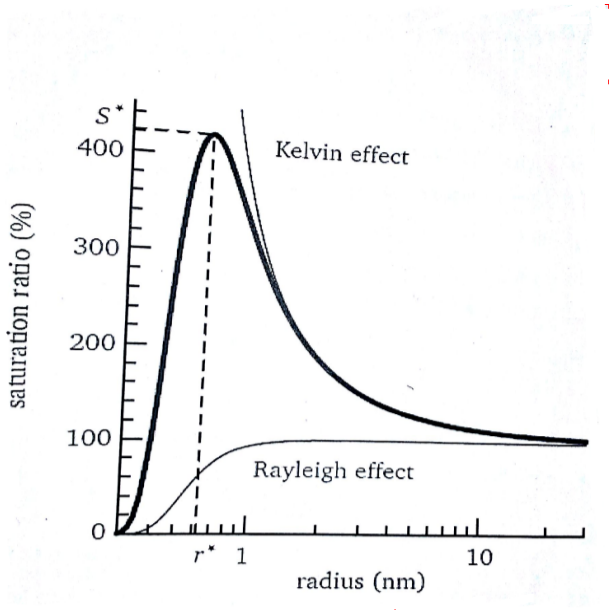

Figure 3.9: Saturation ratio as a function of charged droplet radius for a clean droplet at ${}^0c$ with one unit charge [13].

Cosmic rays generally produce single charge ions in the atmosphere and it has been argued that these may form cloud condensation nuclei. It is clear from the argument above that this is impossible. In order for cosmic ray to influence cloud formation other processes such as ion clustering or charge accumulation have to occur.

Super saturation of 400 percent are fairly easy to achieve in the laboratory. By expanding a vessel which has water and its vapour in it we can achieve very high super saturations: on adiabatic expansion there pressure as well as the temperature will drop. By the Clausius Clapeyron equation the saturation vapour pressurewill also drop and is found to drop more than the pressure itself. This means that vapor that was originally saturated will become supersaturated. This supersaturation in the laboratory is the basis of the Wilson used a

closed chamber. Wilson used a closed chamber with saturated vapour and by expansion made the vapour supersaturation. Any condensation nuclei would rapidly saturate and grow in to drops.

3.6 Droplet growth

If a cloud droplet is activated it is out of equilibrium: the environment is supersaturated with respect to the droplet and the droplet will grow by condensation of vapour onto the droplet. However, this process is limited by the speed with which the condensed vapour is replenished by new vapour from the environment. If there is no such replenishment, the immediate vicinity of the drop will run out of vapour and will become sub saturated. Consequently, the drop would stop growing.

The process that replenishes the vapour is a diffusive flux of water vapour. To a good approximation, this diffusive flux is proportional to the gradient in water vapour density (*Ficks* law). So the vapour flux F_v is

$$F_v = -D\Delta v\rho_v, \tag{3.33}$$

where ρ_v is the mass density of the vapour and the constant of proportionality D is called the diffusion coefficient. The direction of the flux is opposite to the gradient of the vapour density: mass flows from high densities to low densities. If a drop is activated, vapour molecules will condense on the drop thus reducing the vapour density in the immediate vicinity of the drop. This sets up a gradient in vapour density between the immediate vicinity of the drop and the far field. This will lead to a diffusive flux of water vapour towards the drop.

The value of the diffusion coefficient varies with the species being diffused (approximately with the square root of the mass) and it increases with temperature. For diffusion of water vapour at 5^oC we have; $D = 22 \times 10^{-6} m^2 s^{-1}$ with an increase of about $0.15 \times 10^{-6} m^2 s^{-1}$ per 1^0C temperature increase. The origin of the vapour flux is the random motion of the vapour molecules. If there is a density gradient there will be fewer molecules moving from the low density to the high density region than the other way around simply because the low density region has fewer molecules to move to the high density region than the other way around. This discrepancy results in a net motion of molecules from the high density area to the low density area. It is also clear that the discrepancy is proportional to the difference in molecule densities between the high and the low density regions.

Assuming the situation is spherically symmetric, all variables will be functions of the radial coordinate r and of time t. We now make the assumption that away from the drop the density field is constant in time (but not necessarily constant in space). The vapour flux $F_v$is according to $Fick's$fs law

$$F_v = -D\frac{d\rho_v}{d_r}\hat{r} \tag{3.34}$$

with $\hat{r}$ the unit vector in the radial direction. The total mass flux F_m into a shell of radius r then is

$$F_m = -4\pi r^2 F_v \hat{r} = 4\pi D r^2 \frac{d_{\rho_v}}{d_r} \tag{3.35}$$

Note the sign here: if the vapour density increases away from the drop there will be a positive mass flux towards the drop.

F_m is the total vapour mass entering any spherical shell, so it is also the total vapour mass entering the droplet. So the mass growth $\frac{dM_d}{d_t}$ of the droplet is

$$\frac{dM_d}{d_t} = F_m \tag{3.36}$$

The drop grows by condensing the vapour onto its surface. This condensation releases latent heat, which will heat up the drop. In a steady state situation this heat will be conducted away by a total heat flux F_q equal to

$$F_q = -L\frac{dM_d}{d_t} = -LF_m \tag{3.37}$$

with L the latent heat of condensation and F_q the total inward heat flux through a spherical shell. Note that as the drop grows (F_m positive) the heat flux is directed outward (F_q negative). According to $Fouriers$ law, the heat flux is proportional to the gradient of the temperature. Therefore total inward heat flux is, analogous to Eq. 3.35 Flux of vapour through spherical shells towards a drop. In a steady state the flux through all spherical shells has to be equal.

$$F_q = 4\pi K r^2 \frac{dT}{d_r}, \tag{3.38}$$

with K the heat conduction coefficient. The thermal conductivity of air increases with increasing temperature. A typical value of K at 10^oC is $K = 25 \times 10^3 W m^{-1} K^{-1}$. Between $-5^oCand25^oC$, K varies only by about 4%. In a steady state, the total mass flux into a spherical shell is independent of the radius of the shell. Consider the volume between an outer shell of radius r_1 and an inner one of radius r_0, For a steady state the total vapour mass

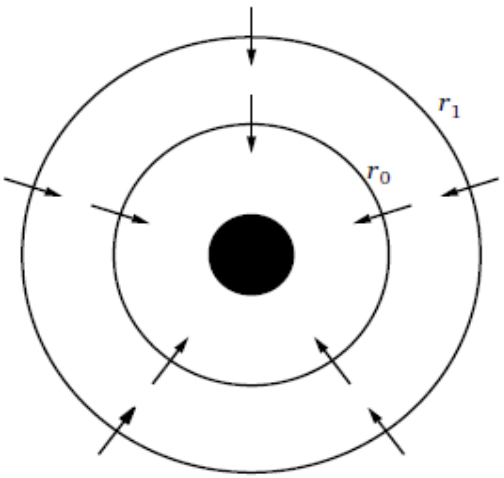

Figure 3.10: Flux of vapour through spherical shells towards a drop. In a steady state the flux through all spherical shells has to be equal.

flux into this volume needs to be the same as the total flux out of this volume. Because this is true for any pair of radii, we have

$$F_m = constant$$

We can combine this constraint with *Ficks* law to find a simple expression for the mass flux. First, rewrite Eq. 3.35 as

$$\frac{F_m}{r^2} = 4\pi D \frac{d_{\rho_v}}{d_r} \tag{3.39}$$

Integrating this expression between the droplet radius $r = rd$ and the far field $r = \infty$ we find

$$\frac{F_m}{rd} = 4\pi D \left(\rho_v - \rho_v, d\right) \tag{3.40}$$

where we have written ρ_v for the vapour density in the far field (that is, the vapour density of the air not too close to the drop) and ρ_v, the vapour density at the drop surface.

The analogous argument holds for a steady heat flux: the total energy between radii r_o and r_1 needs to remain constant, so the heat entering the volume must be the same as the heat exiting the volume. This can only be true if F_q is not a function of the radius. This then leads to

$$\frac{F_q}{r_d} = 4\pi K \left(T - T_d\right) \tag{3.41}$$

with T the air temperature in the far field and T_d the temperature at the drop surface. Substituting these expressions in the steady state condition of Eq. 3.37 we find

$$T - T_d = -\frac{LD}{K}\left(\rho_v - \rho_v, d\right) \tag{3.42}$$

The temperature difference and the vapour density difference are proportional. For large heat conductivity K, the temperature difference will be small; the latent heat can be efficiently transported away from the drop. For small heat conductivity, the latent heat will accumulate on the drop and larger temperature gradients are required to transport the heat away. The precise meaning of large or small will become clear later on.

To calculate the droplet growth we need to know the difference between the vapour density at the drop surface and the vapour density of the air. We assume that vapour near the drop is saturated. Using the ideal gas law we thus have

$$\rho_v, d = \rho_{vs}(Td) = \frac{\rho_s(Td)}{R_v T_d} \tag{3.43}$$

We have suppressed the radius dependence of the saturated vapour pressure so this argument is only accurate for drops substantially larger than the activation radius. It is fairly straightforward to include this radius dependence but this would just make the equations more convoluted without providing any further insight.

Let us for the moment assume the drop temperature Td is the same as the air temperature T. We then have

$$\rho_v, d = \rho_{vs}(T) = \frac{\rho_s(T)}{R_v T} \tag{3.44}$$

Using the relative humidity, the vapour density in the far field can be written in terms of the saturated vapour density,

$$\rho_v = RH_{\rho_{vs}}(T) \tag{3.45}$$

With these expressions the total mass flux Fm in Eq. 40 becomes

$$F_m = 4\pi D r_d \rho_{vs}(RH - 1) \tag{3.46}$$

where ρ_{vs} is evaluated at temperature T. In other words, the droplet growth is due to supersaturation: if the relative humidity is larger than 100% (positive supersaturation) the drop will grow; if it is smaller than 100% the drop will shrink. So for relative humidities below 100% the above equation can be used to calculate how quickly a droplet evaporates. Note

also that the growing droplets extract water vapour from the air and thus reduce its relative humidity; droplet growth by condensation is a self-limiting process.

In the above derivation we took $T_d = T$. As the droplet is heated up by condensation this assumption is not valid. However, assuming that $T_d$and T are not too far apart we can linearize the saturated vapour pressure in Eq. 3.43 around temperature T,

$$\rho_v, d = \rho_{vs}(T_d) = \rho_{vs}(T) + \frac{d\rho_{vs}}{dt}(T_d - T) \tag{3.47}$$

By using the ClausiusClapeyron we have

$$\frac{d\rho_{vs}}{dT} = \left(\frac{L}{R_v T} - 1\right)\frac{\rho_{vs}}{T} \tag{3.48}$$

We therefore find

$$\rho_v, d = \rho_{v_s}\left(1 + \left(\frac{1}{R_v T}\right)\frac{T_d - T}{T}\right) \tag{3.49}$$

The temperature difference between the drop and the far field can be expressed as a vapour density difference by Eq. 3.42 We now get for the vapour density difference

$$\rho_V = \rho_{v_s}, d = \rho_{v_s}\left(RH - 1 - \frac{\rho_v - \rho_v, d}{\rho k}\right) \tag{3.50}$$

where we introduced a conduction density scale

$$\rho k = \frac{KT}{LD}\left(\frac{L}{R_V T} - 1\right)^{-1} \tag{3.51}$$

This equation can be rearranged to find the vapour density difference $\rho_v - \rho_v, d$ in terms of the saturated vapour density ρ_{v_s}. A particularly compact way to write the ensuing expression is

$$\rho_v - \rho_v, d = \rho_r(RH - 1) \tag{3.52}$$

where we have introduced a reduced vapour density ρ_r as

$$\frac{1}{\rho_r} = \frac{1}{\rho_v, d} + \frac{1}{\rho k} \tag{3.53}$$

This density difference then leads to the total mass flux,

$$F_m = 4\pi D r_d \rho_r \left(RH - 1\right) \tag{3.54}$$

As can be seen, Eq. 3.54 is the same as Eq. 3.53 with the far-field saturated vapour density replaced by the reduced vapour density ρ_r. It now becomes clear that latent heating slows the growth of the of the drop because for all ρk we have

$$\rho_r < \rho_{v_s} \tag{3.55}$$

Physically this makes sense: a warmer drop reduces the supersaturation with respect to the far field and thus reduces the vapour diffusion towards the drop. There are two limiting cases of interest. If the heat conduction coefficient K is very large, then any excess heat is immediately diffused away from the drop. We thus expect the heating effect to be small. Indeed, for$K \to \infty$ we have $\rho k \to \infty$ and we find

$$\rho_r \to \rho_{v_s} \tag{3.56}$$

So the mass flux in Eq. 3.54 reduces to Eq. 3.53 the other limiting case is for a small heat conduction coefficient. In this case the latent heat cannot be transported away from the drop and we expect a strong reduction of the supersaturation. Indeed, for $\rho k \to 0$ we have $K \to 0$ and we find

$$\rho_r \to \rho k \to 0 \tag{3.57}$$

so the mass flux to the drop vanishes. Realistic values of the diffusion and conduction coefficients give a value of $\rho k$of about $7 \times 10^{-3} kgm^{-3}$ and it is typically similar or somewhat larger than ρ_{v_s}. The heating effect therefore reduces the growth by a factor of two to four compared to the case without heating. The reduction becomes larger for higher temperatures. In fact, the reduced vapour density is a much weaker function of temperature than the saturated vapour density. The reduced density is very nearly a linear function of the temperature. The approximation,$\rho_r = (2.80 + 0.11T(^0C)) \times 10^{-3} kgm^{-3}$is accurate to within 4% between temperatures of -5^0C and 50^0C. Now we have an expression for the mass-flux, we can calculate the change in droplet radius r_d We have

$$F_m = \frac{dM}{dt} = \rho_l \frac{dV}{dt} = 4\pi\rho_l r_d^2 \frac{dr_d}{dt} \tag{3.58}$$

with ${}_\rho l$ the density of the liquid and V the droplet volume, $V = (\frac{4}{3})\pi r_d^3$. Substituting the expression for the mass flux, we find

$$r_d \frac{dr_d}{dt} = D\frac{\rho_r}{\rho_l}(RH - 1) \tag{3.59}$$

The right-hand side of Eq. 3.59 is a constant at given temperature and relative humidity, so we can integrate the equation to find:

$$r_d t = \sqrt{(r_d^2(0) + 2\wp t)} \tag{3.60}$$

with $\wp = D\frac{\rho_r}{\rho_l}(RH - 1)$ The droplet radius grows as the square root of time. When the relative humidity is smaller than 100%, this equation describes how the droplet radius reduces in time through evaporation.

Square root change of scale with time is typical of phenomena that are driven by diffusive processes. For example, the typical size of a cloud of tracer released by a local source grows with the square root of time and heat penetrates a conductor to a depth that grows with the square root of time.

An interesting property of Eq. 3.60 is that the radii of two droplets that start growing from different initial radii become more and more similar. From Eq. 3.60 we have

$$r_a^2(t) - r_b^2 t = r_a^2(0) - b_a^2(0) \tag{3.61}$$

with r_a and r_b the radii of two different droplets growing in the same background. This is equivalent to

$$r_a^2(t) - r_b^2 t = \frac{r_a^2(0) - b_a^2(0)}{r_a^2(t) + r_b^2 t} \tag{3.62}$$

Because both r_a and r_b will grow with time, their radii will get closer with time.

Note that with changing droplet radius, the mass flux F_m is not constant in time. We then need to ask ourselves whether the steady state assumption leading to $F_m = constnt$ is accurate. Any density fluctuations in the vapour field around the drop will adjust to a steady state through a diffusion process with diffusion coefficient D. The boundary condition change through the changing droplet radius occurs with an effective diffusion coefficient $\wp$. Because $D \gg \wp$.

The vapour field adjusts to a steady state on a much faster time scale than that of the changing boundary (the relevant time scales are $\frac{r_d^2}{D}$ and $\frac{r_d^2}{\wp}$). We can therefore safely assume that the vapour flux is constant for any given droplet radius during the growth process.

Let us now put some numbers in Eq. 3.60 a typical supersaturation at activation is about 2% or less. At 10^oC, the reduced vapour density ρ_r is about $4 \times 10^{-3} kgm^{-3}$. The time taken to grow from a small droplet to a typical rain drop of radius about $1mm$ is several days. Even for growth to modest droplet size of several tens of micrometres we find times of several hours. The typical time scale for rain to form in a convective cloud is very much shorter, perhaps half an hour or less. Diffusive growth cannot be the complete story to go from activation to raindrop.

3.6.1 Droplet growth by collision and coalescence

Cloud droplets will be carried by air currents within the cloud, and if they bump into each other, it is called a collision. However, if they collide then stick together, that is called coalescence. Although this process is important, especially in the tropics, it falls short of being the primary mechanism for the formation of raindrops.

Updrafts in a cloud can transport a droplet upward repeatedly allowing it many opportunities to fall back down through the cloud and collide and coalesce with other droplets. Initially by diffusion, and subsequently by collision and coalescence, tiny aerosol nuclei grow into large water droplets more than 10,000 times their initial size. The collision-coalescence process is often called the warm-rain process, since it is the only way to explain precipitation formation in clouds that remain above freezing.However, it can also occur in cold clouds. Collisions are influenced by gravitational, electrical and aerodynamic forces. Gravitational effects dominate in clouds where large droplets capture small ones.

Collisions require different fall velocities between small and large droplets (ignoring turbulence and other non-gravitational forcing).This is a dominant process for precipitation formation in warm clouds (tops warmer than about -15^0 C). Since the bigger drops fall faster than the smaller drops,they will collect the smaller drops - the bigger drop grows. Droplet fall speed is called its terminal velocity. Need droplets of different sizes for this process to really work.

3.7 Droplet terminal fall speed

Terminal velocity is the constant speed that a falling object has when the gravity force and the drag force applied on the subject reach a balance. Terminal velocity depends on the size of the object: small objects fall slowly and large objectives fall quickly.

Because the steady settling velocity of a droplet as it falls under the influence of gravity through still air (called the terminal fall speed of the droplet) increases with the size of the droplet, those droplets in a cloud that are some what larger than average will have a higher than average terminal fall speed and will collide with smaller droplets lying in their paths. The drag force C_D on a sphere of radius r is as shown figure:

$$F_R = \frac{\pi \mu r C_D R_e}{4} \tag{3.63}$$

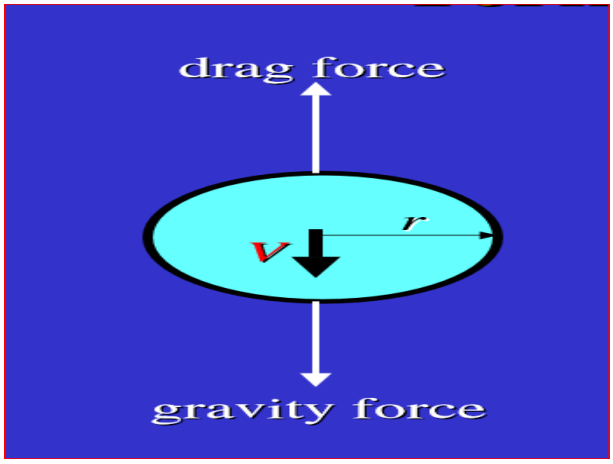

Figure 3.11: Droplet Size.

$$R_e = \frac{2\rho \upsilon r}{\mu} \tag{3.64}$$

Where Reynolds number, ρ fluid density, μ dynamic viscosity. The gravitational force minus buoyancy on a sphere of radius r is

$$F_A = \frac{4}{3}\Pi\gamma^3 g\left(\rho_L - \rho\right) \tag{3.65}$$

Terminal fall speed occurs when $F_A = F_R$.

$$\frac{4}{3}\pi\gamma^3 g\left(\rho_L - \rho\right) = \frac{\pi\mu\upsilon\gamma C_D R_e}{4} \tag{3.66}$$

If

$$\rho_L \gg \rho$$

$$v^2 = \frac{8rg\rho_L}{3\rho C_D} \tag{3.67}$$

$$v = \frac{16r^2 g\rho_L}{3\rho C_D R_e} \tag{3.68}$$

$$\frac{C_D R_e}{24} = 1 \tag{3.69}$$

For very small Reynolds numbers

$$v = \frac{2r^2 g\rho_L}{9\mu} \tag{3.70}$$

If we knew the drag coefficient, then we could find the terminal velocity. The drag coefficient is determined experimentally, with the following results obtained for the terminal velocity of droplets.

3.7.1 Growth due to collection with smaller and uniform droplets

As a droplet falls it may collide with other droplets. The volume swept out by a droplet of radius R falling through a distance of dh is $\rho R_2 dh$. However, if the droplet is falling through a population of smaller droplet each of radius r, then the collection radius of the falling droplet is R because it will touch any smaller droplet of $d = R + r$ from the center of the large droplet as shown figure below. So, the volume swept out by the droplet is

$$d_v = \pi(R + r)^2 dh \tag{3.71}$$

If the larger droplet is falling at its terminal velocity,$u(R)$ then $dh = u(R)dt$ and Eq. (2)

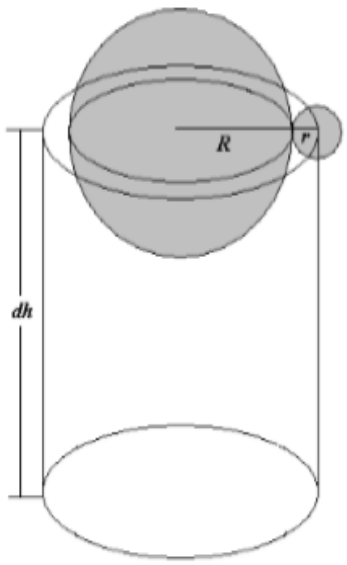

Figure 3.12: Growth due to collection with smaller and uniform droplets.

becomes

$$dv = \pi \left(R + r\right)^2 M_u \left(R\right) \tag{3.72}$$

If we assume that as the large drop falls, any smaller droplets that are at least partially within the volumedV will be collected by or stick to the large drop. Therefore, its mass will

increase by the liquid water content,M of the smaller droplets, multiplied by the volume,dV thus

$$dm = Md_v = \pi (R + r)^2 Mu (R) \tag{3.73}$$

But the smaller droplets are also falling at their terminal velocity,$u(r)$We have to correct for this, because if the small droplets were falling at the same speed of the larger droplet there would be no increase in mass since there would be no collisions. The greater the difference in the terminal Lectures in Clouds Physics velocities, the fewer collisions and the less mass will be collected. The corrected formula is

or, in terms of growth rate

$$\frac{dm}{dt} = \pi (R + r)^2 M [u (R) - u (r)] \tag{3.74}$$

If all the smaller droplets within the volume coalesce (stick) to the falling droplet could be used to predict the rate of mass growth of the falling droplet.

3.8 Collision efficiency

A drop of radius R overtaking a drop of radius r. An object with zero inertia would be swept aside. The collision efficiency E of a droplet of radius r with a drop of radius R is therefore defined as

$$E = \frac{y^2}{(R + r^2)}$$

If $y < (r + R)$ not all drops in the geometrical collision cross section will collide with the collector drop. there is a wake effect and the efficiency can be greater than one. In reality, not all the smaller droplets will collide with or stick (coalesce) to the larger droplet. One reason for this is that, the air flow around the larger drop may force the smaller droplet away as the drops pass. It is also possible for droplets that weren't in the path of the larger droplet to be captured by the wake of the falling droplet. This is known as wake capture.

Even if two droplets collide, they may not actually stick, or coalesce. One reason for this is that there may be a microfilm of air between the droplets that prevents their surfaces from contacting. All the above effects are taken into account by deifying a collection efficiency, E, that is just the ratio of the number of small droplets that actually stick to the larger

drop to the total number of small droplets within the volume. The collection efficiency must be experimentally determined, and is going to depend on many physical factors such as the drop sizes involved, temperature, turbulence, wind shear, etc. Usually E1. With collection efficiency taken into account, Eq. (3.37) for the growth of droplet mass becomes

$$\frac{dm}{dt} = \pi EM\left(R+r\right)^2 M\left[u\left(R\right)-u\left(r\right)\right] \tag{3.75}$$

Equation (3.43) is often written as,

$$\frac{dm}{dt} = K\left(R,r\right)M \tag{3.76}$$

$$K\left(R,r\right) = \pi EM\left(R+r\right)^2 M\left[u\left(R\right)-u\left(r\right)\right] \tag{3.77}$$

where is called the gravitational collection kernel.

3.9 Growth equation in terms of radius

In terms of the change in radius of the larger drop, the growth-rate eq. is;

$$\frac{dm}{dt} = \frac{EM(R+r)^2}{4\rho_L R^2}\left[u\left(R\right)-u\left(r\right)\right]$$

If we assume the smaller droplets have a negligible terminal velocity compared to the larger droplets, then we can use the approximations

$$u(R) - u(r) \cong u(R)$$

$$R + r \cong R$$

to write the growth equation in terms of radius as

$$\frac{dR}{dt} = \frac{EM}{4\rho_L}\frac{dt}{u}(R) \tag{3.78}$$

From the chain rule this can be written as

$$\frac{dR}{dz} = \left(\frac{dR}{dt}\right)\frac{dt}{dz} = \left(\frac{dR}{dt}\right) \Big/ \left(\frac{dz}{dt}\right) \tag{3.79}$$

But d_z/d_t is the rate at which the drop is actually falling, and is the difference of the speed of any updrafts and the terminal velocity of the drop.

$$\frac{\partial R}{\partial z} = U - u(R) \tag{3.80}$$

where U is the updraft velocity. Therefore, Eq. (3.7) becomes:

$$\frac{dR}{dt} = \frac{EM}{4\rho_L}\frac{u(R)}{U - u(R)} \tag{3.81}$$

If the updraft velocity is negligible, then this becomes;

$$\frac{dR}{dz} \cong -\frac{EM}{4\rho_L} \tag{3.82}$$

Eq.(3.47) suggests that the droplet radius should decrease with height since as the droplet falls (z getting smaller) its radius should be increasing through the collision(coalescence) process.

3.9.1 Growth due to collision with smaller droplets of non-uniform size

If there are two different sizes of smaller drops, $r_1 and r_2$, then the growth rate in terms of mass will have a separate term for each smaller droplet size present and will be

$$\frac{dm}{dt} = \pi E_1 M_1 \left(R + r_1\right)^2 \left[u(R) - u(r_1)\right] + \pi E_2 M_2 \left(R + r_2\right)^2 \left[u(R) - u(r_2)\right] \tag{3.83}$$

In general, if there are N different sizes of little drops, then we have;

$$\frac{dm}{dt} = \pi \sum_{i=1 to N} E_i M_i \left(R + r_i\right)^2 \left[u(R) - u(r_i)\right] \tag{3.84}$$

where each drop radius,r_i has its own collection efficiency and liquid water content,$E_i$and Mi. Written in terms of the collection kernel, we have;

$$\frac{dm}{dt} = \sum_{i=1 to N} K(R, r_i) M_i \tag{3.85}$$

If the small droplets are distributed continuously then the summation becomes an integration;

$$\frac{dm}{dt} = \int K(R, r_i) dM \tag{3.86}$$

and we know that;

$$dM = \frac{4}{3}\pi \rho_L r^3 n_d(r) dr \tag{3.87}$$

So that the growth equation for a continuous droplet spectrum is:

$$\frac{dM}{dt} = \frac{4}{3}\pi \rho_L \int K(R, r_i) r^3 n_d(r) dr, \tag{3.88}$$

This Eq. must be solved numerically, and is complicated by the fact that the collection kernel depends on radius of the small droplets, so it must remain within the integral. The collection

kernel is usually determined empirically. If we assume that R>>rand u(R) >> u(r) then we can show that:

$$\frac{dR}{dt} = \frac{EM}{4\rho_L} u(R) \tag{3.89}$$

and

$$\frac{dR}{dz} \cong \frac{-EM}{4\rho_L} \tag{3.90}$$

3.10 Liquid water content (LWC)

The LWC is the measure of the mass of the water in a cloud in a specified amount of dry air. It is typically measured per volume of air ($\frac{g}{m^3}$) or mass of air ($\frac{g}{kg}$). This variable is important in figuring out which types of clouds are likely to form and is strongly linked to three other cloud microphysical variables: the cloud drop effective radius, the cloud drop number concentration, and the cloud drop size distribution. Being able to determine the cloud formations that are likely to occur is extremely useful for weather forecasting as cu mulonimbus clouds are related to thunderstorms and heavy rain.

The liquid water content of a cloud varies significantly depending on the type of clouds present in the atmosphere at a given location. The classification of the cloud is highly related to the liquid water content as well as the origin of the cloud. The combination of these two allows a forecaster to more readily predict the types of conditions that will be in an area based on the types of clouds that are forming or have already formed. Both the precipitation formation and the cooling rates of clouds depend on: 1). The LWC determines the ultimate potential for a cloud to produce precipitation. Generally speaking, unless a cloud generates a liquid-water content in excess of 0.5 ($\frac{g}{m^3}$) , it is unlikely to precipitate.
2). The LWC is important because it determines the rates of shortwave or long wave gradational heating and cooling. How does it calculate? Clouds that have low densities, such as cirrus clouds, contain very little water, thus resulting in relatively low liquid water content values of around 0.03 ($\frac{g}{m^3}$) .

Clouds that have high densities, like cumulonimbus clouds, have much higher liquid water content values that are around 1-3($\frac{g}{kg}$) . Below is a chart giving typical LWC values of various cloud types (Thompson,2007). Various equations are useful in determining LWC and the effects that influence it. One of the most significant variables related to the LWC is the droplet concentration of a cloud, which is the number of water droplets in a volume of cloud,

typically a cubic centimeter,and can be calculated by:

$$LWC = m_w^* \frac{n}{N}$$

where n the droplet concentration in a cloud, N is the total number of water droplets in the volume and m_w^* is the mass of the water in the air parcel.

3.11 Other parameters of macro scale cloud

3.11.1 Cloud temperature

:

The cloud-base temperature indicates the liquid water-producing potential of the cloud. For example, a cloud with a base temperature of $+20^0$C has a cloud-base saturation mixing ratio of -15 ($\frac{g}{Kg}$) , while a cloud with a base temperature of $+40^0$C has a cloud-base saturation mixing ratio of only 5 ($\frac{g}{Kg}$) . Thus, if these two clouds have equal depths, the one with the warmer cloud base has a much greater potential for producing rainfall. Cloud-top temperature is important because the greater the difference between the cloud base temperature and cloud-top temperature, the greater the potential for rainfall.

Further more, if the cloud-top temperature is below 0^0C, then ice is possible. Turbulence is important because it mixes properties of the cloud and interacts closely with the other parameters. The level of turbulence also affects the precipitation processes, due to the formation of higher peak super saturations and to increased interactions among cloud particles of different types and sizes. Turbulence is also likely to affect the radiative properties of a cloud. In a turbulent cloud the fluctuations in liquid water will exist. As a consequence, the radiative emittance and absorptance will differ significantly from that found in a more homogeneous cloud.

3.12 Cloud drop size distribution

The number of droplets per unit volume is called the number density, and is denoted by N. The units of N are m^{-3}. The number density of drops having diameters in the range between D_1 and D_2 is given by the integral:

$$N_{D_{1:D_2}} = \int_{D_1}^{D_2} n_d(D) dD$$

The function n_d(D) is called the drop-size distribution function. The units of nd are m^{-4}. The differential of number density is:

$$dN = n_d(D)dD \tag{3.91}$$

$dN = n_d(D)dD$. This Eq is interpreted as the number density of droplets in the diameter range range between D and D+dD. The probability of randomly selecting a droplet that has a diameter between D_1 and D_2.

$$p(_{D_{1:D_2}}) = \frac{N_{D_{1:D_2}}}{N}\int_{D_1}^{D_2}[\frac{n_d(D)}{N}]dD = \int_{D_1}^{D_2} p_d(D)dD$$

so that nd/N can be thought of as the probability density function for the drop diameter.

$$p(D) = \frac{n_d}{N} \tag{3.92}$$

From probability theory we know that the mean or expected value of a variable x can be found from its probability density function, p(x) via.

$$\overline{x} = \int_0^\infty xp(x)d_x \tag{3.93}$$

Using Eq .(2)and(3),that the mean diameter of the droplets in a distribution is:

$$\overline{D} = N^{-1}\int_0^\infty Dn_d(D)dD$$

Liquid water content in terms of drop-size distribution The mass of a droplet of diameter D is:

$$m = \frac{\pi}{6}\rho_L^{D3}$$

where L is the density of liquid water. The liquid water content of the droplets in this diameter range is just the volume of a drop of diameter D multiplied by Eq. (1)

$$M_{D_{1.}D_2} = \frac{\pi}{6}\rho_L\int_{D_1}^{D_2} D^3n_d(D)dD \tag{3.94}$$

Integrating (8) between diameters D_1 and D_2 gives the liquid water content contributed by droplets in this size range;

$$M_{D_{1:}D_2} = \frac{\pi}{6}\rho_L\int_{D_1}^{D_2} D^3n_d(D)dD$$

The total liquid water content for all drop sizes is:

$$M = \frac{\pi}{6}\rho_L\int_0^\infty D^3n_d(D)dD$$

3.13 Actual drop size distributions

The drop size distributions of actual clouds can be quite complex, and varies from cloud to cloud. Some distributions have more than one peak, while others have a single peak. In general, maritime cumulus clouds have larger drops and broader distributions than do continental clouds. In many instances the drop-size distribution is represented very closely by a form of the gamma distribution, having the form:

$$n_d(D) = aD^2 exp(-bD)$$

where a and b are constant.

At positive (above-freezing) temperatures clouds can naturally be assumed to be droplet clouds. During the fall of snowflakes and crystalline particles, the melting layer usually does not lie more than several hundred meters below the zero isotherm. However, in such cases as the fall of hail, ice particles are also present in the atmosphere at high positive temperatures.

At negative (subfreezing) temperatures one can have droplet, mixed, or crystal clouds. The proportion of drops and crystals (according to mass or according to number of particles) in a mixed cloud generally varies in time and from place to place. Individual parts of a cloud may consist wholly of droplets or only of crystals, while droplets and crystals coexist in other regions.

However,experimental data on the proportions of droplets and crystals in clouds of various forms are practically nonexistent. The only definite thing is that the relative proportion of crystals increases as the temperature drops. At the same time, the frequencies of droplet (that is, consisting just of liquid droplets), crystal (consisting just of crystals), and mixed (containing both droplets and crystals) clouds have been studied in detail, on the basis of data from regular aircraft soundings carried out over several years. In all, observations of more than 41,500 clouds were analyzed.

It was established that the temperature is the main factor determining the probability that supercooled droplets will be present in a cloud. The concentration of ice nuclei, which determines the probability of the appearance of crystals in the cloud, increases sharply with a drop in temperature. The droplet size is the next most important factor: the smaller the droplets, the more likely it is that unfrozen droplets will be present in the cloud at

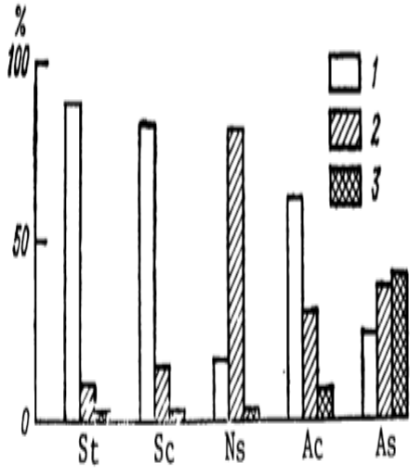

Figure 3.13: Frequencies of phases in clouds of various types at a negative temperature 1) purely droplet supercooled clouds, 2) mixed clouds, 3) crystal clouds [2].

the given temperature. Finally, important roles are also played by the seeding of lower-lying layers with crystals falling from higher layers at a lower temperature, as well as by the secondary multiplication of crystals. The dependence of the phase on the cloud form is actually determined by the temperature regime of the clouds. Figure 5.1 clearly illustrates the frequency of the phases in clouds of different forms in the middle latitudes. In these latitudes high-level clouds are practically always crystal clouds.

3.14 Kinds Of Clouds

In order to recognize and identify clouds it is necessary to classify and name them. Clouds are identified by their development, content and appearance, and their altitude. They are classified into many types and subtypes, but we need be concerned only with the more basic types. We will consider four families of clouds distinguished by their height of occurrence: High clouds, middle clouds, low clouds, and clouds with vertical development. Within the first three families are two main subdivisions:

1. Clouds formed by localized vertical currents which carry moist air upward beyond the condensation level. These are known as cumuliform clouds and have a billowy or heaped-up appearance.

2. Clouds formed by the lifting of entire layers of air, without strong, local vertical currents, until condensation is reached. These clouds are spread out in layers or sheets and are called stratiform.

In addition, the word nimbus is used as a prefix or suffix to indicate clouds producing precipitationresulting in such names as nimbostratus or cumulonimbus. The word fractus is used to identify clouds broken into fragments by strong windssuch as stratus fractus and cumulus fractus. Air stability has an important effect on the type of cloud formation. A stable layer which remains stable through forced lifting will develop stratiform clouds. Cumuliform clouds develop in air that is initially unstable or becomesunstable when it is lifted. A layer of conditionally unstable air which is forced to ascend may first develop stratiform clouds and then develop cumuliform clouds as the layer becomes unstable.

The cumuliform clouds project upward from a stratiform cloud layer. Thus, the type of cloud formation can be used as an indicator of the stability of the atmospheric layer in which the clouds are formed.

3.14.1 High Clouds

High clouds have bases ranging from 16,500 to 45,000 feet. Cirrus, cirrocumulus, and cirrostratus clouds are included in this family. They are usually composed entirely of ice crystals, and this is their most distinguishing characteristic. Cirrus are isolated wisps of thin, feathery cloud up near the top of the troposphere. They are sometimes called mares tails and may have trailing streams of larger ice crystals beneath them.Cirrus clouds are thin, white, feathery clouds in patches or narrow bands. They are composed of ice crystals of varying size. Larger crystals often trail down vertically and have given rise to the name mares tails". Cirrocumulus clouds consist of patches of small, white cloud elements. They may form definite patterns at times, showing small but firm waves or ripples. They are sometimes referred to as mackerel sky. True cirrocumulus are rare and are associated with other forms of typical cirrus at the same level, often changing into other forms of cirrus in a short time. Cirrostratus clouds are thin, whitish veils, sometimes covering the entire sky. Halos around the sun or moon, caused by their ice-crystal composition, frequently identify this cloud type. Cirrocumulus clouds contain small, white individual puffs. They may contain some supercooled water droplets mixed in with the ice crystals. Cirrocumulus is rare and is sometimes called mackerel

Figure 3.14: Cirrus clouds [1].

sky. Cirrus-type clouds may be produced in several ways. Often they are the forerunner of warm-front activity and give advance warning. Sometimes they are associated with the jet stream and usually are found on the south side of the jet. They may also be produced from the anvil tops of thunderstorms. The value of cirrus clouds in fire weather is their advance warning of warm-front activity and their use in indicating high-altitude moisture and wind direction and speed.Cirrostratus is a thin, whitish, transparent cloud layer appearing as a sheet or veil, It generally produces a halo around the sun or moon.

Figure 3.15: Cirrocumulus clouds [1].

3.14.2 Middle Clouds

Middle clouds have bases ranging from 6,500 feet up to 20,000 feet. Altocumulus and altostratus clouds fall into this group. Middle clouds are most generally formed by either frontal or orographic lifting, but may be formed in other ways. Often they are associated with lifting by convergence in upper-air troughs and sometimes develop with thunderstorms.

Altocumulus are white or gray patches, with each individual component having a rounded appearance. Altocumulus may appear as irregular cloudlets or in definite patterns such as bands or rows parallel or at right angles to the wind. Usually, the stronger the wind, the more distinct the pattern. Altocumulus are usually composed of water droplets and often are supercooled. Altostratus appears as a gray or bluish layer or veil with a sort of fibrous texture. It may be made up of supercooled water droplets or a mixture of water droplets and ice crystals. It tends to cover the entire sky, and the sun will shine through dimly as through a frosted glass.Altocumulus are white or gray patches or rolls of solid cloud. They are distinguished from cirrocumulus by the larger size of the cloud elements. Altocumulus clouds are usually composed of water droplets, often supercooled, but may contain some ice cry. As altostratus becomes thicker and lower, the sun becomes obscured. If it becomes dense and low enough, it becomes nimbostratus and takes on a wet and rainy appearance

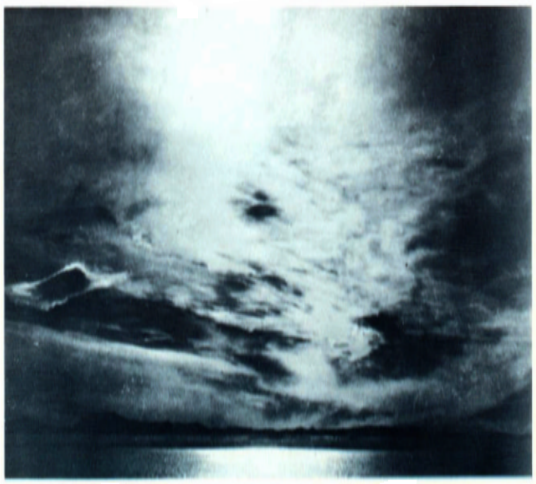

Figure 3.16: Cirrostratus cloud [1].

due to widespread precipitation. If the precipitation evaporates before reaching the ground, it is called virga. Three special types of middle clouds are of considerable importance in identifying weather conditions. The lens-shaped lenticular cloud appears over the ridge and to the lee of mountain ranges. Lenticular clouds indicate waves in the air flow caused by strong winds blowing across the range.

The clouds form in the rising current on the upwind side of the wave crest and dissipate in the downward flow on the other side.The castellans type of cloud consists of cumuliform masses in the form of turrets, usually arranged in lines. They indicate marked instability at high levels, and their occurrence in the forenoon is a warning of possible thunderstorm activity in the afternoon.

Clouds with rounded lower surfaces in the form of pouches or udders are called mamma. They are most common on the underside of cumulonimbus anvils. The pendulous blobs of cloud are sinking into the clear air immediately below because they contain droplets and ice crystals. As the droplets and crystals evaporate, they chill the air in the pendants. By this chilling, they keep the air denser than the surrounding clear air. The process ends when all cloud particles have evaporated. Pilots have reported that the down drafts within mamma are quite weak. However, any adjacent thunderstorms will have zones of marked turbulence.

Altostratus appears as a gray or bluish layer having a fibrous appearance, often associated

Figure 3.17: Altocumulus clouds [1].

with altocumulus. Frequently, it is composed of a mixture of supercooled water droplets and ice crystals. Light rain or snow often falls from it and Nimbostratus is a gray or dark massive cloud layer diffused by more-or-less continuous rain or snow which usually reaches the ground. It is thick enough to blot out the sun. Lower ragged clouds often appear beneath the nimbostratus laye.

Figure 3.18: Altostratus [1].

3.14.3 Low Clouds

The bases of low clouds range from the surface to 6,500 feet. Low clouds include stratus, stratocumulus, and nimbostratus. Nimbostratus clouds form a gray, often dark, layer usually

Figure 3.19: Nimbostratus cloud [1].

accompanied by continuously falling rain or snow. The precipitation usually reaches the ground, but occasionally only virga appears. Nimbostratus usually develops from thickening and lowering alto-stratus.

Stratus fractus or scud clouds often appear beneath the nimbostratus layer. Stratus and stratocumulus are very common and widespread. They usually occur beneath an inversion and are fairly thin, ranging from a few hundred to a few thousand feet thick.

Stratus forms alow, uniform sheet, dull gray in appearance. It is composed of water droplets and does not produce rain, although it may produce drizzle. Fog i s simply a stratus cloud lying on the surface. When a fog layer lifts, as it frequently does during the forenoon, it becomes a stratus layer. In some localities, particularly the west coast, stratus is referred to as high fog. Fog is important in fire weather because of its effect on the moisture content of forest fuels. While fog is forming or persisting, conditions are favorable for fuels to absorb moisture. Fog occurs during calm or light-wind conditions in a stable atmosphere and is formed in several ways. Radiation fog is formed when moist air cools to its dew point at night over a strongly radiating surface.

Some vertical mixing is necessary to produce a layer of fog of significant thickness. Advection fog forms when warm, moist air passes over a cool surface and its temperature is reduced to the dew point. Many fogs are a combination of these two types. Up slope fog forms when moist, stable air is forced to rise along a sloping land surface. This type occurs especially along the western edge of the Great Plains when mT air moves northwestward from the Gulf

of Mexico. Fog may also occur in connection with fronts, particularly in advance of warm fronts where evaporating rain falling through a layer of cold air near the surface saturates the cold air.

The distinction between stratus and stratocumulus is not particularly important. Stratocumulus shows individual rolls or rounded masses, usually soft and gray. It forms when the air is somewhat unstable, whereas stratus forms in stable air. Like stratus, it is composed of small water droplets and may produce light drizzle. Clouds with vertical development include cumulus and cumulonimbus. These are irregularly shaped masses with domes or turrets and have a cauliflower appearance. They usually appear in groups, and individual cloud bases are at about the same altitude. The height of the bases, which is the condensation level depends upon the air temperature and the amount of moisture in the atmosphere.

Cumulus clouds are formed near the top of rising convection columns, and their bases may range in height from a few thousand feet to 15,000 feet or more. Their presence is of special interest in fire weather as an alert to possible convection in the surface layer. They are a common type during the fire season, particularly in mountainous regions. The most common type of cumulus is a small, puffy type occurring during fair weather, called cumulus humili or fair weather cumulus. They appear after local surface heating becomes sufficiently intense to support convection, and dissipate in the late afternoon as surface heating decreases and convection ceases. These clouds have relatively flat bases, rounded or cone-shaped tops, and are usually isolated or in small groups. Their vertical growth is usually restricted by a temperature inversion which makes the tops fairly uniform. Occasionally a single cloud element will develop vertically to some height. True fair weather cumulus clouds, however, remain flat, but their presence indicates local updrafts that may influence fire behavior.

The danger from cumulus clouds is more acute, however, if the air is sufficiently moist and unstable to support their growth into towering cumulus. Virga or rain sometimes falls from the base of large cumulus. The final stage of cumulus development is the cumulonimbus or thunderhead, which is characterized by a flat anvil like formation at the top. The stretched-out shape of the anvil indicates the direction of air motion at that level. The anvil top is composed of sheets or veils of ice crystals of fibrous appearance which are sometimes blown off to form cirrus-type clouds.

Dissipating anvils give the appearance of dense cirrus and are sometimes referred to as

false cirrus. The greater the vertical development of cumulonimbus, the more severe the thunderstorm. Tops of cumulonimbus may extend to altitudes of 60,000 feet or higher and often reach the tropopause. Rain or snow showers usually accompany cumulonimbus clouds, and thunder, lightning, and hail are common. Cumulus and cumulonimbus clouds not associated with frontal or orographic lifting indicate strong surface heating and atmospheric instability from the surface up through the level of the cloud tops. Surface winds are likely to be gusty and increase in speed as the cumulus forms.

Other convection phenomena such as dust devils, whirlwinds, and considerable turbulence may be present. In addition to lightning, strong cold down drafts present a threat from well developed thunderheads. Because of the importance of thunderstorms in fire weather, we will discuss them in detail in the following chapter. Cumulus cloud caps often form atop the convection columns over large forest fires. Their moisture source may be almost entirely water vapor from the combustion process, or it may be water vapor entrained with air through which the column rises. Such clouds occasionally produce showers, but this is quite rare.Stratus is a low, gray cloud layer with a fairly uniform base and top. Usually it does not produce precipitation, although it may cause some drizzle or snow grains. Stratus often forms by the lifting of a layer of fog.

Figure 3.20: Stratus [1].

Stratocumulus clouds consist of gray or bluish patches or layers with individual tolls or rounded masses. They are generally composed of small water droplets and may produce light drizzle.

Cumulus clouds are detached clouds in the form of rising mounds or domes. They are dense, have sharp outlines, and the upper portion often resembles a cauliflower. They are composed of a great density of small water droplets; ice crystals may appear in the tops of

Figure 3.21: Stratocumulus clouds [1].

larger cumulus.

Figure 3.22: Cumulus clouds [1].

Cumulonimbus clouds are heavy and dense with considerable vertical development sometimes reaching the tropopause. The top often takes on the shape of an anvil. Cumulonimbus, often abbreviated to cb, is frequently accompanied by lightning and thunder, rain, sometimes hail, and on occasion a tornado or water spout.

Figure 3.23: Cumulonimbus clouds [1].

Chapter 4

Atmospheric Precipitation

Precipitation is any liquid or frozen water that forms in the atmosphere and falls back to the Earth. As shown in the figure below.

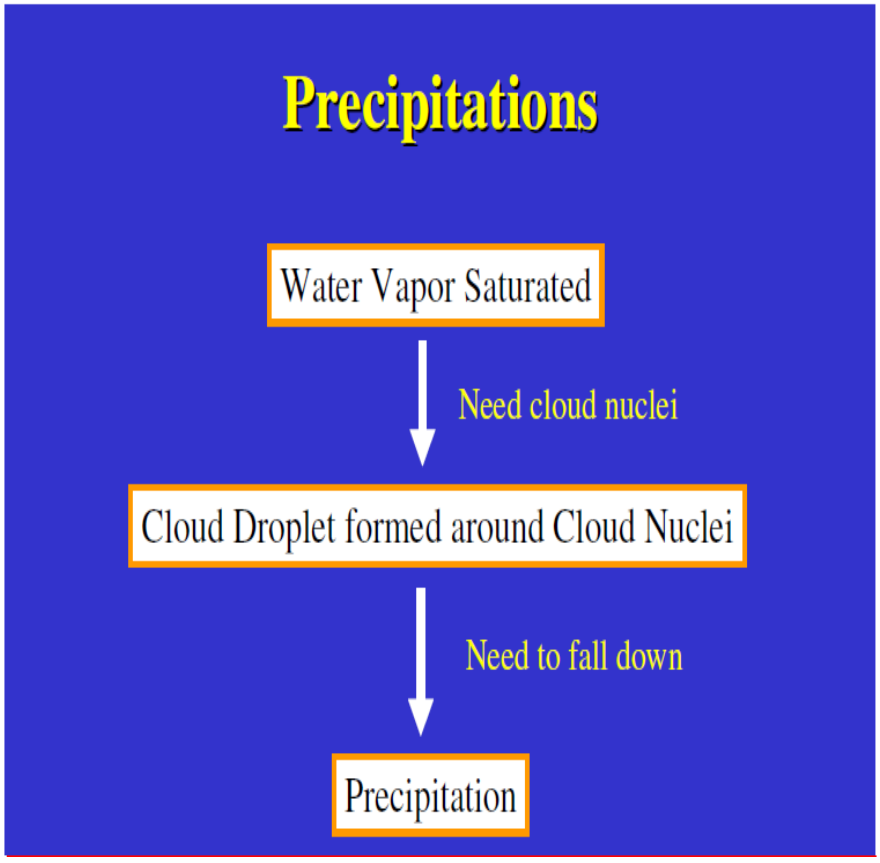

Figure 4.1: Precipitation process.

It comes in many forms like; rain, sleet, and snow. Along with evaporation and condensation, precipitation is one of the three major parts of the global water cycle. It is any product

of the condensation of atmospheric water vapor that falls under gravity. The main forms of precipitation include drizzle, rain, sleet, snow, graupel and hail.

Precipitation occurs when a portion of the atmosphere becomes saturated with water vapor, so that the water condenses and precipitates. Thus, fog and mist are not precipitation but suspensions, because the water vapor does not condense sufficiently to precipitate.

Two processes, possibly acting together, can lead to air becoming saturated: cooling the air or adding water vapor to the air. Precipitation forms as smaller droplets coalesce via collision with other rain drops or ice crystals within a cloud. Short, intense periods of rain in scattered locations are called showers. Precipitation is a major component of the water cycle, and is responsible for depositing most of the fresh water on the planet Mechanisms of producing precipitation include convective, strati form, and orographic rainfall.

Convective processes involve strong vertical motions that can cause the overturning of the atmosphere in that location within an hour and cause heavy precipitation, while strati form processes involve weaker upward motions and less intense precipitation. Precipitation can be divided into three categories, based on whether it falls as liquid water, liquid water that freezes on contact with the surface, or ice.

Mixtures of different types of precipitation, including types in different categories, can fall simultaneously. Liquid forms of precipitation include rain and drizzle. Rain or drizzle that freezes on contact within a subfreezing air mass is called freezing rain or freezing drizzle.

Frozen forms of precipitation include snow, ice needles, ice pellets, hail, and graupel. Some drops need to grow to precipitable size mechanisms: water va pour condensation droplet coalescence droplet coalescence ice processes.[6]

4.1 Kinds of Precipitation:

Precipitation products can be divided into three basic classes depending on their physical characteristics when they strike the earth: Liquid, freezing, and frozen. Rain and drizzle are the two kinds of liquid precipitation.

4.2 Rain

Rain is precipitation that falls to the surface of the Earth as water droplets. Raindrops form around microscopic cloud condensation nuclei, such as a particle of dust or a molecule of

pollution. Rain that falls from clouds but freezes before it reaches the ground is called sleet or ice pellets. Even though cartoon pictures of raindrops look like tears, real raindrops are actually spherical.Rain droplets have to have large enough falling speed in order to overcome the updraft (that produces the rain) to fall to the ground. This means the rain droplets have to GROW to large enough sizes to become precipitation.

How Rain drop Grows?

4.2.1 Growth by Condensation

Condensation about condensation nuclei initially forms most cloud drops. Only a valid form of growth until the drop achieves a radius of about 20 μm due to overall low amounts of water vapor available. Insufficient process to generate precipitation.

4.2.2 Growth in Warm Clouds

Figure 4.2: Growth in Warm Clouds.

Most clouds formed in the Tropics, and many in the middle latitudes, are warm clouds. Those clouds have temperatures greater than 0^0C throughout. The Collision-coalescence

process generates precipitation. This process depends on the differing fall speeds of different-sized droplets. It begins with large collector drops which have high terminal velocities.

4.2.3 Collision

Collector drops collide with smaller drops. Due to compressed air beneath falling drop, there is an inverse relationship between collector drop size and collision efficiency. Collisions typically occur between a collector and fairly large cloud drops. Smaller drops are pushed aside. Collision is more effective for the droplets that are not very much smaller than the collect droplet.

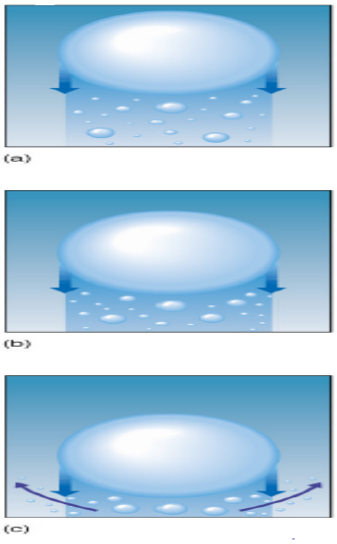

Figure 4.3: Collector drops collide with smaller drops [11].

4.2.4 Coalescence

When collisions occur, drops either bounce apart or coalesce into one larger drop. Coalescence efficiency is very high indicating that most collisions result in coalescence. Collision and coalescence together form the primary mechanism for precipitation in the tropics, where warm clouds dominate.

4.2.5 Cool and Cold Clouds

A portion of most mid-latitude clouds have temperatures below the melting point of ice. Cold clouds are referred to those have temperature below 0^0C throughout and consist entirely of ice crystals, supercooled droplets, or a mixture of two. Cool clouds are referred to those have temperatures above 0^0C in the lower reaches and subfreezing condition above. Cumulonimbus clouds contain both ice (top, fuzzy cloud margins), liquid drops (bottom, sharp margins) and a mix of ice and liquid (middle).

Figure 4.4: Cumulonimbus clouds [11].

4.2.6 Growth in Cool and Cold Clouds

Cool month mid-latitude and high latitude clouds are classified as cool clouds as average temperatures are usually below freezing. Clouds may be composed of:

(1) Liquid water,

(2) Supercooled water, and/or

(3) Ice.

Coexistence of ice and supercooled water is critical to the creation of cool cloud precipitation -the Bergeron Process.

Saturation vapor pressure of ice is less than that of supercooled water and water vapor.

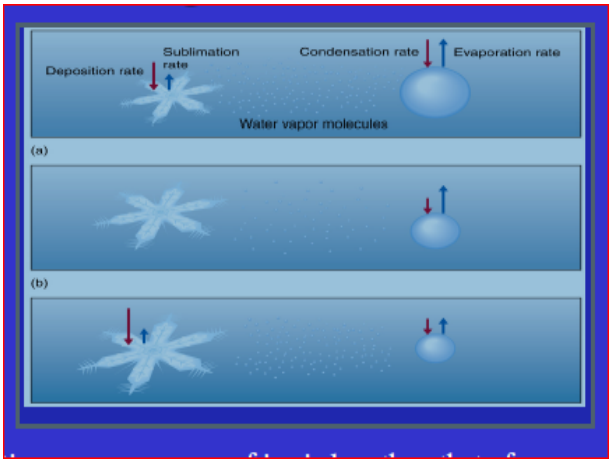

Figure 4.5: The Bergeron Process [11].

During coexistence, water will sublimate directly onto ice. Ice crystals grow rapidly at the expense of supercooled drops. Collisions between falling crystals and drops causes growth through riming and aggregation.

4.3 Hail

Hail forms in cold storm clouds. It forms when very cold water droplets freeze, or turn solid, as soon as they touch things like dust or dirt. The storm blows the hailstones into the upper part of the cloud. More frozen water droplets are added to the hailstone before it falls. Unlike sleet, which is liquid when it forms and freezes as it falls to Earth, hail falls as a stone of solid ice. Hailstones are usually the size of small rocks, but they can get as large as 15 centimeters (6 inches) across and weigh more than a pound. It also formed as graupel is carried aloft in updrafts.

At high altitudes, water accreting to graupel freezes, forming a layer. Hail falls but is eventually carried aloft again by an updraft where the process repeats. The ultimate size of the hailstone is determined by the intensity of the updraft [11].

Reattains = highest frequency of hail events.

Figure 4.6: Concentric layers of ice in hail indicate the cyclical hailstone formation process [11].

4.4 Snow

Snow is precipitation that falls in the form of ice crystals. Hail is also ice, but hailstones are just collections of frozen water droplets. Snow has a complex structure. The ice crystals are formed individually in clouds, but when they fall, they stick together in clusters of snowflakes. Snowfall happens when many individual snowflakes fall from the clouds. Unlike a hail storm, snowfall is usually calm. Hailstones are hard, while snowflakes are soft. Snowflakes develop different patterns, depending on the temperature and humidity of the air.

When snow falls in the form of a ball instead of soft flakes, it is called graupel. This happens when snow is melted and precipitation forms around the snow crystal. Snow requires temperatures at the ground to be near or below freezing less than 0 degrees Celsius (32-degrees Fahrenheit).

Snow that falls on warmer ground melts on contact The difference is mainly one of size and quantity of droplets. Drizzle droplets range in size from about 1/500 to 1/50 inch. Drizzle is formed in, and falls from, stratus clouds, and is frequently accompanied by fog and low visibility. Raindrops range in size from about 1/100 to 1/4 inch. They are much more sparse than drizzle droplets. Rain may come from liquid droplets formed by the coalescence

process in warm clouds, or from melted snowflakes originally formed in cold clouds by both the ice crystal and coalescence processes. The snowflakes melt when they reach air with above-freezing temperatures.

Rainfall intensity may vary from a few drops per hour to several inches in a matter of minutes. Heavier rainfall usually consists of larger drops. Freezing rain and freezing drizzle are formed and fall as liquid drops that freeze on striking the ground. The drops may be above-freezing, but usually they are supercooled and freeze upon striking the ground or other cold objects. This occurs usually with warm-front rain formed in the warm air above the frontal surface, and then supercooled as it falls through the cold air beneath the front.

The temperature at the ground must be lower than 32F frozen precipitation consists of snow, snow pellets, sleet and hail. Snow consists of crystals of ice formed in pure ice clouds or in mixed clouds.

The larger snowflakes are built up by the coalescence process. Air beneath the cloud must be near or below freezing, or the snow will melt before reaching the ground. The heaviest snowfalls occur when the temperature of the cloud portion from which the snow is falling is not much below freezing.

Snow pellets are white opaque grains of ice, usually round. They form when ice crystals coalesce with supercooled droplets, and usually occur in showers before or with snow. They range in size from 1/16 to 1/4 inch.

Sleet consists of transparent hard pellets of ice, about the size of raindrops, that bounce on striking the ground. They are formed by freezing of raindrops or by refreezing of partly melted snowflakes falling through a below-freezing layer of air. Sleet occurs most commonly with warm fronts.

Hail consists of balls of ice ranging in size from 1/5 inch to several inches in diameter. They have layer like structures indicating that they have grown by successive steps. Hailstones apparently begin their growth when supercooled water droplets impinge on ice pellets. The liquid water freezes on the ice pellet to form a layer of ice. This process is repeated until the hailstone falls out of the cloud.

The repetition may be due to the hailstone being caught in strong updrafts and carried upward into the region of supercooled droplets. It is also possible for the process to begin

at very high altitudes, in which case the hailstone grows as it falls through successive concentrations of supercooled water. Hail is associated with thunderstorms and very unstable air. There are two other forms in which moisture from the atmosphere is deposited on the ground. These are dew and frost.

Dew and frost do not fall,but instead are deposited when water vapor condenses or sublimes on the ground or on objects near the ground. Dew forms when air next to the ground or to cold objects is chilled to the dew point of the air, but remains above freezing. A common example is the deposit of water that forms on a glass of ice water. Frost forms by sublimation when the air is chilled to its dew point and the dew point is below freezing.

Dew and frost forming on forest fuels at night can add considerably to the fuel moisture. Radiation - ground fog. Sleet is formed by the freezing of liquid raindrops or the refreezing of partly melted snowflakes as they fall through a below freezing layer of air. Top−Rain arriving at the ground may begin as liquid drops, formed by the coalescence process in warm clouds, which then fall to the ground through warm air.

Bottom−Snow consists of crystals of ice formed in pure ice clouds or in mixed clouds, built up by the coalescence process and falling to the ground through cold air so that the flakes do not melt. Bottom−rain may begin as snowflakes formed by the ice crystal and coalescence processes in cold clouds. The snowflakes subsequently melt as they fall through warm air.

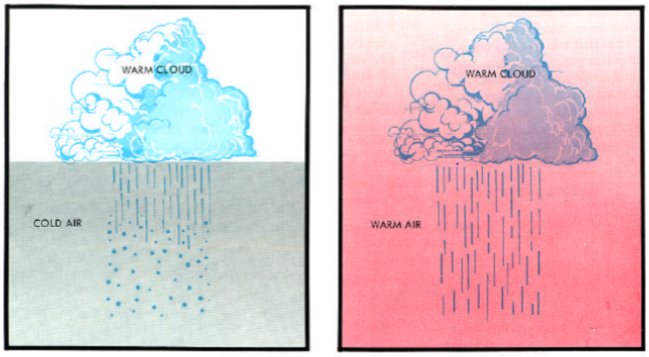

Figure 4.7: Top Sleet [1].

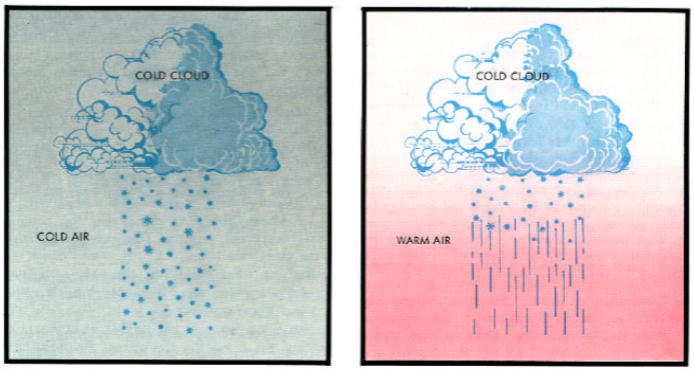

Figure 4.8: Bottom Snow[1].

4.5 Condensation, Sublimation, And Precipitation Processes

We were concerned with the change of state of water from gaseous to liquid or solid forms, and we used for simple examples the impaction of water vapor molecules on a liquid or solid surface. Dew and frost do form that way, but cloud particles are formed in the free atmosphere, and here the process becomes more complicated. Still more complex are the processes of precipitation where cloud particles must grow to a large enough size to fall out by gravity. We are all familiar with condensation and sublimation. We have noticed the condensation of our breath on cold days, and of steam rising from boiling water. We have seen dew formed on grass at night, or on cold water pipes and cold glasses, and have noticed the sublimation of water vapor into frost on cold window panes in winter. For condensation or sublimation to occur in the free air, a particle or nucleus must be present for water-vapor molecules to cling to. These fine particles are of two types: condensation nuclei and sublimation nuclei.

Condensation nuclei, on which liquid cloud droplets form, consist of salt particles, droplets of sulfuric acid, and combustion products. They are usually abundant in the atmosphere so that cloud droplets form when saturation is reached. Sublimation nuclei, on which ice crystals form, consist of dust, volcanic ash, and other crystallline materials. Because of differences in composition and structure, different nuclei are effective at different below-freezing temperatures. As the temperature decreases, additional nuclei become active in the sublimation

process. These The nuclei are not as plentiful as condensation nuclei. Even at temperatures well below freezing, there frequently are too few effective nuclei to initiate more than a scattering of ice crystals.

The small particles that act as condensation nuclei are usually hygroscopic; that is, they have a chemical affinity for water. They may absorb water well before the humidity reaches saturation, sometimes at humidities as low as 80 percent. Condensation forms first on the larger nuclei, and a haze develops which reduces visibility. As the relative humidity increases, these particles take on more water and grow in size while condensation also begins on smaller nuclei. Near saturation, the particles have become large enough to be classed as fog or cloud droplets, averaging 1/2500 inch in diameter, and dense enough so that the mass becomes visible. Rapid cooling of the air, such as in strong upward currents, can produce humidities of over 100 percentsupersaturationtemporarily. Under such conditions droplets grow rapidly, very small nuclei become active and start to grow, and many thousands of droplets per cubic inch will form. With supersaturation even nonhygroscopic particles will serve as condensation nuclei, but usually there are sufficient hygroscopic nuclei so that the others do not have a chance.

As condensation proceeds, droplets continue to grow until they reach a maximum size of about 1/100 inch in diameter, the size of small drizzle drops. The condensation process is unable to produce larger droplets for several reasons. As vapor is used up in droplet formation, supersaturation decreases and the cloud approaches an equilibrium state at saturation. Also, as droplets grow, the mass of water vapor changing to liquid becomes large and the resultant latent heat released in the condensation process warms the droplet and decreases the vapor pressure difference between it and the surrounding vapor. Thus the vast majority of clouds do not produce rain. If growth to raindrop size is to take place, one or more of the precipitation processes must come into play.

An important phenomenon in the physics of condensation and precipitation is that liquid cloud droplets form and persist at temperatures well below freezing. Although ice melts at 32^0F . Water can be cooled much below this before it changes to ice. Liquid cloud droplets can exist at temperatures as low as 40^0F . More commonly in the atmosphere though, cloud droplets remain liquid down to about 15^0F . Liquid droplets below 32^0F . are said to be supercooled. At temperatures above 32^0F . Clouds are composed only of liquid droplets.

At temperatures much below 15^0F. They are usually composed mostly of ice crystals, while at intermediate temperatures they may be made up of supercooled droplets, ice crystals, or both. Why dont ice crystals form more readily.First, the formation of ice crystals at temperatures higher than -40^0 F, requires sublimation nuclei. As was mentioned above, these usually are scarce in the atmosphere, especially at higher elevations. Also, many types of nuclei are effective only at temperatures considerably below freezing. But another reason why vapor condenses into liquid droplets, rather than sub-limes into ice crystals, is that condensation can begin at relative humidities well under 100 percent while sublimation requires at least saturation conditions and usually supersaturation. Given the necessary conditions of below freezing temperature, effective sublimation nuclei, and supersaturation, sublimation starts by direct transfer of water vapor to the solid phase on a sublimation nucleus. There is no haze phase as in the case of condensation.

Once sublimation starts, ice crystals will grow freely under conditions of supersaturation. Since there are fewer sublimation than condensation nuclei available, the ice crystals that form grow to a greater size than water droplets and can fall from the base of the cloud. Only very light snow, or rain if the crystals melt, can be produced by sublimation alone.

Moderate or heavy precipitation requires one of the precipitation processes in addition to sublimation. After condensation or sublimation processes have gone as far as they can, some additional process is necessary for droplets or crystals to grow to a size large enough to fall freely from the cloud and reach the ground as snow or rain.

Cloud droplets, because of their small size and consequent slight pull of gravity,have a negligible rate of fall, and for all practical purposes are suspended in the air. Even drizzle droplets seem to float in the air. Raindrops range in size from about 1/50 inch to 1/5 inch in diameter. Drops larger than 1/5 inch tend to break up when they fall. It takes about 30 million cloud droplets of average size to make one raindrop about 1/8 inch in diameter. There seem to be two processes which act together or separately to cause millions of cloud droplets to grow into a raindrop. One is the ice crystal process and the other is the coalescence process[6].

4.6 The Ice-Crystal Process

We have seen that ice crystals and cloud droplets can coexist in clouds with subfreezing temperatures. For the ice-crystal process of precipitation to take place, clouds must be composed of both ice crystals and super cooleci liquid cloud droplets.In chapter 3 we discussed vapor pressure and saturation vapor pressure at some length, but we considered only saturation vapor pressure with respect to liquid water. The saturation vapor pressure with respect to ice is somewhat less than that with respect to super-cooled water at the same temperature. In the ice-crystal precipitation process, ice crystals grow at the expense of water droplets. Because of the difference in vapor pressure over ice and over water at the same temperature, a vapor pressure gradient exists between supercooled water droplets and ice crystals in mixed clouds. Vapor molecules leave the water droplets and sublime on the ice crystals[].

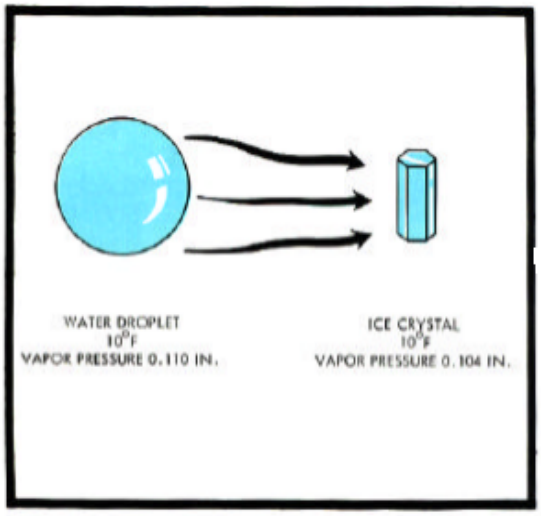

Figure 4.9: Ice-crystal precipitation process [1].

If a cloud containing supercooled water droplets is saturated with respect to water, then it is supersaturated with respect to ice, and the relative humidity with respect to ice is greater than 100 percent. The force resulting from the difference between vapor pressure over water and over ice causes vapor molecules to be attracted to ice crystals, and the ice crystals will grow rapidly. As the ice crystals gather up vapor molecules in the cloud, the relative humidity with respect to water drops below 100 percent, and liquid cloud droplets begin to evaporate.

Vapor molecules move to the ice crystals and crystallize there. Thus, the ice crystals grow at the expense of the water droplets and may attain a size large enough to fall out of the cloud as snowflakes. If the snowflakes reach warmer levels, they melt and become raindrops, This is the ice crystal precipitation process.

4.6.1 Artificial Nucleation

The knowledge that frequently there is a scarcity of sublimation nuclei and ice crystals in supercooled clouds has led to the discovery that precipitation can be initiated artificially. It has been found that silver-iodide crystals, which have a structure similar to ice crystals, can be effective sublimation nuclei in super-cooled clouds at temperatures below about 20F. Silver-iodide crystals can be released in the cloud by aircraft or rockets, or carried to the cloud by convection from ground generators.

Ice crystals can be created in a supercooled (loud by dropping pellets of dry ice, solid carbon dioxide, into the cloud from above. The dry ice, which has a melting temperature of -108F. Cools droplets along its path to temperature lower than 40F. so that they can freeze into ice crystals without the presence of sublimation nuclei. Once crystals are produced, they act as nucleating particles themselves and affect other parts of the cloud. Once formed in a supercooled water cloud, ice crystals may grow by the ice-crystal process and coalescence processes until they are large enough to precipitate. Studies have provided evidence that the artificial nucleation of super-cooled clouds can, under the proper conditions, increase local precipitation significantly.

4.7 Coalescence

In the coalescence process of precipitation, small droplets collide and fuse together to become larger droplets. The process continues until enough droplets are accumulated into large drops so that the large drops fall because of gravity. Snowflakes coalesce into snowflake masses in a similar manner as shown in figure below. Since rain also falls from clouds which are entirely above freezing, there must be a second precipitation process. This is a simple process in which cloud droplets collide and fuse together, or coalesce. Clouds which produce precipitation are composed of cloud droplets of varying sizes. Because of the different sizes, cloud droplets move about at different speeds. As they collide, some of them stick together to form larger

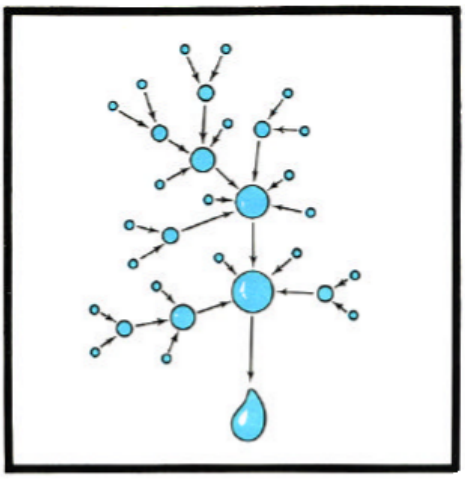

Figure 4.10: The coalescence process of precipitation [1].

drops.

The larger cloud droplets grow at the expense of smaller ones, and actually become more effective in the collecting process as they become larger. As larger drops begin to fall, they tend to sweep out the smaller drops ahead of them. The coalescence process takes place in clouds of both above-freezing and below-freezing temperatures. Snowflakes coalesce with other snowflakes as they fall to form the large clumps which we sometimes observe. They may also coalesce with supercooled water droplets to form snow pellets.

4.8 Measurement Of Precipitation

Precipitation is measured on the basis of the vertical depth of the water or melted snow. Snow, sleet, hail, and other solid forms are also measured on the basis of the depth of the unmelted form. Our common unit of measurement is the inch. The standard rain gage is an 8-inch cylindrical container with an 8 -inch funnel at the top and a measuring tube inside.

The cross-sectional area of the measuring tube is exactly one-tenth that of the funnel top. Thus, if 0.01 of an inch of precipitation falls, it is 0.1 inch deep in the measuring tube. The stick used to measure the precipitation is graduated in inches, tenths, and hundredths, so that 0.01 inch of rain is indicated for each 0.1 inch of stick length. When snow is measured, the

Figure 4.11: A rain gage [11].

funnel and measuring tube are removed, and only the outside cylindrical container is used. Snow caught in the gage is melted and measured in the measuring tube to obtain the liquid equivalent of the snow.

Several types of recording gages that make continuous records of the precipitation are also in use. The tipping bucket gage can be used only for rain. For each 0.01 inch of rain, an electrical impulse is recorded. Another type is the weighing type gage which can be used for either snow or rain. This device simply weighs the snow or rain that is collected. The weight is recorded continuously in inches of water on a chart attached to a revolving drum. The rain gage should be exposed in the open away from large buildings or trees. Low bushes, fences, and walls are not objectionable, provided that the gage is placed at a distance of at least twice the height of the object. The top of the gage should be level [11].

Chapter 5

Conclusion And Summary

In this project we have learned that air becomes saturated either by the addition of moisture, or, more commonly, by cooling to the dew point. In saturated air, clouds form by the condensation of water vapor, which takes place on fine particles called condensation or sublimation nuclei.

Cloud droplets grow to sizes large enough to precipitate by the ice-crystal process, in which water vapor is transferred from evaporating, supercooled liquid droplets to ice crystals where sublimation takes place, or by coalescence of droplets or ice crystals into rain- drops or clumps of snowflakes.

Precipitation falls in the form of liquid rain or drizzle, freezing rain or drizzle, or frozen snow, sleet, or hail. Clouds are classified according to their structure as stratus or cumulus, and according to their altitude as high, middle, low clouds, and those with large vertical development. In the last group are cumulonimbus or thunderstorm clouds.

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DECLARATION

I, hereby declare that this project is my original work and has not been presented for a degree in any other university, and that all sources of materials have been duly acknowledged.

Name: Nigatu Kuchi

Signature:————————- ————————————.

e-mail:-kuchinigatu@gmail.com

Place and time of submission:

Addis Ababa University

Department of Physics

April 11, 2022

This project has been submitted for examination with my approval as University advisor.

Name: Dr. Yitagesu Elfagd

Signature:————————————————.

e-mail:-yitagesuw@gmail.com