

## Artificial satellites

An artificial satellite is a manufactured object that continuously orbits earth or some other body in space. Such a satellite is used to study the universe, help forecast the weather, transfer telephone calls over the oceans, assist in the navigation of ships and aircraft, monitor crops and other resources, and support military activities.

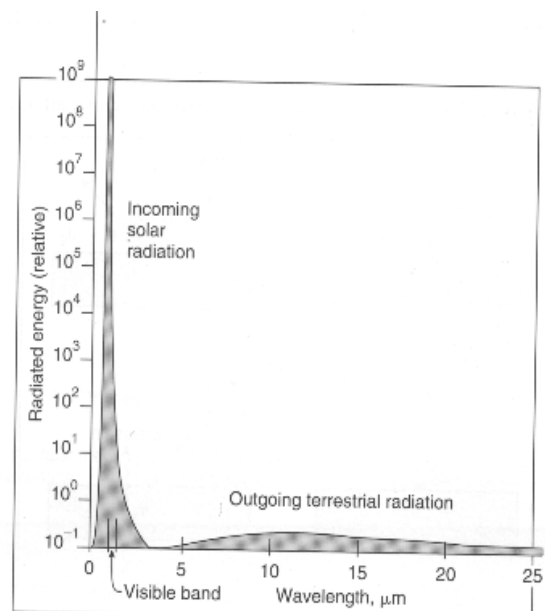
Following are the four main types of satellites (1) high altitude, geosynchronous; (2) medium altitude, (3) sun-synchronous, polar and (4) low altitude. Most orbits of these four types are circular.

1. A high altitude, geosynchronous orbit lies above the equator at an altitude of about 22,300 miles (35,900 kilometers). A satellite in this orbit travels around Earth's axis in exactly the same time, and in the same direction, as Earth rotates about its axis. Thus, as seen from Earth, the satellite always appears at the same place in the sky overhead. To boost a satellite into this orbit requires a large, powerful launch vehicle.
2. A medium altitude orbit has an altitude of about 12,400 miles (20,000 kilometers) and an orbital period of 12 hours. The orbit is outside Earth's atmosphere and is thus very stable. Radio signals sent from a satellite at medium altitude can be received over a large area of Earth's surface. The stability and wide coverage of the orbit make it ideal for navigation satellites.
3. A sun-synchronous, polar orbit has a fairly low altitude and passes almost directly over the North and South poles. A slow drift of the orbit's position is coordinated with Earth's movement around the sun in such a way that the satellite always crosses the equator at the same local time on Earth. Because the satellite flies over all latitudes, its instruments can gather information on almost the entire surface of Earth. One example of this type of orbit is that of the TERRA Earth Observing System's NOAA-H satellite. This satellite studies how natural cycles and human activities affect Earth's climate. The altitude of its orbit is 438 miles (705 kilometers), and the orbital period is 99 minutes.
4. A low altitude orbit is just above Earth's atmosphere, where there is almost no air to cause drag on the spacecraft and reduce its speed. Less energy is required to launch a satellite into this type of orbit than into any other orbit. Satellites that point toward deep space and provide scientific information generally operate in this type of orbit. The Hubble Space Telescope, for example, operates at an altitude of about 380 miles (610 kilometers), with an orbital period of 97 minutes.

## Remote sensing

Remote sensing refers to the science of identification and classification of Earth surface features using electromagnetic radiation as a medium of interaction. It derives information about objects from measurements made from a distance i.e. without actually coming into contact with them.

## Solar radiation and terrestrial radiation



Most remote sensing instruments are designed to detect solar radiation and terrestrial radiation. Electromagnetic radiation (EMR) emitted from sun passes through the atmosphere and is reflected in varying degrees by Earth's surface and atmosphere is detectable only during daylight. about half of solar radiation passes through the atmosphere and absorbed in varying degrees by surface. The graph shows the radiation emitted at different wavelengths.

Sun's visible surface (photosphere) has temperature - 6000K. Energy is radiated from gamma to radio waves. 99% of sun's radiation fall between 0.2 - 5.6 $\mu$ m; 80% - 0.4 - 1.5 $\mu$ m (visible and reflected infrared. Atmosphere is quite transparent to incoming solar radiation. The maximum radiation occurs 0.55 $\mu$ m (visible).

Terrestrial radiation corresponds to energy emitted from the Earth and atmosphere which is detectable both during day and night. The Earth's ambient temperature is 300K. Earth radiates 160,000 times less than the sun. All energy is radiated at (invisible) thermal infrared wavelengths between 4-25 $\mu$ m and maximum emission occurs at 9.7 $\mu$ m.

### EMR-Atmosphere Interactions

EMR travels through space without modification. Diversion and depletions occurs as solar and terrestrial radiation interact with earth's atmosphere. Interference is wavelength selective - meaning at certain wavelengths

Different remote sensing instruments are designed to operate within windows where cloudless atmosphere will transmit sufficient radiation for detection.

EMR interacts with atmosphere in different of ways. It is absorbed and reradiated at longer wavelengths (causes air temperature to rise).

It is reflected or scattered without change to either its velocity or wavelength. It is transmitted in straight-line path directly through the atmosphere.

The important atmospheric windows exploited in remote sensing are 0.3 - 1.1 $\mu$ m (UV, visible, near infrared), 1.5 - 1.8 $\mu$ m, 2.0 - 2.4 $\mu$ m, 3.0 - 5.0 $\mu$ m (Mid infrared), Mid infrared, 8.0 - 14.0 $\mu$ m (thermal infrared) >0.6cm (Microwave).

Most significant absorbers of EMR by the atmosphere are ozone, carbon dioxide, water vapor, oxygen and nitrogen. Absorption of atmospheric gases has maximum influence in wavelengths <0.3 $\mu$ m and minimum impact on wavelengths greater than 0.6cm.

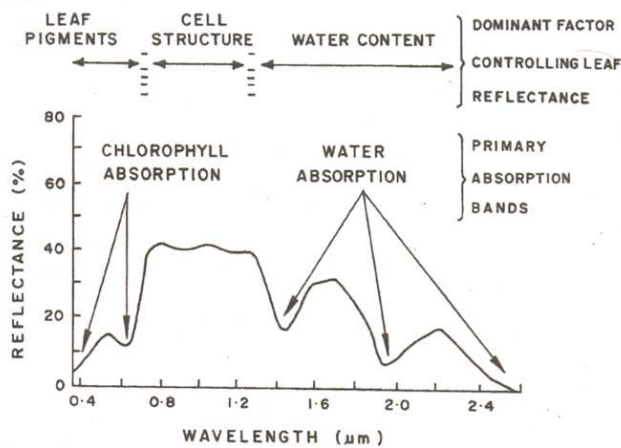
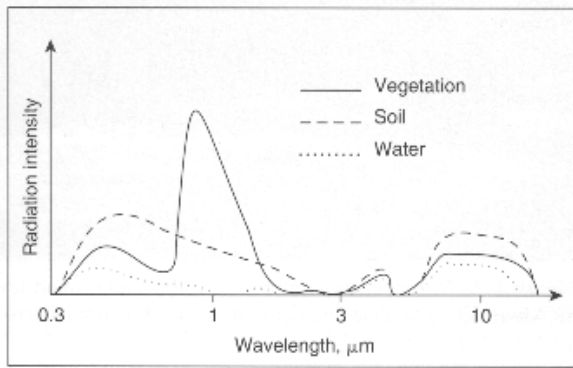
Atmospheric windows become less transparent when air is moist (high humidity). Clouds absorb most of long wave radiation emitted from Earth's surface.

### Spectral Signatures

Every natural and synthetic object reflects and emits EMR over a range of wavelengths in its own characteristic way according to its chemical composition and physical state.

Spectral signatures are the distinctive reflectance and emittance properties of objects/features and their conditions. Remote sensing depends upon operation in wavelength regions where detectable differences in reflected and emitted radiation occur.

**Spectral Response of Some Natural Earth Surface Features**



**Vegetation:** The spectral reflectance of vegetation (Figure 2) is quite distinct. Plant pigments, leaf structure and total water content are the three important factors which influence the spectrum in the visible, near IR and middle IR wavelength regions, respectively.

Low reflectance in the blue and red regions corresponds to two chlorophyll absorption bands centered at 0.45 and 0.65  $\mu\text{m}$ , respectively. A relative lack of absorption in the green region allows normal vegetation to look green to ones eyes. In the near infrared, there is high transmittance of similar magnitude and absorptance of only about five per cent. This is essentially controlled by the internal cellular structure of the leaves. As the leaves grow intercellular air spaces increase and so the reflectance increases.

As vegetation becomes stressed chlorophyll absorption decreases, red reflectance increases and also there is a decrease in intercellular air spaces, decreasing the reflectance in the near IR.

**Soil** Typical soil reflectance curve shows a generally increasing trend with wavelength in the visible and near IR regions. Some of the parameters which influence soil reflectance are the moisture content, the amount of organic matter, iron oxide, relative percentage of clay, silt and sand, and the roughness of the soil surfaces. As the moisture content of the soil increases, the reflectance decreases and more significantly at the water absorption bands.

In a thermal IR image moist soils look darker compared to the dry soils. In view of the large differences in dielectric constant of water and soil at microwave frequencies, quantification of soil moisture becomes possible.

**Water:** Water absorbs most of the radiation in the near IR and middle IR regions. This property enables easy delineation of even small water bodies. In the visible region, the reflectance depends upon the reflectance that occurs from the water surface, bottom material and other suspended materials present in the water column.

Turbidity in water generally leads to increase in its reflectance and the reflectance peak shifts towards longer wavelength. Increase in the chlorophyll concentration leads to greater absorption in the blue and red regions. Dissolved gases and many inorganic salts do not manifest any changes in the spectral response of water.

### Remote Sensing (RS) Applications

1. RS data have been in agriculture, forestry, water resources, landuse, urban sprawl, geology, environment, coastal zone, marine resources, snow and glaciers, etc. Multispectral RS data facilitate identification of crops, estimating their area, inferring the possible yield and hence production forecasts before harvest. Extent of forest cover, their types and density are also detectable.
2. Identification of surface water bodies and their spread, mapping ground water prospect zones, estimating snow cover and possible melt, retreat of glaciers over time period, sedimentation in reservoirs are some applications related to water resources that are routinely being done.
3. Identification of wastelands their status to facilitate their development is another use. Monitoring urban sprawl, preparing development plans and studying impact of urbanisation are some other examples.
4. RS data help in identifying and monitoring flood inundated areas, consequent damage assessment and in relief measures.
5. Monitoring drought and its severity assessment through indicators derived from RS data is another example.
6. Preparation of landslide hazard maps, forest fire prone areas, has been greatly benefited from RS data.
7. Use of thermal data helps monitoring impending volcanic eruption.
8. Ability of microwave radar data to see through clouds has been particularly useful in flood and cyclone related studies since they are generally accompanied by cloud cover.
9. Detection and monitoring of oil spills in oceans is another application.

### Evolution of Remote Sensing in India

Launching of two experimental Earth observation satellites, Bhaskara I and II in 1979 and 1981 carrying both optical and microwave payloads provided the initial thrust to remote sensing in India.

Bhaskara II, weighing 436kg, had a resolution of 1km. Design, development and successful launching of Indian remote sensing satellites (IRS-IA and IB) carrying state-of-the-art

sensors providing data at two different spatial resolutions but in identical spectral bands brought Indian remote sensing to international forefront

The IRS-1A (Indian Remote sensing Satellite), weighing 980kg, was launched from soviet union into polar sunsynchronous orbit of 904km. It had a period of 103 minutes making 14 orbits around the earth a day and covering Indian subcontinent in 22 days.

IRS-IC/ ID carrying a unique set of three sensors, viz., WiFS, LISS-III and PAN camera have made significant impact on the remote sensing scene internationally. Till recently, PAN camera providing 5.8 m resolution data was the highest spatial resolution offered in civilian domain. While, IRS-IA/ IB/ P2/ IC/1D addressed data needs of land observations, IRS-P3 and IRS-P4 carrying ocean colour sensors and microwave radiometer (IRS- P4) provided measurements for the study of ocean biology and atmosphere.

### Composition of Earth's atmosphere

The air we breathe is actually a combination of several gases. Within the first 80 km or so in altitude, the composition is quite constant. This portion is known as the homosphere. (Higher in elevation, the concentrations of each gas vary considerably. The table below shows the composition of the atmosphere (by volume) within the first 80 km.

Constituent	Percent by Volume (Dry Air)
Nitrogen (N <sub>2</sub> )	78.084%
Oxygen (O <sub>2</sub> )	20.946%
Argon (A)	0.934%
Carbon Dioxide (CO <sub>2</sub> )	0.0332%
Trace Constituents	0.0028%

The most significant constituent, Nitrogen, is held at a constant level through the nitrogen cycle. Biological processes remove nitrogen from the atmosphere while decaying matter releases nitrogen. Volcanic eruptions also release nitrogen into the atmosphere. Since nitrogen is chemically stable, it doesn't need to interact with other gases. Meaning that the nitrogen released billions of years ago is still around. That is the reason that nitrogen makes up most of the atmosphere.

The second most abundant gas, Oxygen, is necessary for our survival. Its consumption by animals, decaying matter, and other chemical reactions (Combustion, Oxidation, etc.) is balanced by its production by *photosynthesis*. In the photosynthesis reaction, plants absorb carbon dioxide and release oxygen. Oxygen is a result of one celled organisms. Three billion years ago their byproduct of photosynthesis was oxygen today's oxygen is the result of that build up.

Despite its small contribution to the total volume of air, the amount of carbon dioxide in the atmosphere is critical. Carbon dioxide is a good absorber of infrared radiation emitted from the earth's surface. Its rise in concentration has been blamed for the global warming phenomenon. Animals and the burning of fossil fuels emit carbon dioxide.

The above chart also shows that there are a few trace constituents in the atmosphere. These trace gases include Neon, Helium, Krypton, Hydrogen, Xenon, Methane, Nitrous Oxide, Ozone, and Chlorofluorocarbons (CFCs).

One extremely important gas in the atmosphere is water vapour. Water vapour is highly variable in the atmosphere, ranging from nearly 0% to as much as 4%. This large range can easily be felt by our bodies. (For instance, compare the extremely dry air on a cold, crisp winter's day to the moist, humid, and hazy air on a July afternoon). The amount of water vapor in the air at any given time has a tremendous effect on the weather (such as aiding the development of a thunderstorm!)

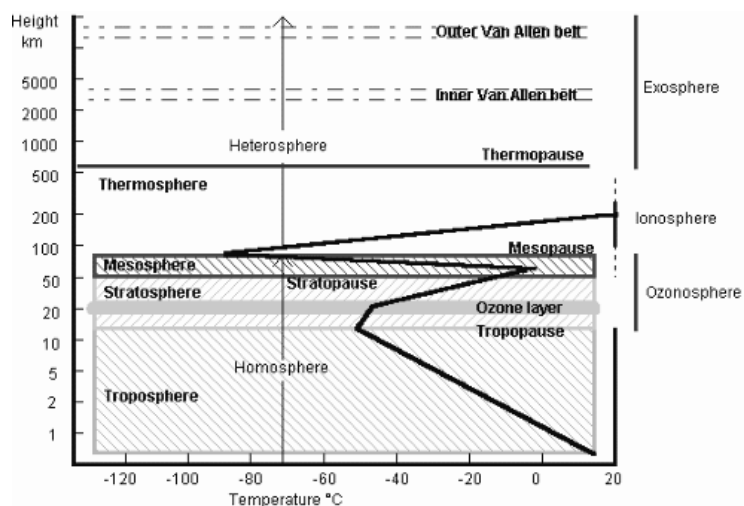
And of course, the air contains thousands (per cubic centimeter) of small suspended particles and pollutants called aerosols. These can be anything from sea salt to smoke particles. Some of these aerosols serve as Cloud Condensation Nuclei (CCN) and encourage the development of clouds and haze.

### Vertical structure of the atmosphere

The changes in the atmosphere with height are results of specific physical conditions which exist on the earth and in its atmosphere. The vertical changes in temperature are important in constraining weather events to the lowest 10-12 km of the atmosphere. The [ozone layer](#), located near 25 km above the earth's surface, causes the temperature to rapidly change in the middle atmosphere. The different regions of the earth's atmosphere are as follows.

#### 1. Troposphere

The troposphere is characterized by turbulent mixing and overturning. This turbulence results from uneven heating of the surface and the atmosphere. Temperature decreases with height in this layer from 20°C to -52°C. This temperature decrease is known as the environmental lapse rate and averages 6.5 °C/km. This layer extends from the surface up to an average altitude of 11 km. This altitude can range as high as 16 km in the tropics to less than 9 km over the poles. This range is due to the temperature differences between the tropics and poles. The warm surface temperatures and turbulent mixing over the tropics help to push the troposphere's boundary upward. The tropopause extends above the troposphere.



#### 2. Stratosphere

The stratosphere lies above the tropopause, extending to a height of about 50 km. The stratosphere is characterized by a strong temperature inversion. The temperature increases from -52°C to -3°C. This is a stable layer with little mixing. As a result, pollutants and other particles may reside in the stratosphere for many years.

A large concentration of ozone ( $O_3$ ) is found in the stratosphere, with a maximum concentration at about 25 km. This "ozone layer" absorbs much of the ultraviolet radiation emitted by the sun. Heat is released as the UV is absorbed, which then heats the atmosphere. This explains why temperature increases with height in this layer. It is interesting to note that the inversion continues beyond the maximum density of ozone. The reason for this is that much of the available UV radiation is already absorbed by ozone above the level of maximum density. Therefore, there is less potential UV that can be absorbed, resulting in lower temperatures near the bottom of the stratosphere. The stratopause lies above the stratosphere.

### 3. Mesosphere

The mesosphere lies above the stratosphere and extends to an altitude of about 85 km. This layer is often referred to as the cold layer, as the lowest readings in the atmosphere are found here. Temperature decreases with height, reaching a minimum average value of  $-90\text{ }^\circ\text{C}$  at the top of the layer. The upper part of the mesosphere contains part of the ionosphere, an electrified region. The mesopause lies above the mesosphere, separating it from the thermosphere.

### 4. Thermosphere

The thermosphere is often considered the "hot layer" because it contains the warmest temperatures in the atmosphere. Temperature increases with height until the estimated top of the thermosphere at 500 km. Temperatures can reach as high as 2000 K or  $1727\text{ }^\circ\text{C}$  in this layer. The air is so thin that a molecule will travel 1 kilometer before striking another molecule. At the top of the thermosphere, molecules will move 10 km before striking another molecule. At this height, many lighter molecules have attained enough velocity to escape earth's gravity into outer space. The region where air molecules escape is known as the exosphere.

## Atmospheric waves

An atmospheric wave is a periodic disturbance in the fields of [atmospheric](#) variables (like [surface pressure](#) or [geopotential height](#), [temperature](#), or [wind velocity](#)) which may either propagate (traveling wave) or not (stationary wave).

Atmospheric waves range from large-scale planetary waves ([Rossby waves](#)) to minute [sound waves](#). Atmospheric waves with periods which are [harmonics](#) of 1 [solar day](#) (e.g. 24 hours, 12 hours, 8 hours... etc) are known as atmospheric tides.

### Causes and effects

The mechanism for the forcing of the wave, i.e. the generation of the initial or prolonged disturbance in the atmospheric variables, can vary. Generally, waves are either excited by [heating](#) or [dynamic](#) effects. Heating effects can be small-scale (like the generation of [gravity waves](#) by [convection](#)) or large-scale (the formation of [Rossby waves](#) by the temperature contrasts between continents and oceans in the [Northern hemisphere](#) winter).

Atmospheric waves transport [momentum](#), which is fed back into the background flow as the wave dissipates. This wave forcing of the flow is particularly important in the [stratosphere](#),

where this momentum deposition by gravity waves gives rise to [sudden stratospheric warmings](#) and the [quasi-biennial oscillation](#).

### Types of waves

The different wave types are [sound waves](#), [gravity waves](#) and [Rossby waves](#). At the equator, mixed Rossby-gravity and [Kelvin waves](#) can also be observed.

**Sound wave** is a disturbance of [mechanical energy](#) that propagates through [matter](#) as a [wave](#). Sound is characterized by the [properties of sound waves](#), which are [frequency](#), [wavelength](#), [period](#), [amplitude](#) and velocity or [speed](#). Sound propagates as [waves](#) of alternating [pressure](#), causing local regions of [compression](#) and [rarefaction](#). Particles in the medium are displaced by the wave and oscillate.

### Gravity wave

These waves are generated in a [fluid](#) medium or on an [interface](#) (e.g. the atmosphere or ocean) and having a restoring force of [gravity](#) or [buoyancy](#). When a [fluid parcel](#) is displaced on an interface or internally to a region with a different [density](#), gravity restores the parcel towards equilibrium resulting in an [oscillation](#) about the [equilibrium state](#). Gravity waves on an air-sea interface are called surface gravity waves or [surface waves](#) while internal gravity waves are called [internal waves](#).

In the earth's [atmosphere](#), gravity waves are important for transferring [momentum](#) from the [troposphere](#) to the [mesosphere](#). Gravity waves are generated in the troposphere by [frontal systems](#) or by airflow over [mountains](#). At first waves propagate through the atmosphere without affecting its [mean velocity](#). But as the waves reach more rarefied air at higher [altitudes](#), their [amplitude](#) increases, and [nonlinear effects](#) cause the waves to break, transferring their momentum to the mean flow.

This [process](#) plays a key role in controlling the [dynamics](#) of the middle atmosphere.

### Rossby (or planetary) waves

These waves are large-scale motions in the [ocean](#) or [atmosphere](#) whose restoring force is the variation in [Coriolis effect](#) with [latitude](#). The waves were first identified in the atmosphere in [1939](#) by [Carl-Gustaf Arvid Rossby](#) who went on to explain their motion.

The special identifying feature of the Rossby waves is its [phase velocity](#) (that of the wave crests) always has a westward component. However, the wave's [group velocity](#) (associated with the energy flux) can be in any direction. In general: shorter waves have an eastward group velocity and long waves a westward group velocity.

Barotropic" and "baroclinic" Rossby waves are defined to distinguish their vertical structure. Barotropic Rossby waves do not vary in the vertical, and have the fastest propagation speeds. The baroclinic wave modes are slower, with speeds of only a few centimetres per second or less.

Rossby waves in the atmosphere are easy to observe as large-scale meanders of the [jet stream](#). When these loops become very pronounced, they detach the masses of cold, or warm, air that become [cyclones](#) and [anticyclones](#) and are responsible for day-to-day weather patterns at mid-latitudes.

Oceanic Rossby waves are thought to communicate climatic changes due to variability in forcing, due to both the [wind](#) and [buoyancy](#). Both barotropic and baroclinic waves cause



variations of the sea surface height, although the length of the waves made them difficult to detect until the advent of satellite altimetry.

Baroclinic waves also generate significant displacements of the oceanic [thermocline](#), often of tens of meters. Satellite observations have revealed the stately progression of Rossby waves across all the ocean basins, particularly at low- and mid-latitudes. These waves can take months or even years to cross a basin like the Pacific.

### Numerical weather prediction

Numerical modelling is the process of obtaining an objective forecast of the future state of the atmosphere by solving a set of equations that describe the evolution of variables (temperature, wind speed, humidity, pressure) that define the state of the atmosphere.

The process begins with analyzing the current state of the atmosphere by taking a previous short range forecast and using observations to amend this forecast so that the best guess of the current true state of the atmosphere is obtained. A computer model is then run to produce a forecast

To produce an accurate weather forecast, precise knowledge of the current state of the atmosphere (the 'initial conditions') is needed. This is achieved by using observations and assimilating those observations into the model. Many thousand observations are received each day from a variety of observing types e.g. satellites, aircraft, ships, buoys, radiosondes and land stations.

Various atmospheric parameters are routinely measured including temperature, wind speed and direction and humidity. Observations are assimilated into the model using a process known as variational analysis.

All numerical models of the atmosphere are based upon the same set of [governing equations](#) which are described here in non-mathematical terms.

**Newton's second Law of Motion** states that the acceleration of a particle is equal to the vector sum of forces acting upon that particle. It is a statement of the Conservation of Momentum principle. The main forces in the atmosphere are the force that acts on air due to differences in pressure and the Coriolis Force. The Coriolis Force (or acceleration) is an apparent acceleration that air possesses by virtue of the earth's rotation. If an air parcel is moving between two points then its path, relative to the surface of the earth, will not be straight but will be curved. The curve will be towards the right in the northern hemisphere and to the left in the southern hemisphere.

**The hydrostatic equation** is an expression relating the variation in pressure with height. In the vertical the two main forces are gravity and the pressure gradient. In fact, the gravitational force is almost exactly balanced by the pressure gradient force i.e. the air's buoyancy, a condition known as hydrostatic equilibrium. The vertical component of the Coriolis Force, although comparable in magnitude with the horizontal components, is negligible when compared against the gravitational and pressure gradient forces. For this reason it is often ignored but our model does include it although in practice it is only significant in regions of strong vertical motion.

**The first Law of Thermodynamics** may be stated as the amount of heat added to a system is exactly balanced by the work done in increasing its volume and the change in internal energy.



$$\therefore \rho = \frac{P}{R_d T} \left[ 1 - 0.378 \left( \frac{P_w}{P} \right) \right] \quad \text{Also } m_w = 18 \text{ and } m_d = 28.9$$

$$\text{or } P = \frac{\rho R_d T}{\left[ 1 - 0.378 \left( \frac{P_w}{P} \right) \right]} = \rho R_d T_v$$

$$\therefore \boxed{P = \rho R_d T_v} \text{ where } T_v = \frac{T}{\left[ 1 - 0.378 \left( \frac{P_w}{P} \right) \right]}$$

$T_v$  is called the virtual temperature. It represents the temperature to which dry air has to be raised in order that it may have the same density as that of moist air at the same pressure.

The density of moist air is less than that of dry air. Thus virtual temperature is always more than the actual temperature.

The relation connecting the quantities  $P, \rho, T$  and  $P_w$  is the equation of state  $P = \rho R_d T_v$ .

The equation of state for moist air can also be written as  $P = \rho R_d T [1 + 0.61q]$ .

where  $q$  is the specific humidity defined as the ratio of the mass of water vapour to the mass of moist air for the same volume of air.

**Hydrostatic Balance:** In still air in which the atmospheric motions are absent, the force of gravity must be exactly balanced by the vertical component of the force due to the pressure gradient.

The pressure at the surface of the earth (atmospheric pressure).

$$P = \frac{F}{A} = \frac{Mg}{4\pi R^2} \quad M \rightarrow \text{average mass of the atmosphere} = 5 \times 10^{18} \text{ kg}$$

$$P = 1.013 \times 10^5 \text{ Nm}^{-2} \quad R \rightarrow \text{radius of the earth} = 6370 \text{ km}$$

With increase of altitude,  $P$  goes on decreasing. Consider a small height of air column  $dZ$  at a height  $Z$  from the surface of the earth. If  $A$  is its area, then

$$dF = \Delta Mg = (\rho A dZ) g$$

$$\text{The decrease in pressure } dp = -\frac{dF}{A} = -\frac{\rho A dZ g}{A}$$

$$\therefore dP = -\rho g dZ$$

$$\text{or } \boxed{\frac{dP}{dZ} = -\rho g} \text{ This represents the condition for hydrostatic balance.}$$

Integrating the above equation between limits  $Z$  and  $\infty$ , the pressure at any point at a height  $Z$  is  $P(Z) = \int_Z^{\infty} \rho g dZ$ .

**Static stability:** A system in atmosphere (part of atmosphere) that does not mix with the surrounding air when it moves is called a **parcel**.

We assume the surrounding air to be in hydrostatic equilibrium and heat conduction in air is comparatively low.

A parcel in motion along the vertical direction tends to return to its original position, then its equilibrium is said to be stable.

If the parcel accelerates and doesn't return to its original position, the equilibrium is said to be **unstable**.

If the net force on the parcel is zero, then the equilibrium is said to be **neutral**.

**Heat Balance of the Atmosphere:**

The diagram below represents the radiation balance for the earth. The mean temperature of the earth almost remains constant.

Let the incoming solar radiation (insolation) = 100 units (1)

16 units are absorbed by ozone of the atmosphere, water vapour in the troposphere and aerosols ----- (5)

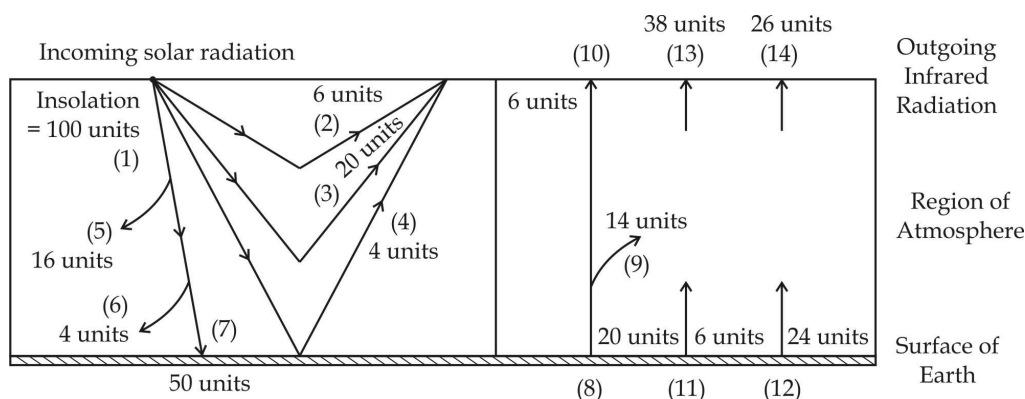
4 units are absorbed by the clouds ----- (6)

6 units are scattered by the air back to space ----- (2)

20 units are reflected by the clouds ----- (3)

4 units are reflected by earth's surface ----- (4)

The balance of 50 units are absorbed by the surface of earth ----- (7)



Of the 50 units absorbed by the earth, 20 units are emitted as long wave IR radiation into atmosphere ----- (8)

Of this 20 units, 14 units are absorbed by  $H_2O, CO_2$  in the atmosphere ----- (9)

Remaining 6 units are transmitted to space ----- (10)

The remaining 30 units are transferred upwards into atmosphere by convection and turbulent processes ----- (11) and (12)

Of this 6 units will be in the form of sensible heat ----- (11)

And 24 units in the form of latent heat ----- (12)

If only atmosphere is considered it absorbs 20 units [(5) + (6)] from sun and 44 units [(9) + (11) + (12)] of energy from the earth's surface. Thus the total of 64 units of this energy is balanced by the IR emitted into space by 38 units [(13)] emitted into space by water vapour and  $CO_2$  and 26 units [(14)] emitted by clouds.

Thus the total of 64 units together with 6 units passing through the atmosphere amounts to a total of 70 units. This is balanced with the absorption of radiation from the sun. Thus the amount of radiation received by the atmosphere and the surface of the earth is reflected back

to space. Thus over all there is heat balance and the temperature of the atmosphere remains constant.

**Greenhouse effect:** The incoming solar radiation passes through the atmosphere and absorbed by the earth's surface. The surface of earth reemits this radiation in the form of Infra red radiation. The constituents of the atmosphere like  $CO_2$ , water vapour and ozone and also clouds reradiate this IR radiation towards earth's surface. Thus after a hot day the temperature is prevented by this absorption from falling rapidly. The radiation gets trapped. This increases the temperature. This effect is called the green house effect.

**ATMOSPHERIC DYNAMICS**

**(1) Equation of Motion:**

Consider a closed surface  $S$  enclosing a volume  $V$  of a non viscous fluid. Let the density of the fluid be  $\rho$  and its velocity  $q$ . (Fluid - parcel - a region of atmosphere (air) in motion).

The mass of the fluid remains unchanged throughout motion i.e., mass = constant or

$$\frac{d}{dt}(\rho dV) = 0 \dots\dots (1) \quad \left[ \rho = \frac{m}{dV}, m = \rho dV \right].$$

The total momentum of the volume  $V$  of the fluid is  $M = \int_V q \rho dV$

[ $\because$  momentum = mass  $\times$  velocity]

Differentiating with respect to time  $\frac{dM}{dt} = \int \left[ \frac{dq}{dt} \rho dV + q \frac{d}{dt}(\rho dV) \right]$  ----- (2)

Putting the condition of (1) in (2) we get,

$$\frac{dM}{dt} = \int \frac{dq}{dt} \rho dV \text{ ----- (3)}$$

Let  $\hat{n}$  be the unit outward normal vector on the surface element  $ds$ . Let  $F$  be the external force per unit mass and  $P$  be the pressure on the area  $ds$ .

The total surface force acting on  $S$  is  $F_T$ .

$$F_T = \int_V F \rho dV + \int_S P(-\hat{n}) ds \quad \text{[ - ve sign is due to the normal inward pressure]}$$

From Gauss's theorem  $\int_S P \hat{n} ds = \int_V \nabla P dV$ .

$$\therefore F_T = \int_V (F \rho - \nabla P) dV \text{ ----- (4)}$$

From Newton's second law of motion, rate of change of momentum = total applied force. Thus comparing (3) and (4)

$$\int \frac{dq}{dt} \rho dV = \int (F \rho - \nabla P) dV \quad \text{or} \quad \int \left[ \frac{dw}{dt} \rho - F \rho + \nabla P \right] dV = 0$$

Both  $S$  and  $V$  are arbitrary. Thus

$$\frac{dq}{dt} \rho - F \rho + \nabla P = 0$$

Dividing by  $\rho$

$$\boxed{\frac{dq}{dt} = F - \frac{1}{\rho} \nabla P} \text{ ----- (5)}$$

Equation (5) is called the Euler's equation of motion.

To get Cartesian equivalent of above equation, consider

$$\vec{q} = (u\hat{i} + v\hat{j} + w\hat{k}), \vec{F} = (X\hat{i} + Y\hat{j} + Z\hat{k}), \vec{\nabla} = \hat{i}\frac{\partial}{\partial x} + \hat{j}\frac{\partial}{\partial y} + \hat{k}\frac{\partial}{\partial z}$$

Simplifying and equating coefficients of unit vectors on either side

$$\frac{du}{dt} = X - \frac{1}{\rho} \frac{\partial p}{\partial x}, \frac{dv}{dt} = Y - \frac{1}{\rho} \frac{\partial p}{\partial y}, \frac{dw}{dt} = Z - \frac{1}{\rho} \frac{\partial p}{\partial z}$$

The general equation of motion of the parcel is,

$$\frac{dq}{dt} = F - \frac{1}{\rho} \nabla P + F_g - 2(\omega \times v')m,$$

where  $F \rightarrow$  Force per unit mass frictional force

$\frac{1}{\rho} \nabla P \rightarrow$  Force due to pressure gradient

$F_g \rightarrow$  Force due to gravity

$2(\omega \times v')m \rightarrow$  Coriolis Force

**(2) Equation of continuity:**

For a given mass  $\rho m = \rho dV$ .

From law of conservation of mass  $\frac{d}{dt}(\rho \delta V) = 0$

$$\text{or } \left(\frac{d\rho}{dt}\right)\delta V + \rho\left(\frac{d}{dt}(\Delta V)\right) = 0$$

dividing throughout by  $\rho \delta V$

$$\frac{1}{\rho} \frac{d\rho}{dt} + \frac{1}{\delta V} \frac{d}{dt}(\delta V) = 0$$

The second term on the LHS of above equation is the three dimensional divergence of the wind velocity vector  $\vec{c}$ .

$$\text{Thus, } \boxed{\frac{1}{\rho} \frac{d\rho}{dt} + \text{div } \vec{c} = 0}$$

It is called equation of continuity.

$\frac{1}{\rho} \frac{d\rho}{dt} = -\text{div } \vec{c}$  i.e., time derivative in the equation expresses the rate of change of the density

with respect to an observer moving with wind velocity.

**(3) Equation of state:**

The relationship between the thermodynamic variables pressure  $P$ , density  $\rho$  and temperature  $T$  is called the equation of state.

It is represented as  $\frac{P}{\rho RT} = 1 + A(T)\rho + B(T)\rho^2 + \dots$ .

where  $A(T), B(T)$  are functions of temperature only and  $R$  is the gas constant. If we neglect functions  $A(T), B(T)$  as a first approximation.

$$\frac{P}{\rho RT} = 1 \text{ or } \boxed{P = \rho RT}$$

It applies only to an ideal gas. By taking into account the volume occupied by the gas molecules and the attractive force between the molecules, the equation of state can be written as  $(P + a)(V - b) = nRT$ .

where  $a$  is a factor that reflects the decreased pressure on the walls of the container as a result of attractive forces between molecules and  $b$  is the volume occupied by the molecules themselves when the pressure is infinitely large.

**(4) Hydrostatic equation:**

The atmosphere of the earth is in the earth's gravitational field. Thus its density decreases with altitude. Vertical motion of atmosphere is generally very small. The force of gravity on a parcel is balanced by the vertical component of force due to pressure gradient. If  $\rho$  is the density and  $P$  the pressure at an altitude  $Z$  measured vertically upwards from the surface of the earth then

$$dP = -g\rho dZ \text{ ----- (1)}$$

For a perfect gas of molecular weight  $M$  at a temperature  $T$ , the equation of state is

$$P = \frac{\rho RT}{M} \text{ ----- (2)} \quad \left[ \because \frac{P}{\rho} = \frac{RT}{M} \right] \text{ } R \text{ is the gas constant per mole}$$

Dividing (1) by (2)

$$\frac{dP}{P} = -\frac{g\rho dZ}{\rho \frac{RT}{M}} = -\left(\frac{Mg}{RT}\right) dZ$$

$$\text{or } \frac{dP}{P} = -\frac{dZ}{H} \text{ ----- (3)}$$

where  $H = \frac{RT}{Mg}$  called the scale height.

Integrating equation (3)

$$\int_{P_0}^P \frac{dP}{P} = -\int_0^Z \frac{dZ}{H} \text{ or } \log \frac{P}{P_0} = -\int_0^Z \frac{dZ}{H}$$

$$\text{or } \boxed{P = P_0 \exp \left[ -\int_0^Z \frac{dZ}{H} \right]} \text{ ----- (4)}$$

where  $P_0$  is the pressure at  $Z=0$ . The quantity  $H$  called the scale height is the increase in altitude necessary to reduce the pressure to  $\left(\frac{1}{e}\right)$  times its original pressure.

The value of  $H = 8.5 \text{ km}$  altitude at  $T = 288 \text{ K}$  is required to reduce the pressure to  $\frac{1}{e} = \frac{1}{2.718} = 0.38$  or 38% of the original pressure.

**(5) First law of thermodynamics:**

Consider the vertical motion of a parcel, (a small quantity of atmosphere) at a pressure  $P$ , temperature  $T$  and specific volume  $V$ .

Also the atmosphere is assumed to be in hydrostatic equilibrium.

From the first law of thermodynamics applied to unit mass of the parcel, we can write

$$dQ = C_v dT + P dV \text{ ----- (1)}$$

[ $\because dQ = dU + dW$   $dQ \rightarrow$  heat absorbed,  $dU \rightarrow$  increase in internal energy ( $C_v dT$ ) and  $dW = P dV \rightarrow$  work done]

$C_v$  is the specific heat at constant volume.

From the equation of perfect gas,

$$PV = \frac{RT}{M} \text{ differentiating } PdV + VdP + R \frac{dT}{M}$$

$$\text{or } PdV = \frac{RdT}{M} - VdP = (C_p - C_v) dT - V dP \text{ ----- (2)}$$

where  $R/M = (C_p - C_v)$  Mayer's equation  $C_p$  is specific heat at constant pressure.

Substituting for  $P dV$  from (2) in (1),

$$dQ = C_v dT + (C_p - C_v) dT - V dP$$

$$\text{or } dQ = C_v dT + C_p dT - C_v dT - V dP$$

$$dQ = C_p dT - V dP$$

For an adiabatic change  $dQ = 0$  [Atmosphere is a bad conductor of heat]

$$\therefore C_p dT - V dP = 0$$

$$\text{or } C_p dT = V dP$$

From the hydrostatic relation  $dP = -g \rho dZ$ .

$$\therefore C_p dT = V (-g \rho dZ) = -g V \rho dZ \quad \left[ \because \rho = \frac{1}{V} \right]$$

$$C_p dT = -g dZ \quad (\text{unit mass})$$

dividing the above equation by  $C_p dZ$

$$\text{we get, } \boxed{\frac{dT}{dZ} = -\frac{g}{C_p}}$$



Thus, the atmosphere is heated by contact with the earth's surface and there is a vertical motion of the atmosphere with a uniform temperature gradient of  $\frac{g}{C_p} = \frac{dT}{dZ} = 10 \text{ km}^{-1}$ .