

Lecture Notes
on
Physical Meteorology
For
Integrated Meteorological Training Course

By

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Objectives: After reading this article, reader will be able to:

- Describe Composition of the Atmosphere
- Explain the vertical structure of the atmosphere
- write the Laws of thermodynamics
- Understand the thermodynamics of dry and moist air
- Write Hydrostatic equilibrium
- Explain the stability of atmosphere
- Write the desirable properties of thermodynamic diagram
- Describe the basics of cloud Physics
- Understand Electromagnetic Radiation

1. Composition of the Atmosphere

The atmosphere is the envelope of gases surrounding the earth and it is bound to earth by the force of gravitation. The atmosphere consists of a mixture of gases within which tiny liquid and solid particles called aerosols also exist. Among the various gases by which atmosphere is made up of, certain gases are permanent gases and certain gases are variable gases. Among the variable gases, the water vapour is the major one which occupies a maximum of about 4% of the volume of a sample of air. However, in spite of its very small concentration, this minor constituent of the atmosphere has a very important role in the atmospheric processes.

Let us discuss the composition of pure dry air first. The major constituents of pure dry air are Nitrogen(78.09%), Oxygen(20.95%), Argon(0.93%) and Carbon dioxide(0.03%). Also the gases like Hydrogen, Helium, Neon, Krypton, Xenon, Ozone are also found in the atmosphere in smaller quantities. In addition to these, trace gases such as Carbon monoxide, Radon, Nitrous oxide, Methane etc. also occur in the atmosphere. Nitrogen and Oxygen make up about 99% of the air but they do not play any active role in the meteorological processes.

Carbon dioxide, the concentration of which is approximately 0.03 % by volume in the lower atmosphere, enters the atmosphere by processes like combustion, human and animal breathing, photosynthesis and volcanic activity. Its concentration varies little in general but in the local scale it shows variations. The concentration is higher in the cities as compared to country side and it is higher very near to the soil.

Ozone is another important constituent of the dry air. The total amount of ozone in a vertical column of the atmosphere is relatively small. If the total amount of ozone is brought down to mean sea level, it will form a layer of thickness of about 0.3 centimeters only. Measurements indicate that, the concentration of ozone increases with altitude and reaches a maximum at about 25 kilometers and then it starts decreasing falling to small amounts at about 50 kilometers. The concentration of ozone falls further and at about 70 kilometers altitude, the concentration is almost nil. Even though the concentration of ozone in the atmosphere is very little, this gas plays a significant role in the atmosphere. Solar radiation in the ultra violet (uv) region is strongly absorbed by ozone in the stratosphere hence the radiative balance and the thermal structure of the stratosphere is highly influenced by the distribution of ozone. In addition to that, absorption of solar radiation in the uv region effectively shields the earth's surface from that harmful radiation thus protects the human and animal life on the earth's surface.

In addition to the gases which constitute the dry air, water vapour is also present in the atmosphere in varying proportions. Its distribution is highly variable in time and space and its concentration decreases normally with altitude. Water vapour enters the atmosphere by evaporation from water bodies and transpiration from plants.

It exists in solid, liquid and vapour stages in the atmosphere and has the capacity to change from one phase to another. The change of phase of water vapour is involved with the absorption or release of latent heat hence it plays a major role in the atmospheric thermodynamic processes. In the radiative processes also, water vapour has a major role to play.

The three minor constituents mentioned above are the most important gases with respect to meteorology. Earth atmosphere system loses energy by terrestrial radiation but carbon dioxide, water vapour and the ozone are very good absorbers of terrestrial radiation and because of this absorption and the resultant emission, a major portion of the energy lost by the earth atmosphere system is returned back to the earth. Thus these minor constituents play a major role in the radiation balance of the earth atmospheric system and they are generally termed as the green house gases.

The composition of the dry air remains generally constant up to the height of about 80 kilometers which indicates the sufficient mixing which takes place in this region. The layer of the atmosphere upto the height of 80 kilometers is known as homosphere. Above 80 kilometers, oxygen begins to dissociate under the influence of ultraviolet radiation. At further higher levels, nitrogen also starts dissociating. However, the quantity of air above these levels is so less that these dissociations will not have much meteorological significances but this process have a significant influence on the propagation of radio waves.

The region above homosphere is known as hetrosphere. Here the gases separates out and exists in individual form with respect to their respective molecular weights. The heavier gases tend to concentrate in the lower levels with the lighter ones lying above. Thus this layer is characterised by decreasing molecular weight with height.

In addition to the gaseous constituents, aerosols; the solid particles and the liquid droplets suspended in the air are in enormous numbers and have their effective radius between 0.005μ and 20μ in the atmosphere. The aerosols can enter into the atmosphere by various processes like dispersion, combustion, photochemical reaction, evaporation of sea spray, industrial activities etc. Dust, smoke, salt particles and condensed water vapour are the major non gaseous materials present in the atmosphere. The concentration of aerosols vary with time and space.

Presence of aerosols in abundance will affect the visibility considerably. It also has some impact in the transmission of the radiation through the air. The aerosols, mainly from industrial origin, in abundance, pose severe health hazards. However, aerosols play a major role in cloud formation, as condensation nuclei or ice nuclei and are thus important meteorologically.

2. Vertical Structure of the Atmosphere

The atmosphere can be divided into different layers according to composition, ionization and the temperature distribution. However, the division with respect to the thermal stratification is the most commonly used one.

In 1950, Chapman divided the upper atmosphere into Troposphere, Stratosphere, mesosphere and Thermosphere. The stratosphere and the mesosphere was bounded at the top by the stratopause and the mesopause. He also suggested a subdivision of the mesosphere into mesocline and mesodecline separated by the region of maximum temperature in the mesosphere at about 50 kilometers called mesopeak.

Goody in 1954, considered the entire region between the tropopause and the temperature minimum near 80 kilometers as stratosphere. He suggested the region from the tropopause to about 32 kilometers as the lower stratosphere and the rest of the stratosphere as the upper stratosphere. The region above stratopause at 80 kilometers is referred to as ionosphere by him.

The system which is most commonly used now a days is associated with the name of Nicolet. It was recommended by International Union of Geology and Geophysics at its Helsinki meeting in 1960 and was approved by World Meteorological Organization's executive committee in 1962. In this system, the lower most layer is the troposphere in which temperature decreases with height and this layer is bounded at the top by an isothermal layer called the tropopause. The stratosphere extends from the tropopause to the temperature maximum near 50 kilometers (stratopause) and the mesosphere lies between the stratopause and the temperature minimum near 80 kilometers (mesopause). The layer above mesopause is referred to as thermosphere in which temperature increases with height.

The vertical divisions of atmosphere according to thermal stratification is given in Figure 1.

The rate of fall of temperature with height is known as lapse rate. Hence when the temperature is decreasing with height, lapse rate is positive and when the temperature is increasing with height, lapse rate is negative. Accordingly, the troposphere and the mesosphere have positive lapse rates and stratosphere and the thermosphere have negative lapse rates.

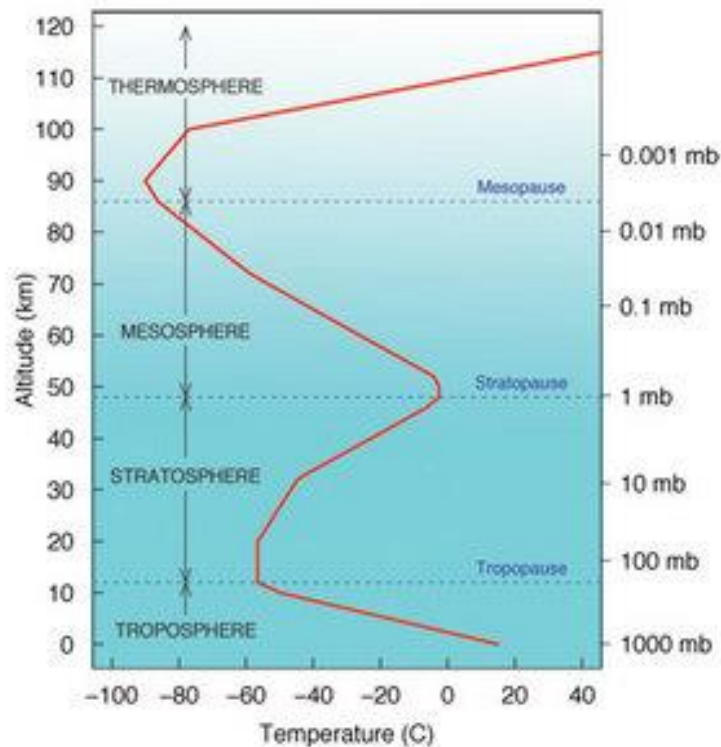


Figure 1 : Vertical Divisions of the Atmosphere

In Figure 1, the altitudes mentioned for termination of different layers is given on an average. But the tropopause in the equatorial region can be found at a height of about 18 kilometers whereas in the polar region, it can be observed at a height of about 8 kilometers itself. Hence lower limits of the stratosphere and thus that of the upper atmosphere depends very much upon the thermal properties of the lower atmosphere at a given time and place and not by definite altitude limits.

The division of the atmosphere into homosphere and the heterosphere according to composition is already discussed in the previous section. As suggested by Chapman(1950), the word homosphere can be used to refer to that part of the atmosphere upto about 80 kilometers with uniform composition and the word heterosphere can be used to refer to further higher regions with varying compositions.

As suggested by Chapman, when the criteria of ionization is referred to, the word neutrosphere can be used to designate the lower part of the atmosphere upto about 60 kilometers where ionization is unimportant and the region above can be referred to as ionosphere where the ionization and thus the concentration of free electrons becomes significant.

In addition to these, the outermost region of the atmosphere, above a few hundred kilometres, where the mean free paths of the molecules are so long that they can escape from the atmosphere is often referred to as exosphere.

3. Laws of thermodynamics

In thermodynamics, a specific sample of matter is referred to as a system and the equilibrium state of such a system is defined with respect to the properties like pressure(p), temperature(T) and volume(V) which are known as the basic variables. In thermodynamics usually we deal with specific volume(α) which is the volume per unit mass. Since density(ρ) is the ratio of mass to volume, specific volume is then the reciprocal of density.

The important laws of thermodynamics are Boyle's law and Charles' law.

3.1. Boyle's Law

Boyle's law states that if the temperature T is kept constant, the pressure p of a given mass of gas is inversely proportional to its volume V .

$$\text{i.e. } p \propto 1/V \quad \text{or} \quad pV = a \text{ constant.}$$

Hence, when the temperature of a given mass of gas is kept constant, its volume increases when pressure decreases and vice versa. This was discovered by Robert Boyle in 1660 experimentally.

3.2. Charles' Law

According to Charles' law, if the pressure p is kept constant, then the volume V of the given mass of gas is directly proportional to its absolute temperature.

Absolute temperature or the temperature in Kelvin = temperature in Celsius + 273.15

Hence according to Charles' law,

$$V \propto T \quad \text{or} \quad V/T = a \text{ constant}$$

3.3. Combination of Boyle's law and Charles' Law

According to Boyle's law, we have $pV = a$ Constant and according to Charles' law, we have $V/T = a$ constant.

Consider a given mass of gas whose pressure, volume and temperature are p_1 , V_1 and T_1 initially. Let the temperature of this sample be kept constant at T_1 and the pressure is increased from p_1 to p_2 . Applying Boyle's law, the volume then reduces to a value of $p_1 V_1 / p_2$. Now keeping the pressure as constant at p_2 , increase

the temperature to T_2 . Then, according to Charles' law, the new volume V_2 will be given by

$$V_2 = V_1 \cdot p_1/p_2 \cdot T_2/T_1.$$

Otherwise,

$$p_2 V_2/T_2 = p_1 V_1/T_1$$

Thus the combination of Boyle's law and Charles' law shows that $pV/T = a$ constant when T is expressed in Kelvin.

In a thermodynamic system, the unit for the mass of a substance is the mol.

One mol of a substance having molecular weight M has a mass of M kilogram (In SI system, we refer to kilogram mole).

Each mole of a gas contains the same number of molecules called Avagadro's number (6.0221415×10^{23}).

The volume occupied by one mole of gas is referred to as its molar volume. It changes with pressure and temperature. In case of standard conditions of pressure (1013.25 hPa) and temperature (273.15 K), the standard molar volume is about 22.414 litres. The standard molar volume is nearly the same for all the real gases.

We have seen that, $pV/T = a$ constant, combining Boyle's and Charles' law.

In this equation, if we put the standard values of pressure (1013.25 hPa), temperature (273.15K) and volume (22.414 litres), we get the value of the constant known as universal gas constant (R^*). It is approximately equal to 8314 joules/mol/K. Accordingly, we can have the ideal gas equation

$$pV = R^*T$$

However, real gases follow this ideal gas equation, only when their pressure is extremely small.

We have the ideal gas equation as

$$pV = R^*T.$$

Or

$$p\alpha M = R^*T \text{ since } \alpha = V/M, \text{ the specific volume.}$$

Thus $p\alpha = R^*/M \times T$ or $p\alpha = RT$ where $R = R^*/M$, the specific gas constant.

This is another form of ideal gas equation in terms of pressure p , α the specific volume, R the specific gas constant and T the temperature.

3.4. Ideal Gas Equation for a Mixture of Gases

For a mixture of gases, the total pressure exerted by the mixture will be equal to the sum of partial pressures exerted by the individual gases, if the mixture alone is occupying the entire volume at the same temperature of the mixture. This is known as **Dalton's law**.

If the mixture consist of n components, then according to Dalton's law, the pressure exerted by the mixture

$$p = p_1 + p_2 + p_3 + \dots + p_n = \sum p_k \text{ where } k \text{ varies from } 1 \text{ to } n$$

We have

$$p\alpha = R^*/M \times T$$

Or

$$p = R^*/M \times T/\alpha$$

For the n th component of the mixture,

$$P_n = R^*/M_n \times T/\alpha_n$$

Applying Dalton's Law and multiplying with α on both sides and rearranging, we have

$$p\alpha = R^*T \times \sum (m_n/M_n) / \sum m_n$$

Thus, $p\alpha = R^*T / \text{Mean } M$

Where, $1 / \text{Mean } M = \sum (m_n/M_n) / \sum m_n$

4. Thermodynamics of Dry Air

Thermodynamics deals with transformation of heat energy into other forms of energy and vice versa.

Energy is the capacity of any system to do work. Since energy is corresponding to the total work which can be done, it is also measured in Joules in SI units. Law of conservation of energy states that in any system, energy can not be created or destroyed, in the absence of nuclear reactions and for velocities which do not approach that of light.

4.1. Specific Heat

The heat energy required to raise the temperature of the system by one degree is called the heat capacity of the system. The heat capacity of the system per unit mass is called specific heat capacity or the specific heat. The amount of heat required to raise the temperature of the system per unit mass by one degree, when the volume occupied by the system is kept constant is called the specific heat at constant volume or C_v . The heat required to raise the temperature of the system per unit mass by one degree, when the pressure of the system is kept constant is called specific heat at constant pressure or C_p . The relation between C_p and C_v is given by the relation

$$C_p - C_v = R, \text{ the specific gas constant.}$$

4.2. First Law of Thermodynamics

If a sample of gas is expanding at constant pressure p , it is doing work on the environment. If dV is the infinitesimal change in the volume and dW is the work done during expansion of the sample, then $dW = pdV$

In case of unit mass, the volume V will be replaced by the specific volume α .

Then work done $dw = p d\alpha$

If the system expands and does work on the environment, the work done is positive. On the other hand, if the system is compressed and thus the work is done by the environment on the system, the work done negative.

According to the law of conservation of energy, when energy is added to any system, the final energy is equal to the initial energy plus the amount of energy added. Some part of the energy added will be utilized to increase the temperature and some part will be utilized to overcome the force of attraction between the

molecules. Both these together constitute the change in internal energy of the system which is represented by du for unit mass.

If the increment of the work done by the system is dw , then, the law of conservation of energy can be written as

$$dh = du + dw$$

Thus, If an increment of energy dh is added to a system, it will be established as the change in its internal energy du and the increment of the work done by the system dw .

This is the first law of thermodynamics. It can also be written as

$$dh = du + pda \quad \text{since } dw = pda$$

In an ideal gas, the intermolecular force of attraction is negligibly small hence the change in internal energy will be established as a change in temperature alone which is not applicable to real gases.

So in case of an ideal gas, if heat is added to a system at constant volume, then the change in internal energy $du = C_v dT$ where dT is the change in temperature and C_v is the specific heat at constant volume.

Thus the first law of thermodynamics can be written as

$$dh = C_v dT + pda$$

If the heat is added to a sample of ideal gas at constant volume, it will be established as the change in internal energy only. Then the first law of thermodynamics reduces to the form $dh=du$.

4.3. Adiabatic Processes

An adiabatic process is one in which no heat is added to or removed from the system. Hence $dh=0$ in such processes. For an adiabatic process, the first law of thermodynamics reduces to the form

$$du+dw = 0$$

or

$$du = - dw$$

From the above equation, it can be seen that adiabatic expansion leads to a decrease in internal energy leading to the cooling of the system whereas adiabatic compression leads to an increase in internal energy leading to heating of the system.

4.4. Poisson's Equation

Poisson equation gives the relation between pressure and temperature in an adiabatic process. It is given by

$$T = \text{a constant} \times p^{\kappa} \text{ where } \kappa = R/C_p$$

This equation can be used to find out the changes in temperature of the sample of gases when it is subjected to pressure changes. For example if the pressure of the sample is changed to 1000 hPa and if θ is the corresponding temperature, then according to Poisson equation,

$$T/p^{\kappa} = \theta/1000^{\kappa}$$

Or

$$\theta = T (1000/p)^{\kappa}$$

The temperature θ is referred to as potential temperature. It is the temperature that a parcel of air would have if it is brought to 1000 hPa adiabatically. Potential temperature is a property of the parcel which is invariant in the adiabatic process. Hence it is a conservative property.

4.5. Alternate Form of First Law of Thermodynamics

First law of thermodynamics is given by the equation

$$dh = du + dw$$

where dh is the energy added or removed from the system, du is the change in internal energy and dw is the work done.

Another form of the same law is given by

$$dh = C_v dT + p da$$

We have the equation of state as $pa = RT$

Differentiating, we get, $p da + a dp = R dT$

Substituting, $dh = (C_v + R) dT - p da$ or $dh = C_p dT - a dp$

since $C_p - C_v = R$, the specific gas constant.

This is another form of the first law of thermodynamics.

4.6. Entropy

We have $p\alpha = RT$ or $\alpha = RT/p$

Substituting, we get $dh = C_p dT - (RT/p) \times dp$

Dividing by temperature T , we get $dh/T = (C_p dT/T) - (R dp/p)$

From this we can define a quantity called the specific entropy (s) from the relation

$$ds = dh/T$$

dh is the amount of heat added to unit mass of gas at temperature T and ds is the change in entropy. Hence specific entropy increases with the absorption of heat by the system and specific entropy decreases with the removal of heat from the system. In case of an adiabatic process, no heat is added or removed from the system hence $dh=0$. Thus, $ds=dh/T=0$ hence there is no change in specific entropy or the specific entropy is a constant. Hence adiabatic processes are isentropic processes.

We have seen that the potential temperature remains a constant in an adiabatic process. This can be proved from the relation between specific entropy and the potential temperature.

We have $\theta = T \times (1000/p)^\kappa$ where $\kappa = R/C_p$

Taking natural logarithms on both sides,

$$\ln\theta = \ln T + R/C_p \ln 1000 - R/C_p \ln p$$

Taking differentials,

$$d(\ln\theta) = d(\ln T) + R/C_p d(\ln 1000) - R/C_p d(\ln p)$$

Multiplying by C_p ,

$$\begin{aligned} C_p d(\ln\theta) &= C_p d(\ln T) - R d(\ln p) \\ \text{i.e. } C_p d(\ln\theta) &= C_p (dT/T) - R(dp/p) \end{aligned}$$

But $C_p (dT/T) - R(dp/p)$ is dh/T or the change in specific entropy ds .

Thus the relation between potential temperature and the entropy is given by

$$ds = C_p d(\ln\theta)$$

We have seen that the adiabatic processes are isentropic processes hence $ds = 0$.

$$\text{Thus, } C_p d(\ln\theta) = 0$$

i.e. $d(\ln\theta) = 0$ or $\ln\theta = \text{a constant}$ or $\theta = \text{a constant}$.

Thus in a dry adiabatic process, potential temperature remains constant or it is conservative property.

5. Thermodynamics of Moist Air

In the previous chapter, we have discussed the thermodynamic properties of dry air. Now we will discuss the various properties of the moist air. Air consisting of dry air and water vapour is called the moist air.

Water occur in three phases in the atmosphere and when it changes from one phase to another, absorption or release of latent heat occur and this corresponds to one of the major energy transformations in the atmosphere. Hence knowledge about the thermodynamics of moist air is very much essential.

Water appears in the atmosphere in the solid, liquid and gaseous phases. The ideal gas law is applicable to the water vapour also whenever it is far from condensation and deposition. The behaviour of water vapour with respect to changes in pressure and temperature can be discussed first.

According to Boyle's law, the volume of a sample of ideal gas varies inversely with its pressure, if the temperature is kept constant. However, in case of condensation and deposition, there exists some change from this rule.

Suppose we have a sample of water vapour at temperature T . Keeping the temperature constant, if the pressure is increased, it gets compressed thus reducing its volume. If the pressure is increased further, a point will reach when with a slight increase in pressure, some of the vapour changes to liquid form. After this stage, even with a slight increment in pressure, the whole vapour will change into liquid form. Thus, during this process, the pressure remains almost a constant, while the entire vapour changes into liquid form. This constant pressure is termed as saturation vapour pressure at that temperature. At this value of pressure, there exists an equilibrium between liquid and gaseous forms of water. Once the vapour totally changes into liquid form, it is not possible to compress the same; hence the volume remains a constant while the pressure increases.

This entire process can be represented in terms of an isotherm for the temperature T on a graph of pressure plotted against specific volume. The isotherm will take the form of a rectangular hyperbola. The same process can be continued with higher values of temperature.

Accordingly, we can define certain temperatures of importance.

It is discussed already that as the vapour is compressed, it changes into liquid. We can also define a state in which the vapour changes into water and ice simultaneously. Thus at this stage, equilibrium exists between gaseous, liquid and solid states simultaneously. This specific state of the vapour is known as triple point. For water, the temperature and pressure corresponding to triple state are 0.01°C and 6.11 hPa respectively.

Another important state is the critical state. At the critical state the distinction between liquid and gas phases disappears. In the case of water, the temperature and pressure corresponding to critical state are 374 ° C and 221000 hPa respectively which never occur in the atmosphere.

The gaseous substances which exist below the critical temperature in the atmosphere are called vapour whereas those which exist above the critical temperature in the atmosphere are called gas.

5.1. Latent Heat

Whenever a unit mass of any substance changes its phase, certain amount of heat is absorbed or released from it which is termed as the latent heat. Latent heat is expressed in Joules/Kg in the SI unit.

The amount of heat required to change one kilogram of ice to one kilogram of water at the same temperature is called latent heat of fusion (L_{iw}).

The amount of heat required to change one kilogram of liquid water to one kilogram of vapour at the same temperature is called latent heat of vapourisation (L_{wv}).

The amount of heat required to change one kilogram of ice to one kilogram of water vapour is called latent heat of sublimation (L_{iv}).

For the changes of phase in the opposite direction, equal amount of latent heats are released.

Thus, $L_{iw} = L_{wi}$; $L_{wv} = L_{vw}$; $L_{iv} = L_{vi}$

Also it can be shown that, $L_{iv} = L_{iw} + L_{wv}$

5.2. Relation between Saturation vapour pressure and temperature

The relation between the saturation vapour pressure and temperature which is below the critical point is given by the Clausius-Clapeyron equation.

$$de_s/dT = L_{12} / T(\alpha_2 - \alpha_1)$$

where L_{12} is the latent heat involved with the transformation of the water substance from phase 1 to phase 2 and α_1 and α_2 are the specific volumes corresponding to those two phases which are in equilibrium at temperature T.

5.3. Equation of state for water vapour

We have the ideal gas equation or the equation of state as $p\alpha = RT$

In case of water vapour, the pressure will be corresponding to the partial pressure exerted by water vapour known as the vapour pressure (e). Hence the equation of state becomes

$$e\alpha_v = R_v T$$

Where $R_v = R^*/M_v$ the specific gas constant for water vapour, R^* is the universal gas constant and M_v is the molecular weight of water vapour. The value of R_v is approximately equal to 461.51 J/Kg/ °K.

5.4. Moisture Variables

In order to represent the amount of water vapour in a sample of air, different moisture variables can be used. They can be discussed one by one as given below.

Vapour pressure (e) : The partial pressure exerted by water vapour is termed as vapour pressure and it can be expressed in the same unit for atmospheric pressure (hPa).

Saturation vapour pressure (e_s) : The partial pressure exerted by water vapour when the space immediately above the surface is saturated is called saturation vapour pressure, at that temperature. It is a function of temperature and its value increases with the increase of temperature and vice versa.

Saturation vapour pressure is always expressed with respect to space below. Hence for water vapour, there can be two saturation vapour pressures for the temperature below freezing point; the saturation vapour pressure with respect to ice and saturation vapour pressure with respect to super cooled water (water can remain in the atmosphere in the same phase, even below the temperatures below freezing point. Such water is known as super cooled water).

The saturation vapour pressure over ice is less than the saturation vapour pressure over super cooled water. Saturation vapour pressure is also expressed in hPa.

Mixing ratio (r) : Mixing ratio for a sample of moist air is defined as the mass of water vapour present in that sample (m_v) to the mass of dry air (m_d) in that sample.

$$r = m_v/m_d$$

In terms of vapour pressure, the mixing ratio can be written as

$$r = 0.622 e/p-e$$

where p is the total atmospheric pressure.

Saturation mixing ratio (r_s) : Saturation mixing ratio is the ratio of the mass of water vapour to the mass of dry air present in a sample of moist air, when the air is saturated. Since the saturation is defined with respect to the surface below, there can be two types of saturation mixing ratio i.e. saturation mixing ratio with respect to ice and saturation mixing ratio with respect to water.

In terms of vapour pressure, the saturation mixing ratio can be written as

$$r_s = 0.622e_s / p - e_s$$

The unit of mixing ratio is Kg/Kg. However, as the amount of water present in a sample of air will be very small as compared to the total mass of the sample, it can be expressed in g/Kg.

Specific humidity (q) : Specific humidity is the ratio of the mass of water vapour present in the sample of moist air to the total mass of the sample.

i.e. $q = m_v / m_d + m_v$

Specific humidity is also expressed in Kg/Kg or g/Kg.

Relative Humidity (U) : Relative humidity is the ratio of the actual mixing ratio of a sample of air to the saturation mixing ratio of the sample, at a given temperature and pressure. Relative Humidity is usually expressed as a percentage.

$$U = (r/r_s) \times 100\%$$

Relative humidity can also be expressed as the ratio of the actual vapour pressure to the saturation vapour pressure for a given temperature and pressure, expressed as a percentage.

i.e. $U = e/e_s \times 100\%$

5.5. Equation of state for moist air

Moist air is the mixture of dry air and water vapour. The equation of state for moist air can be written as

$$p\alpha = R_m T$$

Where p is total pressure exerted by the moist air, α is the specific volume, T is the temperature and R_m is the specific gas constant for moist air.

5.6. Specific gas constant for moist air

Moist air is a mixture of dry air and water vapour. Hence using the formula for the specific gas constant for a mixture of gases we can find out the expression for the specific gas constant for moist air.

Suppose m_d and M_d are the mass and molecular weight of the dry air and m_v and M_v be the mass and molecular weight of water vapour then using the expression for the specific gas constant for a mixture, $R^*/\text{Mean } M$ Where

$$1/\text{Mean } M = \Sigma (m_n/M_n) / \Sigma m_n$$

We have

$$R_m = R^*/\text{Mean } M$$

$$1/\text{Mean } M = (m_d/M_d + m_v/M_v) / (m_d + m_v)$$

$$\text{We have } R_d = R^*/M_d \text{ and } R_v = R^*/M_v$$

Hence,

$$1/\text{Mean } M = (m_d R_d + m_v R_v) / R^*(m_d + m_v)$$

Thus

$$R^*/\text{Mean } M = (m_d R_d + m_v R_v) / (m_d + m_v)$$

Dividing the numerator and the denominator of the R.H.S by m_d and simplifying, we get

$$R_m = (R_d + r R_v) / (1 + r) = [R_d(1 + (R_v/R_d)r)] / (1 + r)$$

where $r = m_v/m_d$, the mixing ratio. If r is very small, the above expression will reduce to $(1 + 1.61r)(1 - r)$

Thus neglecting higher order terms, $R_m = R_d(1 + 0.61r)$

Here we get the expression for specific gas constant for moist air in terms of the specific gas constant for dry air and the mixing ratio.

For moist air, r is always positive. Hence we have $R_m > R_d$

5.7. Virtual temperature

Let us consider two samples of air ; one dry air sample and the other a moist air sample at the same temperature and pressure, the value of the same being T and p respectively.

The equation of state for dry air can be written as $p = \rho_d R_d T$ where ρ_d is the density of dry air.

Similarly, the equation of state for moist air can be written as $p = \rho_m R_m T$ where ρ_m is the density of the moist air.

Equating both the equations above, $\rho_d R_d T = \rho_m R_m T$

Substituting for R_m in terms of R_d and rearranging,

$$\rho_d = \rho_m(1 + 0.61r)$$

r is a positive quantity. Hence we can infer that ρ_d is greater than ρ_m . Thus the density of dry air is more than the density of moist air, at the same pressure and temperature.

But it is to be mentioned that the density of dry air will decrease with the increase of temperature, if the pressure is kept constant. Hence for a moist air sample at a particular temperature and pressure, we can have a sample of dry air with the same pressure and density but with a higher temperature. Let this temperature be denoted by T_v .

Accordingly the equation of state for the dry air and the moist air sample can be respectively written as

$$p = \rho_d R_d T_v \quad \& \quad p = \rho_m R_m T$$

Thus, $\rho_d R_d T_v = \rho_m R_m T$

Substituting for R_m in terms of R_d and rearranging, we get $T_v = T(1 + 0.61r)$

Thus the virtual temperature can be obtained from the temperature T and the mixing ratio r . Since r is a positive quantity, $T_v > T$

Hence the equation of state for moist air can be written as

$$p = \rho_m R_d T_v$$

or $p \alpha_m = R_d T_v.$

Thus, virtual temperature is the temperature that a moist air sample should have, if its pressure and specific volume are equal to that of a sample of dry air.

5.8. Isobaric processes of moist air

Physical processes in which the pressure remains constant are termed as isobaric processes. Isobaric processes corresponding to moist air are the cooling or warming of moist air at constant pressure.

The cooling of the moist air at constant pressure can be carried out by adding water vapour by evaporation or by not allowing the water vapour to enter or leave the sample of air.

Warming at constant pressure can be done by the condensation of water vapour present in the sample and the resultant release of latent heat.

According to these processes, certain characteristic temperatures can be defined.

In the process of cooling of the sample of moist air without allowing any water vapour to enter or leave the sample, the mixing ratio of the sample remains a constant. If the cooling process is continued like this, a particular temperature will be reached in which the air sample becomes saturated with the available amount of water vapour itself. This temperature is known as the dew point temperature.

In the process of cooling of the air parcel by adding water vapour, the cooling is occurring due to availing of the latent heat from the sample for evaporation to occur. In this case the temperature which is eventually reached when the sample gets saturated is known as the wet bulb temperature.

The third type of isobaric process is one in which the water vapour present in the sample condenses at constant pressure. During condensation, latent heat is released which will eventually warm the sample. The temperature reached when all the water vapour in the sample gets condensed is known as equivalent temperature. However, there is no practical mechanism available in the atmosphere to condense all the water vapour present in the air sample at constant pressure.

5.9. Adiabatic processes of moist air

We have discussed about the adiabatic processes of dry air in some earlier section. In an adiabatic process, the pressure and temperature of the air sample are related by the Poisson equation.

The moist air can be of two types; unsaturated or saturated. For an unsaturated air parcel, the Poisson equation is applicable if the mixing ratio of the sample remains

constant. This is possible when no water vapour condenses from or evaporates into the air sample and no exchange of water vapour occurs between the sample and its surrounding because of eddy transport.

In such condition, the Poisson equation for the dry air is applicable for the moist air also but the value of κ will be given by R_m/C_{pm} where R_m is the specific gas constant for moist air and C_{pm} is the specific heat at constant pressure for moist air.

But R_m/C_{pm} is approximately equal to $(R_d/C_{pd}) (1-0.2 r)$ Where R_d is the specific gas constant for dry air and C_{pd} is the specific heat at constant pressure for dry air and r is the mixing ratio.

Since the mixing ratio r is a very small quantity as compared to the terms, there is not much difference between the values of R_d/C_{pd} and R_m/C_{pm} . Hence the Poisson equation for the adiabatic process of dry air itself can be considered for the adiabatic process of moist unsaturated air.

5.10. Adiabatic processes of saturated moist air

The saturated moist air consists of dry air and water vapour, where the water vapour content is the maximum possible at that temperature. If such an air sample expands adiabatically, water vapour condenses and due to that latent heat is released. However, the adiabatic process for the air sample is applicable for the system as a whole and not for the individual constituents.

Before going into the details of the adiabatic process of moist saturated air, we have to define the reversible and irreversible physical processes.

A reversible process is one, which after passing through several stages of the process and finally returned back to its original stage, will retain all the properties of the sample. An irreversible process is one in which the initial properties of the sample gets changed when it is brought back to the initial stage after passing through several stages. The adiabatic process of saturated air can be reversible or irreversible depending upon the way it is carried out.

It is already mentioned that adiabatic expansion of saturated air leads to its cooling resulting in the condensation of the water vapour present in that. Suppose the condensation products are retained in the parcel. Here the rising and expansion of the parcel will lead to condensation of the water vapour. Once the parcel is brought back, it is compressed and warmed but this heating will be utilised for the evaporation of the condensed products. Hence, the air sample retains all its properties and will consists of the same material when it is brought back to its initial level. Such process can hence be considered as reversible.

In the second method, the condensation products are allowed to leave the parcel as soon as they form hence they fall out of the parcel once they form. Thus the sample will change its mass and composition. In this case the adiabatic expansion leads to cooling and condensation of the products which are dropped out. During its return journey, it is compressed and warmed but this warming can not be utilised to evaporate the condensed products since they have left the parcel already. Hence on reaching the initial level it will have a higher temperature and thus its thermodynamic properties are also changed hence this process is not reversible but it is irreversible.

In the second case, the condensation products while falling out will carry some heat with it hence the process can not be considered as strictly adiabatic. Hence it is referred to as pseudo adiabatic.

We have seen that in a dry adiabatic process, the entropy and potential temperature are conserved. But in pseudo adiabatic process, since there is an exchange of heat between the sample of air and the surrounding, dh is not zero. Hence in a pseudo adiabatic process, entropy and potential temperature are not conserved.

6. Hydrostatic Equilibrium

In atmospheric physics, we deal with cloud formation and precipitation for which moist air has to ascent, cool and condense to form clouds. Hence vertical movement of moist air is very much essential for the cloud formation. In the atmosphere, there can not be any vertical acceleration, once there is a complete balance between the force of gravity and the pressure gradient force arising due to the decrease of pressure with height.

6.1. Force of gravity

Combining the effects of gravitation by which any object (here a sample of air) is attracted by earth towards its centre (as per the Newton's law of universal gravitation) and the effect of centrifugal force arising due to the rotation of the earth, the force of gravity is obtained. Both are vector quantities.

The force of gravitation is acting towards the earth's centre whereas the centrifugal force is acting outwards.

The centrifugal force is given by $\Omega^2 R$ where Ω is the angular velocity of the rotation of the earth and R is the distance of the object from the axis of rotation.

The shape of the earth is such that, it is flattened at the poles and it has an equatorial bulge. Hence the value of R is maximum at the equator and it is minimum (zero) at the poles. Hence gravity which is the vector sum of the gravitation and the centrifugal force is minimum at the equator (approximately 9.789 m/s^2) and maximum at the poles (9.832 m/s^2). The gravity is always acting perpendicular to the surface below.

6.2. Geo potential and Geo potential meter

Any object will possess some energy due to its position in the gravitational field of earth and it is known as gravitational potential energy. In atmosphere, we have to consider the object as the sample of air.

If a sample of air of mass m kilogram is at a height of z meters from the mean sea level in the gravitational field of earth with the acceleration due to gravity as $g \text{ m/s}^2$ (which does vary much with height) then the gravitational potential energy possessed by the sample will be equal to mgz joules.

Then the gravitational potential energy per unit mass will be equal to gz joules/kg. The gravitational potential energy per unit mass is called geopotential (ϕ).

The unit of geopotential is Joules/kg. WMO later adopted another unit for the geopotential known as geopotential meter (gpm). It is defined in such a way that

$$1 \text{ geopotential meter} = 9.8 \text{ Joules/Kg}$$

Hence the geopotential meter is defined as the gravitational potential energy possessed by unit mass of air at a height of one meter from the mean sea level, where the value of the acceleration due to gravity is 9.8 m/s^2 .

Using the above relation, geopotential can be expressed in geopotential meters as given below.

$$\Phi = gz/9.8 \text{ gpm}$$

It is known that the value of g varies with latitude but it does not vary very much from 9.8. Hence from the above expression, it can be concluded that, the value of ϕ expressed in geopotential meters is approximately equal to the altitude of the air sample in meters.

6.3. Pressure in a fluid at rest

Pressure is defined as the force per unit area. In the atmosphere, we have to consider the force of gravity. The force of gravity corresponding to a mass m of air sample is nothing but the weight of that sample. Hence the force of gravity on a surface of unit area in the atmosphere is equal to the weight of the overlying air per unit area.

Let us consider a vertical column of air of unit cross sectional area with density assumed as a constant through out the column.

We have density $\rho = \text{mass}/\text{volume}$ or $\text{mass} = \rho \times \text{volume}$

If the depth of the surface of unit cross sectional area from the top of the atmosphere is h then the volume of the air column will be given by $1 \times h = h$.

Then mass of the air column will be ρh .

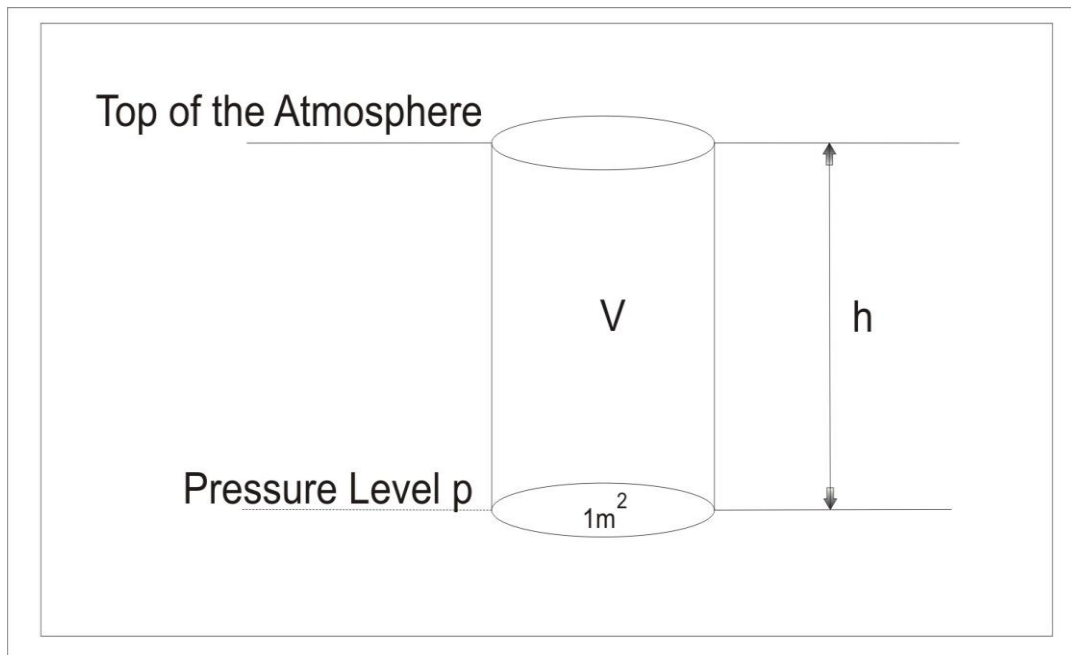


Figure 2: Pressure exerted by Air Column of Unit Cross Sectional Area

Accordingly, the weight of the air column or the force of gravity is ρgh .

Then the weight of the air column per unit cross sectional area or the pressure exerted by the air column per unit area is given by ρgh where ρ is the density, g is the acceleration due to gravity and h is the depth of the air column from the top of the atmosphere.

Thus the pressure acting on a surface of unit area in the atmosphere is given by the above expression. The unit of pressure is Newton/meter². From the above expression it is understood that as we move up in the atmosphere, the pressure decreases with height since the depth of air column goes on decreasing as we move up.

6.4. Hydrostatic equilibrium

Since the pressure goes on decreasing with height, there exist a vertical pressure gradient directed vertically upwards from high pressure to low pressure and the force arising due to this decrease of pressure is the vertical component of pressure gradient force occurring in the atmosphere.

We have already seen that the force of gravity is acting vertically downwards. If the vertical component of the pressure gradient force is completely balanced by gravity in the atmosphere, then the atmosphere is in hydrostatic equilibrium. In case of large vertical accelerations like that in thunderstorm and tornado etc., the hydrostatic equilibrium gets disturbed.

The mathematical expression for hydrostatic equilibrium can be obtained in the following way.

For this, let us consider a small cylindrical column of air of unit cross sectional area. Within this column, let us take two levels from the top of the atmosphere, level 1 and level 2 respectively corresponding to pressure p_1 and p_2 which are at a depth of h_1 and h_2 from the top of the atmosphere and which are at a height of z_1 and z_2 from the mean sea level.

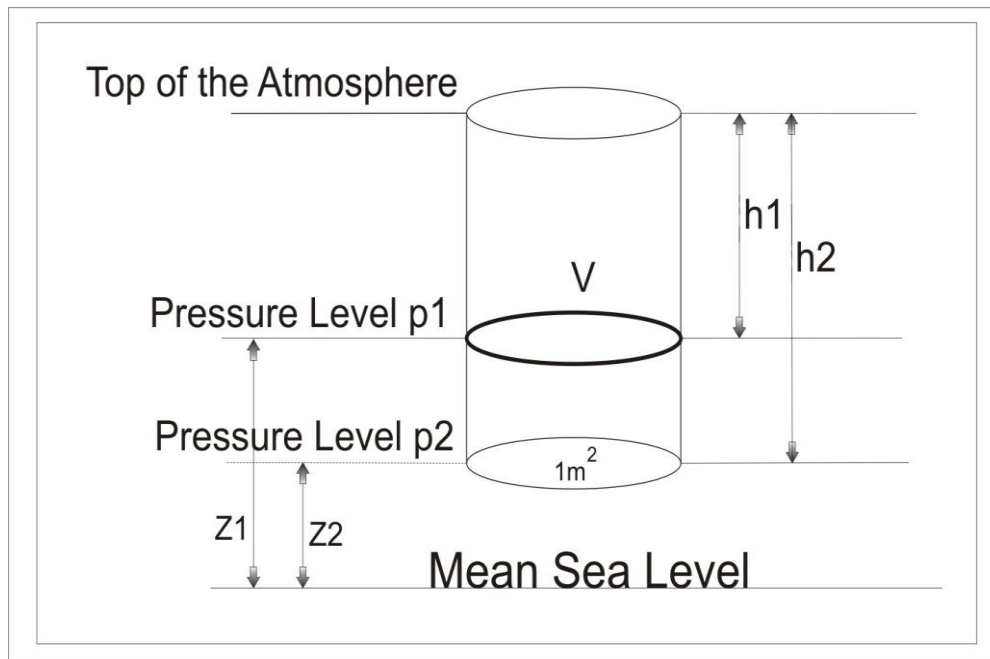


Figure 3: Increment of Atmospheric Pressure in terms of Increment of Altitude

Then $p_1 = \rho gh_1$ and $p_2 = \rho gh_2$

Then $p_2 - p_1 = \rho g(h_2 - h_1)$

But $h_2 - h_1 = z_1 - z_2$

Hence $p_2 - p_1 = \rho g(z_1 - z_2)$ or $p_1 - p_2 = - \rho g(z_1 - z_2)$ or $dp = - \rho g dz$

in differential form where dp is the increment in pressure corresponding to dz , the increment in altitude.

Here p_1 is the pressure corresponding to the upper level and p_2 is the pressure corresponding to the lower level. Hence the minus sign indicates that as we move up in the atmosphere, the pressure decreases.

6.5. Thickness of an atmospheric layer

We have the hydrostatic equation as $dp = -g dz$

$$\text{Or } g dz = -1/\rho dp$$

Thus, $g dz = -\alpha dp$ where α is the specific volume.

We have the equation of state for moist air as $p\alpha = R_d T_v$ or $\alpha = R_d T_v/p$

Thus we have $d\phi = -R_d T_v dp/p$

Now Let us take two levels in the atmosphere, the lower level corresponding to pressure p_2 which is at a height of z_2 from mean sea level and the upper level corresponding to pressure p_1 which is at a height of z_1 from mean sea level. Then the geopotential corresponding to these two levels will be ϕ_2 and ϕ_1 respectively.

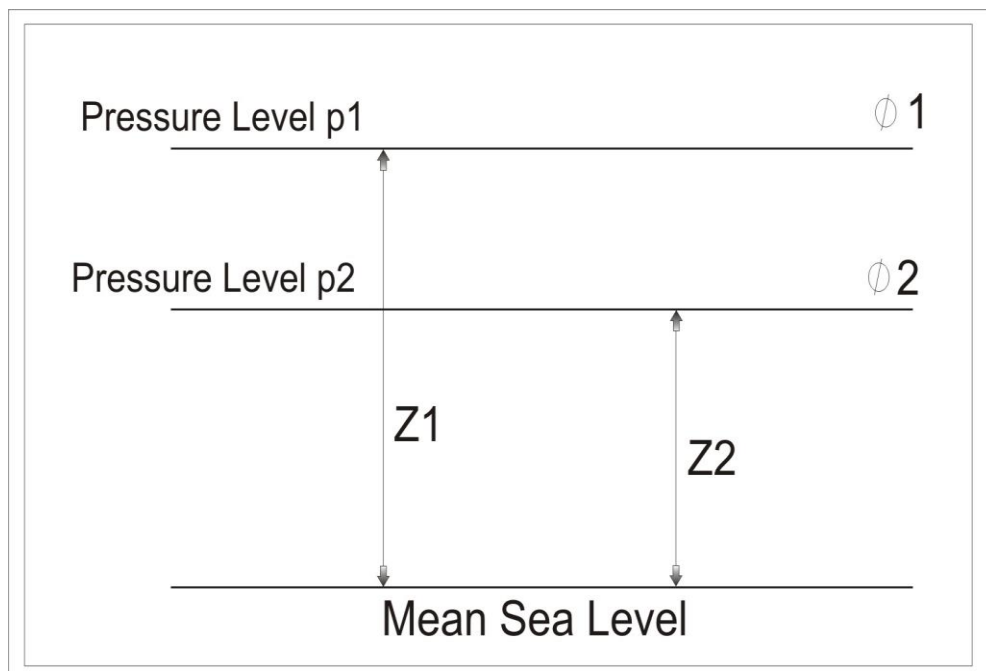


Figure 4: Geopotential Thickness of an Atmospheric Layer

Integrating the above expression between the limits ϕ_2 and ϕ_1 on the LHS and p_2 and p_1 in the RHS, we get,

$$\Phi_1 - \Phi_2 = R_d T_v \ln (p_2/p_1)$$

This gives the geopotential thickness of the layer between pressure level p_1 and p_2 . It can be determined using the above equation if the mean virtual temperature of the layer is known. The geopotential thickness will be numerically equal to the actual thickness approximately.

6.6. Reduction of pressure to mean sea level

For comparison of pressure between different points, it should be measured with respect to some standard reference level. In Meteorology, we use the standard level as the mean sea level and the station level pressure is reduced to mean sea level for plotting on the weather charts and for its analysis and comparison. This procedure can be explained now.

We have the equation for the thickness of the air column between two pressure levels explained above. In this if we treat p_2 as the mean sea level pressure and p_1 as the station level pressure, and then the expression for the mean sea level pressure can be obtained as follows.

We have

$$\Phi_1 - \phi_2 = R_d T_v \ln (p_2/p_1)$$

$$\text{Thus, } \ln (p_2/p_1) = (\Phi_1 - \phi_2)/R_d T_v$$

$$\text{Or } p_2/p_1 = \exp[(\Phi_1 - \phi_2)/R_d T_v]$$

$$\text{Thus, } p_2 = p_1 \times \exp[(\Phi_1 - \phi_2)/R_d T_v]$$

Thus using the station level pressure p_1 and the geopotential thickness between the station and the mean sea level, the mean sea level pressure can be determined if the mean virtual temperature of the fictitious air column between the surface and the mean sea level is known.

Generally, the temperature taken from the thermometer attached with the barometer is used for the computation.

7. Stability of the atmosphere

We have already learnt about the hydrostatic equilibrium of the atmosphere which is the balance between the vertical component of the pressure gradient force and the gravity. In case of hydrostatic equilibrium, there can not be any vertical acceleration. But in case of weather phenomena like turbulence and convection, there will be vertical movement of the air hence we have to study the forces which come into play in case of such phenomena in which the hydrostatic equilibrium gets violated.

Before going into the details of the stability or instability of the vertically moving air, the details regarding the lapse rate can be looked into. It is known that the temperature in the troposphere normally decreases with height. The rate of decrease of temperature with height is known as the lapse rate or the environmental lapse rate (ELR).

i.e.
$$\text{ELR}(\gamma) = -dT/dz$$

In case of moist air, the lapse rate of the virtual temperature (T_v) is considered rather than the actual temperature (T).

$$\gamma \text{ i.e. ELR}(\gamma_v) = -dT_v/dz$$

When the temperature is falling with height, the lapse rate is positive. In the troposphere, occasionally there can also be layers in which the temperature increases with height. This is known as temperature inversion. In an inversion layer the lapse rate will be negative.

Even though the lapse rate varies in the atmosphere, it can be approximated as a constant for a layer. If p_0, T_{v0}, z_0 and p, T_v, z are the pressure, virtual temperature and the altitude at the bottom and top of a layer respectively, and γ_v is the constant lapse rate for the layer then,

$$T_v = T_{v0} - \gamma_v(z - z_0)$$

7.1. Adiabatic Lapse Rate

Within the environment, we may consider certain parcels of air which is moving up or down adiabatically. Since the movement of the parcel is adiabatic, the rate of fall of temperature of the parcel will be different from that of the environment and that corresponds to the parcel lapse rate. The vertical movement of the parcel and whether the same will be stable or unstable can be determined by comparing the lapse rates of the parcel and that of the environment.

Since the parcel is moving adiabatically, there is no exchange of heat between the parcel and its surroundings. This condition is strictly not possible in the free atmosphere hence we assume that the energy exchange is negligible which can be considered as possible in short period processes.

But the lapse rate of the parcel will depend upon the thermodynamic properties of the parcel.

First of all let us consider a sample of dry air which is moving vertically upwards in the atmosphere. Once it is moving upwards in the atmosphere, it enters the regions of low pressure aloft hence it expands and thus cools. The rate of fall of temperature the parcel has with height is referred to as dry adiabatic lapse rate (DALR) denoted by γ_d .

i.e. $\gamma_d = -dT/dz$ for the parcel which is moving adiabatically.

We can get an expression for γ_d from the first law of thermodynamics for adiabatic processes.

$$C_p dT - \alpha dp = 0$$

But we have $dp = -\rho g dz$ from the hydrostatic equation.

Substituting, we have

$$C_p dT - (-g dz) = 0$$

$$-dT/dz = g/C_p$$

Substituting the value of g ($9.80665 \text{ J/Kg/}^\circ \text{K}$) and C_p ($1005 \text{ J/Kg/}^\circ \text{K}$) for the dry air, We have the value for dry adiabatic lapse rate (γ_d) as 9.8° per kilometre.

It was already discussed earlier that the adiabatic processes of unsaturated moist air is similar to the adiabatic process of dry air, till the parcel gets saturated with the ascent.

Next we will consider the adiabatic process of a saturated air parcel. The rate of fall of temperature that a parcel of saturated air will have with height is known as saturated adiabatic lapse rate or SALR. It is denoted by γ_s .

In case of saturated air parcel, the vertical ascent and the resultant expansion and cooling leads to the condensation of the water vapour. This condensation leads to the release of latent heat which will partly compensate the cooling due to adiabatic expansion. Hence the rate of fall of temperature with height in case of a saturated air parcel will be less than that in case of a dry air parcel. Hence, we can conclude that SALR is less than DALR.

The expression for SALR is given by

$$\gamma_s = g/C_p \{ (1+ Lr_s/R_dT) / (1+ 0.622 L^2r_s/C_pR_dT^2) \}$$

Where g is the acceleration due to gravity, C_p is the specific heat at constant pressure, L is the latent heat of condensation, r_s is the saturation mixing ratio, R_d is the specific gas constant for dry air and T is the temperature.

We know that the dry adiabatic lapse rate γ_d is given by g/C_p . Hence from the above expression it is understood that γ_s is a fraction of γ_d . Hence it is once again shown that γ_s is less than γ_d .

7.2. Equilibrium Conditions

If the forces acting on an object are balanced, it will be in a state of equilibrium. When such an object is subjected to any displacement with the application of an external force, they respond differently and accordingly three types of equilibrium can be defined.

If the displaced object comes back to its original position, then it is in stable equilibrium. If the object continues to move further, it is in unstable equilibrium. If the displaced object remains at its new position without moving further or coming back, then it is in neutral equilibrium.

In Meteorology, the object of interest is a sample or parcel of air which is originally in hydrostatic equilibrium. Its behaviour of movement after it is given a small vertical displacement determines its hydrostatic stability or the static stability. In order to determine the static stability of a parcel, the parcel method is used.

7.3. The Parcel Method

The parcel method is used to determine the static stability of a parcel of air. In this we make the following assumptions first of all.

1. No heat is added to or removed from the parcel hence its movement is strictly adiabatic.
2. During its movement, the parcel does not mix with the environment hence it retains its identity.
3. When the parcel moves there is no compensatory movement occurring in the environment.

The movement of the parcel under adiabatic conditions can be justified since the air is a very poor conductor and the mixing of the parcel with the environment is very slow thus the heat exchange due to that is insignificant. Also the temperature

changes due to adiabatic cooling and warming are highly significant with respect to the temperature changes due to radiative processes.

With these assumptions, we consider a parcel of air, initially at rest is given a small vertical displacement.

7.4. Expression for Vertical Acceleration of the Parcel

Let p, ρ, T_v be the pressure, density and virtual temperature for the environment and p', ρ' and T_v' be the pressure, density and virtual temperature for the parcel. It is assumed that the parcel is initially in hydrostatic equilibrium.

Hence, $-g - 1/\rho'(dp'/dz) = 0$ for the parcel.

Suppose this parcel is given an initial push by which it starts moving up (thus having a vertical acceleration) and due to the fall in pressure it expands adiabatically.

Since the pressure of the parcel is always getting adjusted with that of the environment, we can consider $p' = p$

Hence the above equation becomes,

$dw/dt = -g - 1/\rho'(dp/dz)$; but $dp/dz = -g\rho$ from the hydrostatic equilibrium equation.

Thus, $dw/dt = g(\rho - \rho')/\rho'$

If the parcel is assumed to be lifted from the initial level zero (mean sea level), then its new position with respect to mean sea level after the removal of the force will be z , the altitude. Also the parcel and the environment will have the same temperature initially.

Applying this, the vertical acceleration of the parcel in terms of lapse rate of the environment (γ) and adiabatic lapse rate of the parcel (γ') will be given by

$dw/dt = gz(\gamma - \gamma')/T_v$ where T_v is the virtual temperature of the environment at altitude z .

The sign of dw/dt is determined by the difference between γ and γ' .

If $\gamma > \gamma'$ $dw/dt > 0$ then the parcel is unstable

If $\gamma < \gamma'$, $dw/dt < 0$ then the parcel is stable

If $\gamma = \gamma'$, $dw/dt = 0$ then the parcel is in neutral condition.

7.5. Application of parcel method

The parcel method for the determination of the stability of an air parcel is already discussed in the previous section. Now we will consider the same for the determination of the stability condition of the different types of air parcels.

Let the atmosphere be in hydrostatic equilibrium in the beginning with a certain Environmental Lapse Rate (γ). The pressure, temperature and density of the parcels are all the same as that of the environment in the initial stage.

Now, suppose the parcel is given a small upward displacement by some external agency. Then it will expand and cool adiabatically. If the parcel is dry or unsaturated, it will cool as per DALR(γ_d) and if the parcel is saturated, it will cool as per SALR(γ_s).

After certain vertical displacement by the application of external force, if the parcel is at higher temperature than the surrounding then it will accelerate further upwards. On the contrary, if the temperature of the parcel is less than that of the surrounding, then it will be forced downwards.

After the initial displacement three types of equilibrium can occur. The parcel is in stable equilibrium if it tends to return to its original position. If the parcel tends to move away from its original position further, it is in unstable equilibrium. If the parcel does not move in either direction from the displaced position, it is in neutral equilibrium.

By the application of parcel method, for the dry air as well as for the unsaturated air parcels, the equilibrium stages is given by

$$\gamma < \gamma_d \quad \text{stable equilibrium}$$

$$\gamma > \gamma_d \quad \text{unstable equilibrium}$$

$$\gamma = \gamma_d \quad \text{neutral equilibrium}$$

For a saturated air parcel, the equilibrium states are given by,

$$\gamma < \gamma_s \quad \text{stable equilibrium}$$

$$\gamma > \gamma_s \quad \text{unstable equilibrium}$$

$$\gamma = \gamma_s \quad \text{neutral equilibrium}$$

When $\gamma > \gamma_d$, it implies that $\gamma > \gamma_s$ also since $\gamma_d > \gamma_s$.

Hence, $\gamma > \gamma_d$ is the condition for absolute instability.

Also when $\gamma < \gamma_s$ it implies that $\gamma < \gamma_d$ also since $\gamma_s < \gamma_d$.

Hence, $\gamma < \gamma_s$ is the condition for absolute stability.

But on most of the occasions, the environmental lapse rate (γ) lies between the dry adiabatic lapse rate (γ_d) and the saturation adiabatic lapse rate (γ_s). **This is known as the conditional instability.**

Thus $\gamma_s < \gamma < \gamma_d$ is the requirement for the conditional instability of the parcel of air.

In this case, the atmosphere is stable for dry air or moist unsaturated air parcels but it will be unstable for the moist saturated air parcels.

7.6. Latent instability

If the atmosphere is in conditional state, then any unsaturated air parcel lifted will return to its original level. However if the mechanism by which it is forced to ascent is sufficient enough to make it saturated before the forced lifting ceases, it may be able to reach a level where it can become unstable. This is due to the fact that the unsaturated air parcel is cooling at the rate of DALR but once it becomes saturated its cooling will be as per SALR due to the release of latent heat. Hence the rate of cooling during ascent gets reduced and at some level, its temperature will become equal to that of the environment. Further ascent of the parcel will make it warmer than the environment and hence it will become unstable as far as it remains warmer than its surrounding. Once all the water vapour available in the parcel gets condensed out, it will become dry and will start cooling as per DALR. Due to this increased rate of cooling, at some level it will again become cooler than the environment after which it can not remain in unstable equilibrium.

The atmospheric layer in which an unsaturated air parcel becomes unstable by forced ascent is said to have latent instability.

The level at which an unsaturated air parcel becomes saturated by ascent is called Lifting Condensation Level (LCL). The level at which the temperature of the parcel become equal to that of the environment during ascent is called Level of Free Convection (LFC). After LFC, the temperature of the parcel will be warmer than the environment. The level at which the parcel temperature again becomes equal to that of the environment after all the water vapour available in the parcel gets condensed out is called the Equilibrium Level (EL).

Hence atmosphere in a conditional state can release instability when there is a mechanism of sufficient strength to make forced ascent of moist unsaturated air to its saturation point, preferably sufficient enough to lift the parcel upto LFC level. Secondly, there should be sufficient amount of moisture present in the parcel so that the parcel becomes saturated by forced ascent at lower level itself. Moreover, significant amount of water vapour in the parcel can keep it warmer than the

surrounding for more time above LFC, till the whole water vapour gets condensed out.

It can be seen that till LFC, the parcel is cooler than the surrounding hence for its ascent, energy is to be consumed by the parcel in terms of some mechanism. Hence the area corresponding to this process refers to the negative area on a thermodynamic diagram.

However after LFC till EL the parcel is releasing energy to the surrounding hence the area corresponding to this process is referred to as positive area on a thermodynamic diagram.

According to the size of the positive and negative areas on a thermodynamic diagram, two types of latent instability can be defined. If the positive area is greater than the negative area, there is real latent instability and if the negative area is greater than the positive area there is pseudo latent instability.

For weather developments, the requirement is that of real latent instability.

Surface heating, orography, low level convergence and frontal ascent are the various mechanisms present in the nature by which an unsaturated moist air parcel can be lifted to saturation.

7.7. Potential Instability

In the earlier sections, the stability criteria for different types of parcels have been discussed. Now the stability criteria for a layer of air which is lifted as a whole can be considered. This is referred to as potential instability or the convective instability.

A potentially unstable layer is one which becomes unstable after the entire layer has been lifted to saturation. For the layer to be unstable there should be sufficient temperature difference between the bottom and top of the layer, due to the release of latent heat.

It is observed that, layers with latent instability often overlap or coincide with the layers of potential instability. A layer with real latent instability can be potentially stable.

For real latent instability atmosphere needs to be conditionally unstable and there should be very high amount of relative humidity. But the potential instability does not depend directly upon these two factors as such.

8. Thermodynamic diagrams

Thermodynamic diagrams are graphical diagrams which are used to study the various thermodynamic processes occurring in the atmosphere. It is a tool by which the energy transformations that occur during different thermodynamic processes can be studied. In such diagrams, the various processes to which an air parcel can be subjected to like isobaric, isothermal, dry adiabatic, pseudo adiabatic etc can be represented in terms of isolines. Thermodynamic diagrams can also be referred to as aerological diagrams.

In meteorology, the temperature and humidity etc. are determined with respect to pressure levels. In aviation also, the data with respect to pressure level are more convenient to use than the data with respect to the height. Since the logarithm of pressure is closely related to the altitude, in thermodynamic diagrams, usage of temperature and logarithm of pressure is more convenient.

8.1. Desirable properties of a thermodynamic diagram

A thermodynamic diagram should have the following properties.

1. The area enclosed by the lines representing any cyclic process should be proportional to the energy exchange or work done during the process.
2. As far as possible, the lines representing the fundamental process should be straight.
3. The angle between the lines of equal temperature (isotherms) and lines of equal potential temperature (dry adiabats) should be as large as possible.
4. In the lower levels of the atmosphere, the angle between the dry adiabats and the pseudo adiabats should be as large as possible.

8.2. Tephigram

Tephigram is the thermodynamic diagram which is commonly used in India. This diagram was introduced by Sir Napier Shaw. Here the coordinates are the temperature T and entropy ϕ hence the name Tephigram. Since entropy and potential temperature are related by the expression .

$dh/T = \phi = C_p d(\ln\theta)$, potential temperature θ is also used along with entropy on the y axis.

We have the relation

$$C_p d(\ln\theta) = C_p d(\ln T) - R d(\ln p)$$

Multiplying with T,

$C_p d(\ln\theta) \times T = C_p dT - RT/p dp = C_p dT - \alpha dp = dh$, as per the first law of thermodynamics.

When entropy and temperature are used as the coordinates for the diagram the area enclosed by any cyclic process on this diagram will be proportional to the total energy exchanged during the process, according to the above expression. Thus the tephigram satisfies the first condition for a thermodynamic diagram.

On a thermodynamic diagram, the vertical lines are lines of equal temperature (isotherms) and the horizontal lines are lines of equal potential temperature or dry adiabats. The straight lines slanting towards left are the lines of equal mixing ratio or isohygrics. The lines of equal pressure or isobars are logarithmic curves rising towards right which are almost straight in the higher levels. The pseudo adiabats are significantly curved and they are rising towards left. Thus on a tephigram, the dry adiabats, isotherms and isohygrics are straight lines whereas the isobars and the pseudo adiabats are curved. Hence the tephigram follows the second criteria desirable for a thermodynamic diagram.

The angle between dry adiabats and isotherms on a tephigram is 90° which is sufficiently large enough thus the tephigram follows the third criteria also.

In a tephigram, there is significant angle between the dry adiabat and the saturated adiabat in the lower levels of the atmosphere, thus the fourth criteria is also met in a tephigram.

8.2.1. Plotting and interpretation of Tephigram

On a tephigram, the actual temperature and the dew point temperature with respect to different pressure levels as reported in radio-sonde data are plotted. From the data plotted, the following quantities can be derived.

Mixing ratio (r) : Mixing ratio for any pressure level can be found out from the mixing ratio line passing through the dew point temperature of that level.

Saturation mixing ratio (r_s) : Saturation mixing ratio can be found out with reference to the mixing ratio line passing through the dry bulb temperature for any level.

Once these two quantities are found out from the tephigram, the relative humidity for that level will be given by the relation $RH = (r/r_s) \times 100\%$.

In any level if the dew point and the temperature lines are close by, it shows that the actual mixing ratio is very close to the saturation mixing ratio indicating very high

relative humidity. On the contrary, if the dew point and dry bulb temperature curves are far apart, that indicates lower values of relative humidity in that level.

Virtual temperature (T_v) : We have the relation between the virtual temperature and the dry bulb temperature as $T_v = T(1 + 0.61r)$ where T is the dry bulb temperature in Kelvin and r is the mixing ratio. Since the mixing ratio is given in grams in a tephigram, the value of T_v can be approximately equal to $T + r/6$ where r is the mixing ratio value derived from the tephigram.

Potential temperature (θ) : For finding out the potential temperature for any pressure level, move along the dry adiabat from the dry bulb temperature of that level to reach the 1000 hPa level. The temperature in ° K at the 1000 hPa level is the potential temperature for that level.

Lifting Condensation Level (LCL) : To find out the lifting condensation level of an air parcel for any pressure level, from the dry bulb temperature value of that level draw a dry adiabat and from the dew point temperature value of that level, draw an isohyric line. The point (corresponding pressure level) of intersection of the two lines will give the lifting condensation level for that parcel.

Convective Condensation Level (CCL) : From the LCL, extend the isohyric line further to meet the dry bulb temperature curve and the level of intersection of the isohyric line with the environmental curve corresponds to the convective condensation level.

Level of Free Convection (LFC) : From the LCL, draw the pseudo adiabat to meet the dry bulb temperature curve and the level at which the pseudo adiabat from the LCL meets the dry bulb temperature curve corresponds to the level of free convection or LFC.

Normand's Theorem : The Normand's theorem states that the dry adiabat from the dry bulb temperature and pseudo adiabat from the wet bulb temperature and the isohyric line from the dew point temperature will intersect at a point known as the Normand's point. The Normand's point usually coincides with the LCL.

Wet bulb temperature (T_w) : Using the above theorem, if we descend from LCL along the pseudo adiabat to reach the initial pressure level, the temperature corresponding to the point will be the wet bulb temperature (T_w) for that pressure level. Instead of the initial pressure level, if we descend till 1000 hPa, the temperature corresponding to that will be the wet bulb potential temperature (θ_w).

Equivalent temperature : From the lifting condensation level move along the pseudo adiabat towards the left side boarder of the tephigram (to assume the condensation of all the water vapour present) and then from there move along the

dry adiabat to reach the initial pressure level to get the *equivalent temperature* (T_e) for that level. On the other hand, if the dry adiabat is extended till the 1000 hPa level, the corresponding temperature received will be the equivalent *potential temperature* (θ_e).

Stability of layers : Stability or instability of an atmospheric layer can be found out by comparing the alignment of the environmental curve with respect to the dry adiabat and pseudo adiabat for that layer.

If the dry adiabat and pseudo adiabat for a layer are lying on the left side of the environmental curve, the layer is absolutely stable.

If the dry adiabat and pseudo adiabat are lying to the right of the environmental curve, then there is absolute instability in that layer.

If the environmental curve is lying between the dry adiabat and pseudo adiabat, then there is conditional instability in that layer.

Latent Instability : In order to find out the latent instability of a parcel of air from any pressure level, draw a dry adiabat from the dry bulb temperature of that level and draw isohyric from the dew point temperature of that level, to find out the LCL which is point of intersection of the above mentioned lines. Then from the LCL, draw the pseudo adiabat to represent the ascent of the parcel further. Where this pseudo adiabat cuts the dry bulb temperature curve that level corresponds to the LFC. After the LFC, the pseudo adiabat will be lying on the right side of the environmental curve and it may or may not cut the dry bulb temperature curve further. If the pseudo adiabat cuts the dry bulb temperature curve once again, it is the equilibrium level.

The area enclosed between the dry bulb temperature curve and the parcel curve, lying to the left of the dry bulb temperature curve is the negative area and the area lying to the right of the dry bulb temperature curve is the positive area.

If the positive area is found to be greater than the negative area, then, there is real latent instability and if the negative area is greater than the positive area, there is pseudo latent instability present in the atmosphere.

Potential instability : In order to find out the potentially unstable layers, the lapse rate of the wet bulb temperature with reference to the SALR needs to be looked into. The layers in which the lapse rate of the wet bulb temperature is greater than the lapse rate corresponding to the pseudo adiabat for that layer, then that layer can be considered as potentially unstable.

9. Basics of Cloud Physics

Air is referred to be saturated when the partial pressure of water vapour reaches the saturation vapour pressure or when its relative humidity is 100%. If air is cooled further, the actual vapour pressure exceeds the saturation vapour pressure and then the relative humidity becomes more than 100 %, if condensation is not occurring. Condensation does not take place until the water vapour has a suitable surface on which condensation can occur. These are referred to as condensation nuclei. If the condensation is occurring on a liquid water surface itself, the process is known as homogeneous nucleation. If the condensation nuclei is other than water surface, it is termed as heterogeneous nucleation. The homogeneous nucleation require very high amount of super saturation.

The condensation nuclei with affinity for water vapour is known as hygroscopic nuclei (example: salt particles) and those which does not possess such property is known as non hygroscopic nuclei (example:dust particles). On hygroscopic nuclei, the condensation can start even before the relative humidity reaches 100%.

The condensation nuclei can be divided into three categories according to their size. The nuclei with radius less than 0.1μ are called Aitken nuclei. If the radii exceed 1.0μ they are termed as giant nuclei. Nuclei with radii between 0.1μ and 1.0μ are termed as large nuclei.

It is possible for the water to remain in liquid phase without freezing even at temperature below freezing point and it is termed as super cooled water. The super cooled water will spontaneously freeze once the temperature falls below $-40\text{ }^{\circ}\text{C}$. However, for the super cooled water with temperature above $-40\text{ }^{\circ}\text{C}$ to freeze, it has to come into contact with some ice forming nuclei or ice nuclei. It is a solid particle which when come into contact with a mass of super cooled water will initiate growth of an ice crystal about itself.

9.1. Formation of clouds

Moist air needs to be lifted through vertical motion for its cooling and resulting condensation.

Gradual upward motion leads to the formation of sheet or layer clouds or stratiform clouds covering large areas in the sky.

Vigorous vertical currents produce cumuliform clouds which are usually separated from one another by clear space. The vertical motion can be of the following types.

9.1.1. Mechanical turbulence

Airflow near the earth's surface is getting deformed by friction due to the presence of buildings, trees, hills etc. The flow becomes turbulent due to that and in the upper part of this turbulent layer stratiform clouds can form, if the air is sufficiently moist.

The cloud formed by mechanical turbulence is stratus, a sheet cloud. However, the upper/lower surfaces will develop wave like structure and also breaks in between due to condensation in the ascent and evaporation in the descend. This give rise to the formation of stratocumulus.

If the turbulent mixing is in the higher level, it can give rise to the formation of high level stratocumulus or altocumulus.

9.1.2. Thermal turbulence or convection

Convection develops when the earth's surface is heated. Due to convection, air moves upwards. The air will rise till it remains warmer than the surrounding. On reaching a stable layer, the cloud top will spread out horizontally. Rapid local ascent give rise to formation of cumuliform clouds. Usually in such clouds, the vertical development will be more than the horizontal extension.

Isolated convective clouds of limited extent are called fair weather cumulus. Their vertical extent will be insufficient to form precipitation.

If the vertical currents are terminated due to the presence of an inversion layer above, stratocumulus cloud will develop.

If instability is present throughout a large vertical depth, cumulonimbus clouds form.

9.1.3. Orographic ascent

When moist air is forced to ascend over mountain ranges, clouds will develop as the air cools adiabatically due to the forced ascent.

When there is no sufficient moisture, no clouds will form. Large instability will give rise to the formation of cumulonimbus. Formation of cumulus or stratus clouds can also occur depending upon the instability created by the forced ascent. Due to orographic ascent, clouds form on the windward side with clear condition on the lee side.

Sometimes, the clouds form on the top of the mountain, thin at the edges and thick at the centre, shape similar to lense. These are known as ***lenticular clouds***.

9.1.4. Slow widespread ascent

This type of ascent is associated with convergence with respect to low pressure systems or with the frontal ascent. The cloud formation associated with this has large area coverage and it may last for several days also and it is associated with the combination of various types of clouds.

9.2. Fog

Fog is a cloud on the ground. It is composed of water droplets or ice crystals. For fog to occur, water vapour needs to be changed to liquid water by condensation or to ice by deposition. Hence water vapour saturation is needed in the layer close to the ground for occurrence of fog.

For fog formation, relative humidity needs to be close to 100% near the ground. However, in industrial areas, due to the presence of large amount of condensation nuclei, fog can occur even when the relative humidity is less than 80%.

Saturation of the air near the ground can be due to various processes. It can be due to cooling of moist air, by addition of water vapour or by mixing of air.

9.2.1. Cooling of moist air

Fog is mainly formed by this mechanism in which the layer of the atmosphere near the earth's surface is cooled.

Vertical motion of air through an upslope is one mechanism by which moist air can be cooled. Due to this, fog forms on the hill sides. The gentle slope needs to extend quite far a distance to produce sufficient cooling.

Turbulent heat transport is another mechanism by which air can be cooled. When there is an inversion present in the atmosphere, eddy heat transport will be downward from the layer above. However, in case of very light wind, supply of heat from above to the ground will not be sufficient to compensate the heat loss from the ground. Hence the boundary layer will continue to cool enhancing fog formation near the ground.

Radiation from the lower layers of the atmosphere is another mechanism by which air near the ground can be cooled. Moist air radiates heat more than dry air. Hence when water vapour is present sufficiently in the layer above the ground, it can lose heat to the ground as well as to the layer above. If this moist layer is having dry layer aloft, then the radiation from the lower layer can escape to space, leading to its cooling and saturation. This will give rise to the formation of fog.

When warm moist air moves over cold surfaces, the cooling of the lowest layers of the air will give rise to the formation of fog known as advection fog. In summer season, warm moist air moving over cool ocean surfaces or water bodies will give rise to the formation of fog. In winter season, the warm moist air moving over cold land surfaces will give rise to the formation of fog.

9.2.2. Addition of water vapour

Addition of water vapour by evaporation to the lower layers of the atmosphere can also give rise to the formation of fog. Evaporation can be from the earth's surface or it can be by evaporation of falling raindrops also. Fog may not develop immediately after evaporation and addition of water vapour but may develop subsequently when the air gets cooled. Evaporation of rain from a warm air mass to the cold air mass below in a front will give rise to the formation of fog in this manner.

9.2.3. Mixing of air

If two parcels of air near saturation at different temperatures are mixed together, fog can develop. Mixing of parcels of saturated or nearly saturated parcels at different temperatures and moisture contents can produce supersaturated mixture, which in presence of enough condensation nuclei can give rise to fog formation. For fog formation by this mechanism, high relative humidity, large temperature gradient and a mixing mechanism like turbulence should be present.

10. Electromagnetic Radiation

Energy is getting transferred from one body to another by means of electro magnetic waves and this transfer can occur with or without the presence of a material medium. Radiation is the form of energy transfer without medium and the energy transferred like this is known as electromagnetic radiation.

The electromagnetic spectrum consists of radio, infrared, visible, ultraviolet, X and gamma radiation. The radiation can be referred by its wavelength (λ) or the frequency (n) so that $c = n\lambda$ where c is the constant speed with which the radiation travels in a vacuum ($c = 2.997925 \times 10^8$ m/sec).

In order to represent the quantity of radiation transfer, different radiometric quantities can be defined. The radiant flux is the rate of flow of radiant energy which is the power emitted, received or transferred in unit time. Since joule (J) is the unit of energy, the unit of radiant flux is Watts (W).

Another radiometric quantity is the radiant intensity. Let us consider a source of radiation. If we consider an infinitesimal element of the source, the rate of flow of energy emitted by that element or the flux of radiation from that element can be found out. The three dimensional angle subtended by this area at the centre of the source of radiation is known as solid angle. If we divide the flux of radiation from the element of the source by the solid angle subtended by the element, the intensity of radiation can be received. Since the unit of solid angle is steradian, the unit of intensity is Watts/steradian.

Radiance: Let us consider an infinitesimal area of a source emitting radiation. Then the radiance of the energy getting emitted at a particular direction will be given by the quotient of the intensity of radiation in that direction divided by the orthogonal projection of that elemental area perpendicular to that direction. Its unit is Watts/steradian/Square meter.

Emittance: Emittance of radiation is the ratio of the radiant flux to the area of the element emitting radiation. Its unit is Watts/square meters.

It is to be mentioned that Emittance is π times the radiance.

Radiation is distributed over a range of wavelengths. Radiation pertaining to one wavelength is referred to as monochromatic radiation. Spectral density of any radiometric quantity is the quotient of this quantity over an infinitesimal range on either side of a given wavelength by the range of the wavelength.

10.1. Laws of Radiation

Kirchoff's Law : Let us consider an object which is emitting energy in a particular direction and simultaneously absorbing energy falling upon the same from the same direction. The monochromatic intensity of emission in a given direction (E_λ) can then be defined as the radiant energy emitted at a particular wavelength λ per unit area per unit time per unit wavelength which can be expressed in Joules/square meter/sec/micron. The monochromatic fractional absorption or the absorptivity (a_λ) which is a non dimensional quantity can be defined as the ratio of the radiation which is absorbed to the total incident radiation for a particular wavelength and direction. Kirchoff's law states that for all bodies, the ratio of the intensity of emission to the fractional absorption is a function of wavelength and temperature only.

Thus, $E_\lambda/a_\lambda = f(\lambda, T)$. This ratio is known as the emissivity. From the equation, it can be seen that, if the fractional absorption vanishes then the emitted intensity also vanishes. Otherwise the emission can occur only in wavelengths where absorption occurs. If the fractional absorption is a constant but less than one, the emission will be less than the maximum possible. An object with such property is known as a grey body. If the fractional absorption is complete or when its value is equal to one, then the emission will be the maximum possible

Or $E_\lambda = f(\lambda, T)$. This type of object absorbs all the radiation falling on it and emits the maximum radiation possible at all wavelength. Such a body is known as a black body.

Since $f(\lambda, T)$ can be considered as the basic characteristic of black body radiation, efforts were made to find out the explicit form of the same. Accordingly, Stefan – Boltzmann has shown that the integral of $f(\lambda, T)$ over all directions and all wavelength gives the total flux of energy which is given as

$$F = \sigma T^4$$

where T is the temperature and σ is the Stefan-Boltzmann constant. The above equation represents the Stefan-Boltzmann Law. In the plot of $f(\lambda, T)$ against λ for a particular temperature, the area under the curve represents the total radiation possible at that temperature.

From the observations, it was found that the black body radiation was maximum at some wavelength which depends upon the temperature. In 1893, Wein has shown that the wavelength of maximum emission (λ_{\max}) is a function of the temperature of the emitting body in such a way that $\lambda_{\max} \times T = a$ constant where the temperature is in degree Kelvin. This is known as **Wein's displacement law**. From the application of this law, it is seen that the wavelength corresponding to the maximum black body radiation from the sun is centred around 0.48μ in the visible spectrum, considering

the average temperature of the sun as 6000° K. For earth, the wavelength corresponding to the maximum black body radiation is centred around 10μ, considering the average temperature of the earth to be 288° K.

In 1900, Plank obtained a semi empirical expression for $f(\lambda, T)$, from the quantum theory. He has shown that,

$$E_{\lambda} d\lambda = (c_1 \lambda^{-5}) / (e^{c_2/\lambda T} - 1) d\lambda \text{ where } c_1 \text{ and } c_2 \text{ are constants.}$$

This is known as **Plank's law**. From Plank's law, the black body radiation is found to be a function of wavelength and temperature. The black body radiation curves for different temperature values shows that, the emission approaches zero for very small and very large wavelengths for each value of temperature and the emission is maximum for each temperature at a particular wavelength.

Beer's Law : When a monochromatic beam of radiation of intensity I_{λ} pass through a thin layer of the atmosphere dx , then a small increment of the same gets absorbed. Due to the absorption of the energy, there will be some emission from the layer also. Thus the absorption and the corresponding emission together contribute towards the change in the intensity of radiation. If the emission at that wavelength from the layer is considered to be very small, the change in the intensity will be due to the absorption only (neglecting the impact of reflection and scattering).

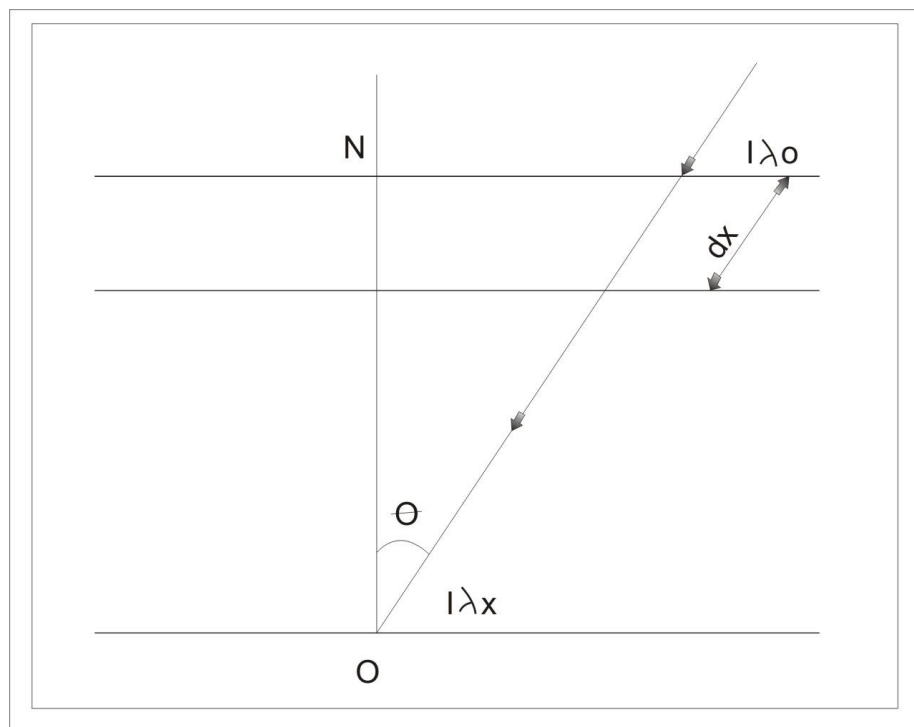


Figure 5: Transfer of Radiation through an Atmospheric Layer

Thus we have,

$dI_\lambda = -a_\lambda I_\lambda$ where a_λ is the absorptivity. It is to be mentioned that the absorptivity (a_λ) is proportional to the density of the air ρ and the distance travelled by the beam dx .

$$a_\lambda = k_\lambda \rho dx \text{ where } k_\lambda \text{ is the absorption coefficient of the air.}$$

Thus,

$$dI_\lambda = -k_\lambda I_\lambda \rho dx$$

or

$$dI_\lambda / I_\lambda = -k_\lambda \rho dx$$

Thus the fractional decrease of intensity owing to absorption is proportional to the mass per unit area of the absorbing material (ρdx) through which the beam passes through. This quantity is known as the optical path of the layer. If $I_{\lambda 0}$ and $I_{\lambda x}$ are the initial and final intensities before and after passing through a finite distance x , integration between these limits on both sides of the equation leads to the expression that

$$I_{\lambda x} = I_{\lambda 0} e^{-\int k_\lambda \rho dx}$$

This is integral form of the Beers law of absorption. The same law is valid for the scattering also but with a change of absorption coefficient k_λ to scattering coefficient s_λ .

The factor $k_\lambda \rho dx$ is often referred to as $d\tau$ and in terms of that the Beer's law can be written as

$$I_{\lambda x} = I_{\lambda 0} e^{-\tau}$$

the significance of τ is that when its value is equal to unity, the intensity of the beam gets reduced by a factor $1/e$, after its passage through the absorbing medium.

In case of passage of the radiation through the atmosphere, it is convenient to use the vertical distance z rather than the path x . The solar radiation is not vertical to the surface below but is from a zenith angle θ . Hence the actual path travelled by the beam dx can be written in terms of the vertical distance dz as

$$dx = dz \sec\theta$$

Thus the Beer's law can be written in terms of vertical distance as

$$I_{\lambda z} = I_{\lambda 0} e^{-\int k_\lambda \rho \sec\theta dz}$$

Mass per unit area of the absorbing material along the vertical distance between two levels of the atmosphere (ρdz) is known as **optical depth**. The unit for optical depth as well as for optical path is kilogram/square meters.

It is to be mentioned that, emission from the layer was neglected as compared to absorption during the derivation of the Beer's law of absorption. But the case is not the same always especially when we consider the emission of longwave radiation from the atmospheric layers due to the absorption of solar radiation by those layers. Using Kirchoff's law, The emitted energy is given by

$$E_{\lambda} = a_{\lambda} f(\lambda, T)$$

Hence the net change in the initial intensity will be given by

$$dI_{\lambda} = -a_{\lambda} I_{\lambda} + a_{\lambda} f(\lambda, T) = a_{\lambda} (f(\lambda, T) - I_{\lambda})$$

This is the Schwarzschild's equation or the equation of radiative transfer which takes care of the absorption as well as the resultant emission for which the radiative beam is subjected to while its passage through an atmospheric layer.

10.2. Solar Radiation

Radiation from the sun is the predominant source of energy for all the processes taking place on the planet earth. The source of this energy is the nuclear fusion process taking place in the earth's interior in which hydrogen is converted into helium. The energy getting produced in the interior travels to the surface of the sun by radiation and convection and from there it gets radiated into the space from the outer layers of the sun.

The luminous region of the sun is called the photosphere. It consists of hot gases under pressure and in various stages of ionization. Eventhough the photosphere appears very smooth, telescopic examination reveals various features of the photosphere. The darker and hence cooler regions appearing on the photosphere are called sunspots whereas the brighter and thus the hotter regions are called the flocculi. Above the photosphere lies the reversing layer which consists of the cooler gases. Absorption of the energy in this layer produces the Fraunhofer lines.

If the photosphere is blocked from sight during eclipse etc. layers lying further outside can be seen. First layer observed just outside which appears reddish is called the chromosphere. Above the chromosphere lies the outer layer of the sun known as corona. It is silvery white in colour and consists of extremely thin gases which extend outwards for millions of kilometres. The most spectacular activity of the chromosphere is the solar flares which is a short lived intense eruption during which

a large area of the region becomes several times brighter within a shorter period. It slowly subsides and within a few hours, the region returns to its original brightness.

10.3. Important features of solar radiation

The radiation emitted by the sun is mostly confined to the wavelength range between 0.15μ to 4.0μ . The spectral distribution of this energy is about 9% in the ultraviolet, 45 % in the visible and 46% in the infrared. The spectral distribution of solar energy is closely related to that of a black body at a temperature of about 6000°K .

In order to express the amount of energy received from the sun, a reference quantity called solar constant can be defined. It is the amount of energy received per unit area per unit time on a surface normal to the radiation which is situated at the outer limits of the atmosphere, if the earth is located at the mean distance from the sun (about 1.4968×10^8 kilometers). The approximate value of solar constant is $1.40 \times 10^3 \text{W/m}^2$.

Applying the laws of radiation it is possible to define the effective and the colour temperature of the sun which is different from its average temperature.

The effective temperature of the sun is obtained from the Stefan-Boltzmann law, by assuming the sun as a blackbody to determine the temperature it would have in order to conform to the experimentally determined value of solar constant.

According to Stefan-Boltzmann's law, the total energy emitted per second in all directions from the sun is equal to $4\pi R^2 \sigma T_e^4$ where R is the radius of the sun and T_e is the effective temperature. If r_M is the mean distance between the earth and the sun and S is the solar constant, the energy being received at the surface of a sphere concentric with the sun will be $4\pi r_M^2 S$.

Hence we have,

$$4\pi R^2 \sigma T_e^4 = 4\pi r_M^2 S.$$

$$T_e^4 = (r_M/R)^2 \times (S/\sigma)$$

The mean distance between the sun and the earth (r_M) = 1.4968×10^8 kilometers

Radius of the sun (R) = 6.96×10^5 kilometers

Solar constant (S) = 1.40×10^3 watts/square meters

Stefan Boltzmann constant (σ) = $5.67 \times 10^{-8} \text{J/m}^2/\text{K/Sec}$

Substituting these values, the effective temperature of the sun is obtained and it is about 5800°K .

The colour temperature of the sun is obtained by the application of Wein's displacement law.

According to Wein's displacement law, $\lambda_{\max} \times T = a$ constant where λ_{\max} is the wavelength corresponding to the maximum intensity. For solar radiation, $\lambda_{\max} = 0.474 \mu$. The value of the constant is $0.2898 \times 10^{-2} \text{ m } ^\circ \text{ K}$. Applying these values, we get the color temperature of the sun to be about $6100 ^\circ \text{ K}$.

10.4. Depletion of Solar Radiation

The amount of radiation received at the top of the atmosphere and the radiation received at the earth's surface are not the same since the solar radiation while passing through the atmosphere gets modified due to the processes like absorption, scattering, and reflection.

10.4.1. Absorption

The fine absorption lines and bands seen in the solar spectrum is due to the absorption of radiation by gases in the sun's atmosphere and also due to the absorption of radiation in the earth's atmosphere. If we consider the absorption in the earth's atmosphere, the most important one is the absorption of ultra violet radiation by ozone in the stratosphere. The major absorption bands for ozone are centered around 0.29μ . The amount of radiation absorbed by ozone is relatively very small and contribute to about 2% of the total radiation only. Other than ozone, water vapour and carbon dioxide also absorb solar radiation to a lesser extent and it is occurring in the near infrared region of the electromagnetic spectrum.

Dust particles, smoke and other aerosols also absorb solar radiation but its impact will vary depending upon the quantity as well as the properties of those particles.

10.4.2. Scattering

Scattering involves merely a change in direction of the propagation of the radiation when it comes into contact with scattering particles. There is no transformation of energy in scattering as it happens in case of absorption. However depletion of radiation occurs in the direction of incident radiation. This is due to the fact that some of the energy gets dispersed backwards and sideways.

The change in intensity (I) of radiation pertaining to a wavelength λ is given by

$dI_\lambda = -S_\lambda I_\lambda dx$ where S_λ is the scattering coefficient pertaining to the wavelength λ and dx is the path travelled by the beam of radiation.

For particles as small as air molecules this dispersion is known as **Rayleigh scattering**. In this case, the ratio of the radius of the scattering particle to the wave length of the incident radiation is small.

Scattering coefficient of Rayleigh scattering is given by

$S_\lambda = 32 \frac{\pi^3 (n_\lambda - 1)^2}{3\lambda^4 N}$ multiplied by the ratio of the actual density of dry air to the standard density of dry air.

Here N is the number of molecules for unit volume under standard conditions of pressure and temperature and n_λ is the refractive index at wavelength λ under standard conditions.

From this equation it is seen that in case of Rayleigh scattering the scattering coefficient is inversely proportional to the 4th power of the wave length. In the visible range the wave length of blue color is about 0.4 μ where as the wavelength of red color is about 0.8 μ . Due to that the scattering coefficient is very large for the blue color than for the red color. Hence the blue color scatters more than the red color. Due to this, the sky appears blue in color.

In case of rising and setting sun, sun is at a far away distance at the horizon from the observer. The shorter wavelengths of the visible radiation gets scattered once it passes through the atmosphere, but the red color will not get scattered that much. Hence rising and setting sun appears red in color.

For particles larger than air molecules such as dust particles the scattering coefficient is given by

$$S_{d\lambda} = \beta \lambda^{-\gamma}$$

Where β is a parameter proportional to the density of the scattering particle and γ is the parameter which goes on decreasing with increase in particle size. On the average the value of γ is about 1.3 (normal particle size) and if the air is highly polluted with larger particles, the value can even come down to 0.5 or less. This is applicable in case of dust storms or volcanic eruptions.

In case of very large particles such as those occurring in fog where the particle size is sufficiently large, the value of γ may be equal to zero also. Here the dispersal of the radiation is equally effective for all the wavelengths. Then it is known as diffuse reflection. Large dust particles, water droplets and ice crystals come under the category of particles which can cause diffuse reflection. The diffused reflection accounts for the white color of fog.

In case of haze or scattering from cloud, the ratio of the diameter of the scattering particle to the wave length will be of the order unity (one) or greater. This is referred to as **Mie scattering**.

The solar radiation coming from the solid angle of the sun's disc, received on a surface perpendicular to the axis of the solid angle is called direct solar radiation. The vertical component of the direct solar radiation as received on a horizontal surface is referred to as downward direct solar radiation.

The amount of solar radiation received on a horizontal surface from the solid angle 2π of the hemisphere of the sun facing the earth with the exception of the solid angle subtended by the sun's disc is known as diffuse solar radiation.

The downward direct solar radiation and the diffuse solar radiation as received together on a horizontal surface from the solid angle 2π of the sun is referred to as the global solar radiation.

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