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UTTAR PRADESH RAJARSHI TANDON OPEN UNIVERSITY

(Established by ... of 1999)

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PGD-ESD-04
Understanding the Environment

FIRST BLOCK : Introduction to the Environment
SECOND BLOCK : The Atmosphere
THIRD BLOCK : Hydrosphere

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PGD-ESD-04

UNDERSTANDING THE ENVIRONMENT

BLOCK 1

INTRODUCTION TO THE ENVIRONMENT

Unit 1
Introduction to the Environment

Contents

Introduction

The first step towards understanding the environment is knowing the definition of the environment. This includes recognizing the concept of the environment, defining its limits and sizes and identifying the types of environments in nature. The natural world where we all live on planet earth is confined to a very thin sphere around the globe. This region is known as the biosphere. When observing nature at work, naturalists have recognized distinct functional units of living organisms in the biosphere. The ecologists refer to these functional units as ecosystems.

1.1 Concept of the environment

Everyone talks about the environment. But what do we really mean when we talk about the environment.? Reference to dictionaries indicate different meanings. According to Merriam Webster (1974) the environment is defined in four different ways as given below:

The environment is

- the objects and circumstances by which one is surrounded
- the complex climatic, edaphic and biotic factors that act upon organisms or an ecological community and ultimately define its form and survival.
- an aggregation of social cultural conditions that influence the life of an individual or a community
- an artistic or theoretical work that involves or encompasses the spectator.

From the above, it is apparent that the meaning of the term environment is ambiguous and imprecise. It varies with circumstances in which the individual defining it exists.

In very broad terms, we can say the environment of a living organism is its "surrounding'. In general it considered to include the following:

- Life supporting physical factors, like landscape moisture, temperature, soil texture and the circumstance deriving from them like the climate and the fertility of soil.
- Biological factors arising from interactions with other organisms like competition, predation etc.

However, in referring to the concept of environment, the general practice has been to refer to it in relation to man. Hence in addition to the life supporting physical and biological factors, the environment will include social and cultural norms set by human society.

The concept of the environment has changed over the years depending on the human perceptions and concern about environmental problems and environmental quality. Human society has always accepted that the environment is critical for human life. In the years before 1950s environmental problems were almost non existent. Hence to people in those days the environment were the conditions at home or in their work places. In the years that followed the circumstances under which human society existed changed. For eg. a series of catastrophes like congenital deformation in babies due to thalidomide, Torrey Canyon spilling of oil along France's picturesque northern coast, death of fish and other organisms in thousands in Swedish lakes due to long range air pollution (acid rain) from western Europe, occurred. These together with the publication of Rachel Carson's Silent Spring and Garret Hardins's Tragedy of commons led to fresh concern about the balance between human life and the environment in the world over.

By the end of 1960, the concept of the environment among human society evolved to include the complex interactions between man's activities and all components of the natural environment. People were considered as root causes of problems related to destruction of wildlife, land and soil degradation, water pollution etc. At the Stockholm conference on Human and Environment in 1972, the state of the environment was reviewed and steps were taken to mitigate pollution and destruction of natural resources. A major step forward in this direction was taken in 1984, when United Nations Environmental program (UNEP) jointly with World Wide Fund for Nature and the International Union for Conservation for Nature published the "World Conservation Strategy". This document was effectively a prospectus for environmental conservation with a sharp focus on the need to protect nature and natural resources. However the strategy largely ignored the question of development and the need for food and other resources for humans.

In the 1980s poverty became a particular challenge as population growth in the developing world. In 1984 a leak from Union Carbide plant left 3000 people dead and 20,000 injured in Bhopal, India. In the same year up to one million people starved to death in Ethiopia. In 1986, the worlds worst nuclear accident happened as a reactor at the Chernobyl nuclear plant exploded in Ukraine, Republic of Soviet Union. Beside these, the British researchers located a hole in the ozone layer for the first time in 1985. They also found and that species extinction was threatening biodiversity.

With the above problems it was apparent that the environment was integral to the overall process of development. Attempts were made to integrate environmental concerns with the economic growth and development. The initial perception that the environment was a constraint to economic activity changed. People became aware that only by respecting the functioning of ecosystems that it is possible to promote economic development in a healthy and sustainable manner. This issue was particularly important for developing nations, which need to keep promoting economic activities in order to improve the living standards of their people.

Thus, in keeping with the above concerns, the concept of the environment changed to include ecological, economic, aesthetic and ethical concerns in modern times. Solutions for environmental problems were based on ecological, economic, aesthetic and ethical knowledge related to the problem. Attempts were made to promote sustainable development in most instances where natural resources were being exploited to meet the needs of the people.

Sustainable development is defined as the development that meets the needs of the present without limiting the potential to meet the needs of future generations (World Commission on Environment and Development, 1987) UN conference on Environment and Development (Earth Summit) held in Rio de Janeirio, Brazil in 1992 endorsed the implementation of sustainable development. After the Summit, sustainable development took on a life of its own, forcing its way into the deliberations of bodies ranging from city councils to international organizations.

At present there are three points on which there is general agreement with regard to environment. They are

- the environment is a common concern to both industrial and developing countries, although the problems resulting from poverty and affluence are different.
- the solution of global environmental problems can only be achieved through international cooperation.
- integration of economic growth and environmental protection must be done according to sustainable development approach.

1.2 Limits and sizes of the environment

From the above description, the environment is now considered a common concern in the world over. Then the question in our mind is, is it always a common concern? This leads us to identify the limits and sizes of the environment.

It has been customary for us to define the limits and thereby the size of the environment based on the extent the environmental problem impinges on human society.

Hence, at one end is the mega - environment that surrounds the planet earth. We call it the global environment and includes the ozone layer and phenomena like global warming. Similarly, if we go down the scale we could identify continental environment, regional environment, community environment, local or micro environment. Fir eg. prevalence of Malaria in the tropics could be considered a regional environmental problem allowing us to consider it on a regional scale. The extinction of species due to habitat change could be considered on a local or micro level environment scale. However limits and sizes of the environment will vary with changes in the extent of the problem. Environmental problems identified as the regional, will become a global problem in the years to come, eg. HIV

1.3 Types of Environments

Identification of the limits and sizes of the environments allows us to recognize different types of environment in nature.

For instance if you consider a fish living in a natural pond, its **external environment** will be the water in the pond in which it primarily inhabits. The water would contain nutrients, oxygen and other organisms that the fish require to sustain it's life. As opposed to the external environment, the body cavity within the fish provides an **internal environment** quite separate from the outside environment. The body surface acts as an exchange barrier between the internal and the external environment of the fish. The internal environment is relatively stable as compared to the external environment. However illness and injury or even environmental stress can upset it. But when the cause of the upset is removed, the internal environment comes back to its original condition.

The pond in which the fish inhabits is a **natural environment** . The abiotic factors of the pond, like light, temperature, depth, nutrients, dissolved gases will provide the life supporting chemical and physical factors for the fish. The other living organisms inhabiting the pond, like bacteria, insects, worms, molluscs, tadpoles, frogs, submerged vegetation etc could be food for the fish or they could provide nutrients that dissolve in water on which the fish depends upon indirectly. Examples of such natural environments on land include, forests, grasslands, savannah, deserts, etc. In any one of these natural environments, the climatic, physiological, edaphic and biotic factors interact with each other and influence the form and life of a person or a community.

As you are aware such natural environments have been encroached by man to cultivate his crops and to construct industrial zones and cities. Hence we could identify yet another type of environment which is called the **Man made environment.** In such environments man does not usually depend on natural sources for his requirements for food, water and shelter. For instance, in the city water is not taken directly from a stream like in a village. Water is filtered and purified and then used for drinking and other purposes. The waste water and garbage are not disposed off locally but are carried for treatment or for dumping to remote places away from the city.

1.4 Significance of the environment for life

Whatever the type of environment organisms inhabit, they all need life supporting elements for their survival on earth. These include air they breath, food and water they take in, and shelter either as natural enclosures (like caves and tree holes) or as artificial dwellings (like houses). Primarily all these elements are provided by the environment in the form of land, soil, vegetation and climate.

We make use of the land to cultivate our food and commercial crops. Soils provide nutrients needed for the growth of plants. The land form determines the soil types found in any one area and soil itself varies from place to place. Some soils are rich in nutrients and others are lacking in them. Those soils lacking nutrients need the addition of fertilizer.

Vegetation provides food for living organisms and also shelter. Vegetation at higher elevations on land acts like a sponge and absorbs excess water precipitating on land. They form the drainage basin for rivers, which provide water for many living down stream.
Climate and short term weather changes are characterized mainly by wind, temperature, pressure and rainfall and are determined by the properties of the atmosphere. Air in the atmosphere provides living organisms with oxygen, without the which survival of most of the living organisms will be threatened. The dramatic climatic changes, such as droughts, cyclones, hurricanes, tornadoes and global warming will have massive impacts on land, soil, vegetation and animals.

Because of the dependence of man on natural components in the environment they are called resources. There are different types of resources that man depends on and they include both living and non-living natural components. Non- metal elements, minerals, air, water, solar energy, tidal and wind power, plants, animals, micro-organisms are few of the natural resources.

Many have tried to classify these resources in various ways For eg., resources may be classified depending on their scarcity or abundance, whether they are widespread or localized in occurrence and whether they are renewable or non renewable. The most frequently adopted system is the last classification where renewable resources include those available at least in the human life time, for eg. solar energy, tidal energy and wind energy. Comparatively the most obvious examples of non renewable resources include fossil fuels like coal, oil and gas. These materials are available in limited quantity and from the human point of view are not renewable. Between these examples of renewable and non - renewable resources lie other resources that are difficult to categorize. These include plants, animals, micro-organisms, soil, minerals, air, water etc.

From the above description you came to know about the concept of the environment, different types of environments and about the significance of the environment for life. Next we will find out where these environments exists on planet earth.

1.5 Concept of biosphere and ecosystem

Biosphere is part of the earth occupied by living organism. This biosphere is restricted to a very small zone around the earth. Since living organisms need essential elements for survival like air, water and land the biosphere includes parts of the atmosphere, hydrosphere and lithosphere. (**Fig. 1.1**). When the concept of the biosphere was first proposed it was considered to be the earth's integrated living and non living life supporting system. Although it was proposed as early as 1920, it was only in the recent times that it has been widely adopted and used.

The integration of living organisms and the non living life supporting system mentioned in the concept of the biosphere has occurred in many ways. For instance, the biosphere has

- linked the lower atmosphere (trophosphere) with the lithosphere.
- provided a vehicle for the transfer of chemicals via the bio-geochemical cycles.
- played an important role in the water cycle.
- affected rates and patterns of weathering within the lithosphere
- contributed to the global energy system.

You will come to know more about the above processes in later units. Here we will outline variations that took place to the concept of the biosphere, since it was first proposed.

Originally the concept of the biopshere was applied to the earth's surface where plants and animals made their home. In recent times the biosphere has been extended by Gaia hypothesis to include parts of the atmosphere and subsurface geology that were previously thought of as nonliving.

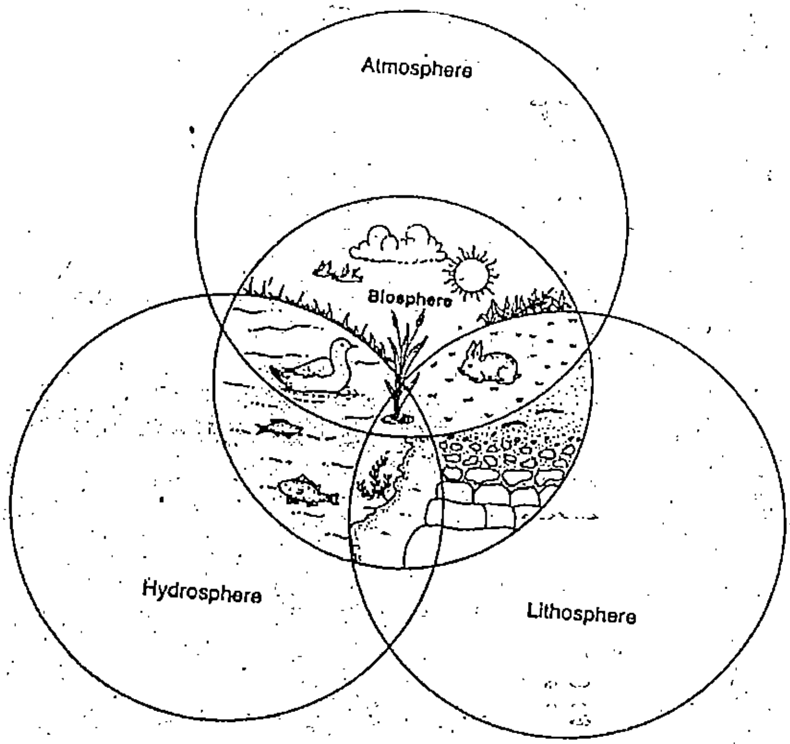


Fig. 1.1 Idealized scheme of the biosphere

For centuries scientists have viewed the earth and its environmental systems as a sort of mechanical machine, driven by physical forces like volcanoes, rock weathering and the water cycle. It was clear that organic activities played important role in some environmental systems, such as the biogeochemical cycles. However, quite recently biological factors were seen as secondary to physical and chemical ones.

A revolutionary new theory was put forward by James Lovelock in 1970s. He called it the Gaia hypothesis, after the Greek Earth goddess. The theory was revolutionary because it treated the earth as a single living organism, in which biological, chemical and physical factors all played important roles. Lovelock argued that the earth's living and nonliving systems form an inseparable whole, regulated and kept adapted for life by living organisms themselves. He sees Gaia as a complex entity involving the earth's biosphere, atmosphere, oceans and soil and constituting a feedback system which seeks an optimal physical and chemical environment for life on this planet.

Even though Lovelock regarded the biosphere as a "single organism" (called a super organism by some scientists), looking closely at plants and animals in the biosphere, Naturalists observed groups of plants and animals in the biosphere arranged in an orderly manner. Two concepts emerged from their observations which led to the use of the term "ecosystem" to describe the complex interactions between living organisms and their non living surroundings.

The first concept was that plants and animals formed a natural association, each with distinctive members. Just like morphological data allowed systematists to assign species to a hierarchy of taxonomic groups, detailed studies of the ecological distributions of plants led to the classification of biological communities.

The second concept was the realization that organisms are linked, both directly and indirectly by means of their feeding relationships.
Arising from these, the concept of the ecosystem was formulated. Basically an ecosystem is the sum of all natural organisms and the nonliving life supporting substances within an area. It was considered as an open system with a series of major inputs and outputs and these effectively "drive" the internal dynamics of the system.

The ability to recognize distinctive ecosystems in the biosphere gave ecologists a convenient scale with which to consider plants and animals and their interaction. This is because it is more localized and thus more specific than the whole biosphere. A variety of natural ecosystems are found in the biosphere and you will come to know about them and about their components and their functioning in later units.

Summary

The concept of environment is ~~the~~ surrounding of an organism, including its life supporting physical and biological factors. In the context of man this concept includes the social and cultural norms set by the human society as well.
The limits and the size of the environment depend on the extent of environmental problem and this vary from the largest global environment to smallest local or micro level environment.
The different types of environments include external environment, internal environment of an organism and the natural and man made environment in which the species survive.
Biosphere is the region on earth where all living organisms survive. An ecosystems is the basic functional unit in nature, defined by ecologists. Ecologists help to understand the complex relationships between living organisms and their surrounding.

Objectives

After reading this unit you should be able to

- Outline how the concept of the environment has evolved over time.
- State the types of environments and provide examples for the types from your country.
- Describe the concept of the biosphere.

Unit 2
The lithosphere

Contents

Introduction

From the previous lessons you learnt that the environment is made up basically of living and non-living components. The living component includes flora, fauna and micro biota where as the non-living component consists of three spheres, namely Lithosphere, Hydrosphere, and Atmosphere.
In this lesson, you will study the structure and composition of the Earth, components of the lithosphere, properties of soil, types of soils (classification of soil), soil biota and soil fertility and how all these have contributed to make a good quality soil.

2.1 Structure and composition of the Earth

The Earth may be divided into three zones, **the crust, mantle and the core. The crust** represents less than 1% of the earth's total mass and only about 0.5% of its radius. It does not however, have a uniform thickness, varying from an average of about 35 Km in the continental regions to about 5-10Km under the oceans. The high points of crust are continually being worn away by the processes of weathering and erosion. The low points are being infilled with the debris generated by these destructive processes. As this has been happening since the formation of the crust between $3.9 - 4.5 \times 10^9$ years ago, one might expect it to be smooth and yet it is not. This is because the crust is subjected to earth building processes that are generated within the planet.

The term tectonics refer to the major structural features of the crust and the processes that form them. According to the widely accepted theory of plate tectonics, the surface of the earth is divided into a series of essentially rigid plates of lithosphere. The fact that the asthenosphere (soft region of the mantle below lithosphere) can deform allows these plates to move each other. The plates meet at plate boundaries, three major types of which are recognised: transform boundaries, divergent boundaries and convergent boundaries. A transform boundary occurs when two plates move along a transform fault without either moving apart or colliding. When plates move alternatively, divergent boundaries are formed. When two plates moving towards one another meet, a convergent boundary is formed (Fig.2.1).

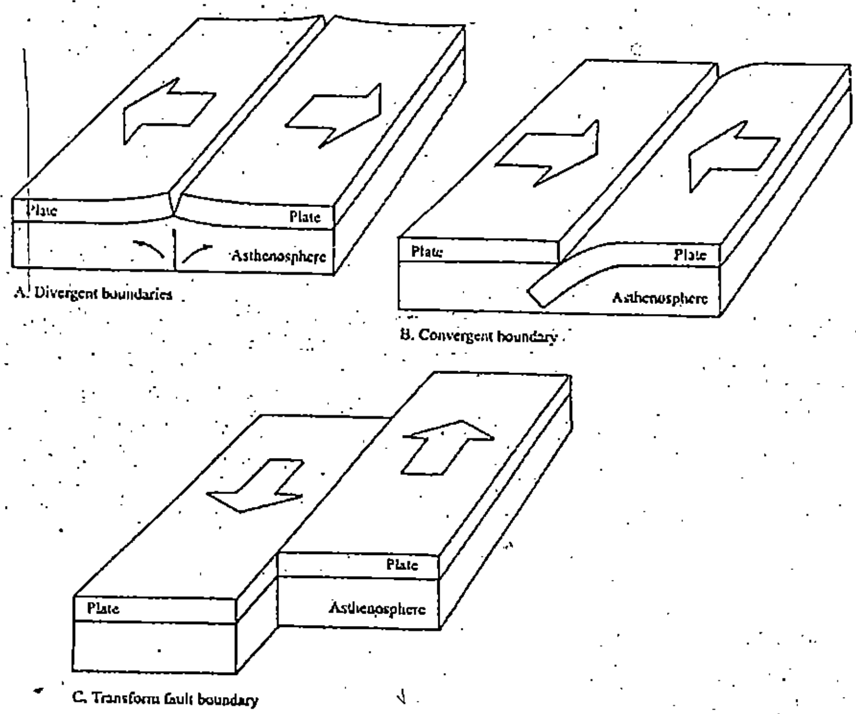


Figure 2.1: Different types of plate boundaries

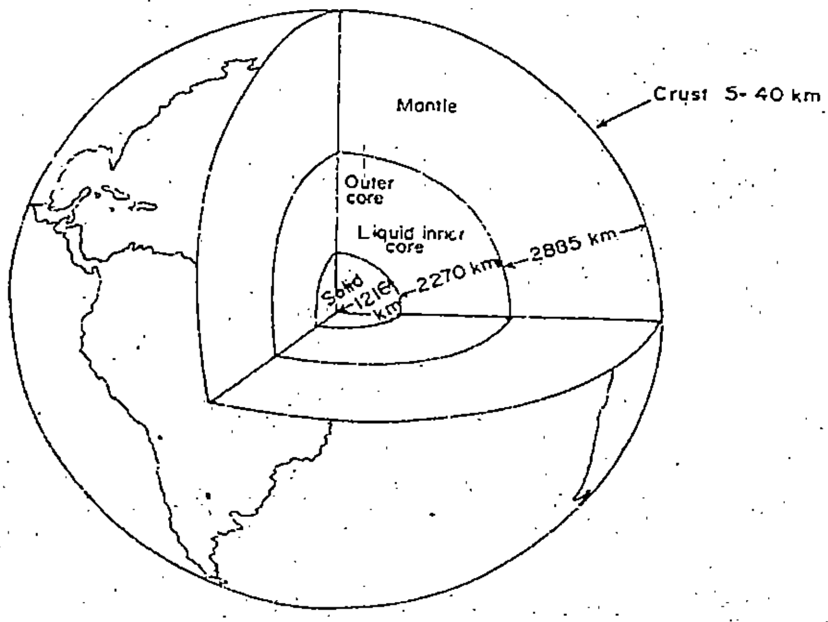


Figure 2.2: The major components of the Earth

The thick, dense rocky matter that surrounds the core is called the **mantle.** The mantle covers the majority of the earth's volume. This is basically composed of silicate rock rich in iron and magnesium. The mantle is less dense than the core but denser than the outer crust layer (Fig.2.2).

The most interior region of the earth is referred to as earth's **core** and is divided into outer and inner core. The inner core area is solid and rich in iron and the outer core consists of liquid iron at a temperature of 2500^0C.

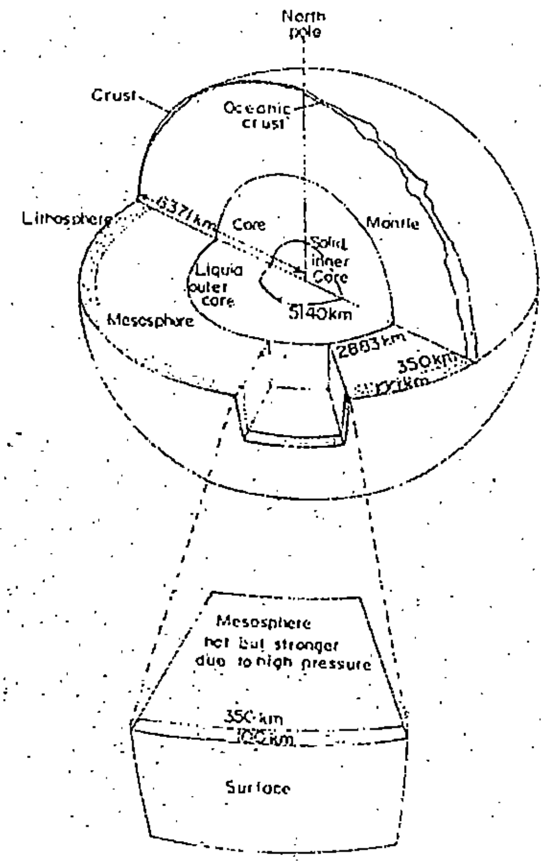


Figure 2.3: Three dimensional view of the lithosphere

The crust and the upper part of the mantle form a coherent solid layer known as the **lithosphere** (Fig.2.3).

2.2 Components of the Lithosphere

The lithosphere is divided into the lithosphere plates that constitute earth's outer surface. There are about seven or eight large plates and many small ones. The interactions occurring at plate boundaries are responsible for the production and destruction of oceanic crust, the rifting and collision of continents and the formation of mountains, volcanoes and earthquakes. The basic lithosphere components of the earth are **rocks, sediments and soils.**

Rocks

The non-living, naturally occurring solid materials that form the crust are called rocks. Rocks are consolidated units of the earth's crust made of minerals. Rocks are the parent material from which sediments and soils are developed.

There are three types of rocks.

I. Igneous rocks

II. Sedimentary rocks

III. Metamorphic rocks

Igneous rocks are formed by the solidification of molten rock material called magma that has its origin in the mantle or in the lower part of the crust. Such rocks are made up of interlocking crystalline minerals with few voids. As a result, they are generally impervious and show considerable mechanical strength.

Sedimentary rocks are formed by hardening of sediments obtained in one or more of the following ways;

- from the fragments produced during weathering
- as a result of dissolved material
- as a consequence of biological activities

Sedimentary rocks that form from the solid debris of erosion and weathering are said to be elastic rocks; other sedimentary rocks are non-elastic.

Metamorphic rocks are formed by the alteration of existing rocks by the action of extreme heat and/or pressure and/or permeating hot gases or liquids. The action of heat alone causes thermal metamorphism.

The rocks undergo weathering. Weathering occurs by the interaction of physical processes, chemical processes and the biological processes. During physical weathering, large rock fragments are broken into smaller pieces a solely mechanical process. These physical processes are usually accompanied by chemical processes which produce changes in the nature and the composition of rocks. The chemical processes take place mainly on the surface of the rock essentially in the presence of water. Therefore, the intensity of chemical weathering is more in tropics than in temperate countries. The living organisms largely contribute for further decomposition of rocks. They secrete various kinds of exudates which help for the solution of rocks. All these biological processes require water and activities of living organisms are totally dependent on the availability of water. However the end result of all these is the formation of soil and sediments.

Sediments

Sediments are the substances that are deposited after transportation which may or may not be altered by weathering. When sediments are hardened into rock like formations such as limestone, sandstone, they are called sedimentary rocks (as explained earlier). When they are not hardened, they will remain as sediments.

There are basically four agents that are responsible for transporting material from one place to another. These agents are **wind, water, ice and gravity.** Wind action is capable of shifting material. Wind moved material may be grains or sand of variable sizes that could collect in low swell or steep slopes forming sand dunes. Wind can also shift volcanic ash and soils may be formed from the break down of those materials. These soils are of little agricultural importance.

Sediments that have been transported by flowing water can be deposited in the bottom of streams, lakes, rivers and oceans. With the flow of water, lot of nutrients is also carried by the sediments and therefore, these sediments are rich in nutrients.

In the polar region of earth, snow accumulates and the pressure from its weight changes snow to ice. During summer, the ice melts and the water flowing from the melting ice carries the sediments. These sediments finally build up and form natural parent material from which soil may develop. Fragments of rock debris detached from heights above and carried down the slopes to be deposited at the base of hill due to gravity may also be the parent material for the development of soils. These deposits are not of great agricultural importance because of their unfavourable physical and chemical characteristics.

Soils

Soils are of different types with varying physical and chemical characteristics; e.g. sandy soil, clayey soil, saline soil, acidic soil, soil rich in mineral nutrients. All these characteristics affect the plants that grow on it and are important in determining the use to which the land could be used. Mineral soils consist of four major components; namely mineral matter, organic matter, water and air. For a soil that shows optimum conditions, the normal distribution of above components are given below in (Fig.2.4)

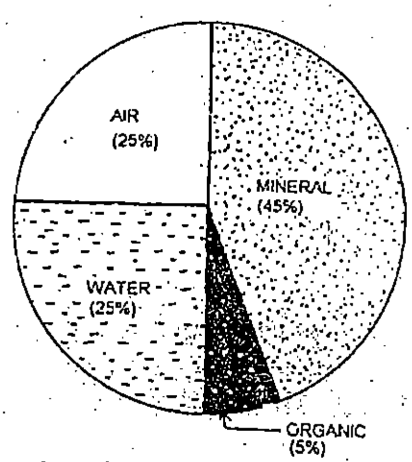


Figure 2.4: The normal distribution of four major components in mineral soil

Mineral matter is received by weathering of rocks and minerals. These particles are very important in maintaining proper soil structure and porosity. In addition these particles provide some nutritional requirements as well. Soil organic matter is an accumulation of partially decayed and partially synthesized plant and animal residues. The organic matter is very important as a source of nutrients for plants and soil organisms, as a soil structural manipulator and as a storehouse for water. Humus is also a part of the organic matter. In brief, soil organic matter controls chemical, physical and biological properties of soil and thereby, helps in better plant growth. Soil water is held within soil pores with varying degrees of tenacity depending on the amount of water present. The water is used by plants for various metabolic activities and to maintain rigidity. In addition, water acts as a solvent in soil to make soil solution which is the medium for various ions in soil. Soil air is also an important constituent in the soil environment. It helps to have good soil structure and provides necessary air for respiration of soil organisms and roots. The content and composition of soil air differs from the air in the atmosphere. Soil air is present in soil pores and not continuous. Soil air is generally more humid and CO_2 content is higher(O_2 content is lower than the normal air). Soil water is the factor which determines the content of soil air.

Soil is considered as a biological laboratory with a multitude of living organisms. Every type of natural soil has a varied population of living organisms both plants and animal in nature. They vary in size ranging from large rodents, worms and insects to minute bacteria. The activities of soil organisms range from largely physical disintegration of plant residues by insects and worms to the complete decomposition of these residues by smaller organisms such as bacteria and fungi. These are the processes which release several of the essential elements from organic combinations to the soil. The soil organisms are therefore involved in a continuous turnover of organic material including humus.

Natural soil is not a simple substance. Soil is a complex component derived from rocks and minerals and comprises of rock fragments, mineral matter, organic debris, water, air and living organisms in different quantities. To understand the complex nature of soil, let us examine the features of a soil profile.(Fig:2.5) Soil profile is a downward section through the topmost layer of the soil exhibiting its horizontal layers.

Soil formation depends on the characteristics of the parent rock, the climate, time, topography and the vegetation. Due to the changes of these factors, different soils are formed in different environments. Soil is the base for plant growth and it also provides necessary nutrients.

Soils show different characteristics such as texture (sandy, clay etc.), structure (arrangement of soil particles), colour, consistence (condition of soil with varying quantities of water) and porosity. These are very important in using soil for various purposes. These characteristics will be discussed in detail later.

Soil as an abiotic component in the environment provides nutrients to plants. It has billions of micro and macro organisms and they interact with the abiotic components of soil and make it a living environment. As a result of these interactions, two important processes; **decomposition and mineralization** take place in the soil environment. Without these processes, life could not survive on this earth as most of our resources are limited in the environment.

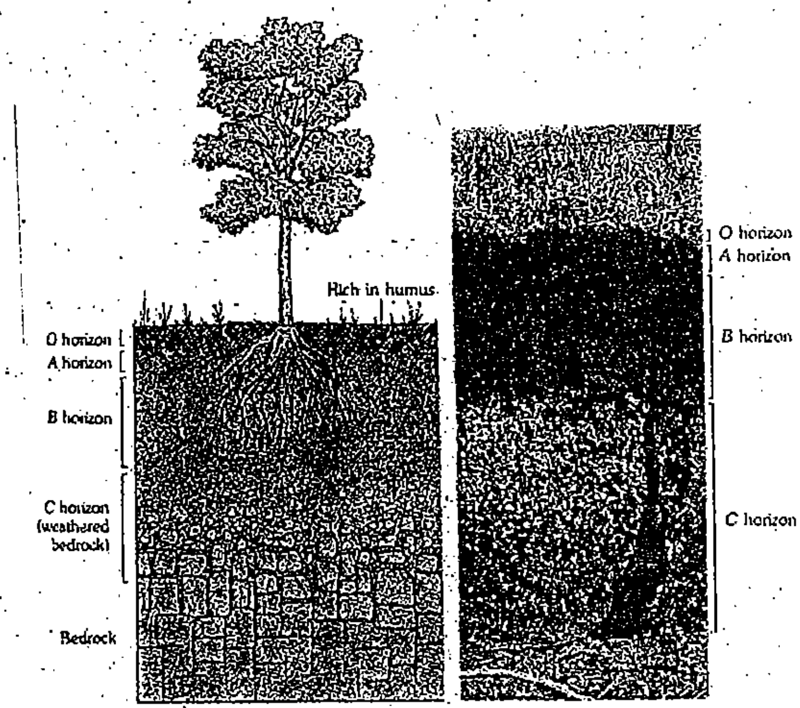


Figure 2.5: The Typical Soil profile

Therefore, cycling of these materials is very essential for the survival of life on earth. The soil acts as a medium for life. It provides the base for anchorage of plants, necessary moisture air and required nutrients for plant growth.

2.3 Properties of soil

We learnt that soil is made up of particles and organic matter having a net work of channels filled with air and water. Therefore, a mineral soil is a porous mixture of inorganic particles, decaying organic matter, air and water. However, as discussed earlier soil is not homogeneous because the individual soil particles are typically clustered together in the form of aggregates which vary widely in size.

From the point of view of crop production soil quality is determined by the degree of its fertility. In other words, a good quality soil is a fertile soil. A fertile soil is a soil in which plants grow well and produce a satisfactory yield. Soil fertility or soil quality in relation to crop growth is determined mainly by its physical, chemical and biological properties. Some properties may influence soil quality to a greater extent than others.

The variable composition of soil constituents gives rise to certain physical properties. These are of vital importance to man to gauge its suitability for various purposes such as how easily a farmer is able to dig or plough his fields; whether it is dry or moist; how a clod of soil looks like when crushed by hand; whether the soil in a field is a hard pliable mass or a loose pliable one. All these factors are determined by the physical properties of the soil. The rigidity and supporting power the soil can offer to plants, the moisture storage capacity, the degree of drainage, the ease of penetration by roots, the degree of aeration and the retention of plant nutrients, are all intimately connected with the physical conditions of the soil. It is therefore vitally important that students studying soil should be well acquainted with the physical nature of different soils and their physical properties.

A large number of chemical properties also influence soil quality. Some chemical properties such as pH, electrical conductivity, cation exchange capacity and base saturation play a major role in deciding fertility of soil. The biological properties of soil largely affect on the quality of soil. The soil enriched with more useful organisms would be obviously of a better quality than one with lesser number of useful organisms. Similarly a soil infested with parasitic organisms would be a soil of low quality even if all the other properties are satisfactory.

Thus, for a soil to be in good quality, its physical, chemical and biological properties should be at satisfactory levels. Therefore, awareness about soil properties would be of immense use for a farmer to decide his associations with soil. To get a complete picture about lithosphere, properties of soil will be an important part.

2.3.1 Physical properties of soil

Let us now learn about the properties of soil beginning with physical properties. There are about eight physical properties and you will study all of them in this session.

1. Soil colour

The colour of soil is a very apparent feature. It is perhaps not one of the most important properties of soil, but it gives an indication of the nature of the soil, and is thus of service to both farmers and soil scientists. Organic matter content, drainage conditions and aeration are soil properties related to colour, which are of interest to farmers.

Some farmers associate the dark brown and black soils with a high level of fertility. **The dark colours are usually due to high contents of humus or organic matter in the soil.** If soils are poorly drained, there is usually a greater accumulation of organic matter in the surface layers, thus giving it a very dark colour. The lower soil layers, which contain very little organic matter on the other hand, are of a light grey colour indicating the poorly drained conditioned. **Dark colour need not always be a sign of high fertility.**

Generally, in temperate (cool) climates, dark coloured soils are relatively higher in organic matter than light-coloured soils. In well-drained soils, the colours usually range from very pale brown, through intermediate browns, to very dark brown or black, with the increase in organic matter. The most stable part of the decomposed organic matter is **humus** which is darker than less well decomposed plant remains. As organic matter is neither all of the same colour nor the only colouring material in soil, soil colour by itself is not an exact measure of this important constituent. Well-drained soils having high content of organic matter are browner and less black under tropical environments compared to those under cool conditions. Yet, the dark-clays of the tropics which include some of the blackest soils of the world, seldom contain as much as 3% organic matter. Dark-coloured soils low in organic matter may contain compounds of iron and humus, elemental carbon and compounds of manganese.

Soils high in organic matter content under tropical conditions are browner and less black, than those having high organic matter under temperate conditions. The very dark clayey soil abundant in the tropics are not very high in organic content, but are dark due to compounds of iron, elemental carbon and manganese, in addition to humus.

The red colour of a soil is due to the presence of iron compounds, usually unhydrated iron oxides. It has been found that partly hydrated iron oxides and manganese dioxides also contribute to the red colour. Since unhydrated iron oxides are relatively unstable under moist conditions and cannot exist in poorly drained soil, red colour usually indicates good drainage and aeration. Well-developed red coloration indicates that such soils are relatively old or at least that the soil material has been subjected to relatively intense weathering for a long period.

The yellow colour is also largely due to hydrated iron oxides. Very often yellow soils develop under imperfect drainage conditions. Yellow colour in deeper horizons usually indicates a somewhat more moist soil climate than red colours. Other factors being the same, yellow colour is more common than red colour in regions of high humidity. Well drained yellow soils owe their colour to the fact that small amounts of organic matter and other colouring material such as iron oxides are mixed up with large amounts of whitish sand.

The brownish colour of a soil would be mainly due to the presence of less ferro-magnesium or iron bearing minerals in the parent material. On the other hand, brownish colour could result in soils containing relatively large amounts of iron oxides, including organic matter. Brown colour also could mean a very young and immature soil.

Gray or whitish soils result from several substances namely, quartz, kaolin and other clay minerals, $CaCO_3$, gypsum, various salt and compounds of ferrous ions. Light colours or white also indicate a very low content of organic matter and iron in the soils. In arid and semi-arid areas, soil may be white or nearly white because of the very high content of $CaCO_3$, gypsum or other salts. The failure of the light-coloured soils to accumulate organic matter usually indicates an environment unfavourable for plant growth and the existence of microorganisms.

The red colour of a soil is mostly due to the presence of unhydrated iron oxides, while the yellow colour is due to the presence of hydrated iron oxides. Brownish colour is mainly due to large amounts of iron oxides and organic matter. Younger soils are also very often brown in colour.

2. Mechanical analysis of soil

If you make a closer observation of a soil sample, you will realise that it is composed of a variable mixture of large pebbles and stones, coarse sand, fine sand, something resembling flour (silt); lumps or clods of varying sizes which are actually also clusters of soil particles (slay), plant roots, a dark substance spread throughout the soil mass, which is universally called **humus,** dried leaves and twigs, and organisms such as ants, earthworms or insects. Not all of these substances are properly a part of soil in a physical sense. **Only the mineral matter and other particles which are less than two millimetres (2mm) in diameter are considered as soil.**

Activity 1

What is the reason for choosing soil particles with diameter less than 2 mm for experiments, conducted in the lab?

It has been customary to describe soils in terms of the **proportions of particles of different sizes** that they contain. This basis of characterisation developed because particle size is an obvious characteristic and is also related to soil behaviour and plant response. Particle size alone has no known direct effect other than that of the mechanical impediment offered by large stones that deflect the passage of roots or prevent the vertical emergence and growth of aerial parts. However, particle size influences soil characteristics indirectly, by affecting the porosity, drainage, aeration, consistence, etc.

Although the nature and properties of coarse and fine soil particles differ considerably, there is no sharp natural division of any kind at any particular size. For purposes of experimentation and classification, however, some arbitrary boundaries have been established. Two systems are commonly used.

1 The International System, and
2 The Scheme used by the U.S. Department of Agriculture.

The latter is now consistent with the International Systems, but makes more separations (Table 1). The word **soil separates** is preferred to soil particles at this stage to accommodate the fact that some of the separations are not of particle nature.

Table 2.1 - Size limits of soil separates from two schemes of analysis (Adapted from U.S. Department of Agriculture handbook no. 18 - Soil Manual - 1952)

1.1.1 U.S.Dept.Agri.Scheme		1.1.2 International scheme	
Name of separate	diameter(range) (mm)	Name of separate	diameter(range) (mm)
Very coarse sand	2.0 - 1.0	Coarse sand (I)	2.0 - 0.2
Coarse sand	1.0 - 0.5	Fine sand (II)	0.2 - 0.02
Medium sand	0.5 - 0.25	Silt (III)	0.02 - 0.002
Fine sand	0.25 - 0.10	Clay (IV)	Below 0.002
Very fine sand	0.10 - 0.05		
Silt	0.05 - 0.002		
Clay	Below 0.002		

The mechanical composition of a soil is the "weight percentage of the mineral matter that occurs in each of two or more specified size fractions". Usually, the soil is separated into at least three size fractions, namely: **sand, silt and clay.** The process by which the mechanical composition is determined, is known as **mechanical analysis**.

3. Soil texture

The proportion in which sand, silt and clay occur together in any soil will determine the 'feel' of the soil, or more correctly, the **texture** of the soil. It refers to the fineness or coarseness of the soil. More specifically, **soil texture is the relative proportions of different size groups or separates.** The rate and extent of many important physical and chemical processes which occur in soils, are governed by texture, because it determines the amount of surface on which the reactions can occur. In the laboratory, the texture of a soil is determined by the process of mechanical analysis which we learnt about earlier. In the field, the texture is determined by the feel of the moist soil when rubbed between the thumb and forefinger. Any one of us should be able to distinguish between a sandy, a loamy or a clayey soil according to the feel we experience when rubbing the moistened soil between our fingers. The sand particles are gritty. The silt has a floury or talcum-powder feel when dry, and is only moderately plastic and sticky when wet. Persistent cloudiness and stickiness is generally imparted by clay. The soil scientist, however, has to more precious use more precious terms that describe texture more accurately. For this purpose, several basic soil-textural classes have been developed based on the proportions of sand, silt and clay in the soils which convey some indication of their physical properties. The class names have originated through several decades of soil study and classification. Three broad, yet fundamental groups of soils are recognised:

1 **Sands**

2 **Loams**

3 **Clays**

Sands - The sand group includes all soils in which the sand separates make up 70% or more of material by weight.
Loams - Loams are mixtures of sand, silt and clay in about equal proportions.
Clays - A soil that has at least 35 - 40% of the clay separate.
Within each of these three main groups, a number of additional classes have been recognised (Fig.2.6).

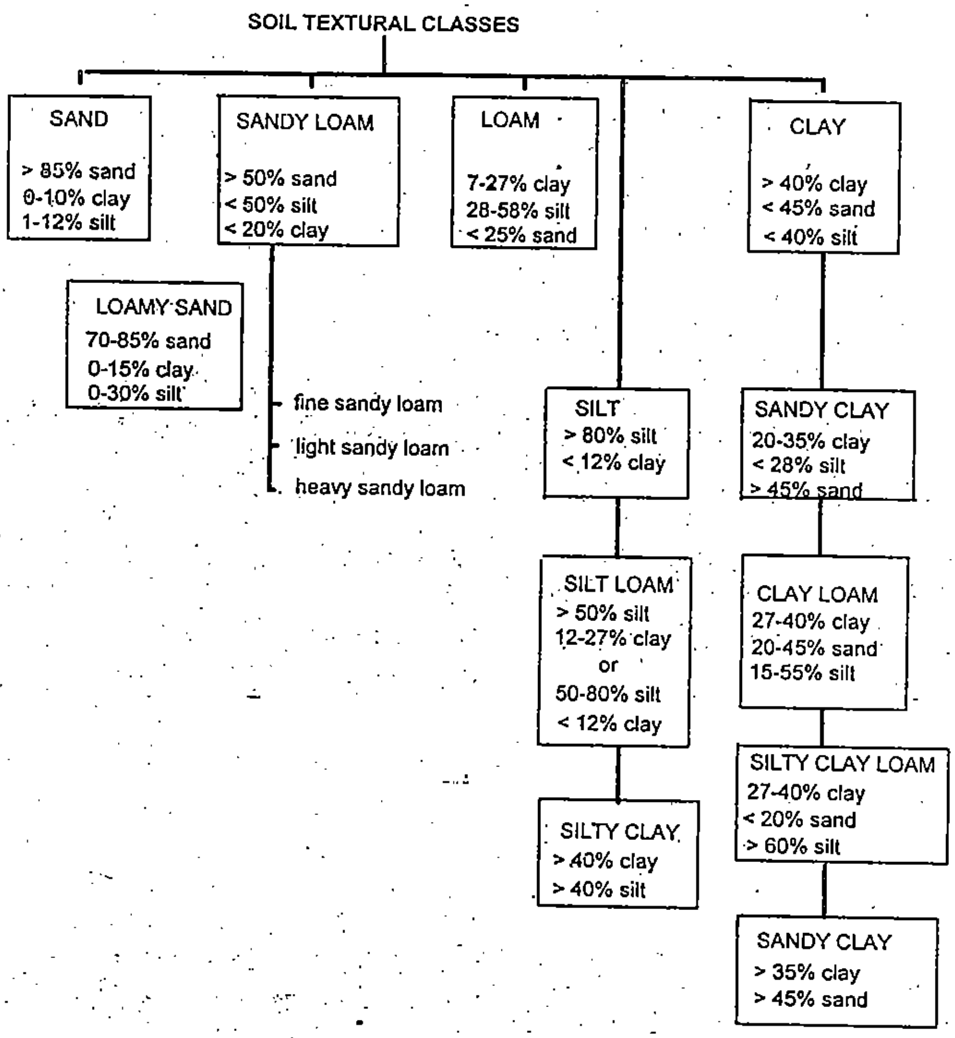


Figure 2.6: Percentages of sand, silt and clay in the principle textural classes of soil.

Activity 2

Determine the texture of your home garden soil and compare it with that of your neighbour's soil. Comment on the differences.

One point has to be clearly understood in assigning soils to different classes. **The soils are designated to soil classes and named according to the separate or separates which contribute most to their characteristics.** This does not mean that a class is necessarily named after the separate which is present in the largest quantity. It takes a very large quantity of the coarser particles to exert as much influence on soil properties as a comparatively small quantity of the finest particle, clay. Clay is the most potent separate in imparting its properties to a mixture of separates, and hence the adjective clay is added to the class names of many soils which contain a higher percentage of the other separates than they do of clay.

4. Soil structure

The term **texture** refers to the relative proportion of individual particles in a soil mass, but **when the arrangement of the particle is being considered, the term structure is used.** In other words, soil structure describes the aggregation of primary soil particles (sand, silt and clay) into compound particles, or clusters of primary particles.

Soil structure obviously modifies and influences texture in regard to moisture and air relationships, availability of plant nutrients, action of microorganisms and plant growth. It is at once apparent that soil conditions and characteristics such as water movement, aeration, bulk density and porosity, will be much influenced by structure. In fact, the important physical changes imposed by the farmer in ploughing, cultivating, draining, liming and manuring his land are structural rather than textural.

In a soil-profile you may find that the profile is dominated by a single structural pattern. However, more often, a number of types of aggregations are encountered as one moves from horizon to horizon. We shall return to this aspect later.
The structural forms of soil that have been developed from one or both of two possible non-structural forms are

1 single-grained, and
2 massive.

If each particle in a soil functions as an individual without attachment to any other particle, then this condition is called **single-grained.** Loose sand is a good example of this type. Some of the particles in many soils exist in this single-grained condition, but to find all the particles so functioning is unusual. The binding influence of organic matter often modifies this original form by building up weak aggregations. If the colloidal material (such as clay) in the soil is considerable, it makes the soil denser and upon drying, it forms very large, irregular and featureless clods. This condition is called **massive.** When this soil is too moist, the cementing material of the aggregation is so soft that the aggregates are destroyed. The particles readily slide over one another, smaller ones slipping in between larger ones to create a compact mass dominated by a massive condition. The soil is then said to be **puddled.**

Activity 3

Define the term soil structure.

These two extremes are the theoretical boundaries of the eight structural types commonly recognised in soil profiles. These are described as **platy, laminar, columnar, prismatic, blocky, nut-like, granular and crumb.** A brief description of each of these structural types and a schematic drawing is given below (Fig.2.7). For convenience most soil text-books consider these eight as sub-types of four primary types of structure, namely: plate-like, prism-like, block-like and spheroidal.

Activity 4

List the major structural types of soil and explain their main characteristics.

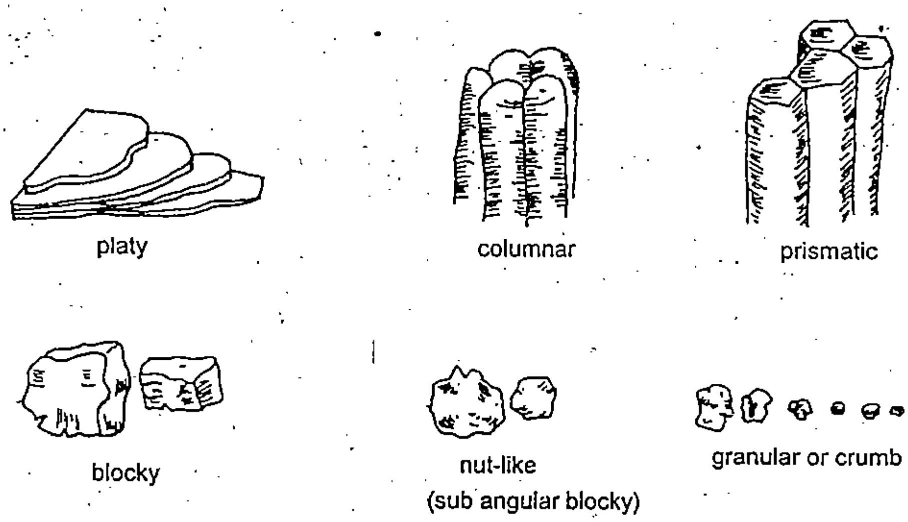


Figure 2.7: Diagrammatic representation of various soil structural types.

5. Soil consistence

An important physical property of soil that should receive our attention is **soil consistence.** This is a term used to describe the physical condition of a soil at various moisture contents as evidenced by the behaviour of that soil towards mechanical stresses or manipulations. Although consistence and structure are inter-related, structure deals with the shape, size and definition of natural aggregates that result from variations in the forces of attraction within a soil mass, whereas **consistence deals with the strength and nature of such forces themselves,** as influenced by soil moisture. Terms that are commonly used to describe soil consistence are as follows: **loose, friable, firm, soft, harsh, plastic and sticky.** These and other terms for consistence are used to describe soil consistence under three standard moisture contents **dry, moist and wet.**

The consistence of soil material when dry is characterised by resistance to pressure, a tendency to crush to powder or to fragments and an inability to cohere again when pressed together.
Consistence when wet is usually measured with very moist soil or slightly above "field capacity".

Consistence when moist is determined at a moisture content approximately midway between air-dry and field capacity. At this moisture content, most soil material may exhibit a form of consistence characterised by a tendency to break into smaller masses rather than into powder.

Activity 5

What is the main difference between soil structure and soil consistence?

6. Soil Porosity

The portion of a given volume of soil, which is not filled with solid matter, is termed **pore space.** This obviously depends on both the texture and structure of the soil and also on the shape of the particles. A high content of organic matter leads to much pore space. In soils containing little silt and clay, the total amount of pore space is small, even though the individual spaces are large. In soils containing a high proportion of silt and clay, the relative volume occupied by pores is large, and if these soils are well aggregated, the total volume of pore space may be equal to or even greater than that occupied by the solid particles. This is because space exists not only between soil grains, but also between aggregates. Therefore, **medium-textured soils high in organic matter** and porous aggregates have a **higher pore space per unit volume** than sandy soils. We shall return to this aspect a little later.

The amount of pore space in a soil is expressed as a percentage of the total soil volume. The term **porosity refers to the total pore space** in a soil rather than to the size of the individual pores.

Pore spaces are important because they contain air and moisture. Air and moisture play just as much a part in the crop-producing capacity of a soil, as the mineral and organic matter. The amount of pore space in a given volume of soil has a great influence on its weight.

Sandy soils usually have less pore spaces or porosity compared to a fine-textured loam-clayey soil. Yet, our everyday experience tells us that water can move much faster through a sandy than a clayey soil.
The explanation for this lies in the differences in size of the pores that are found in each soil.

The total pore space in a sandy soil may be low, but a large proportion of it is composed of large pores which are very efficient in the movement of water and air. The percentage of the volume occupied by small pores in sandy soils is low, which accounts for their low water-holding capacity. In contrast, the fine-textured silty and clayey loam soils have more total pore space and a relatively large proportion of it is composed of small pores. The result is a soil with a high water-holding capacity. Water and air move through such soils with difficulty because there are fewer large pores. Thus, we see that the size of pores in soil may be as important as the total amount of pore space.

Two types of individual pore spaces are therefore recognised in soils: **macro** and **micro.** While no sharp demarcation occurs, the macro pores abundant in sandy soils characteristically allow the ready movement of air and water percolation. In contrast, the smaller pores normally tend to be filled with water and allow only slow capillary movement. Air movement in such capillary pores or micro pores is greatly impeded. For the growth of many economic plants, an equal distribution of macro and micro pores is desirable.

7. Aeration

In a well-aerated soil, gases are present, both in sufficient quantities and in the best proportions which encourage optimum rates of essential metabolic processes, for the growth of organisms, specially the higher plants.
A soil with satisfactory aeration should have:

1. Sufficient space devoid of solids, and
2. ample opportunity for the easy and ready movement of essential gases into and out of these spaces. Water relations largely control the **amount of air spaces** available.

Under actual field conditions, two situations can give rise to poor aeration in soil i.e.,

1 When the moisture content is excessively high, with little or no room left for gases, and
2 When the exchange of gases with the atmosphere is not sufficiently rapid where the concentrations of the major soil gases, O_2, in particular may not be kept at desirable levels.

The latter may often occur even when sufficient total air space is available. Excess moisture very often leads to waterlogged conditions. This may be temporary, but nevertheless, often seriously affects plant growth. Such a situation is frequently found in poorly drained, fine texture soils which have a minimum of macro pores through which water can move rapidly. It may also occur in soils that are normally fairly well drained, if the rate of water supply at the soil surface is high. Low areas in a field of even, flat ground with depressions, in which water tends to stand for a short while, are good examples of this condition. Such complete saturation of the soil with water can be disastrous for certain plants even for a short time. Plants, which have previously been growing under conditions of good soil aeration, are more susceptible to damage from flooding, than plants which have been growing on soils where poor aeration has prevailed from the very start.

Activity 6

List the conditions which help to develop good and bad aeration in the soil.

Inadequate interchange of gases between the soil and the free atmosphere above it, affects:

1 the rate of biochemical reactions influencing the soil gases, and
2 the actual rate at which each gas moves in or out of the soil.

Obviously, the more rapid the usage of O_2 and the corresponding release of CO_2, the greater will be the necessity for the exchange of gases. Factors such as temperature, organic residues, etc., which markedly affect these biological reactions, are of considerable importance in determining the air status of any particular soil.

There is little doubt that most gaseous interchanges in soil occur by **diffusion**. Thus, a higher concentration of O_2 in the atmosphere, which is often the case, will result in a net movement of this gas into the soil. An opposite movement of CO_2 (and water vapour) often occurs simultaneously since the concentration of this gas is generally higher in the soil than in the atmosphere. The data on composition of soil air at various locations in the world have clearly shown that CO_2 concentration is higher and O_2 concentration is lower in the topsoil as compared to that of the atmosphere.

There is also a general inverse relationship between O_2 and CO_2 contents, that is, O_2 decreases as CO_2 increases.
Compared to the atmosphere, soil air is usually much higher in water vapour content, being essentially saturated except at or very near the surface of the soil. The concentration of gases such as methane(CH_4) and hydrogen sulphide (H_2S) which are formed by organic matter decomposition is somewhat higher in soil air.
The anaerobic decomposition of organic material is much slower than the aerobic decomposition. Furthermore, the products of decomposition are also different.

8. Soil temperature

Plant growth as well as chemical and biological activities in the soil are greatly influenced by soil temperature. Plant growth usually becomes slow at about 15-18°C. Nitrification does not begin until the soil temperature reaches about 18°C and reaches a maximum between 45-50°C.
The optimum temperature for seed germination, as might be expected, varies widely, being low for certain crops and high for others. The chemical processes and activities of microorganisms which convert plant nutrients into available forms are also influenced by temperature.

The amount of heat absorbed by soils is determined primarily by the quantity of effective solar radiation reaching the earth (Fig.2.8). The latter in any particular locality depends fundamentally upon its climate. In addition, the amount of energy entering the soil is affected by other factors such as colour of soil, slope and the vegetative cover of the site under consideration. **Dark soils are well known to absorb more energy than light coloured ones. Red and yellow soils are also known to show a more rapid temperature rise than those that are white.** In the Northern Hemisphere soils which are located on southern and south-eastern slopes warm up more rapidly in the morning, than those on flat land or on the northern slopes. The reason here is that they are more nearly perpendicular to the sun's rays and hence, a maximum amount of radiant energy strikes a given area. The farmers of the Northern Hemisphere therefore select soils with a southern or south-eastern exposure for the growing of early season (after winter) vegetables and fruits. Whether the soil is bare or covered with vegetation is a factor that markedly influences the amount of insulation received. Bare soils warm up quickly and cool off more rapidly than those covered with vegetation or with artificial mulches. The cooling effect of forests is universally recognized.

Activity 7

What is the relationship between soil temperature and biological activities in the soil?

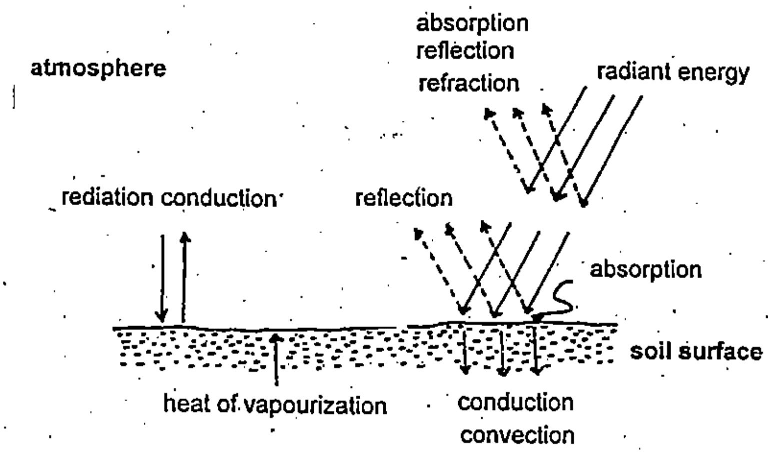


Figure 2.8: A diagram showing the acquisition, loss and movement of heat in soils.

2.3.2 The Chemical Properties of soil

The study on chemical properties is very important from the point of view of crop production as it gives the basic idea about how chemical nature of soil helps to have good fertile soil.

1. pH

One of the most important physiological characteristics of the soil solution is its reaction. The presence of microorganisms and development of higher plants depend on the chemical environment of soil and that is why the study of soil reaction is so important in the study of the lithosphere. There are three types of soil reactions.

1 acid.

2 alkaline

3 neutral.

Instead of stating in a vague manner that a particular solution is feebly acidic, alkaline or neutral, the acidity or alkalinity can be expressed on the scale of acidity or alkalinity. This scale is called pH scale. The units of this scale are called pH values. Therefore soil pH depends on the acidity or alkalinity of soil. Measurement of soil pH and its effects are therefore an integral part of studying soil ecology. The term pH is defined on the logarithm of the reciprocal of the hydrogen (H^+) activity.

$$pH = -\log [H^+]$$

Soil pH therefore, is a measure of the concentration of the hydrogen ions in soil solution. Further soil pH can also be considered as a measure of the acidity of the soil solution because acids are compounds that dissociate in water to produce hydrogen ions.

You are already aware that the neutral point of pH is 7.0. A solution is acidic if the pH is below 7 to 1 and alkaline if it is above 7 to 14. For an example, a solution of pH 5 has a concentration of 10^{-5} m hydrogen ions, and one of pH 8 has a concentration of 10^{-8} m hydrogen ions or 10^{-6} m hydroxyl ions.

Soil pH can be measured on the extracted soil solution, or on a suspension of soil in water. The pH should strictly be measured when the soil is suspended in a solution of about the same composition as the soil solution. Commonly, however, the pH is measured on a suspension in water. The buffering of soil pH is due to several soil properties. They are cation exchange capacity, clay minerals, humus hydrated aluminium and iron oxides, aluminium ions, solubilization of soil minerals and reaction of acids with calcium and magnesium carbonates.

Activity 8

Why do you think that soil pH is very important?

The element required in large amounts from soil are nitrogen and potassium.Nitrogen can be taken up by plants

High pH tends in particular to affect the plant adversely by reducing the availability (solubility) of manganese and iron to the root system. Phosphorus availability is also reduced due to formation of calcium phosphates and low pH. High soil acidification tends to affect the plant adversely through increased availability of aluminium and also manganese. The pH is also affected by the microbial communication of the soil. The most acid-tolerant S-oxidising bacteria can grow at pH 1 even while the most alkali-tolerant streptomycetes can grow at pH 10. In general, most of bacteria are more prefer to pH among 7 and most of fungi are prefer to the pH range of 2 to 7. Some actinomycetes can grow at the pH between 7 to 10.

Activity 9

What are the contributions made by higher plants towards the pH of soils?

2. Electrical conductivity

Electrical conductivity (EC) of a soil suspension is used to estimate the concentration of soluble salts in the soil. The soluble salts consist predominantly of cations Na^+, Mg^{++} and Ca^{++} and the anions Cl^-, SO_4^{--} and HCO_3^-. Soluble fertilisers may also contribute other ions such as K^+, NH_4^+ and NO_3^-. High electrical conductivities corresponding to high concentration of soluble salt in the soil are undesirable for the growth of most plants. The electrical conductivity values increase with increasing temperature and must be corrected if not measured at 25°C. An approximate correction can be made by increasing the values by 2% for each degree that the ambient temperature is below 25°C and decreasing them when the temperature is above 25°C.

There is no clear relationship between EC (1:5 soil/water) and total soluble salts due to the different ionic conductivities of the various salts. An approximate value for percentage total soluble salts may be obtained by multiplying the EC at 25°C (dsm^{-1}) by 0.34 (1 dsm^{-1}=1 m mho cm^{-1}).

Activity 10

What do you understand by the term electrical conductivity of soil?

3. Redox potential

One important chemical characteristic of soils related to soil aeration is the reduction and oxidation states of the chemical elements in the soils. Non-photosynthetic biological activity in the soil (e.g. fungi, bacteria, soil animals) derives energy from the oxidation of reduced substrate, which may be either organic (Heterotrophic metabolism) or inorganic (chemo-qntotropic metabolism) in nature. If a soil is well-aerated, oxidised states such as those of ferric ion (Fe^{+3}) manganese (Mn^{+4}) nitrate (NO_3^-) and sulphate (SO_4^{-2}) dominate. In poorly drained and poorly aerated soils, the reduced forms of such elements are found for example ferrous (Fe^{+2}), manganese (Mn^{+2}), ammonium (NH_4^+) and sulphides (S^{-2}). The presence of these reduced forms is an indication of restricted drainage and poor aeration. An indication of the oxidation and reduction status of systems is given by the oxidation-reduction potential or redox potential (Eh). It provides a measure of the tendency of a system to reduce or oxidise chemically and is usually measured in volts or millivolts. If Eh is positive and high, strong oxidising conditions exist. If it is a low and even negative, elements are found in reduced forms. Eh can be measured using the following equation.

$$Eh = E_0 + \frac{0.059}{N} \log \frac{[O_x]}{[Red]}$$

Eh = Electrode/Redox potential (mv)
E_0 = Standard potential of the system
N = Number of electrons in the system
$[O_x]$ = Electrons lost
[Red] = Electrons gained
In a well-drained soil, the Eh is in the range of 400-700 mv. As aeration is reduced, the Eh declines to a level of about 300-350 mv when gaseous oxygen is depleted.

Activity 11

What is the importance of studying the redox potential of soil?

4. *Cation Exchange and Cation Exchange Capacity(CEC)*

Clay colloids are normally negatively charged and they attract positively charged ions-cations or bases which are also replaceable. Therefore, they are called exchangeable bases or exchangeable cations and the phenomenon is known as **cation exchange**. Cation exchange reactions are reversible. Therefore in general the cation exchange is the interchange between a cation in solution and other cation on the surface of any active material such as clay or organic matter.

Activity 12

List the methods which help to create negative charges in colloids.

The cations which are held on the surface of soil particles are subject to exchange with other cations held in soil solution. For example a magnesium ion held on the colloidal surface is subject to exchange with two H^+ ions in the soil solution. Thus, soil colloids are focal points for cation-exchange

reactions, which have profound effects on soil-plant relations. This process is called cation exchange.
A substantial amount of organic and mineral acids is formed as the organic matter decomposes. The hydrogen ions thus will tend to replace the exchangeable calcium of the colloidal complex.

$$\text{Ca Micelle} + 2H^+ \longrightarrow \text{H Micelle H} + Ca^{+2}$$

The reaction takes place fairly rapidly and the interchange of calcium and hydrogen is chemically equivalent.
It is important to note that where sufficient rainfall is available to leach the calcium, hydrogen ions are entering and calcium and other bases are being forced out of the exchange complex into the soil solution.
In regions of low rainfall, the calcium and other salts are not easily leached from the soil as indicated. That is it prevents the removal of bases from the exchange sites, thereby keeping the soil neutral or above in reaction.
Now you may realise how the interaction of climate, biological processes and cation exchange help to determine the properties of soils.
The Cation Exchange Capacity (CEC) is expressed in terms of moles of positive charges absorbed per unit mass of soil. In other words, it is the total numbers of milli-equavalents of exchangeable cations which 100 g of soil contains. A soil has a cation exchange capacity of 1meq/1mg is capable of adsorbing 1 mg of H^+ or its equivalent per 100g of soil. The advantage of using equivalent is that all ions can be expressed in term of milli-equivalents.

Table 2.4 Cation exchange capacity of some soil colloids.

Fraction	meq/100grm of soil
Humus	200
Montmorillonite	80-150
Hydrousmica or illite	10-40
Kaolinite	03-15

Cation exchange reactions are reversible. This cation exchange can be measured quantitatively by describing the cation exchange capacity (CEC), Simply it is defined as the sum total of the exchangeable cations that a soil can adsorb.

By the standard methods, all the adsorbed cations in a soil are replaced by a common ion, such as potassium or ammonium, then the amount of adsorbed potassium or ammonium ions is determined.

The cation exchange capacity of given soil is determined by the relative amounts of different colloids in that soil and by the CEC of each of these colloids. Sandy soils have lower CEC than clay soils because the coarse textured soils are commonly lower in both clay and humus content. Cation exchange capacity is increased with increasing pH. At very low pH values (acidic conditions), CEC is generally low. Therefore CEC should be determined at a pH of 7.0 or above. Exchangeable cation are generally available to both higher plants and microorganisms. By cation exchange, H^+ from the root hairs and microorganisms replace nutrient cations from the exchange complex. The nutrient cations are forced into the soil solution where they can be assimilated by the adsorptive surface of roots and soil microorganisms.

Activity 13

How do you relate CEC with soil fertility?

5. Percentage base saturation of soils

Two groups of adsorbed cations have opposing effect on soil alkalinity. Both hydrogen and aluminium ions contribute to the concentration of H^+ in the soil solution. Adsorbed H contributes directly to the H^+ concentration. Al^{+3} does this indirectly through hydrolysis as shown below.

$$Al^{+3} + H_2O \longrightarrow Al(OH)^{+} + H^{+}$$

$$Al(OH) + H_2O \longrightarrow Al(OH)_2 + H^{+}$$

$$Al(OH)_2 + H_2O \longrightarrow Al(OH)_3 + H^{+}$$

Most of the other cations called exchangeable bases act to neutralise the soil acidity.

The proportion of the cation exchange capacity occupied by these bases is called the % base saturation (i.e. the extent to which adsorption complex of a soil is saturated with exchangeable cations other than H and Al). It is expressed as the percentage of the total cation exchange capacity. Thus, if the percentage base saturation is 80, 4/5th of the CEC is satisfied by bases and the other by H^+ and Al^{+3}.

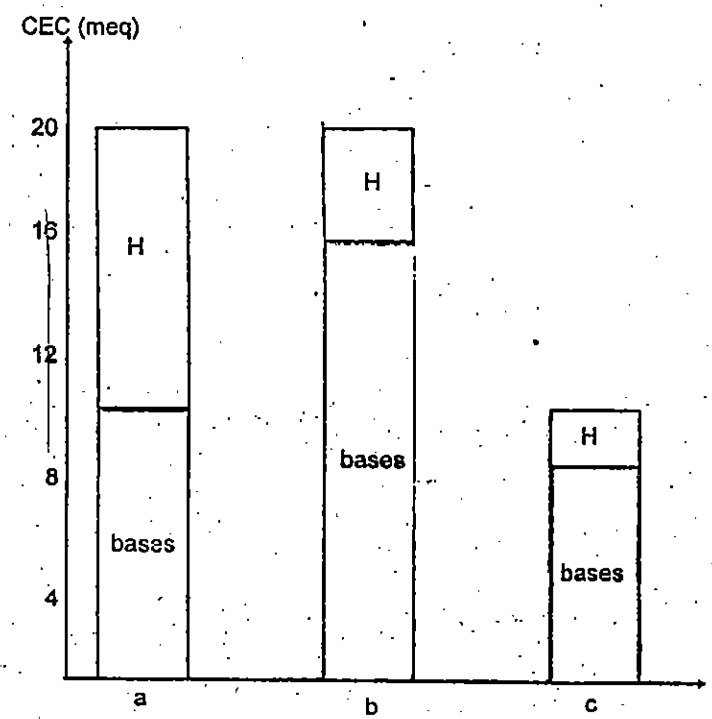


Figure 2.9: A diagrammatic illustration of the % base saturation in different soils.

The diagrammatic illustration shows the percentage base saturation in

(a) a clay loam

(b) the same soil limed

(c) a sandy loam

The clay-loam has a CEC of 20 m.eq. while the sandy loam has a CEC of 10 meq. With liming the percentage base saturation has increased from 50 to 80, while the CEC remains the same.

Activity 14

What is the importance of base saturation in soil?

6. Anion exchange

The adsorbed anions are commonly present in smaller quantities than the cations because the negative charges generally predominate on the soil colloid, especially in soils of temperate regions. The positive charges associated with hydrous oxides of iron and aluminium, some 1:1 - type clays, and amorphous materials, such as allophane, make possible the adsorption of anions. In turn, these anions are subject to replacement by other anions, just as cations replace each order. The anion exchange that takes place, although it may not approach cation exchange quantitatively, is important as a means of providing readily available nutrient anions to higher plants.

In anion exchange, like cation exchange, equivalent quantities NO^-_3 and Cl^- are exchanged, the reaction can be reversed and plant nutrients can be released for plant adsorption. Although simple reactions such as this are common, note that the adsorption and exchange of some anions including phosphates, molybdates, sulphates etc are somewhat more complex.

7. Organic carbon

Plant residues, as litter, branches, root detritus and exudates, comprise by far the largest fraction of the carbon entering the soil, although microbes and animals also contribute to soil carbon supply under natural conditions, the tops and roots of the trees, shrubs, grasses and other native plants annually supply large quantities of organic residues. Even with harvested crops, one tenth to one third of the plant tops commonly fall to the soil surface and remain there or are incorporated into the soil. Naturally, all of the roots remain in the soil. In the top 10 cm, soil usually contains about 1-3% carbon in organic compounds; there is often more in soil under grass or trees, forests and up to about 30% in peat.

These organic materials are decomposed and digested by soil organisms, and become part of the underlying soil by infiltration or by actual physical incorporation. Accordingly, the residues of higher plants provide food for soil organisms, which in turn create stable compounds that help maintain the soil organic levels.

Animals usually are considered secondary sources of organic matter. As they attack the original plant tissues, they contribute waste products and leave their own bodies as their life cycles are consummated. The soil animals such as earthworms, termites and ants also play on important role in the translocation of soil and plant residues.

2.4 Classification of soils

Soils are very variable in their characteristics and are as diverse as the landscape of the world itself. It is therefore necessary to classify them in a systematic and orderly arrangement based upon their differences and similarities. The details of soil classification is essentially dynamic in nature and keeps on changing as the knowledge and understanding of the soil increases. The scientific systems of soil classification are of a very recent origin and were developed mainly in Russia and USA.

The modern classification systems are based upon the study of the soil profile, which is the ultimate product, and reflection of all the soil forming factors and processes. The soils are classified into well-defined categories such as orders, sub-orders, great soil groups and sub-groups. These give a general understanding of the soils over large areas indicating their worldwide relationships. The lower categories are families, series, types and phases, and are more important in recognizing local differences and assessing productive capacities of soil for its use in agriculture for other forms of land-use.

Many soil scientists agree that classification systems that have been developed mainly in temperate climates may not be fully applicable to the soils developed under tropical and sub-tropical climates. In tropical Asia, scientific studies on soils have been undertaken for many decades by India, Sri Lanka, Indonesia and Thailand. Recently, soil surveys have been undertaken in Vietnam, Philippines, Malaysia etc. As a result the knowledge about the soils of our region is rapidly increasing. Based upon the information available from various tropical countries, a tentative scheme for classification of tropical and sub-tropical has been suggested by a scientist named Middleberg as the 1949 system.. However with recent advances about classification of soil, the most accepted method at present is 7th approximation and is described below.

Soil Classification- 7th Approximation

The soil scientists of the US Department of Agriculture have developed and evaluated through several stages or approximation. The 7th approximation has been officially adopted as the comprehensive soil classification system. Efforts have been made in India and in some other tropical Asian countries to test this system for adoption and modification if any, under the local conditions. Several important features are obvious in this new system of classification which makes it attractive.

These are as follows:

(i) the primary bases for identifying classes of the systems are the properties of soils, as found in the field - properties that can be measured quantitatively.

(ii) the nomenclature employed especially for the broader categories give a definite connotation easily understood in many languages since Latin or Greek root words are the bases for the names. For example, the name "Aridsols" indicates a soil order, and suggests dry conditions and includes soils of desserts.

(iii) it permits classification of soil rather than forming processes; it permits classification of soil of unknown genesis since only the knowledge of their properties are needed.

(iv) it permits greater uniformity of classification as applied by a large number of soil scientists. Differences in interpretation of bow a soil was formed, do not influence its classification under this new system.

(v) however, it should not be assumed that the new system ignores soil genesis. Since soil properties the basis for the new system, are often related directly to soil genesis, it is difficult to emphasize soil properties without at least indirectly emphasizing soil genesis as well.

It is not possible to generalize with respect to the kinds of soil properties used as criteria for soil classification. A few examples of criteria used include moisture, temperature, colour, texture and structure of the soil. Chemical and mineral properties such as contents of organic matter, clay, iron and aluminum oxides, pH% base saturation, and the presence of salts are also considered. Soil depth and presence or absence of certain diagnostic soil horizons is also among the most significant of the properties used as a basis for classification. Brief mention will be made of certain diagnostic horizons, to illustrate this point.

Categories of the new soil classification system

The order category is based largely on the morphology, but soil genesis is an underlaying factor. A given order includes soils whose properties suggest that they are not too dissimilar in their genesis. As an example, soils developed under grassland vegetation have the some general sequence of horizons and are characterized by a thick dark surface horizon high in bases. They are thought to have been formed by the same general genetic processes and are thereby usually included in the same soil order.

Sub- Orders are divisions or orders which emphasize genetic homogeneity. Thus wetness, climatic environment, and vegetation are characteristics which help determine the sub-order in which a given soil found.

Within a sub-order, diagnostic horizons are used to differentiate the **great soil groups**. Soils in a given group are considered as having the same kind and arrangement of horizons in their profile.

Sub groups are sub divisions of the great soil groups. The typical or central concept of a great group makes up one sub group . Other sub-groups have characteristics that integrates between those of the central concept and those of another great group.

The **family** category is not so well defined. However, properties important for growth of plants are used to differentiate families. This category permits the grouping of members of sub-groups having in common similar characteristics such as texture, mineral content, Ph, soil temperature and soil depth.

The series category is essentially a collection of soil individuals uniform in differentiating characteristics and in arrangement of horizons.

Distribution of soils in the tropical far east

A considerable part of the tropical Far East is occupied by soils in which the overriding pedogenic factor is characteristic of the local rather than the regional environment. Some have a well – developed soil profile, but factors such as location in poorly drained part of the landscape or situation on an unusual parent material, for example, a pure limestone, has exerted a predominant effect.

Weathering is strong in the humid tropics, being favoured by both warmth and moisture. The high rainfall frequently wets the soil throughout, with a surplus to penetrate to the ground – water. Thus constituents which can be carried in solution are continually removed; this is the process of leaching. The temperature and water content of a well-drained tropical soil favour a high level of biological activity.; including the continuos production of organic matter by the vegetation and its rapid degradation by macro-and micro-organisms in or on the soil. High levels of weathering, leaching and biological activity control the properties of tropical soils.

The strong weathering in the humid tropics leads generally to deep soils (2-15m or more). The silicates of igneous and metamorphic rocks are usually decomposed to a much greater depth than soil). For the same reason fresh rock fragments and grains of weatherable minerals are rarely found in soils of the humid tropics, thus excluding an important source of slowly available plant nutrients.

Complete weathering of all common minerals, except quartz, to kaolinite clay and to iron and aluminium oxides and hydroxides leads to soils richer in clay than would be the case in an otherwise similar temperate environment.

It is hard to separate weathering from leaching because weathering is much assisted by the removal of soluble products. The most soluble products are base cations such as sodium, potassium, calcium, and magnesium, which are depleted to very low levels in the great majority of soils in the humid tropics. This does not produce extreme acidity, a pH in water between 4.5 and 5.5 is typical, but leads to high exchangeable aluminium in the less- weathered soils which still have an appreciable cation – exchange capacity.

High exchangeable aluminium, specially it forms a large proportion of the exchangeable cations, is associated with aluminium toxicity, and this leads to the stunting of root growth in many tropical soils. Exchangeable aluminium is also very important in making phosphate insoluble in tropical soils.

Deep, complete weathering combined with leaching results in soils of low fertility in stable landscapes. N steeper slopes erosion and landscaping is constantly truncating the soil profile and exposing little – weathered parent material ;as probably for example at Sogeri, Papau. Transferred materials such as colluvium and alluvium, which are infertile in stable landscapes, may retain useful weatherable minerals where they originate in strongly dissected terrains.

Silica is very slightly soluble in water and is slowly leached, and even quartz suffers some attack. In time this has the effect of removing slit and fine sand particles from the soil, so that the mineral fractions of well-developed soils in the humid tropics have low silt/clay ratios and consist mostly of kaolinitic clay and coarse sand composed of quartz and ironstone nodules. Silica dissolved in the soil solution can combine with gibbsite (aluminium hydroxide) initially formed from felspars and other silicates to produce kaolinite, a process known as resilication. This may be one reason why soils derived from basic igneous rocks low in silica and containing no quartz are normally rich in gibbsite while soils derived from granite (with quartz) do not normally contain much gibbsite. Once formed, the kaolinite is very stable. The micaceous minerals common in shales may also weather to kaolinite, but the process may be relatively slow compared with the weathering of igneous rocks, partly because weathered shale subsoil tend to be impermeable.

The kaolinite nature of the clay of most well – drained, well-developed soils of the humid tropics has several important consequences. Such clay does not swell and shrink much on wetting and drying, so that prominent cracking defining a well- developed blocky or prismatic structure is rare. On the other had, kaolinite unmixed with humus or other clay minerals is hard to disperse in pure water; it flocculates to stable, fine aggregates so that subsoils are permeable, relatively resistant to erosion, and easily penetrated by roots.

Kaolinite has low action – exchange capacity especially at a low pH. Therefore, in soils of the humid tropics, nutrient-holding capacity is mainly a function of humus content and is very low where humus content is low, as is usual in subsoils, in fact in many cases more than half of the adsorbed bases in I m of soil are in the top 0.25m. Humus tends to both reduce the phosphate –

fixation capacity of associated mineral colloids and to release phosphate as it decomposes. It follows that the phosphate available to plants is also concentrated near the surface, falling to negligible values in the subsoil. These facts help to explain why roots are concentrated in the surface soil and a few in subsoils, even though many tropical subsoils are easily penetrated physically.

While differences in fertility are relatively slight the soils being described varie appreciably in colour, texture, and sesquioxide content and in the distinctness and nature of layering seen in the profile. As has been indicated they have usually been subdivided in soil classifications based on these characteristics, but it should be noted that none of the schemes are easy to apply, since, wherever the limits are set, intergrade soils and the data required to apply any specified criterion are often lacking. Two main groups are found. Typically, as in lowland Malaya, ultisols occur on young surfaces and relatively quartzose materials while oxisols occur on old surfaces and basic materials. In south – east Asia generally, ultisols are much more extensive, occupying about 55 percent of the region (acrisols 51 and ultisols 4 percent) according to FAO/UNESCO (1979), who mapped oxisols as dominant over only 4 percent (although detailed studies suggest this is an underestimate). On the steepest slopes ultisols pass to inceptisols., specially tropics, resembling European brown earths These are cambisols in the legend of the FAO/UNESCO map, on which they cover about 5 percent of south – east Asia, and are most extensive in mountainous areas.

Ultisols (red – yellow podzolic soils)

The distinguishing feature of utisols is the so-called "argillic' or 'textural B' (B) horizon in the sub soil. This has at least 1.2 times as much clay as the topsoil, and not infrequently has more than twice as much. The structure of the horizon is produced by fissures a few centimetres apart, given a medium to coarse, blocky structure with clay coatings on the surface of the fissures, which appear in this section to have uniformly fine particles with parallel orientation; features suggesting transport in suspension from higher horizons. Bright yellow brown or reddish colours are usual in the argillic horizon, often becoming paler above, for example, throughout Malaysia, and in the non-volcanic parts of Indonesia. Ultisols with pale, often greyish, colours throughout the profiles are common in Burma, Thailand, and the countries bordering the lower Mekong river, but only occasionally in the wetter parts of south – east Asia). In Africa, ultisols are found on Upper and Middle Pleistocene erosion surfaces within the humid tropical belt. They are replaced on the much more extensive older surfaces and peneplains by oxisols In the young landscapes of south – east Asia affected by extensive recent mountain building, old erosion surfaces and peneplains are rare (Baillie and Ashton 1983) and ultisols are therefore much more extensive than oxisols – the reverse of the situation in Africa.

Ultisols have, by definition, a low base content (less than 35 percent saturated in the subsoil). Acrisols (FAO/UNESCO 1974) are less than 50 percent saturated. South – east Asian utisols are particularly poor in nutrients and often have high levels of exchangeable aluminium (Buurman and Dai 1976; Buurman, Rochimah, and Sudihardjo 1976). If forest is removed they are highly susceptible to erosion.

The surface horizons of ultisols, having relatively sandy texture and weak structure unstabilized by clay coatings are especially vulnerable. However, Kenworthy (1971) suggested that surface run-off under undisturbed forest is small and occasional, perhaps no more than 3 to 5 percent of the 2000 mm mean annual rainfall on steeply sloping land in Malaya. The loss from virgin lower montane rain forest at the Cameron Highlands, Malaya, was only 24.5m3 km-2year-1. this increased ten and thirty times under tea and vegetable gardens respectively (Shallow 1956 in Daniel and Kulasingam 1974), and also increased greatly after logging (19.3). Since the topsoil commonly contains 80 percent or more of all the plant – available mineral nutrients in the profile, erosion losses seriously prejudice forest regeneration (20.2)

Oxisols (ferralsols, latosols)

Oxisols (Soil Survey Staff 1975) are characterized by the presence of an oxic horizon in the subsoil and usually by the absence of an argillic horizon. The oxic horizon has a low cation – exchange capacity (less than 10 milliequivalents at soil pH and less than 16 at ph 7 per 100 g of clay) and the clay has a low activity (that is, does not disperse appreciably when shaken in plain water). It lacks significant amounts of weatherable materials (2,1 lattice clays, feldspars, micas, glass, or ferromangnesian minerals). Very sandy soils are excluded from the oxisol order.

Typical oxisols in south – east Asia are generally confined to intermediate to basic igneous rocks, for example, basalt (Fig II.4) although some are known on other parent materials, especially old alluvium, in stable landscapes. Oxisols on basic igneous rocks have high sesquioxide contents, often with gibbsite as well as ferric oxide minerals present, and they have a characteristic strong coloration, frequently dark red. The clay content is high (50-95 percent) but stickiness is low, so that the subsoil is friable and easily penetrated by roots.

Oxisols are strongly weathered, and in humid climates are strongly leached. They have a low base content and a very low capacity to retain cations, lower than ultisols. A high proportion of the cations held is usually aluminium and in the tropical Far East oxisols are nutrient poor. High permeability and resistance to dispersion make oxisols very much less prone to erosion than ultisols. On the other hand ready drainage of most pores limits the retention of water. Water held against gravity but available to plants is usually about to 10 percent by volume in oxisols, regardless of texture. In ultisols texture matters more, available water being low in sandy soils, but generally ultisols have higher available water capacities than oxisols, and andoslos much higher Integrated to inceptisols are also common on steep slopes.

Laterite

"Laterite" formation implies the redistribution and concentration of sesquioxide within the profile. Features which may be broadly grouped under this heading are widely distributed in the zonal soils of south – east Asia., both ultisols and oxisols. They are rare in the wettest areas, become commoner in climates which are seasonally or absolutely drier and are most common in the regions of monsoon climate outside the rain – forest zone proper (II.II).

The commonest and most widespread form of laterite is a subsoil layer of compact clay with bright reddish and pale yellow or grey motting, lacking any hardening. An alternative name for unhardened but hardenable laterite is plinthite. In seasonally dry areas similar material may be found near the surface as a continuos layer or as blocks with the reddish phase hardened (Fig. II.5); unhardened mottled clay commonly occurs below. In Malaya such profiles are commonest in Malacca (hence their designation - for example, in Fig. II.3, and II.6 – as the Malacca soil series) and in Kedah. It is notable that these are regions with a slightly more marked dry season than much of Malaya; this emphasize the association between widespread laterite and seasonally dry climates. The mottled clay impedes rooting and hardened laterite prevents it, so such soils are considered relatively infertile although they are not more weathered than comparable "non- lateritic" soils, Other "lateritic" profiles are those with ironstone gravel (Fig. II.6), either developed *in situ* or transported, and those with cementation due to the seepage of iron – rich water at a spring line as in Fig. II.3. General accounts of laterite have been given by Sivarajasingham, Alexander, Cady and Cline (1962) and Prescotte and Pendleton (1952). Occurrences in Indonesia have been described by Mohr and van Baren (1954) and in Malaya by Panton (1957) and by eyles (1967) fig. II.6). The greatest extent is in Indonesia, associated with oxisols, particularly in south – east Kalinatan, and also with ultisols. Hardened laterite is also important in oxisols in Vietnam and Cambodia, in ultisols in Thailand, and in alfisols in Laos.

Soils on volcanic ash.

Volcanic activity has been very extensive on many of the islands of south – east Asia, and most particularly on the Sunda group (Sumatra, Java, Bali, Lombok, Sumbawa, and Flores). Java alone has 112 volcanoes, of which 34 have been active since 1600. Where volcanoes are formed by magma relative low in silica most of the discharge is lava with a subordinate quantity of volcanic ash, but where the magma is high in silica, ash predominates. Volcanic ash is principally composed of pumice debris, that is to say, shreds of silicate glass, although crystals of igneous minerals usually occur in it also.

The soils first formed on recent lavas are usually tropepts and resemble brown earths. With time these evolve in the lowland to give oxisols (latosols), which are described in II.3. Andosols are the initial product of weathering of volcanic ash in most climatic zones, their distinctive properties depending on the presence of allophane. Allophane is an X-ray amorphous association of hydrous silica and alumina.

2.5 Soil biota and soil fertility

As described earlier in this lesson, soil fertility is defined as the ability of a soil to supply essential elements for plant growth without a toxic concentration of any element. In order to achieve this target, the soil should be in optimum condition as explained before with respect to its components and properties. In addition, most of the essential elements in the environment are limited and their circulation is one of the limiting factors in relation to the plant growth. To make this process to occur, soil organisms play a key role as decomposers in the soil environment. Therefore, a study on soil organisms is very essential to understand the proper path of this cycling process without which no plant will be able to survive on this earth. It is generally accepted that it is the microorganisms in the soil which play the most important role in the release of minerals and CO_2 for plant growth. Knowledge on different types of microorganisms is important to understand soil fertility.

Kinds of microorganisms in soil

Soil is the home of a great variety of microorganisms than any other environment. One of the most striking features of the soil microflora is, its diversity. Almost any soil sample will consist of fungi, bacteria, cyanobacteria, algae and viruses belonging to innumerable genera and species. It is the relative proportions of the different groups that vary with the environment. The relative proportions of various kinds of organisms in soils are indicated in Fig. 2.10.

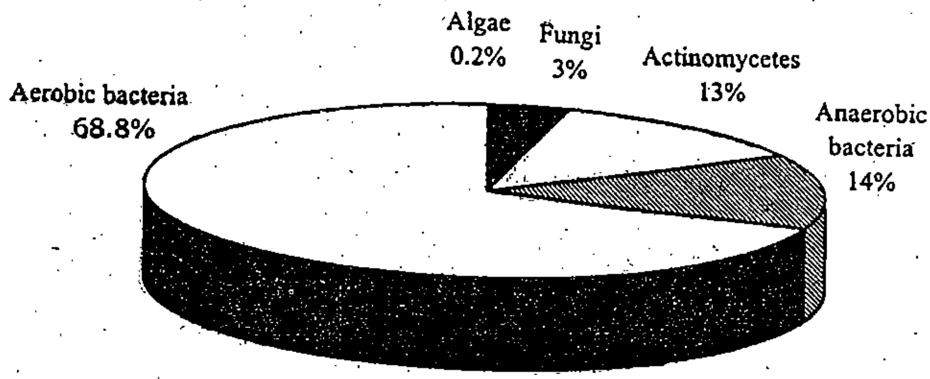


Figure 2.10: The relative proportions of various kinds of microorganisms found in soil.

The microorganisms isolated from soil can be placed in two broad divisions: the **indigenous or autochthonous** species that are true residents of the soil and the **invaders** or **allochthonous** organisms. Indigenous populations may have resistant stages and endure for long periods without being active metabolically. This is discussed in detail in section 1.5. During favourable conditions they participate in the bio-chemical functions of the community. Allochthonous species, by contrast, enter the soil with precipitation, diseased tissues, animal manure or sewage sludge. They may persist for sometime in a resting form but do not participate in a significant way in ecologically significant transformations or interactions. Soils may harbour pathogenic microorganisms depending on the nature of animal and plant tissues and waste it contains. These microorganisms may cause infections in man.

Microorganisms colonise only a part of the available surface area of the soil as shown in Fig. 2.11. It is seldom possible to determine accurately the organisms in soil. The types of microorganisms especially fungi and bacteria found in soils will largely be a reflection of the methods used to cultivate them.

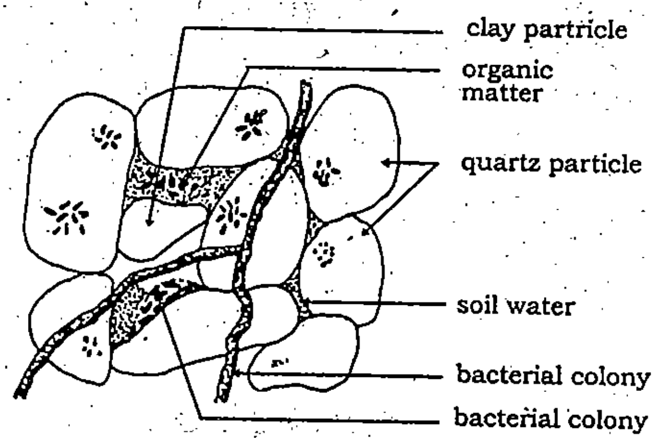


Figure 2.11 - Diagrammatic representation of microbial colonisation on soil particles.

Bacteria

The most abundant group of organisms in soils are bacteria. A grain of fertile soil may contain more than 10^9 bacteria. They are of great importance in various soil processes such as cycling and transformation of carbon, nitrogen, sulphur and other minerals. They are responsible for much of the decomposition of cellulose, protein, pectin etc. Some are harmful either competing with plants for nutrients or causing diseases of plants and other soil organisms. Many bacteria are rod shaped. Many species are flagellated and capsulated. The presence of a large number of capsule-forming bacteria is thought to have the effect of improving the crumb structure of the soil by cementing together mineral particles and humus.
A number of *Bacillus* species in most soils are spore-formers. Most bacteria are **heterotrophs** including those which can use many organic compounds such as sugars, cellulose, chitin, organic acids, alcohols and hydrocarbons. Some are **autotrophs** which include nitrifiers, sulphur oxidisers and iron bacteria. Many are **aerobes** and some are **anaerobes**. Depending on the soil temperature they can be **mesophilic** or **thermophilic**.

Most common species are those of *Bacillus, Clostridium, Arthrobacter, Pseudomonas, Rhizobium, Azotobacter, Nitrobacter*. Actinomycetes are present in surface soil and also in the lower horizons to considerable depths. In abundance, they are second only to true bacteria. They occur as conidia or as the vegetative hyphae. Actinomycetes include species of *Nocardia, Streptomyces* and *Micromonospora* (Fig. 2.13). Actinomycetes are abundant in neutral or alkaline soils but they cannot tolerate acid conditions. Water logging and anaerobic conditions limit the development and spread of actinomycetes in soils.

Bacteria tend to grow as individuals or microcolonies (often less than 10 cells) on the surfaces of soil particles and roots. Clay and organic colloids adsorb bacteria particularly in the presence of tri- or poly- valent cations. On the roots, bacteria are concentrated along the epidermal cell boundaries possibly where exudation is greatest. Actinomycetes actively grow through the soil from a food base.

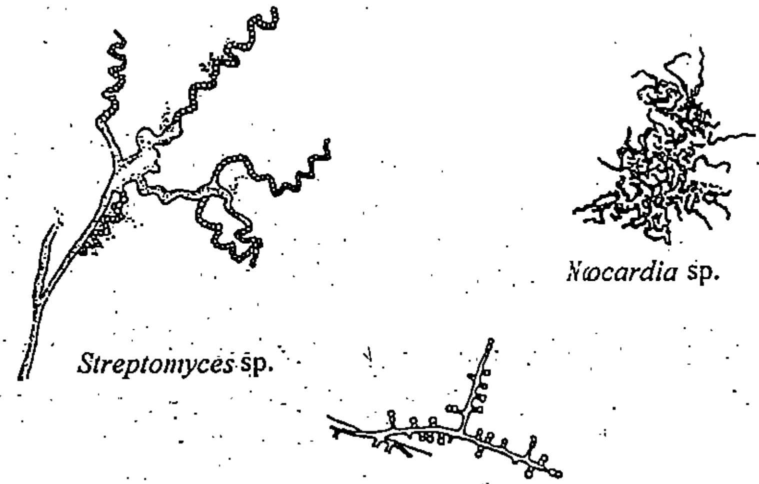


Figure 2.13 - Some soil actinomycetes (x 1000).

Cyanobacteria such as *Calothrix*, *Nostoc* and *Oscillatoria* (Fig.2.14) are often primary colonisers and play a key role in the transformation of bare rocks to soil. As they die and decay the environment becomes more suitable for higher plants. Because of their capacity to utilise simple inorganic compounds, these microorganisms seems to be among the earliest living forms in surroundings where life has been eliminated by natural or artificial agencies. Some species form continuous mats on the soil surface under damp conditions to an extent that distinctive visible blooms appear at the surface. Extensive blooms bind soil particles together into large aggregates. Flooded paddy field is an environment where cyanobacteria could have a great agronomic significance. The microbiological action may be associated with the fixing of atmospheric nitrogen, the release of oxygen or the excretion of products stimulating plant development. Among the more common inhabitants of these water logged areas are species of *Anabaena*, *Calothrix*, *Oscillatoria*, *Nostoc* and *Tolypothrix* (Fig 2.14). This group is of special interest because many are nitrogen fixers and thereby contribute to soil fertility.

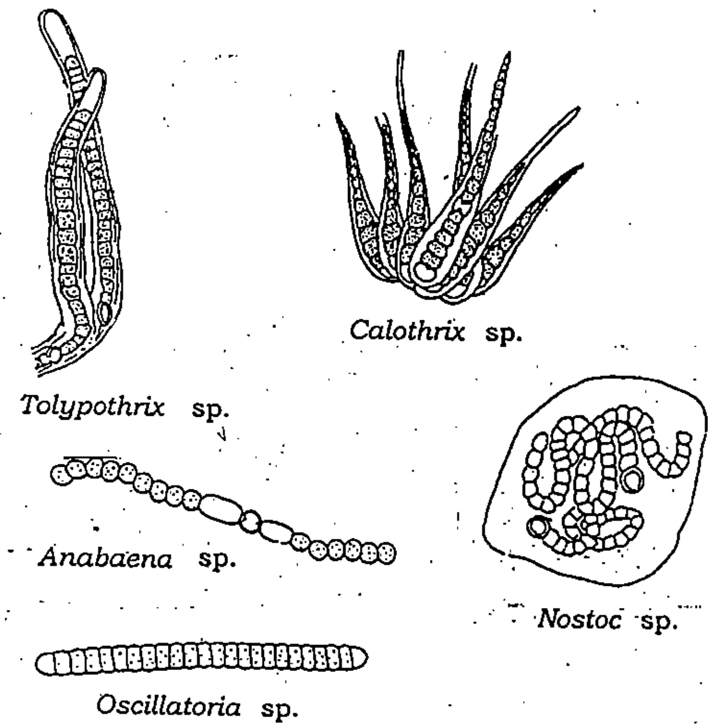


Figure 2.14 - Some common cyanobacteria in soil (x 400)

Fungi

Fungi are of equal importance to bacteria in contributing to soil processes and plant nutrition in neutral and alkaline soils. They usually tolerate acid conditions better than bacteria and for this reason are more important than bacteria in acid soils.

Hundreds of different species of fungi inhabit the soil. They are being most abundant near the soil surface where oxygen is readily available. They are present as both mycelia and as spores. Both, the numbers of propagules and diversity of fungal species decrease with increasing depth in soil. Fungi follow the distribution of organic matter and thus are more numerous in the litter zone. A large quantity of the mycelium present in the lower soil horizons is usually dead.
Soil fungi are of many different types, ranging from microscopic kinds to the toadstools with their large and complex fruit bodies. Some of the more common species of micro fungi are *Penicillium, Mucor, Rhizopus,*

Cladosporium, *Fusarium*, *Aspergillus*, *Trichoderma*, *Cephalosporium*, and *Curvularia* (Fig. 2.15). Soil also generally possesses a distinct yeast flora consisting of species not common in other environments e.g., *Aureobasidium pullulans*. Species of yeasts characteristic of leaf surfaces may also be washed down by rain into the upper layers of soils. Best known macrofungi are the basidiomycetes which include mushrooms, toadstools, puff balls, stinkhorns, birds-nest fungi and others. Most of them are saprophytes playing an important role in litter and wood decay. Soil also contains mycorrhizal fungi.

Most soil fungi contribute to the essential processes of decomposing complex organic constituents of plant tissue in soil such as cellulose, lignin and pectin. This aspect is discussed in detail in lesson 3 of this volume. Some are harmful causing economically important root diseases. There are also some that attack and parasitise soil animals particularly the nematodes and protozoa (the predaceous fungi). Fungi also contribute to the improvement of the physical structure of soil. Fungal mycelia form extensive networks around soil particles. These have an important function in binding the soil particles together, helping to maintain a good soil structure for plant growth. It gives the soil a crumbly structure.

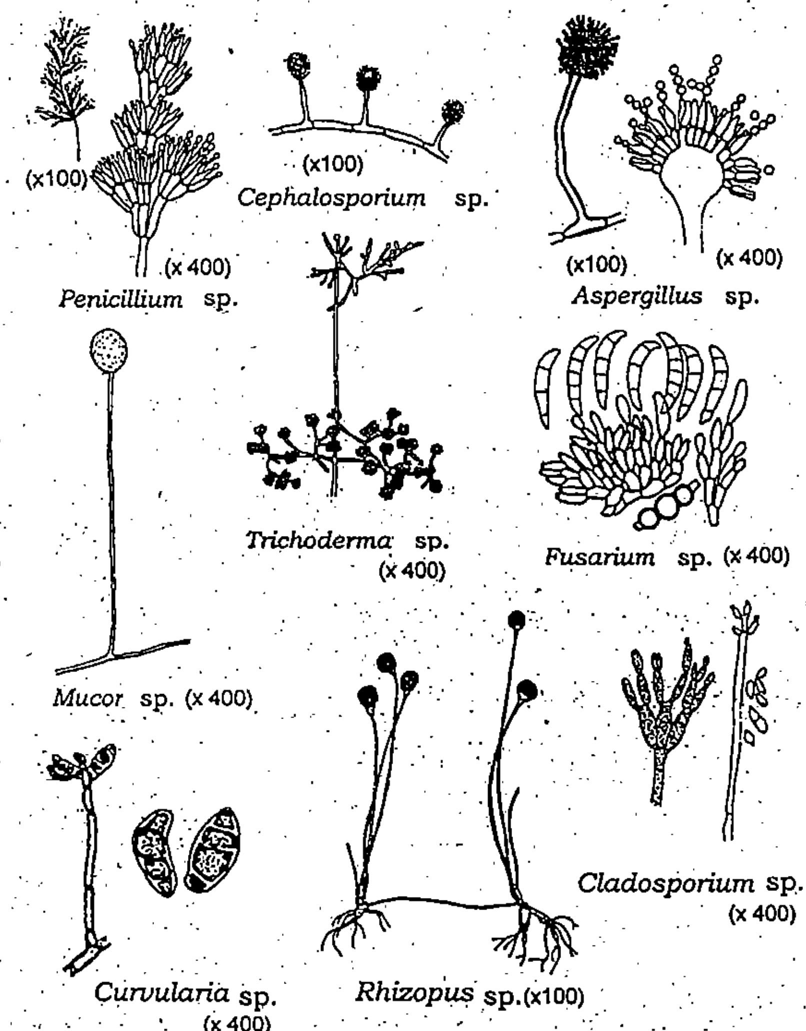


Figure 2.15 - Some common soil fungi.

Fungi can actively grow through the soil from food base. While they may originate in an organic particle the hyphae can ramify through the air spaces. They do not grow as discrete compact colonies as seen in pure cultures. Instead hyphae grow, come upon a food base, cover the surface of food base by branching and as the substrate is exhausted hyphae grow off again. The original hyphae may have died and lysed. Sporulation occur on the soil or litter surface or in the large pore-spaces. Fungal spores, like bacteria could be adsorbed to soil particles.

Algae

Algae are found most abundantly either on the soil surface or just below the surface, provided that the soil is sufficiently moist. This is also because they need light for photosynthesis. In desert, denuded and other barren soils algae contribute to the accumulation of organic matter in the soil. They also have the ability to corrode and weather rocks. Green algae are the group most commonly represented in the soils. Species include *Hormidium, Pleurococcus, Chlorella, Chlamydomonas* and diatoms such as *Pinnularia, Navicula* (Fig.2.16).

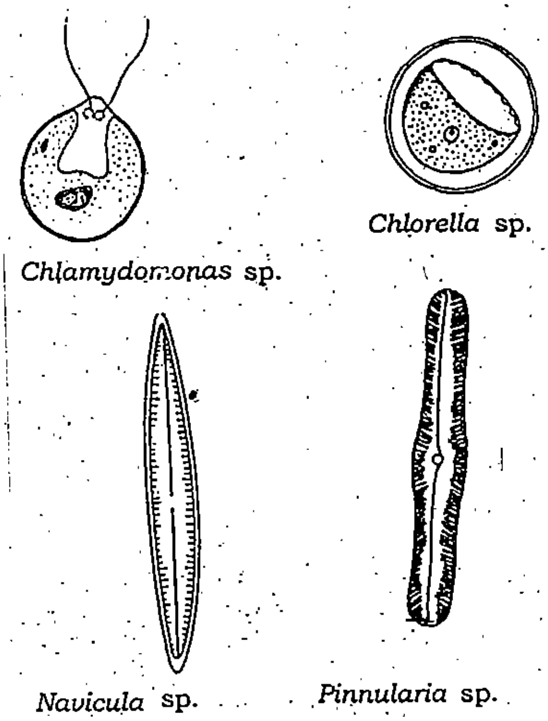


Figure 2.16 - Some common soil algae (x 400).

In general sense, the algae cannot be considered as contributing appreciably for soil fertility. One of the major algal functions in soils is an outcome of their photoautotrophic nutrition. This function is in the generation of organic matter from inorganic substances. Algae also contribute to soil structure and erosion control. They bind together soil particles and by forming surface bloom reduce erosion losses.

Unicellular algae occur singly in water films. Filamentous algae may occur in almost single species as discrete colonies which can be seen with the naked eye.

Viruses

Since viruses are obligate intracellular parasites, they do not 'live' or multiply in soils. They only survive by some means in the soil. Some of the more stable plant viruses e.g., tobacco mosaic virus may remain infective in the soil for several months or longer. There are root-infecting plant viruses which are actively transmitted through the soil by their vectors such as nematodes and fungi (e.g., tobacco rattle virus by nematodes and tobacco necrosis by fungi). These viruses persist for varying periods of time in the soils through their vectors. Very little information is available on the occurrence or survival of animal viruses in soils. There are a number of types of bacteriophages or bacterial viruses which can infect the bacterial species living in soils. Whenever a bacterial species is present in soil, its phage also can usually be found. There are also viruses which infect and cause lysis of cyanobacteria, soil fungi, algae and protozoa.

Protozoa

Soil protozoa are classified on the basis of their means of locomotion. Some move about by virtue of flagella (e.g., *Tetramitus*) others by means of short cilia (e.g., *Colpoda*), and a third group by temporary organelles known as pseudopodia (e.g., *Biomyxa*, *Euglypha*). Some protozoans are illustrated in Fig. 2.17.

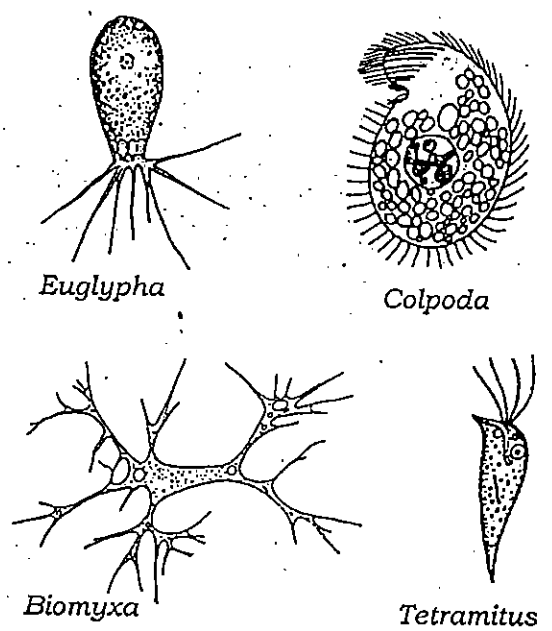


Figure 2.17 - Some common soil protozoa (x 400).

These organisms do not constitute a large portion of the biomass of the microbial community. They are found in greatest abundance near the surface of the soil particularly in the upper 15 cm. Generally the population is most dense where bacteria are especially numerous in the soil profile. The vast majority of these protozoa are on **phagotrophic** i.e., directly feed upon microbial cells or other particulate matter. The ingested food particle is surrounded by a vacuole where digestion takes place.

Activity 14

To what major groups do soil microorganisms belong?
List the characteristic features of the major groups of soil microorganisms.

Up to now, we learnt that microorganisms are very important and useful in decomposition process and also the end product of the decomposition process is the organic matter. From many centuries ago, it had been observed that the productivity of a soil was more or less directly related to the amount of organic matter in that soil. It is common knowledge that farmers all over the world, in designating soils which they consider highly fertile, usually select the dark coloured ones. They make this choice because experience has taught them that such soil are usually more productive than the light coloured ones, and hence in selecting dark coloured soils, farmers are involuntarily acknowledging the value of organic matter in soil.

In general, we can begin with the statement that organic matter exerts a controlling influence on soil properties, including productivity, and without it, the surface layers of the earth's crust could hardly be correctly designated as soil. The organic matter content of soil is one of the most important and also most easily exhausted resources. Soil organic matter from living or dead plant and animal residue is a very active and important portion of the soil. **It is the nitrogen reservoir, it furnishes large portion of soil phosphorus and soil sulfur; and protects soil against erosion; and it loosens up the soil to provide better aeration and water movement.**

Sources of soil organic matter

In addition to the remains of higher plants, the bodies of small organisms such as worms, insects, moulds, fungi and bacteria contribute to the organic matter of most soils in appreciable quantities.

Man influences the amount of organic matter in soils a great deal by the addition of it into soils in many forms. **Crop residues** are important sources of organic matter added back to soil. The leaves, stems, roots and mature plant material such as straw, stalks of cereals, stems of crops. etc. are ploughed back into farmlands purposely. They decomposed slowly and little nitrogen or other plant nutrient elements are released within a long period. **Green manure** plants are also important sources of man introduced organic matter in to soils.

Many plants are nowadays grown for cover and green manuring purposes. They aid in conserving soil organic matter by virtue of the fact that they decrease soil erosion losses: they conserve plant food-nutrient elements by reducing leaching losses; when turned under, they increase the supply of soil organic matter and with leguminous green manure crops, the total supply of nitrogen is increased, provided that the legumes have been inoculated with N-fixing bacteria.

Composts of plant residues are also excellent sources of organic matter, if they have been well prepared and if sufficient nitrogen, phosphorous and lime have been used to bring about rapid decomposition. When compost material is added to soil, they leave considerable amount of organic matter in the soil and some of the nutrients they contain are readily liberated in unavailable form.

Another source of organic matter to soil comes from the man's use of **organic fertilizers**. Dried blood, faecal matter, cotton seed meal, sewage sludge, etc. are common components of organic fertilizers.

Activity 15

List the reasons to say that green manure is very important source of organic matter in soil.

2.6 Beneficial effect of organic matter on soil characteristics and productivity

There is no doubt that highly decomposed forms of organic matter, if present in appreciable quantities have a marked influence on **the chemical, physical and biological properties** of mineral soils. However, a liberal quantity of humus aids in maintaining a desirable structural condition in all classes of soils.

2.6.1 Physical functions of organic matter in soil

Many of the useful effects of organic matter on soil are purely physical. Organic matter **increases the water-retaining power** of soils, **decreases water runoff losses, improves aeration** especially on the finer-textured soils, and produces a better **soil structure** (tilth) by **encouraging granulation**. All these effects will undoubtedly reduce the damage that might be done by water or wind erosion, to a particular soil type. Organic matter serves to bind sandy soils which are easily taken away by wind, thus reducing wind erosion.

Humus as a part of organic matter assumes colloidal properties and as such, has a very **high adsorptive capacity** for **water** and often acts like a sponge. Soil water is also retained in the small pores or air spaces in between the well granulated soil particles. In the more sandy soils, these spaces frequently are too large for maximum water retention, and **humus tends to partially fill** these large spaces and make them a **more effective size for water holding**. The humus also tends to pull the sand particles together, **thereby increasing water retention.** In clay soils, the pore spaces between the mineral particles frequently are too small for the greatest moisture storage. Organic matter improves this condition by forcing soil particles apart, thus increasing the ability of clay soils to retain water. Thus, you will realize that the water-holding capacity of all mineral soils can be **increased by raising the humus level.**

You may recall that 'tilth' refers to the physical condition of a soil. A soil in good tilth is easily cultivated, loose and mellow, and is characterized by being well granulated. Sandy soils are not well granulated because they lack sufficient binding material to hold the particles together. Organic matter added to sandy soils **promote granulation** greatly by binding sand particles together into cluster, thereby improving the **tilth of these soils.** Clay soils are also poor in tilth if the organic matter content low, because of clodding. The thorough mixing of organic matter in these soils greatly decreases clodding while **promoting granulation**. Organic matter makes clay soils less sticky, enables them easier for roots to penetrate. The degree of soil aeration which is so vital for root growth and which is determined to a large extent by the structure or tilth of the soil, is greatly improved by the presence of organic matter.

The **dark colour** of soils is also due largely to organic matter. It is known that dark-coloured surfaces **absorb radiant energy** more than lighter coloured surfaces. Thus, dark - coloured soils could **warm up faster than lighter** - coloured soils which may be similar in all other respects, including the moisture content. It must be apparent to you that under certain conditions, as in crop-sown fields, this would **hasten seed germination** and a more **rapid early growth of plants** which may be highly desirable.

Activity 16

How do you explain the high adsorptive capacity for water of humus?

2.6.2 Chemical functions of organic matter in soils

Organic matter improves the soil chemically by serving as a 'storehouse' or supply of plant nutrients. As the organic materials are decomposed, the plant - food elements contained in them are **gradually released.** Most of the soil nitrogen is held in organic form. Nitrogen cannot be held in soils in humid regions to any great extent, except in organic combination. The organic -N undergoes conversion to NO^{-3} under normal soil conditions and in the absence of growing plants most of the NO^{-3} may leach out. Decomposition of organic matter favours the release of plant-food elements from the soil minerals. Various **organic and inorganic acids are produced** in soils when organic matter decays, and they have a very **pronounced dissolving effect on soil minerals.** Obviously, the carbonic acid that results from the dissolving of CO_2 is a very efficient dissolving agent for soil minerals.

We have learnt in the preceding sections of this lesson that after the organic matter has undergone considerable decay, it usually assumes a colloidal state. The **colloidal properties** thus exhibited as explained earlier, have an important **influence on the soil productivity.** The organic-colloids with their great capacity for **cation-exchange** hold valuable plant nutrients in available from. Furthermore, these organic colloids have a strong ability to absorb or hold the constituents of fertilizers and nutrients released form soil minerals, thus **decreasing their rate of loss by leaching.** The colloids may also **act as buffers** in the soil, retarding the process by which changes in soil reaction (acidity or alkalinity) are produced.

Activity 17

Explain the high CEC of humus with reasons?

2.6.3 Biological functions of organic matter in soils

Organic matter improves the soil for the **growth of micro organisms.** It serves as a **source of food and energy** for the majority of **the soil micro-organisms.** There are some evidence that certain organic -N compounds can be directly absorbed by micro organisms as well as higher plants readily. For example, some amino acids such as alanine and glycine can be absorbed directly. The beneficial effects of even exceedingly small amounts of organic compounds added to soil might be explained by the presence of **growth-promoting substances**. It is more than likely that vitamin-like compounds are developed as **organic decay proceeds** and that these may at times stimulate both higher plants and micro organisms. It can be said again that, without the presence of organic matter to **supply food and energy** for the **soil microbes**, the plant food elements in organic matter could not be changed to usable forms.

Detrimental effects of soil organic matter in soil

Let us now explain some harmful or detrimental effects of organic matter in soil.

Many benefits of organic matter in the soil are counterbalanced in certain situations by detrimental influences. Organic matter is an energy and **carbon source for many disease organisms**, ensuring their longer survival in soils. Excessive amounts of organic matter are physically difficult to incorporate into soil and hinder easy planting.

Numerous plants contain or produce **phytotoxins** which make them undesirable as organic matter; so any plant material should not be used to incorporate into the soil. Unfortunately the problem cannot always be avoided because the decomposition of many crops produces such toxins. **Alleopathy** is also active. Alleopathy is any direct or indirect harmful effect of one plant on another through the production and liberation of toxic or inhibiting chemical compounds into the environment.

Activity 18

Suggest some more problem associate with soil organic matter.

Amount of organic matter and its distribution in representative soils

Although highly important in soils, organic matter makes up only a very small fraction of the total weight of mineral soils. The actual amount of humus found in any soil depends on texture of surface and sub soil horizons, topography, drainage climatic factors, and native vegetation. The amount can vary very widely: it can range from a trace in the very sandy desert soils to 12-15% or more in soils under grassland. Generally speaking, coarse-textured soils (clays, clay loams, etc.) because the sandy soils did not originally support dense vegetation, are more subject to leaching losses and, rapid decomposition usually occurs in them because of their better aeration.

Humus is usually concentrated in the uppermost soil layers and diminishes rapidly in the sub soil. This is readily explained by the fact that most of the organic residues in both cultivated and uncultivated soils are incorporated in or deposited on the surface. This increases the possibility of organic matter accumulation in the upper layers.
Since most of the organic matter is found in the topsoil, any erosion that occurs will result in significant organic matter losses. Furthermore, it is the topsoil that is disturbed by various tillage operations, which greatly increase the loss in organic matter and nitrogen.

Summary

Lithosphere is a major component of an environment. The major non-living components are soil, rocks and sediments. Whereas, living components makes up basically of microorganisms. Soils by being the major non-living components exhibits various properties. The knowledge of their proportions are immensely useful to understand the functions of soil. The living components of soil play a major role in maintaining the quality of soil. Soil in the upper crust of the earth and as the productive part of the Lithosphere, should be at a particular condition with which all plants are able to grow well.

Objectives

- After studying this session, you should be able to explain
- the structure and composition of the earth
- different components of the lithosphere
- characteristics of soil in relation to plant growth
- how microorganisms help to make good quality soil
- the contribution of organic matter to improve the properties of soil

Uttar Pradesh
Rajarshi Tandon Open University

PGD-ESD-04

UNDERSTANDING THE ENVIRONMENT

BLOCK 2

THE ATMOSPHERE

Unit - 1
The Atmosphere

Unit 3

The Atmosphere

Contents

Introduction

The atmosphere can be described as the envelope of air surrounding the Earth and bound to it by virtue of the Earth's gravitational attraction. Life exists on Earth mainly because of the presence of this envelope of gases around it. The atmosphere helps us in many ways. It maintains the temperature on the Earth's surface within limits tolerable to plant and animal life. It filters out high energetic harmful radiations coming from the sun, provides life-giving oxygen to animals and human beings, and carbon dioxide to plants and trees. It supports the formation of clouds and rain maintaining the hydrological cycle, which provides water to plants and animals. The presence of the atmosphere also makes the earth a place surrounded with sound, and makes flying, both by animals and machines, possible. As such, the atmosphere plays a key role in life on Earth.

The atmosphere is a dynamic entity. The physical parameters that describe its behaviour, such as the pressure, temperature, and relative humidity vary from day to day at a given place and also from place to place at a given time. Such changes are brought about by the 'chaotic' nature of the atmosphere, is a result of the uneven heating of the globe by solar radiation, due to the inclination of the Earth's axis to the plane of its orbit around the sun. The changes in pressure causes winds to develop, which carry air masses from one place to another. The upward movement of air parcels results in their expansion, which in turn causes the temperature to drop. Similarly, downward movement of air causes temperature to rise. These temperature fluctuations also cause the rainfall patterns to change. An understanding of the behaviour of the atmosphere therefore would require a knowledge of not only the radiative, dynamical, chemical and physical processes that take place within it, but also its interactions with land, vegetation and the oceans.

3.1 Composition of Air

The composition of air depends on the moisture content in it, which could vary between 0-4%, depending on the location and the season. We will first consider the atmosphere in general and then some special features of the moist atmosphere.

3.1.1 Atmospheric Constituents

The atmosphere comprises mainly two gases, nitrogen and oxygen, which contribute 99% to its composition. The balance 1% comprises a large number of gases, some inert, and some chemically active. The individual gases present in the atmosphere are referred to as its constituents. The major and the inert constituents have long lifetimes, while the chemically active gases have short lifetimes. Among the inert gases present are Argon, Helium, Neon, Krypton and Xenon.

The photo-dissociation or photo-ionization of the major constituents such as nitrogen, oxygen, carbon dioxide and water vapour produces atomic constituents such as O, N, and H. The interactions of these atomic species among themselves and with the major gases produce a large number of molecules. In addition, gases of terrestrial origin such as N_2O, CH_4 and many industrial gases such as cholo-fluoro-carbons (CFC) add to the composition of the atmosphere. The combustion of fossil fuels and biomass also add a variety of gases including CO_2, CO, SO_2 and many organic compounds. The atmosphere also has a large number of heavy metals in very minute quantities.

Long-lived Constituents

The turbulence and chaotic nature of the atmosphere, particularly in the lower atmosphere, keeps the long-lived gases well mixed. As such, their composition or the mixing ratios remains the same as at ground level, up to about 100 km. The region up to this he
ight is homogeneous, and is referred to as the **Homosphere.** Above this height, these gases separate out according to their own molecular weight and have their own profiles under diffusive equilibrium. The concentration of a constituent gas in the atmosphere is generally expressed as a volume-mixing ratio (r_v), which is defined as,

$$r_v = N_i/N_d, \qquad 1$$

where, N_i is the number density or concentration of the i[th] gas, and
N_d, is the number density or concentration of dry air.

The corresponding mass mixing ratio, r_m, could be then expressed as,

$$r_m = \frac{M_i N_i}{M_d N_d} = \frac{M_i r_v}{M_d} \qquad 2$$

where, M_i and M_d are the respective molecular weights.

The ground level concentrations of the long-lived gases, in terms of their volume mixing ratios, are given in **Table 3.1**. The so-called greenhouse gases, CO_2, N_2O, CH_4 and CFCs, which control the Earth's climate, also have long life in the atmosphere. These will be discussed later in this unit.

Constituent	Composition (%)
Nitrogen	78.08
Oxygen	20.95
Carbon dioxide	0.036
Argon	0.934
Neon	1.82 (-3)
Helium	5.24 (-4)
Krypton	1.14 (-4)
Xenon	8.7 (-6)

Note: 1.0 (2) means 1.0×10^2

Source: Ratcliffe (1960)

Table 3.1 Composition of major and inert constituents present in the dry atmosphere

The atmosphere, due to gravity, has higher density at ground level than at higher altitudes. At sea level, the density is 1.225 kg/m^3 corresponding to a standard temperature of 15 °C or 288 K. The density has an exponential variation with height. A variation of a quantity, where the rate of variation at a given point is proportional to the value of the quantity at that point, is called an exponential variation. Its characteristic feature is that, for a given interval, the quantity gets multiplied by the same amount.

The density of the atmosphere (ρ) at a height z can be expressed by the Eqn. 3, as

$$\rho \quad = \quad \rho_o \, e^{-z/H}; \qquad 3$$

where, ρ_o is the density at sea level,
H is a constant known as the scale height, and
e = 2.71828, a mathematical constant called the exponential quantity.

The derivation of the above equation, and also the calculation of H, will be shown in the next section. H has a value of 8.42 km at ground level. This means that for every 8.42 km altitude, the pressure drops by a factor 1/e or 0.37. Generally, near the equator, density decreases by a factor of nearly 1000 for every 50 km altitude. As such, 99.9% of the mass of the atmosphere lies below this height. The variation of the atmospheric density with height up to 100 km is shown in **Fig. 3.1** for an atmosphere with a constant temperature.

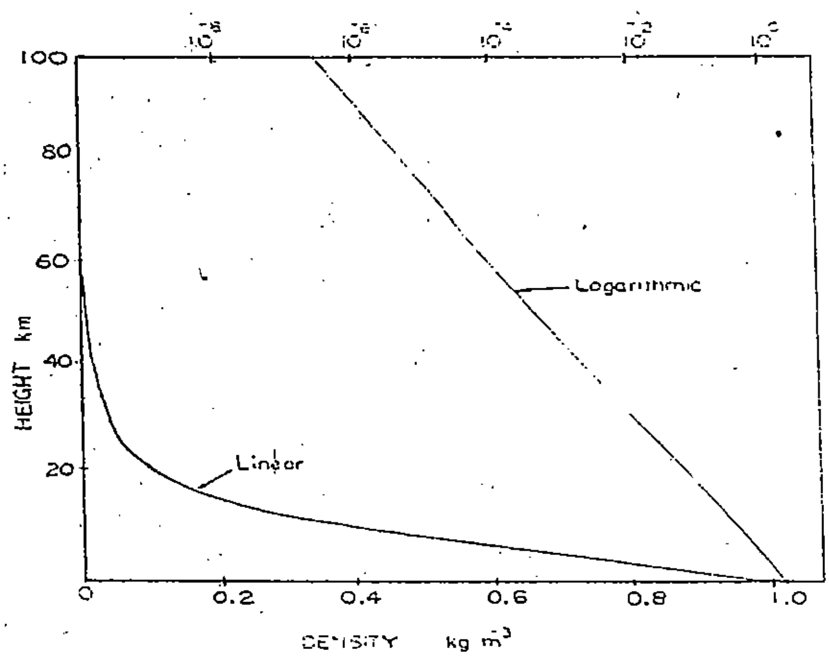


Fig 3.1. Variation of density with height in the atmosphere

The number density, or concentration, of dry air, N_d, is given by,

$$N_d = (n_d / V)\, N_o, \qquad 4$$

where, n is the number of gram-molecules present in a volume V (m^3), and
N_o is the number of molecules per gram-molecule, known as the Avagadro's Number, which has a value 6.025×10^{23}

Since n_d can be expressed as m_d/M_d, where m_d is the weight of dry air parcel of volume V, and M_d is the molecular weight of dry air (28.97), and the density of dry air, $\rho = m_d/V$, Eqn. 4 becomes,

$$N_d = \rho \times N_o/M_d \quad \text{(molecules/m}^3\text{)} \qquad 5$$

Substituting the values for the quantities in Eqn. 5, the number density of dry air at sea level is obtained as $N_d = 2.548 \times 10^{25}$ molecules per m^3. The number densities of individual gases could then be obtained by multiplying the above value by the respective mixing ratios.

Short-lived Constituents

The chemical reactions between the atomic species O, N and H, and the other constituents produce a large number of compounds such as NO, OH, CO, O_3, NO_2, NO_3, H_2, HO_2, H_2O_2, and many others. Their concentrations are not more than a few parts per million; yet they play an important role in controlling the energy balance and climate as well as creating a safe environment for people to live. One such important minor constituent is ozone (O_3) as it absorbs harmful UV-B radiation preventing it from reaching ground level. We will read more on this later. Some others, such as OH, act as scavengers in the atmosphere, removing polluting gases. Most of these constituents are chemically active, and hence have short lifetime. As such, their concentrations are highly variable, determined by local chemistry and dynamics, rather than by a fixed mixing ratio. Their concentrations in the upper atmosphere are highly variable with the hour of the day, seasons and the latitudes. Some of their typical concentrations at ground level and at a height of 25 km are given in **Table 3.2.** Some of these have been determined experimentally, while others were based on model calculations.

Minor constituent	Concentration expressed in ppm by volume	
	At ground level	At 25 km
O_3	4 (-2)	8-10
H_2O	1-4 (4)	3-5
H_2	6 (-1)	6 (-1)
OH	3 (-8)	1-2 (-6)
HO_2		8-10 (-5)
H_2O_2	3 (-3)	3-5 (-5)
NO	5 (-6)	1-2 (-3)
NO_2	4 (-5)	2-3 (-3)
N_2O	3 (-1)	2-3 (-1)
HNO_3		5-6 (-3)
CO	1-2 (-1)	
CH_4	2	0.9-1
SO_2	5 (-5)	

Note: 1.0 (-1) means 1.0×10^{-1}

Source: Jacobson (1999)
WMO (1985)

Table 3.2 Concentration of some minor constituents in the atmosphere

3.1.2 The Moist Atmosphere

The atmosphere contains moisture to varying degrees depending on the location and time of the day and season. The amount of moisture present in air determines whether a location habitable or not. The amount of moisture in air at ground level in the tropics where the temperature is high could be as much as 4-5%. The presence of moisture reduces the composition of other gases. For example, the composition of N_2 and O_2 drops to 74.13% and 19.94%, respectively, when the moisture content increases to 5%. Moisture also acts as a carrier of heat energy from the oceans to higher altitudes.

Relative Humidity

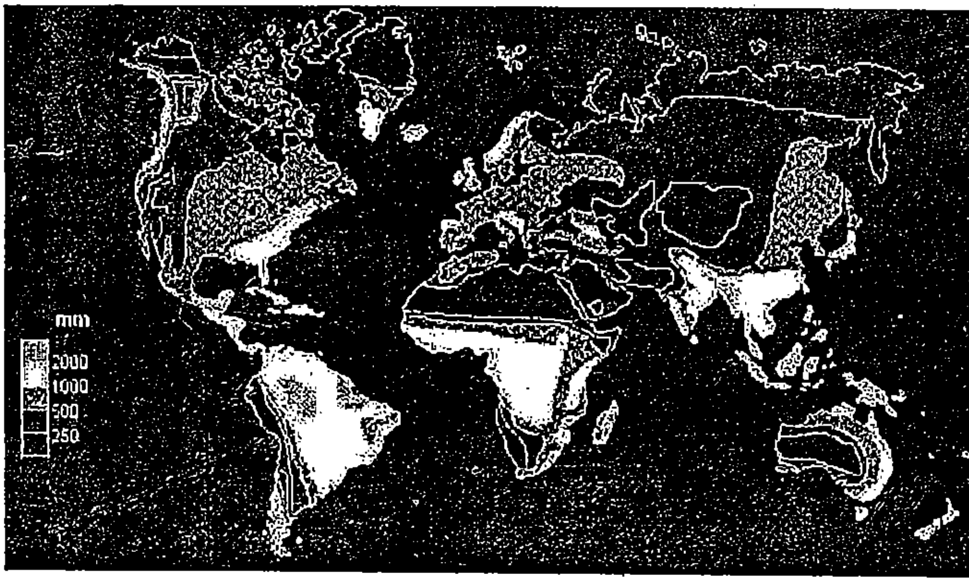
At a given temperature, there is a maximum amount of moisture that the air can hold. Such air is said to be saturated with moisture. Once air is saturated with moisture, if the temperature drops, the excess moisture condenses on any surface. Air that is unsaturated at a higher temperature will become saturated at a lower temperature. Water vapour in air acts as any other gas, exerting its own partial pressure, known as the vapour pressure.

The moisture content in air is generally expressed as a ratio of the actual moisture content and the moisture content required to saturate the air at the same temperature, and this ratio is called the **relative humidity**. Since the vapour pressure is proportional to the moisture content, relative humidity is also expressed as a ratio of the vapour pressure of air to the saturated vapour pressure. Within a geographical region, the relative humidity can vary from place to place and between daytime and night time. Relative humidity suitable for comfortable living is in the range 60% - 70%. Air conditioning of a room reduces both the temperature and the relative humidity to a comfortable level.

Precipitation

The atmosphere receives water vapour through the process of evaporation from oceans and other water bodies. This water vapour diffuses upwards, becomes cooler and ultimately reaches the saturation point. Cooling beyond this stage leads to condensation around dust particles known as aerosols acting as nuclei. The latent heat of evaporation is then released to the atmosphere, which has a value of 2,510 J/g at 0°C and 2,259 at 100°C. This is one of the main sources of energy that drives the dynamics of the lower atmosphere. This heat is balanced by the outgoing terrestrial radiation from the atmosphere. A mass of condensed water particles is generally described as a cloud. They appear from about a few kilometres upwards within the troposphere.

Precipitation appears either as rainfall or snow, depending on the temperature of the location. Countries may receive an annual rainfall as high as 10,000 mm or as little as 100 mm. The highest rainfall totals occur near the equator in the tropics, where the strong heating by the Sun creates significant vertical uplift of air, and the formation of prolonged heavy showers and frequent thunderstorms. The global distribution of the annual mean rainfall is shown in **Fig. 3.2.** Any variation of the rainfall, particularly long spells of droughts induced by various factors such as global warming or El-nino phenomenon have a direct bearing on the lives of people. The world is classified into various climatic zones depending on the annual rainfall and its distribution, humidity and temperature. More on this topic will be discussed under Climate Zones later in the unit.



Source: www.doc.mmu.ac.uk/aric/eae/climate/

Fig. 3.2 Global distribution of annual mean rainfall

3.1.3 Aerosols

Aerosols are particulates of terrestrial origin that remain suspended in air. Some are emitted direct into the atmosphere, while others are formed within the atmosphere from gaseous precursors. The former includes soil dust particles, dust from industries, fly-ash from power plants, carbonaceous particles from fossil fuel and biomass combustion, sea spray and volcanic eruptions. Among the latter are sulphates, nitrates, and carbonates, derived from respective precursor gases such as SO_2, ammonia, oxides of nitrogen and organic compounds.

Particles with diameters in the range 0.1 to 1.0 μm are referred to as aerosols in the accumulation mode. They could hydrate to diameters up to 2 μm with smaller particles coagulating to form large particles. Accumulating mode aerosols form the majority of cloud condensation nuclei. Typical diameters of aerosols of dust origin are in the range 2-4 μm, while those of salt sprays cover a wide range extending from 0.05 to 10 μm.

Aerosols alter the cloud formation process by increasing droplet and ice particle concentrations. They also decrease the precipitation efficiency of warm clouds. Aerosols have rather short lifetime in the atmosphere, and their effectiveness in the above processes depend on their size and composition. Since aerosols affect radiative processes, cloud formation and chemical processes in the atmosphere, projecting their influence on the climate is a very complex task, and a proper understanding of the processes involved demand complex computer simulation.

3.2 Atmospheric Pressure

The gravitational pull of the Earth keeps the atmosphere bound to it without allowing it to escape into outer space. The weight of the atmosphere is therefore felt at the surface as a force. The force per unit area exerted against a surface by the weight of the air above that surface, is called the atmospheric pressure. It is equivalent to the weight of a vertical column of air extending above a surface of unit area to the outer limit of the atmosphere. Humans and animals do not feel this pressure as it is counter-balanced by a pressure developed internally. In this section, we will discuss the variation of the atmospheric pressure with height, and its distribution on the surface

3.2.1 Variation with Height

The pressure at a point depends on the weight of air above it. Hence, the variation of the pressure is directly proportional to the variation of the atmospheric density. The variation of pressure with height is described by the **hydrostatic equation.** It is derived by considering the pressure difference between two layers in the atmosphere separated by an increment of the height denoted by *dz*. The pressure difference is equal to the weight of air lying between the two layers. This can be expressed, according to the definition of pressure, as:

$$dp = -\rho g.dz, \qquad 6$$

where, *d* indicates the differential (or the increment),

p is the pressure,

ρ is the density of air,

g is the acceleration due to gravity, and

z is the vertical height

This differential equation is known as the hydrostatic equation. The density of air can be expressed in terms of pressure using the gas equation,

$$pV = nRT, \qquad 7$$

where, n is the number of gram-molecules,

V is the volume of a parcel of air in m^3

T is the absolute temperature of air in Kelvin (K), and

R is the gas constant (8.3145×10^{-2} m^3 mb /mole.K)

Since n can be expressed as m/M_d, where m is the weight of the parcel of air in kg, and M_d is the molecular weight of dry air (28.97), and $\rho = m/V$, Eqn. 7 becomes,

$$p = \rho RT/M \qquad 8$$

The substitution of Eqn 8 in Eqn 6 yields,

$$dp/p = -(Mg/RT)dz. \qquad 9$$

When integrated between sea level and a height z, the pressure can be expressed by the exponential equation,

$$p = P_0 e^{-z/H}, \qquad 10$$

where, P_0 is the sea level pressure, and

H is a constant referred to as the "**scale height**" given by H = RT/Mg.

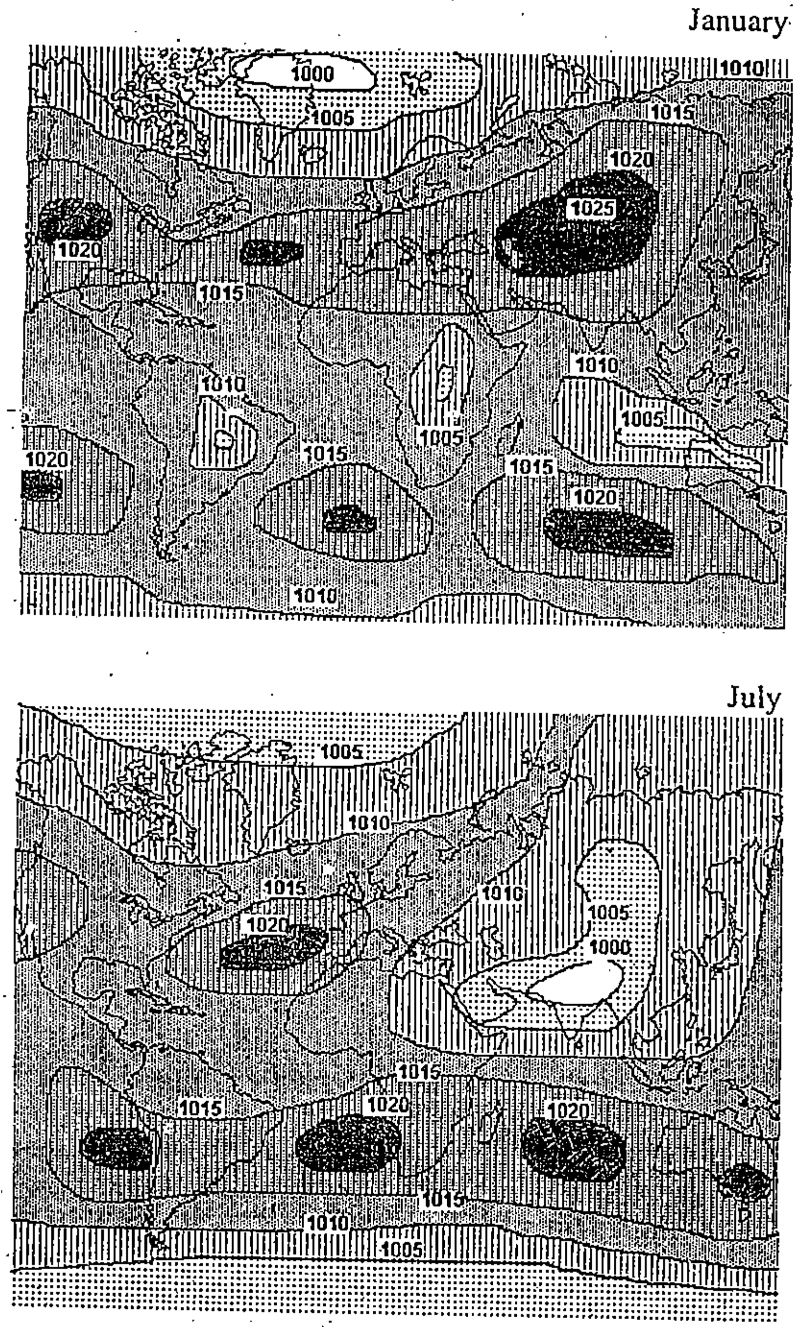
H is however, not a true constant as the temperature in the atmosphere has a certain variation. Assuming a constant temperature or an isothermal atmosphere, the atmospheric pressure can be described by the above exponential equation. For a temperature of 288 K, which is the average temperature at sea level, the scale height has a value of 8.42 km. The variation of pressure with height will have the same exponential variation as that given for density variation shown in **Fig. 3.1.**

3.2.2 Global Distribution of Pressure

The atmospheric pressure is measured using a barometer. It is basically a manometer filled with mercury. The height of the column of mercury with a vacuum above it, which balances the weight of air, indicates the pressure At sea level, the global mean pressure is 760 mm of mercury, or one atmosphere. It is also measured in terms of another unit called "**bar**", defined as 10^5 Pascal (Pa) or Newton per sq. metre (N/m^2). The unit often used in weather forecasting is the millibar, which is equivalent to 100 Pa or 1 hPa. The surface air pressure varies both spatially and temporally. The average sea level atmospheric pressure is 1.01325 bar, but it may vary between 1000 and 1040 mbar, depending on the location and time. The standard sea-level pressure is taken as 1013 mbar. This is also equivalent approximately to a force of 1 kilogram per square centimetre.

The continents and oceans significantly influence the major pressure belts that develop from the circulation patterns of the atmosphere. Land gains and loses heat much more quickly than seawater. Consequently, the large landmasses of the Americas and Asia become much warmer in summer and much colder in winter. The extra surface heat in summer generates a continental region of low pressure, whilst in winter, the colder descending air gives rise to dominant high pressure anticyclones. In summer the development of continental low pressure significantly influences the pattern of monsoons that affect the weather of India and southern Asia.

Generally, high pressure predominates at about 30° N and S, and low pressure predominates at high latitudes and in the tropics. Also, the winter hemisphere pressure is higher over land than over oceans and the summer hemisphere pressure is higher over the oceans than over land. The low pressure area found near the equator is known as the **'equatorial trough'**, and sometimes called the **'doldrums'**. The global distribution of surface pressure in January and July are shown in **Fig. 3.3.** These pressure differences give rise to winds, both vertically and horizontally.



Source: www.doc.mmu.ac.uk/aric/eae/climate/

Fig. 3.3 Global distribution of mean surface pressure in January and July

3.3 Solar Radiation

Under this section, we will learn about the nature of solar radiation, its interaction with the atmosphere including the energy balance and the radiative processes of radiation in the atmosphere such as absorption and scattering.

3.3.1 Solar Spectrum

The sun emits different types of radiations including gamma rays, X-rays, ultra-violet rays, visible rays or what we call light, infrared radiation and radio waves. All these are electromagnetic waves having the common property of travelling in free space with the speed of light. These radiations can be described in two ways: one as waves and the other as particles or 'packets' of energy. In the wave mode, we can visualise radiation travelling in the form of waves, each wave having a distinct wavelength, which is the distance between two adjacent crests or troughs. Each type of radiation has a distinct wavelength range as given in **Table 3.3**

Band	Wavelength Range	Energy per photon (eln.volt)
Gamma rays	< 0.01 nm	$>1.24\times10^5$
X-rays	0.1 - 10 nm	1.24×10^5 - 124
Extreme UV	10 - 100 nm	124 - 12.4
UV	100 - 400 nm	12.4 - 3.10
Visible	400 - 760 nm	3.10 - 1.60
Near IR	0.76 - 10 μm	1.60 - 0.12
Far IR	0.01 - 1.0 mm	0.12 - 1.24×10^{-3}
Microwaves	1.0 - 300 mm	1.24×10^{-3} - 4.1×10^{-6}

Table 3.3 Radiation bands in the solar spectrum

In the particle mode, radiation travels in the form of energy packets known as **'photons'** or **'quanta'**. The energy contained in one photon is inversely proportional to the associated wavelength, which means that shorter wavelengths will have more energy than waves with longer wavelengths. Thus gamma rays, X-rays and extreme UV radiation are highly energetic and are lethal to humans. Fortunately, these radiations get totally absorbed high in the atmosphere by oxygen and nitrogen gases and therefore do not reach ground level. The energy levels of different radiation bands are also shown in Table 3.3.

The energy content in a monochromatic or single-colour beam of unit cross-section is given by the product of the energy per photon and the number of photons present within a unit wavelength interval. In X-rays and UV, energy per photon is high, but in the visible range, the total number of photons is high. Though visible radiation contains less specific energy, because of the large number of photons in this band, it has the highest total energy content.

The solar irradiance closely matches that of a black body at a temperature of 5,900 K. A black body emits radiation in a continuous spectrum with a peak at a wavelength, which is inversely proportional to the temperature. Therefore, a hot body emits more short wave radiation while a cold body emits more long wave radiation. The distribution of energy contained in the radiation per unit area per unit wavelength interval at the top of the atmosphere is illustrated in **Fig. 3.4** for the entire spectrum. Maximum energy is radiated in the visible band around 500 nm, and almost all the energy emitted is contained below 4 µm. This range of wavelengths is referred to as solar short wave radiation.

The radiations of wavelengths above about 320 nm reach ground level with little attenuation. These include near UV, visible, IR and microwave radiations. However, various gases present in the atmosphere such as oxygen, ozone, carbon dioxide and water vapour, absorb partially different wavelength bands as shown in Fig. 3.4. As a result, the spectrum received at sea level is very different to that at the top of the atmosphere. Further, other radiative processes such as reflection and scattering also cause attenuation of the radiation before reaching the ground

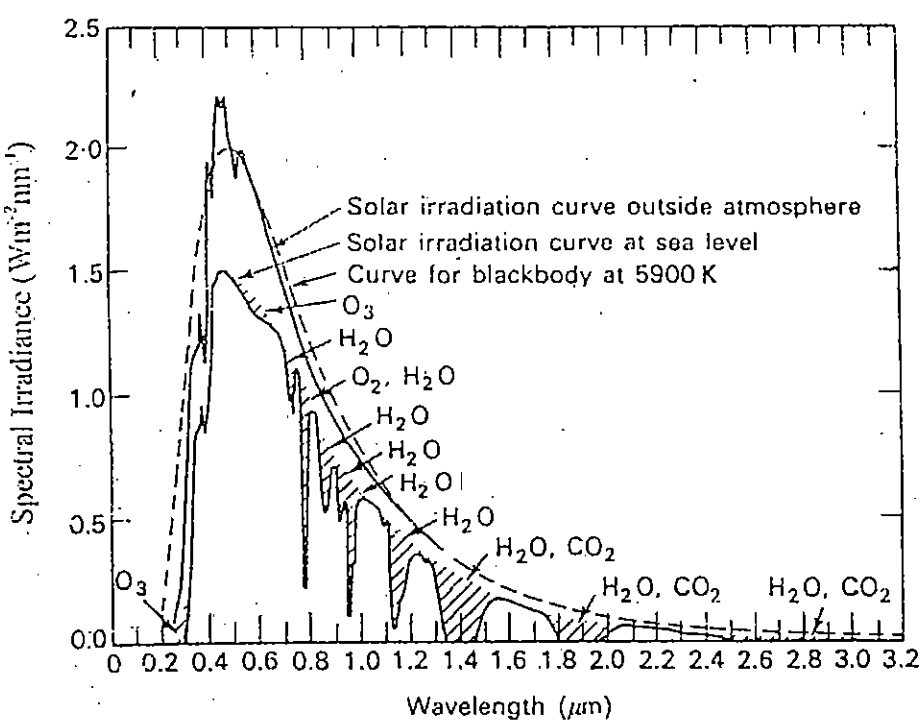


Fig. 3.4 **Distribution of energy in the solar spectrum**

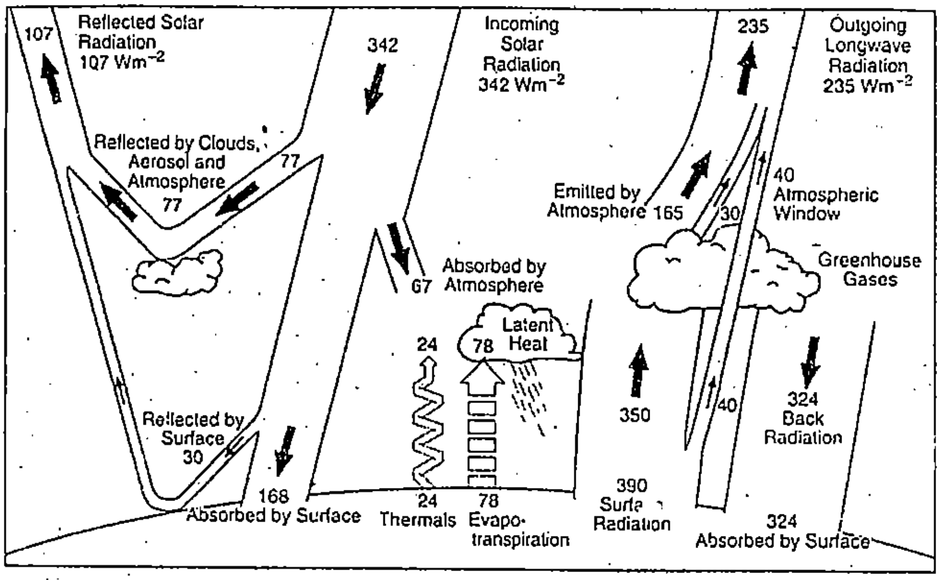
3.3.2 Energy Balance

Average Energy Exchanged

The total energy content in solar radiation per unit area is known as the '**Solar Constant**' and has a value of 1370 Watt/m^2 at the top of the atmosphere. Except for visible radiation and near infra-red, other radiations get absorbed in the atmosphere and only a fraction is filtered to the ground. As a result, the total radiation energy received at ground level with over-head sun in the tropics is about 1000 Watt/m^2. However, the incident value averaged over day and night is much less. The total radiation intercepted by the Earth is equal to the Solar Constant (S) multiplied by the cross sectional area. This amount of energy is radiated out from the total surface so as to balance the average energy received. Since the Earth's surface has an area equal to 4 times the cross sectional area, the average energy received or radiated out by the Earth is given by S/4 or 342 W/m^2.

When averaged over a day, about 20% (67 W/m^2) of the incoming radiation gets absorbed by various atmospheric gases. Another 22% (77 W/m^2) gets reflected and scattered from the clouds and aerosol particles. Out of the balance 58% (198 W/m^2) reaching the surface, about 9% (30 W/m^2) get reflected from the surface including land and oceans. Only the remaining 49% (168 W/m^2) gets absorbed by the Earth and contribute to its heating. Thus, out of the 342 W/m^2 incident on the atmosphere, the net radiation received is only 235 W/m^2. This amount has to be balanced by an equal amount of out going long wave radiation emitted by the earth. The various processes that contribute to the energy balance is illustrated in **Fig. 3.5.** A significant component is the re-emission of 342 W/m^2 of heat energy towards the Earth by the greenhouse gases present in the atmosphere.

Aerosols play an important role in the energy balance in the atmosphere, through absorption and scattering of radiation. The carbonaceous aerosol belongs to two types; those with organic carbon and those with black carbon. These two types have different optical and radiation absorbing properties, with black carbon particles having higher absorbing properties than the other. Combustion of fossil fuels and biomass are the sources of both organic and black carbon aerosols, with the former emissions being 5-6 times that of the latter.



Source:IPCC (2001)

Fig. 3.5 Radiative processes that contribute to the energy balance

3.4 Radiative processes in the Atmosphere

When solar radiation is incident on the atmosphere, part will get transmitted and the balance reflected. The component traversing through the atmosphere will get partly absorbed and partly scattered. Only the balance reaches the ground. In this section, we will discuss these phenomena.

3.4.1 Absorption in the Atmosphere

The absorption of energetic radiation by atmospheric gases gives rise to several phenomena: heating of the atmosphere, breaking down of the gases into new substances, electrifying certain parts of the atmosphere etc. Each of these processes depend on the energy content of radiation and the absorption properties of the constituent gases in the atmosphere.

Effect of X-Rays and UV Radiation

X-rays and EUV radiations are the most energetic radiations. Molecular nitrogen and oxygen totally absorb them above about 60 km. Because of their high energy content, they knock off an electron from the molecules, leaving the remaining portion positively charged. These are called 'ions'. Hence the region of the atmosphere above about 60 km up to about 500 km where such positively charged ions are present is known as the '**ionosphere**'. This region plays an important role in long-distance short-wave radio communication. The ionosphere has the ability to reflect to Earth any short-wave radio signal incident on it. This makes it possible for a radio signal to travel round the Earth through multiple reflections between the Earth and the ionosphere.

The UV spectrum is divided into three bands, UV-A, UV-B and UV-C based on their degree of penetration into the atmosphere. The UV-A band (315-400 nm) is the least energetic band. This radiation is relatively harmless and reaches the ground level. The UV-B band (280-315 nm) is more energetic and is damaging to both plant and animal life. It gets absorbed by ozone and partly reaches the ground. The most energetic band, UV-C (100-280 nm), which is lethal, gets totally absorbed by nitrogen and oxygen before reaching the ground level.

Major Absorbing Gases

Oxygen gas is a good absorber of UV radiation up to about 180 nm and absorbs all of it by the time the radiation reaches a depth of about 35 km. Between 180 and 200 nm, the absorption efficiency of oxygen drops by a factor of more than 5,000 and oxygen becomes ineffective in absorbing any radiation beyond 200 nm. The absorption of UV radiation by oxygen causes its photo-dissociation yielding highly reactive atomic species, some of which are in the excited states. These excited atoms subsequently return to the ground state through the emission of radiation, which can be detected from ground level. These emissions are called airglow and provide much useful information about the upper atmosphere.

In the range from 200 to 320 nm, the absorption efficiency of ozone is about a million times more than that of oxygen. Therefore, despite being a trace gas, ozone becomes the major absorber of radiation in this wavelength range. Wavelengths up to about 280 nm are totally absorbed by ozone before reaching a depth of about 15-20 km. This causes heating in the stratosphere forming the bulge in the temperature profile (see Fig. 3.10). The UV-B band between 280 – 315 nm, is absorbed partially by the ozone layer depending on the total ozone density. Beyond 315 nm, ozone also becomes a poor absorber. Visible radiation above 400 nm could therefore reach ground level without much attenuation.

Absorption Cross Section

When electromagnetic radiation is incident on the surface of any substance, the amount absorbed depends on the wavelength (λ), molecular structure of the substance and the level of incident radiation. The incremental absorption when traversing a path length of dl can be expressed by the equation,

$$dI(\lambda) = -I(\lambda).\sigma_i(\lambda).n_i.dl, \qquad 11$$

where, d is the differential,

$I(\lambda)$ is the intensity of radiation of wavelength λ,

$\sigma_i(\lambda)$ is the absorption cross-section of the i^{th} substance at wavelength λ,

n_i is the number density of the i^{th} substance, and l is the path length

This yields the following expression for the variation of intensity of a monochromatic radiation while traversing through an absorbing medium:

$$I(\lambda) = I_o.\exp(-\sigma_i(\lambda).\int n_i.dl) \qquad 12$$

Thus, a monochromatic radiation incident on the atmosphere subject to absorption by a single gas, will attenuate during its passage through the atmosphere, according to the above formula. When there are several absorbing gases present, the contribution of each of the gases will have to be computed and summed. The above expression will then reduce to:

$$I(\lambda) = I_o.\exp\sum_i(-\sigma_i(\lambda)\int n_i.dl). \qquad 13$$

To determine the total irradiance at the surface, the above intensity has to be summed up over the entire spectral band that is incident on the surface. The plot of $\sigma(\lambda)$ vs λ for a given substance is referred to as its absorption spectrum, and is a signature of the substance

Depth of Penetration

According to Eqn. 12, an absorbing gas will reduce the intensity of radiation to a factor of 1/e or 37% of its initial value at a height when $\sigma_i(\lambda).\int n_i.dl = 1$. Since the attenuation is exponential, the radiation gets totally attenuated within a few kilometres below, and this height gives the depth of penetration of that particular wavelength. **Fig. 3.6** shows the depths to which different wavelengths of solar radiation could penetrate into the atmosphere. The cut-off wavelength of solar radiation reaching the ground lies in the UV-B band, the exact wavelength being controlled by the column density of ozone and the solar angle. While wavelengths above 315 nm reach the ground at more or less full intensity, wavelengths below it arrive at reduced intensity getting completely cut off before reaching 280 nm.

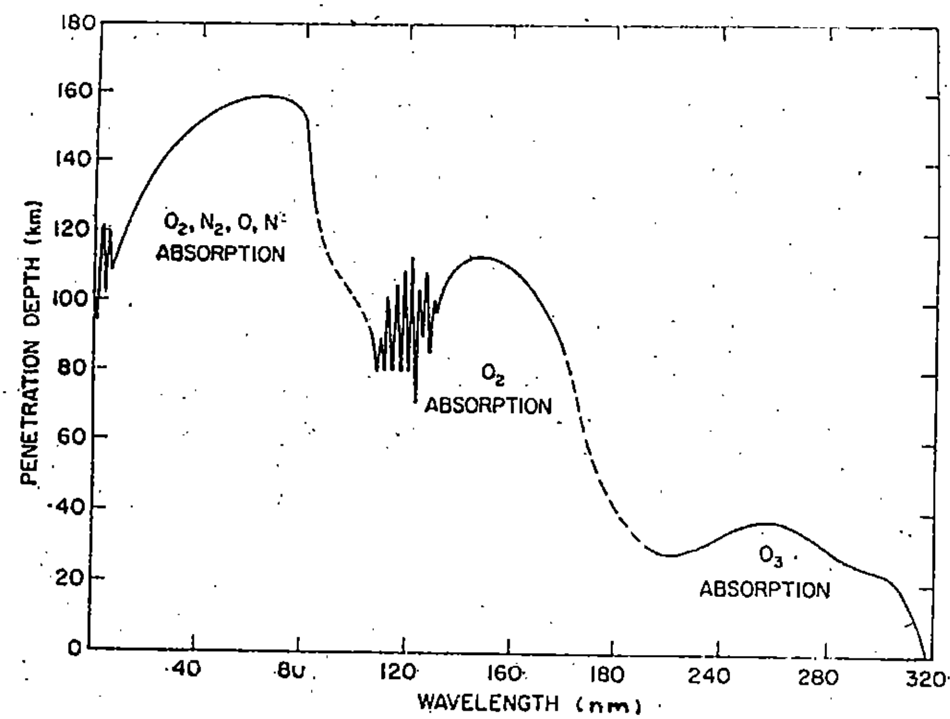


Fig. 3.6 Depth of penetration of solar radiation into the atmosphere

3.4.2 Natural Global Warming

The gases that were previously referred to as greenhouse gases, which are naturally occurring in the atmosphere, have the ability to absorb radiation selectively, and cause warming of the Earth. This enables the Earth to maintain its global mean surface temperature at the present value of 15 °C. This is referred to as the natural global warming. The atmospheric gases, which take part in the global warming process are called **radiative gases**, and their contribution to global warming is called **radiative forcing.**

Greenhouse Gases

The gases naturally present in the atmosphere that have this radiative forcing property include CO_2, CH_4, N_2O, H_2O and O_3. They have low absorption cross sections at shorter wavelengths but high cross sections at longer wavelengths. These radiative gases have the capability to permit most of the solar radiation which are of shorter wavelengths to penetrate the atmosphere and heat the ground, while absorbing the terrestrial radiation of longer wavelengths, thus disallowing such radiation to escape and cool the atmosphere. In addition to these naturally occurring gases, there are many gases introduced through human activities, causing enhanced global warming. This topic will be discussed later.

The above radiative process taking place in the atmosphere is similar to what takes place inside a greenhouse used in temperate countries to grow crops in the cold seasons. In a greenhouse, the transparent roofing, either glass or plastic, permits short wave solar radiation to penetrate it but not the outgoing terrestrial long wave radiation, which gets absorbed by the roofing sheet. This results in the chamber maintaining a temperature several degrees above the cold ambient temperature, helping plant growth. Hence, the radiative gases playing the role of the chamber roof are also called greenhouse gases.

Greenhouse Effect

The enhancement of the temperature of a greenhouse chamber due to the absorption of out-going terrestrial radiation by the chamber roof, is the origin of the greenhouse effect. The Earth's surface temperature is much cooler than that of the sun. Hence, as mentioned before, the energy radiated out from the Earth is in the form of long wave IR radiation with wavelengths 5-100 μm. This radiation gets absorbed by the greenhouse gases present in the troposphere and referred to earlier. Part of the absorbed radiation is emitted back to the Earth, and part to outer space, as shown in Fig. 3.4. This process reduces cooling of the Earth, and gives rise to additional warming, causing surface temperature to increase. This results in higher outgoing radiation, which then would match the incoming radiation.

This phenomenon of increasing the surface temperature caused by the presence of certain gases naturally in the atmosphere with selective absorption properties is referred to as the greenhouse effect as applicable to the atmosphere. If there had been no greenhouse effect occurring in nature, it is estimated that the global average temperature would have been -18 °C or 33 °C below the present mean temperature. Such a low global mean temperature would not have sustained life on Earth.

3.4.3 Scattering

Much of the radiation received on the Earth's surface is scattered radiation rather than direct. Though radiation comes from the sun, the entire sky appears visible because of the scattered light. Particles of all types including gas molecules, dust, aerosols, water droplets and ice particles of size from sub-micron to several hundred microns cause scattering.

The direction of scattering depends on the ratio of particle size to wavelength of the incident radiation. Very small particles such as gas molecules scatter radiation in all directions while large particles tend to scatter radiation more in the forward direction. These two cases are described as **Rayleigh scattering** and **Mie scattering**, respectively. In Rayleigh scattering, the intensity of the scattered radiation in a particular direction is inversely proportional to the 4^{th} power of the incident radiation.

If we consider the visible radiation, the wa elengths that get scattered away from the sun are the shorter wavelengths or blue light. This makes the sky during day time appear blue, while during dawn and dusk, when the direction of scattering is more in the direction of the sun, the sky appears red. In the case of Mie scattering where the particle size is comparable to or larger than the wavelength, the scattering is not wavelength dependent. Such scattering takes place in clouds and fog, resulting in their colour appearing as white or grey.

Aerosols also scatter incoming radiation. Particles with diameters below 1 μ m scatter more light per unit mass and have a longer lifetime than the larger particles. Aerosols smaller than 1 μm containing sulphates, nitrates or organic carbon scatter radiation efficiently, while those containing black carbon absorb radiation efficiently. Scattering aerosols lower the surface temperature during daytime by decreasing the solar radiation incident on the ground. The scattering of the incoming solar radiation by aerosol particles, particularly the sulphates, plays an important role in climate change by reducing the intensity of radiation reaching the Earth's surface resulting in off-setting the global warming process partly.

3.5 Seasonal and Daily Variations

Climate and weather, which depends on the heating of the atmosphere, in turn, depend on the amount of solar radiation received at any given instant, and this varies with the season and hour of the day. In this section, we will discuss how seasonal and daily variations in solar radiation influence the atmosphere.

3.5.1 Seasonal Variations

Solar Declination

Seasons manifest in the Earth's environment because of its elliptic orbit around the sun with its axis inclined to the plane of its orbit or the ecliptic plane. The Earth's equator is tilted 23.45 degrees with respect to the ecliptic. The angle between the sun's rays and the Earth's equatorial plane is known as the **solar declination.** It also corresponds to the latitude at which the sun is directly overhead at midday. Declination values are positive when the sun is north of the equator (March 21 to September 23) and negative when the sun is south of the equator. Its maximum and minimum values are +23.45 degrees and -23.45 degrees. The variation of declination with the day of the year is shown graphically in **Fig. 3.7.**

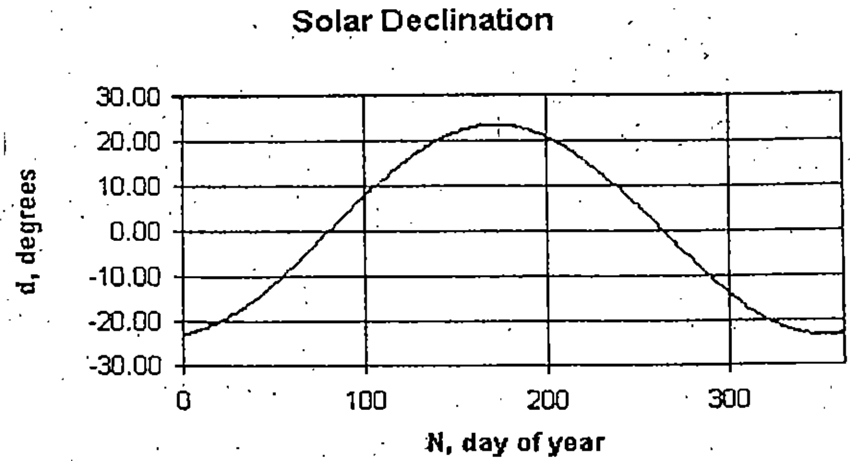
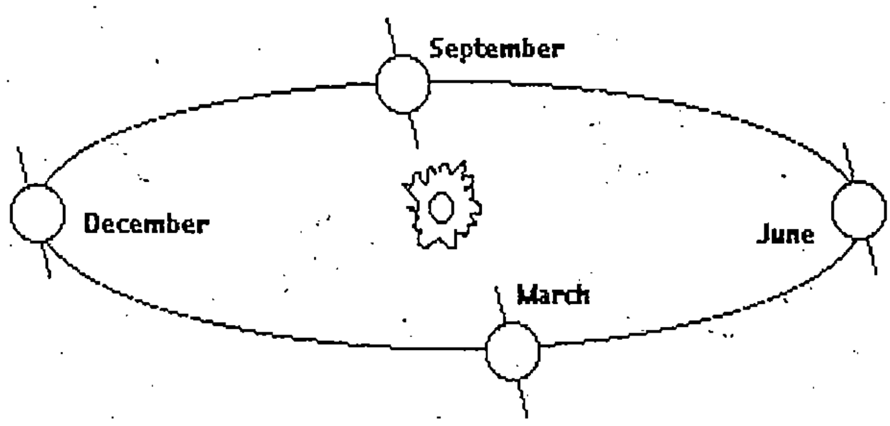


Fig. 3.7 **Variation of solar declination during the year**

The Seasons

The exposure of one hemisphere to the sun more than the other for several months of the year gives rise to the seasons. The hemisphere tilting towards the sun receives more radiation with longer daylight hours and the corresponding season is called the summer. When it is summer for one hemisphere, it is winter for the other with less radiation received and short daylight hours. On June 21st, the northern hemisphere is tilted the maximum of 23.45 degrees towards the sun. It is also called the **summer solstice** for the northern hemisphere. On the same day, the southern hemisphere is tilted 23.45 degrees away from the sun, when it is **winter solstice** for the southern hemisphere. Similarly, on December 21st, it is the winter solstice for the northern hemisphere and summer solstice for the southern hemisphere. The portion of the earth illuminated in different seasons is illustrated in **Fig. 3.8.**



Source: www.doc.mmu.ac.uk/aric/eae/climate/
Fig. 3.8 Position of the Earth in its orbital path

At the summer pole, the sun never sets, shining all 24 hours, while at the winter pole, the sun never rises being in dark all 24 hours. The hours when solstices occur year after year are not exactly the same, but have a small cyclic variation of less than one day with a periodicity of several years.

On March 21st and September 21st, the suns rays are in line with the equatorial plane and the solar declination is zero. These are called the autumn (or fall) and spring equinoxes and the sun passes directly over the equator. The latitudes corresponding to the maximum declination of the sun in the northern and southern hemispheres are called the **Tropics of Cancer and Capricorn,** respectively.

The seasonal variation with longer daylight hours during the summer has significant implications on the growth of crops, particularly the annual crops. A given crop has a better yield in temperate countries as opposed to tropical countries, because of the longer exposure to solar radiation and corresponding photosynthesis. Seasons also influence the flowering of crops.

3.5.2 Daylight Period

As the sun rises and reaches noon in local time, the angle of the sun to the zenith varies from 90° to a minimum. This minimum angle at noon depends on the latitude of the location and the day of the year. In the northern hemisphere, within the latitude equivalent to the declination of the sun and the equator, the sun reaches overhead on two days of the year, at noon. Away from this latitude, sun does not reach overhead even at summer solstice noon, but reaches only a minimum value, depending on the latitude. The duration of daylight hours therefore depends on the solar zenith angle. It can be expressed by:

$$\cos \chi = \sin \phi * \sin \delta + \cos \phi * \cos \delta * \cos \tau \qquad 14$$

where, χ is the zenith angle,
ϕ is the latitude,
δ is the solar declination, and
τ is the hour angle (15° per hour from noon).

The daylight hours or the photoperiod at a given location and a given day of the year can be obtained from the above equation, by setting $\chi = 90°$, which yields,

$$\cos \tau = -\tan \phi * \tan \delta \qquad 15$$

Since τ gives the deviation from the noon, the total day light hours will be twice that given by τ. However, the actual duration is slightly longer than what is given by this expression because of the refraction of rays coming from the sun after sunset or before sunrise. The daylight hours corresponding to summer and winter solstices for different latitudes, obtained from the above expression, are given in Table 3.4.

Latitude (Deg)	Winter Solstice		Summer Solstice	
	Hrs	Min.	Hrs/mon	Min
90	0		~6 mon	
80	0		~4 mon.	
70	0		~ 2 mon.	
60	5	33	18	27
50	7	42	16	18
40	9	08	14	52
30	10	04	13	56
20	10	48	13	12
10	11	25	12	38
0	12	00	12	00

Source: Campbell (1979)

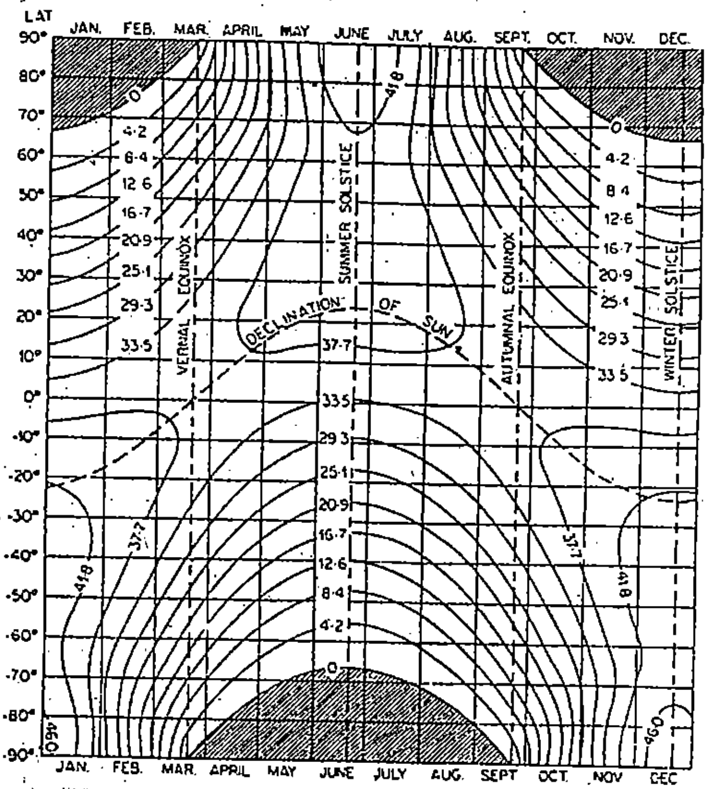
Table 3.4 Daylight hours at different latitudes at summer and winter solstices

3.5.3 Variation of Solar Irradiance

The total daily average solar irradiance incident on a horizontal surface at the top of the atmosphere, E_o, is obtained by integrating the above expression from sunrise to sunset, which yields,

$$E_o = S*\{H_d(\sin\phi * \sin\delta) + \cos\phi * \cos\delta * \cos H_d\} \qquad 16$$

Using the value for S as 1353 W/m^2, the maximum value obtained for E_o is corresponding to south pole during winter, which is 574 W/m^2 or 49.6 MJ/m^2. The daily integrated solar irradiance at the top of the atmosphere for different latitudes and times of the year are given in **Fig. 3.9**. According to this figure, the daily irradiance at the equator at summer solstice is only 33.5 MJ/m^2, while it is more than 42 MJ/m^2 at the north pole. This is because of the 24 hour daylight hours at the poles when it is only 12 hours at the equator. There is also a slight asymmetry between the summer values in the north hemisphere and those in the southern hemisphere, the latter being higher than the former. This is because the Earth, in its elliptic orbit round the sun, is closest to it on December 21st.



Source: Campbell (1979)

Fig. 3.9 Daily solar irradiance at different latitudes and times of the year

In order to determine the solar irradiance at the ground level, the absorption of radiation by the intervening atmosphere, as described in Section 3.4.4 has to be taken into consideration. The intensity after traversing through the atmosphere given by Eqn. 13 for one wavelength has to be integrated both over the entire spectral range that is incident on the ground and also for all the absorbing gases in the atmosphere. Since the spectral intensities at different wavelengths, as shown in Fig. 3.3, or the absorption cross sections of different gases cannot be expressed in an analytical form, this integration has to be performed numerically. However, this value is highly variable because of the unpredicted variation of cloud cover as well as the presence of other substances such as aerosols, which attenuate the incident radiation. Direct measurements of the irradiance are carried out using various types of instrument. Such measurements show that almost 50% of the radiation received at the top of the atmosphere is lost while traversing through the atmosphere, as shown in the energy balance in Fig. 3.4.

The amount of radiation that gets lost in the atmosphere increases with its path length through it. Hence, the path length controls the amount of radiation received at different latitudes of the Earth at a given instant. Near the Equator, the path length and hence the attenuation is a minimum, and the intensity becomes a maximum. At higher latitudes, the path length is long even in the summer and hence the radiation incident is low. However, the long duration of the day compensates for the loss of intensity at high latitudes.

3.6 Temperature Profile

The temperature of the atmosphere is an important factor that determines the climate of the Earth. Factors that control the temperature, its variation with height and stratification of the atmosphere into different regions based on temperature will be discussed here.

3.6.1 Ground level Temperature

The atmospheric temperature depends on several factors: the absorption of solar radiation by the Earth and the atmosphere, its loss into outer space by terrestrial radiation and the gas laws which determine its distribution with altitude. The temperature near the surface is determined by the long wave heat radiation emitted by the Earth after being heated by short wave solar radiation. As mentioned previously, the greenhouse gases in the atmosphere helps in maintaining the current global mean temperature at 15 °C. Scientists have found that this temperature has remained within half a degree over the last thousands of years.

In the tropics, the surface temperature is more or less constant throughout the year. The ground level temperature in the tropics has a small variation in the range 25-35 °C year round, except in the deserts, where the temperature may even reach values as high as fifties. The surface temperature shows a wide variation with the seasons at high latitudes, which could vary from about 25 – 30 °C in summer to about –30 or –40 °C in winter.

The heating of the Earth's surface takes place only when it receives solar radiation, but it cools all the time due to emission of terrestrial radiation. Hence, during winter months, when the polar regions are in complete darkness during the entire day, the surface temperature could drop to very low values. This is more prominent in the south pole because there is hardly any mixing of polar air with air in the warmer regions due to the air circulation in the southern hemisphere being more circumpolar than in the northern hemisphere. The difference in the surface topography in the two hemispheres plays a role here.

The surface temperature also shows a diurnal variation, with a maximum temperature attained a few hours after the noon, and the minimum just before sunrise. At the tropics, the diurnal variation could be in the range 5 – 10 °C, depending on the elevation, while in the temperate regions, it could be even more, depending on the season.

3.6.2 Stratification into Regions

Among the planets, the temperature profile of the Earth is unique, in having several stratified layers characterized by different temperature gradients. These are described below, and shown in **Fig. 3.10.**

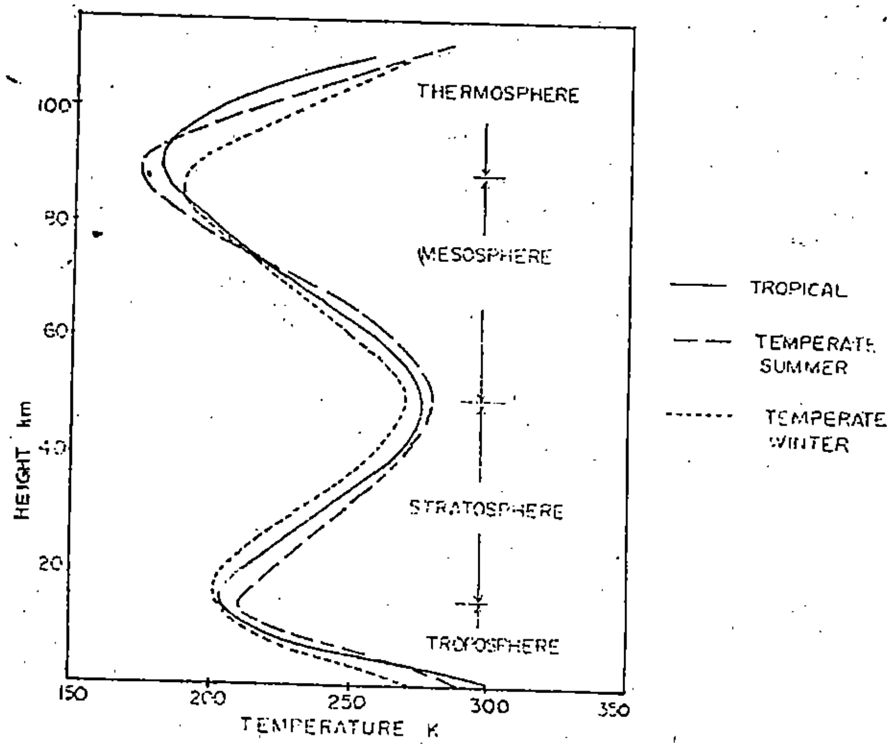


Fig. 3.10. Variation of temperature with height in the atmosphere

Troposphere

In the region closest to the surface, the temperature has a negative gradient up to a height range 8 - 15 km, with the lower values at high latitudes and higher values at low latitudes. This region is known as the **troposphere**. The temperature gradient is about 6-8 °C per km. The height range with the minimum temperature is known as the **tropopause**, where the temperature is about 200-220 K at low latitudes. The drop in the pressure away from the surface causes the temperature to drop at higher altitudes, as well as the drop in the intensity of terrestrial radiation away from the ground contributes to the decline in the temperature above the surface. Most of the phenomena responsible for the weather at ground level, such as cloud formation and winds, take place in this region.

In the troposphere, there are a large number of polluting gases present, emitted when fossil fuels and biomass are burnt for energy generation in industries, power plants and vehicles. The reactive organic compounds present in a polluted atmosphere enhance the conversion of NO to NO_2. Solar radiation at ground level photo-dissociates NO_2 producing O, which will combine with O_2 to form ozone. This results in formation of atmospheric smog.

Stratosphere

The region above the troposphere, known as the **stratosphere,** has a positive temperature gradient up to about 50-55 km. The temperature rise in this region is due to the heating through the absorption of 200 – 300 nm solar UV radiation by ozone The temperature profile however, reaches a peak around 55 km, because the loss of heat at greater heights is smaller than that at lower heights. This peak temperature is around 260-280 K, which is slightly less than that near the surface, and the corresponding height range is known as the **stratopause.** The stratosphere is characterized by extreme low moisture and the presence of the ozone layer. Ozone is concentrated into a layer with a peak density of about $5x10^{18}$ m^{-3} or a mixing ratio of about 6 ppm around 25 km in low latitudes.

Mesosphere

The temperature above the stratosphere has a negative gradient up to about 80-85 km, and this region is known as the **mesosphere**. The height range where the temperature reverses again is called the **mesopause**. The minimum temperature attained here is around 180-200 K. The absence of a suitable medium to absorb UV and EUV radiation filtering to these heights results in the declination of the temperature in the mesosphere. The mesosphere is characterized by the presence of negative ions along with free electrons and hydrated positive ions. The ionisation in this region is caused mainly by solar Lyman alpha line at 121.6 nm ionising NO, resulting in the formation of the D-region of the ionosphere.

Thermosphere

Above the mesopause is the **thermosphere**, where the temperature has a positive gradient with temperatures reaching values in excess of 1000 K at heights above 120 km. The absorption of solar extreme UV radiation and X-rays by oxygen and nitrogen at these heights causes the temperature to increase. These ionizing radiations are responsible for the formation of E and F regions of the ionosphere, through the ionisation of O_2 and N_2.

The intensities of EUV and X radiations have a strong dependency on solar activity, which has a 11-year cycle. Hence, the heat absorbed and the temperature in turn, also vary with the solar cycle. The maximum thermosphere temperature varies between 1000 K and 2000 K, corresponding to solar minimum and solar maximum, respectively.

3.7 Transport Mechanisms

It was mentioned at the beginning of this unit that the atmosphere is a dynamic entity. The uneven heating, fluctuations in the pressure and temperature cause movements of air in the atmosphere, both in the planetary scale and the micro-scale. Some of these transport mechanisms are given, along with their scale dimensions, in **Table 3.5**, and described below.

Description	Scale Dimension	Examples
Molecular scale	< 2 mm	Molecular diffusion, molecular viscosity
Micro-scale	2 mm – 2 km	Eddies, plumes, exhaust emissions, cumulus clouds
Meso-scale	2 – 2000 km	Gravity waves, thunderstorms, tornados, local winds, urban air pollution
Synoptic Scale	500 – 10,000 km	Weather fronts, tropical storms (cyclones), hurricanes
Planetary Scale	>10,000 km	Rossby (Planetary) waves, Global wind systems

Source: Jacobson (1999)

Table 3.5. Description of different atmospheric transport mechanisms

3.7.1 Diffusion

There are two types of diffusion processes, molecular and eddy diffusion. The rate of diffusion of one constituent within another is proportional to the divergence of its concentration. The constant of proportionality is called the diffusion coefficient.

Molecular Diffusion

This is the smallest scale movement, taking place at molecular level. This applies particularly to minor-species introduced from an external source or through internal reactions. The molecular diffusion coefficient could be obtained from the kinetic theory of gases, as:

$$D_m = 3\,\{kT/2m^*\}^{1/2}/\{8[X]d\}, \qquad 17$$

where, k is the Boltzman's Constant (1.3807×10^{-23} kg m^2 /(s^2 K molecule)

m* is the reduced mass of the molecules

d is the mean molecular diameter.

The values of D_m for most gases at standard temperature and pressure lie between 1.4 and 2.1 $\times 10^{-5}$ m^2/s. Molecular diffusion becomes important as a transport mechanism only above 100 km where the density is low.

Eddy Diffusion

We often see plumes of smoke discharged from stacks move as a whole and diffuse into the surrounding air after travelling some distance. Such diffusion is referred to as eddy diffusion. These eddies may have scale lengths ranging from several metres to several kilometres. The eddy diffusion coefficient, D_e could be determined by making observations carried out on specially released vapour trails at lower altitudes, and by observing the diffusion of nuclear debris at higher altitudes. The typical values determined for D_e are about 1 m^2/s in the lower stratosphere and about 500 m^2/s in the mesosphere. Transport due to eddy diffusion is determined by using the same equations as for molecular diffusion.

3.7.2 Boundary Layer, Convection and Advection

The boundary layer is that portion of the lower atmosphere influenced by the earth's surface which responds to surface forcings with a time scale of about an hour or less. Generally, movements of air parcels vertically in the atmosphere due to pressure or buoyancy differences are known as **Convection,** and movements taking place horizontally are known as **Advection.**

Free convection is a predominantly vertical motion due to buoyancy created by thermal turbulence. It is the mass movement of molecules containing energy resulting from density differences. Thermal turbulence occurs when solar radiation heats the ground differentially between shaded areas and sun-lit areas. Radiation incident on the ground heats it and transfers heat to the air molecules in contact with the ground through thermal conduction. The heated air rises buoyantly and expands. The void is filled with cool air from shaded areas, which in turn gets heated and rises. The resulting convection is thermal. The boundary layer is under free convection when the vertical motion is due to thermal turbulence.

Forced convection is vertical motion produced by mechanical turbulence, created by winds blowing over hills and objects protruding from the surface. Mechanical turbulence is the effect of eddies of different sizes, and mixes energy and material vertically and horizontally. Forced convection also takes place when horizontal winds converge and rise. The boundary layer is under forced convection when the vertical motion is due to mechanical turbulence.

3.7.3 Prevailing Winds

A prevailing wind is the wind that blows most frequently across a particular region. Different regions on Earth have different prevailing wind directions, which are dependent upon the nature of the general circulation of the atmosphere and the latitudinal wind zones. The direction of wind is measured in terms of where it is coming from. A northerly wind blows air from north to south. A southwesterly wind blows air from the southwest to the northeast.

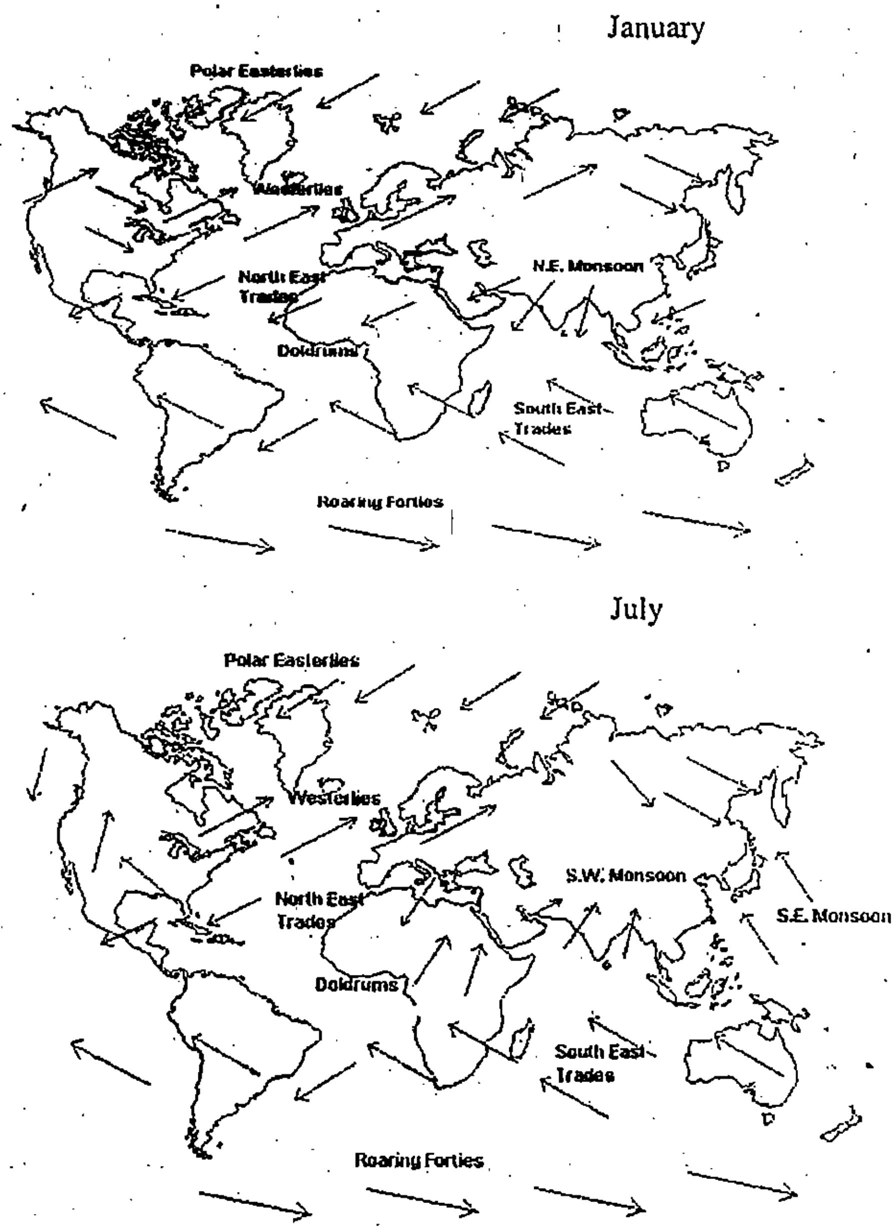
In general, the prevailing winds across the Earth may be identified as shown in **Table 3.6**, although variations arise due to the positions and differential heating rates of the continents and oceans. The rotation of the Earth about an axis, which is inclined to the direction of radiation, makes the heating of the Earth asymmetric with respect to the Equatorial plane. This uneven heating gives rise to convective currents travelling from west to east or westerlies near the tropopause. In addition, convective currents in the meridional plane transport excess heat from the tropics to the deficit areas in mid and higher latitudes.

Latitude	**Direction (from)**	**Common Name**
90-60°N	NE	Polar Easterlies
60-30°N	SW	Southwest Antitrades
30-0°N	NE	Northeast Trades
0-30°S	SE	Southeast Trades
30-60°S	NW	Roaring Forties
90-60°S	SE	Polar Easterlies

Source:http://www.doc.mmu.ac.uk/aric/eae/Climate/

Table 3.6 Prevailing wind zones

The convective air rises from the equatorial belt, and travels towards the poles where there is subsidence of air. However, because of the rotation of the earth, due to what is called **Coriolis force**, air moving towards the north pole in the northern hemisphere deflects to the right, and air moving towards the south pole in the southern hemisphere deflects towards the left. Due to frictional forces, winds turn only by about 45°, and give rise to N-E winds in the northern hemisphere and S-W winds in the southern atmosphere. These winds, which flow near 20° – 30° latitudes, were given the name of **'trade winds'** or 'traders' as they helped traders to sail across oceans centuries ago.



Source: www.doc.mmu.ac.uk/aric/eae/climate/

Fig. 3.11 Prevailing wind patterns in January and July.

The pressure differences in air close to the surface give rise to regional winds. With a definite variation in the latitudinal as well as longitudinal differences in pressure, except for disturbed days, the wind regime too adopts a corresponding pattern flowing in both equatorial or polewards and northern eastward or southern westward. These wind patterns are shown in **Fig. 3.11** for the months of January and July.

3.7.4 Monsoon Winds

Monsoon is caused by land-sea temperature differences due to heating by the sun's radiation. In winter, the continental landmass cools rapidly resulting in extremely low temperatures over central Asia. As temperature drops, atmospheric pressure rises and an intense high pressure system (anticyclone) develops over Siberia. Cold air flows out of Siberia as north-westerlies and turns into north-easterlies on reaching the coastal waters of China before heading towards Southeast Asia. During the summer months however, the large landmasses of Asia and the Indian subcontinent heat up, generating a seasonal continental region of low pressure. Airflow reverses and wind blows southwesterly across the Indian Ocean, accumulating considerable moisture, which is deposited as heavy rainfall during the wet season from May to September. Monsoon in South and South-East Asia is characterized by the SW winds flowing from May to September, and NE winds flowing from December and February, each year.

3.7.5 Tropical Cyclones

Low-pressure areas originating in the tropics, little away from the equator, which travel westwards with the prevailing winds often give rise to strong winds. They are called cyclones in the Asian region, hurricanes in Central America, and typhoons in the Far-East. High moist air is lighter than low moist air, as a heavier air molecule is replaced with a lighter water molecule. Therefore, when low-pressure areas are created for some reason or other, air with high moisture content could get into these areas first. In the process, the moist air gets cooler and the moisture gets condensed, releasing the latent heat and providing energy to build up a vortex. With the prevailing winds, this vortex of air, with diameter of several hundred kilometres, moves westwards gathering momentum and speed as it does so.

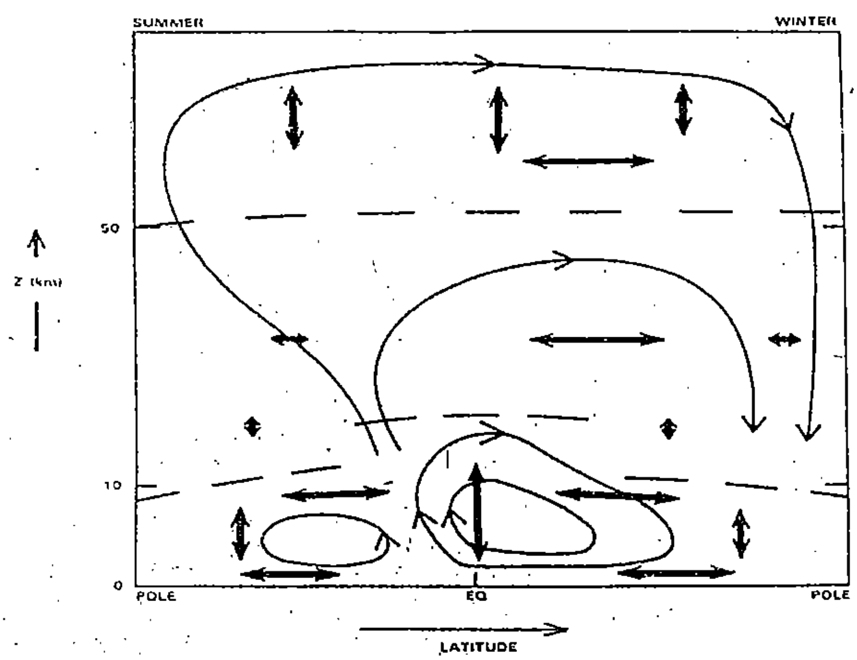
Generally, such storms originate over the oceans, and once the storm reaches the land, it results in much damage to property and even lives. There is also a tendency of cyclone storms to turn northwards as it approaches land, as evident in many events that took place in the past over the Bay of Bengal. Fortunately, the movement of cyclones can be closely monitored using satellite images, and precautionary measure taken in time.

3.7.6 General Circulation Models

Modelling the transport systems in the atmosphere is a very complex problem. A realistic model needs to consider all dynamical behaviour, photochemical and also radiative processes, including their feedback mechanisms, and coupling to oceans and land. The simulation of all the atmospheric processes in a computer requires, first the expression of these processes mathematically in the form of partial differential equations by applying the basic physical laws, and then their solution numerically.

These models represent the large-scale circulations and interactions of the atmosphere in three-dimensional form. Current atmosphere models are solved spatially over a three-dimensional grid of points on the globe with a horizontal resolute typically of 250 km and some 10-30 km in the vertical. The integration is carried over 30 min steps. The physical laws are written for each cell in the form of differential equations, and they are integrated simultaneously to determine the values of the parameters of interest. Among the outputs are temperature changes due to changes in radiative forcing, wind patterns and rainfall pattern changes.

In view of the large number of grid points and the number of parameters, the execution of these models requires super computers. In this exercise, an initial set of values is assumed, and the integration is carried through different periods of time until the values converge to a stable set. Some of the results obtained for the movements in the troposphere, stratosphere and the mesosphere are illustrated in **Fig. 3.12**, which shows the zonally averaged transport processes. The single arrows show the mean circulations, and the double arrows show the quasi-horizontal and vertical diffusion.



Source: WMO (1985)

Fig. 3.12 Zonally averaged transport processes in the atmosphere.

3.8 Weather and Climate

3.8.1 Definitions

Weather

The fluctuating state of the atmosphere, characterised by atmospheric parameters such as temperature, pressure, humidity and precipitation that occur from hour-to-hour or from day-to-day is generally known as **'weather'** or it can be defined simply as the state of the atmosphere at any given time and place. As such, weather itself is a dynamic entity, and changes with time. These weather systems arise mainly from instabilities in the atmosphere caused by its uneven heating and are governed by chaotic non-linear dynamics. As a result, these variations cannot be predicted with any certainty beyond a few days into the future. The Meteorology Departments or National Weather Bureaus in a country give forecasts on possible rainfall and temperature and their values expected during the day are referred to as weather bulletins. Such bulletins may cover specific geographic areas, large or small.

Climate

The weather, averaged over a period of time and over a specific geographical region, is referred to as the **'climate'** of that region. Thus, climate is a more static entity, but changes from place to place, depending on latitude, distance to the sea, vegetation, presence of mountains and other geographical factors. Climate also varies with time; from season to season, year to year, decade to decade or on much longer time-scale. It is sometimes defined as the average weather. A travel brochure describing a country's geography may include a write up on the country's climate, giving information on the average temperature for each season and average rainfall data. The climate is essentially a long-term averaging process. The Climate System, according to IPCC 1995 Report on the Science of Climate Change is defined as "the totality of atmosphere, hydrosphere, biosphere, and geosphere and their interactions"

3.8.2 Climate Zones

A number of climate zones or belts, characterized by different temperatures and rainfall distributions, can be traced between the equator and the pole in each hemisphere. A brief description of these zones is given below.

Tropical Climates

Centered roughly on the equator is the tropical or equatorial zone. Much of the equatorial belt within the tropical climate zone experiences hot and humid weather, with temperatures in the twenties and thirties of degrees, with the coldest month above 18°C, and relative humidity percent in the nineties. There is abundant rainfall, with annual values in the range 5,000 – 10,000 mm, due to the active vertical uplift or convection of air that takes place there, and during certain periods, thunderstorms can occur every day. Nevertheless, this belt still receives considerable sunshine, and with the excessive rainfall, provides ideal growing conditions for luxuriant vegetation. The principal regions with a tropical climate are the Amazon Basin in Brazil, the Congo Basin in West Africa and SE Asia.

Desert Climates

At about 30° north and south of the equator is a subtropical climate belt of generally dry descending air, associated with high atmospheric pressure and clear skies leading to desert conditions. Deserts are areas where the rainfall is too low to sustain any vegetation, except very scanty scrub. The annual rainfall is less than 250 mm, and some years may experience no rainfall at all. The hot deserts are situated in the subtropical climate zone where there is unbroken sunshine for the whole year due to the stable descending air and high pressure. The intense heat and lack of rainfall is typical of the desert climate, which is commonly found in the subtropical zone. Here, maximum temperatures of 40 to 45°C are common, although during colder periods of the year, night-time temperatures can drop to freezing or below due to the exceptional radiation loss under the clear skies.

Savannas

Between the subtropical and equatorial zones trade winds blow, north-easterly in the Northern Hemisphere and south-easterly in the Southern Hemisphere. These regions are much drier than the equatorial zone, but receive more rainfall than the desert climates. They have a single short rainy season when the Sun is nearly overhead, whilst the rest of the year is dry. These regions are often characterized by savannas, scrub and grassland, which blossoms during the rainy season and die off during the prolonged dry season.

Temperate Climates

In the mid-latitudes around 50° to 60° north and south there is a belt of cyclonic low pressure, where the climate is usually temperate. The low-pressure areas arise from the convergence of cold polar easterly winds and warm subtropical westerly antitrades. The precipitation tends to develop along warm and cold fronts, where cold air from the polar easterlies forces the warm, moist air of the westerlies to rise, which, on cooling, releases the moisture as clouds and ultimately rain and snow. Temperate climates are those without extremes of temperature and precipitation (rain and snow). The changes between summer and winter are generally invigorating without being extreme.

Polar Climates

The polar-regions are perpetually covered by snow and ice throughout the year. In these regions, the temperature rarely rises above freezing. During the long polar nights, which last six months at the poles, temperatures can fall to extremely low values, as low as -80°C. Polar climates tend to be dry because the descending air is cold and lacks significant moisture, precluding the formation of clouds and snowfall. Some polar regions receive less than 250 mm of precipitation each year. At the highest latitudes in the polar-regions, the cold air sinks producing high atmospheric pressure producing dry, icy winds that tend to radiate outward from the poles

3.8.3 El-Nino Phenomena

El Niño, the Spanish name for "Christ Child", is the name given to the occasional development of warm surface waters in the Pacific Ocean along the coast of equatorial South America. El Niño occurs roughly every 2 to 7 years, usually around Christmas, and lasts usually for a few weeks or months. Sometimes an extremely warm event can develop that lasts much longer.

The formation of El Niño is linked with the cycling of a Pacific Ocean circulation pattern known as the **El Niño Southern Oscillation** or ENSO. In a normal year, low atmospheric pressure develops over northern Australia and Indonesia, with high pressure over the Pacific. Consequently, winds over the Pacific move from east to west. The easterly flow of the trade winds carry warm surface waters westward bringing rainstorms to Indonesia and northern Australia. Along the coast of Peru and Ecuador, cold deep water wells up to the surface to replace the warm water that is pulled to the west.

El Niño has the capacity to influences atmospheric wind patterns worldwide as well as flooding in Peru and Ecuador, and drought in Indonesia and Australia. The possible impacts include a shifting of the jet stream, storm tracks and monsoons, producing unseasonable weather over many regions of the globe. During the El Niño event of 1982-1983, some of the abnormal weather patterns which were observed included drought in Southern Africa, Southern India, Sri Lanka, Philippines, Indonesia and Australia; heavy rain and flooding in Bolivia, Ecuador, Northern Peru and Cuba; and hurricanes in Tahiti and Hawaii. Because El Niño may influence the mid-latitude Northern Hemisphere jet stream, even the weather in Europe and North America can be influenced. The most recent El Niño episode in 1997 and 1998 brought record high winter temperatures to many areas in Europe including the UK. Globally, 1998 became the warmest year on record.

3.8.4 Climate in South Asia

Monsoons

The South Asian region is characterized by a tropical monsoon climate. Differences in rainfall are of primary significance in defining the climate of the region. The most important feature is the seasonal alteration of atmosphere flow patterns associated with the monsoon, which is due to the seasonally modulated excess heating of the Asian landmass in summer and the excess cooling in winter compared to the adjacent oceans. Two monsoon systems operate in the region: the southwest or summer monsoon (June-September) and the northeast or winter monsoon (December-April). The rainfall during the summer monsoon largely accounts for the total annual rainfall over most of South Asia (except over Sri Lanka where rainfall of the winter (Northeast) monsoon is dominant) and forms a chief source of water for agriculture and other activities.

The monsoon rainfall in South Asia is characterized by large spatial and temporal variability. The arid, semi-arid region encompassing Pakistan and Northwest India receive monsoon rainfall as low as 50 mm while parts of Northeast India and the west coast receive over 1000 mm. This region also features large year-to-year variations in the rainfall frequently causing severe floods/droughts over large areas.

Extreme Events

There are two major anomalous regions: the arid and semi-arid parts comprising of large areas of Pakistan and north-western Indian states of Rajasthan, Punjab, Haryana and Gujrat which experience frequent droughts, and the eastern Himalayan sub-region, fed by the Ganga-Brahmaputra-Meghna river system, which are subjected to frequent floods. In India, during the period 1871-2000 there were 22 drought years and 19 flood years. There had been three cases of prolonged drought condition, viz., 1904-05, 1965-66 and 1985-87. Such cases cause great calamity.

Similarly, there had been two cases of prolonged flood conditions, viz., 1892-94 and 1916-17. Studies indicate a clear relationship between the occurrence of droughts (floods) in South Asia with the El Niño (La Niña) events in the east Pacific Ocean. It has been observed that, during the period 1856-1997 there were 30 El Niño years in which the averaged monsoon rainfall over India was 7% below normal; in 10 out of these 30 cases, drought conditions prevailed over India. Two years featured flood conditions (1878 and 1983). During the same period there were 16 La Niña years, 9 of which featured flood conditions over India.

The cyclonic storms originating in the Bay of Bengal and blowing towards India and Sri Lanka has been a frequent occurrence. Around 16 cyclonic disturbances occur each year, of which about 6 develop into cyclonic storms. During 1965 – 1990, on an average 2.3 severe cyclones with hurricane force winds have taken place.

3.9 Enhanced Global Warming

3.9.1 Anticipated Climate Change

Earlier in this unit, we considered the Greenhouse Effect and Greenhouse Gases, and learnt that the Greenhouse Effect helps the Earth to maintain its global mean temperature at the present value of 15 °C. In this Section, we will discuss more on this topic, as it has been observed that during the last century, the concentrations of these Greenhouse Gases have increased significantly, causing an enhancement of the global warming and a corresponding increase in the global mean surface temperature by about 0.6°C. Such studies have been undertaken by an expert body called the Inter-Governmental Panel on Climate Change (IPCC) established by the UN.

Scientists have expressed fear that if the emissions of these gases continue to increase without any control, the Earth's global mean temperature will also increase correspondingly by several degrees. Based on palaeo-data, scientists predict that even a few degrees rise in the global mean temperature could bring about a drastic change in the climate, resulting in adverse impacts on both the terrestrial environment and the humans. Sea-level rise is an important consequence of enhanced global warming. A combination of climate change and sea level rise would affect all socio-economic sectors including agriculture, forestry, coastal zones, wetlands, water resources, energy generation, human settlements, tourism and also would increase the occurrence of extreme events such as droughts, floods and cyclonic storms.

3.9.2 Increase in Radiative Gas Concentration

There are two types of radiative gases: those occurring naturally in the atmosphere, and those introduced in recent times through various industrial applications.

Naturally Occurring Gases

As mentioned before, the three naturally occurring radiative gases are carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O). In addition, H_2O and O_3 present in the atmosphere also contribute to the Greenhouse Effect. Each such gas has a specific global warming potential (GWP), which is expressed relative to that of CO_2, which is taken as unity. During the past several decades, there has been a significant increase in the concentration of these three radiative gases due to various human activities, resulting in enhanced global warming.

Carbon dioxide is emitted mainly by combustion of fossil fuel and biomass for energy generation, and due to changes in land use, particularly in the tropics. Methane is produced mainly due to agricultural activities including paddy cultivation, rearing of ruminant animals, disposal of waste and also from gas fields. Various human activities such as biomass burning, application of nitrogen containing fertilizers and fuel combustion contribute to atmospheric N_2O. Their estimated pre-industrial concentrations and the current concentrations are shown in **Table 3.7**, along with their global warming potentials.

Radiative Gas	Pre-industrial concentration (ppm)	2000/1998 concentration (ppm)	GWP for 100 yr time horizon	Contribution to Global Warming %
Naturally Occurring Gases				
Carbon dioxide	280	369	1	63.8
Methane	0.70	1.76	23	19.2
Nitrous oxide	0.27	0.32	296	5.7
Industrial Gases				
CFC 11	0	268(-6)*	3,800	10.0
CFC 12	0	533(-6)	8,100	
HCFC 22	0	132(-6)	1,500	0.82
HCFC 123	0	11(-6)	90	
HFC 134a	0	7.5(-6)	1,300	
HFC 152a	0	0.5(-6)	140	
Sulphur hexafluoride	0	4.2(-6)	22,200	0.08
Perfloromethane	40(-6)	80(-6)	5,700	0.29
Perfluoroethane		3.0(-6)	11,900	

Source: IPCC (2001)

Table 3.7 **Pre-industrial and current concentrations of radiative gases**

Industrial Gases

Among the second type are the halocarbons such as chloro-fluoro-carbon (CFC) compounds, hydro-chloro-fluoro-carbon (HCFC) compounds, hydro-fluoro-carbon (HFC) compounds, perfluoro-carbon (PFC) compounds and sulphur hexafluoride (SF_6). A common property of all these gases is their long residence time in the atmosphere extending from several years to several centuries. CFCs and HCFCs are used as refrigerants, foam blowing agents and aerosol propellant and cleansing fluid. However, in view of their ability to deplete the stratospheric ozone layer resulting in many adverse impacts, their consumption has already been phased out in developed countries, and are being phased out within this decade in other countries. HFC was introduced in recent times as a substitute for CFCs. Their concentration levels are also shown in Table 3.8. Though these gases are found in extremely small quantities, because of their high GWP values, they contribute significantly to radiative forcing.

3.9.3 Emission Scenarios

In order to estimate the future changes in the climate arising from enhanced global warming, it is necessary to first estimate the possible emissions caused by human activities in the future. These would include emissions from energy generation by fossil fuel combustion, agricultural activities and land use. The energy consumption itself depends on several socio-economic factors, such as population growth and economic prosperity of nations. If we were to make projections of emissions up to the end of this century, no one could say definitely what the population and the economy of nations would be at this period of time. The best we could do is to make certain assumptions and develop several possible scenarios.

Emission Storylines

The IPCC has previously published two assessment reports based on six future emission scenarios developed in 1992 using economic development projections of the World Bank. The best estimate scenario out of these was referred to as IS92a. In 2000, the IPCC adopted a Special Report on Emission Scenarios (SRES) in which 35 possible emission scenarios were considered. The Third Assessment Report (TAR) of the IPCC published in 2002, made us of these scenarios in their draft form. The set of scenarios comprised four basic cases, described as 'storylines', designated A1, A2, B1 and B2. Each storyline describes a set of demographic, social, economic, technological and policy options that would reflect possible situations in the future. Under each storyline, several variants were considered, designated A1B, A1C etc.

Emission Projections

Once the population growth and energy consumption from different sources are known, it is possible to work out the GHG emission values expected in 2100, under each of the scenario families. Their emission levels, along with that of SO_2, which was an additional gas included in the TAR, are shown in **Table 3.8.** The range of values under each family reflects the values for variants under each family. The two fossil fuel dominated alternatives, A1C and A1G gives higher GHG emissions, and the scenario A1T gives the lowest, than the scenario A1B. The current emission levels are also shown here.

The emissions of Methane and nitrous oxide were based on the population increases and the agriculture intensities needed to provide adequate food for the people, as these two gases are emitted mostly from agricultural activities. The energy related methane emissions were based on the energy consumption scenarios.

Scenario	CO_2 (GtC)	CH_4 (Mt CH_4)	N_2O (MtN)	SOx (MtS)
A1B	13.5-17.9	289-510	5.8-9.7	27.6-47.0
A1C	36.7-32.1	392-672	6.1-6.2	83.3-49.8
A1G	30.8-30.3	289-436	5.9-6.2	27.4-40.5
A1T	8.7-4.3	291-274	4.8-5.4	27.4-20.2
A2	28.2-34.4	549-1069	8.1-16.5	60.3-92.9
B1	5.7-6.4	236-377	5.4-9.3	15.5-25.9
B2	13.3-19.1	465-607	6.9-12.0	33.3-47.9
Current	6.4	600	16	88

Source: IPCC (2001)

Table 3.8 Emission projections in 2100 under scenarios considered in the IPCC Third Assessment Report.

3.9.4 Temperature Rise Projections

Climate Sensitivity

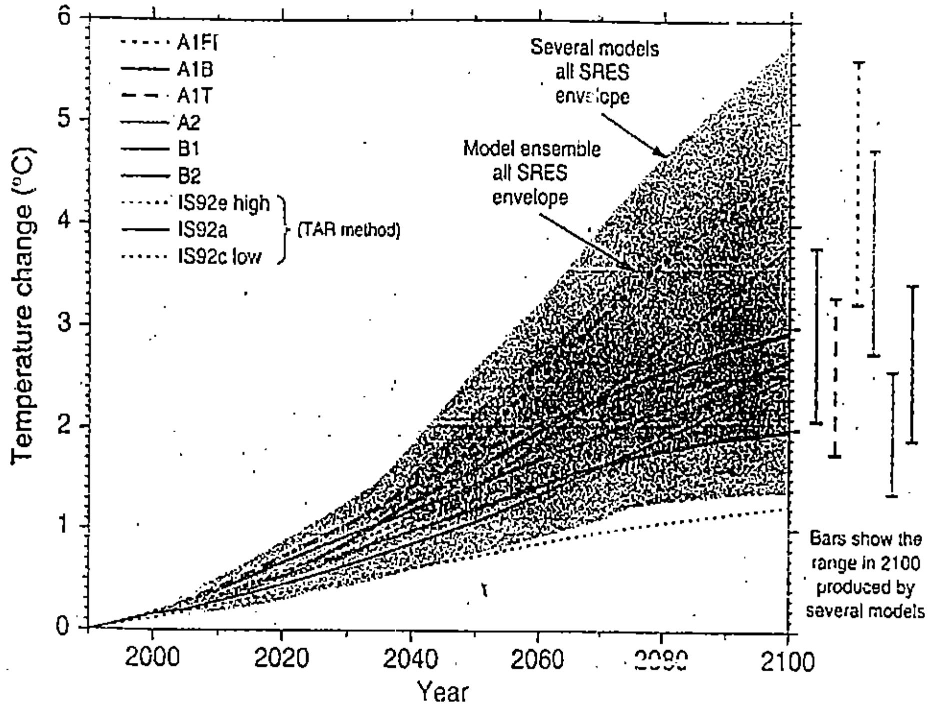
In recent years, with the availability of powerful computers, both in memory and speed, scientists have been able to develop computer models to study the Earth's climate system, using the General Circulation Models (GCM) described earlier. One parameter that was calculated using these models was the rise in surface mean temperature due to doubling of CO_2 concentration, denoted by $\Delta t/\{2x[CO_2]$. This is equivalent to the Earth's thermal capacity and is referred to as the **climate sensitivity** of the Earth system. Even with these complex models, only an approximate value could be obtained, because of the difficulties encountered in expressing all physical processes mathematically to simulate them on a computer. Hence, a range of possible values is worked out, representing the minimum, the best estimate and the maximum. These were estimated to be 1.7, 2.8 and 4.2 $^{\circ}C/\{2x[CO_2]$, respectively, as given in the TAR.

Temperature Projections

The global mean temperature rise could now be computed using the range of emissions and the climate sensitivities discussed above. Climate change simulations were assessed up to 2100. The GCM in which the atmosphere-ocean coupling was considered, was used in these projections.

The SRER 1998 has considered 3 new scenarios, only a selected few have been considered here. These are the four marker scenarios A1B, A2, B1 and B2, and two others from the A1 family representing different energy technology options; A1FI and A1T. Corresponding to the mean climate sensitivity of 2.8 $^{\circ}C/\{2x[CO_2]$, these six emission scenarios give a set of values for temperature rise, ranging from 2.0 °C to 4.5 °C.

The projections corresponding to the extreme climate sensitivities of 1.7 and 4.2 $^{\circ}C/\{2x[CO_2]$, were in the range 1.4 °C - 5.6 °C. The full set of values corresponding to all SRES scenarios and the climate sensitivities lies within the range 1.4 °C and 5.8 °C. The different projections obtained for different emission scenarios are shown in **Fig. 3.13.** These projected increases in temperature, particularly the high values, may not be realised if the developed countries adopt new measures to mitigate GHG emissions in the future.



Source: IPCC (2001)

Fig. 3.13 Projections in global mean temperature due to anticipated global warming

3.9.5 Precautionary Measures

The Framework Convention on Climate Change

The topic of enhanced global warming and its adverse impacts globally has drawn the attention of scientists and policy makers in the seventies and the eighties, and as a result, the UN adopted the Framework Convention on Climate Change (FCCC) in 1992 at the Earth Summit meeting held in Rio de Janeiro. The main objective of the FCCC was to achieve stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system. The Convention required 38 industrialized countries, listed in an Annex, to reduce their GHG emissions back to 1990 levels by 2000. However, this was a measure to be taken totally voluntarily by each country. Fears have been expressed by developing countries and others that this measure alone would not achieve the objectives of the Convention.

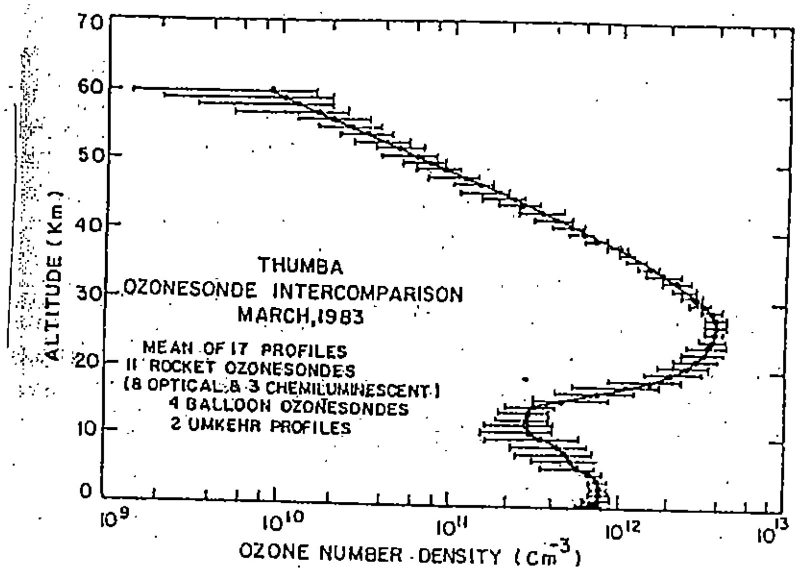
The Kyoto Protocol

One of the adverse impacts of global warming is the rise in sea level, which could affect severely the large number of low-lying small island states. With their initiative, the Parties to the FCCC, in 1997, adopted a legally binding protocol, Kyoto Protocol to UNFCCC, which made it mandatory for the 38 countries to reduce their GHG emissions below their 1990 levels by specified amounts, during the 5-year period 2008-2012. The overall reduction of emissions is expected to be at least 5% below 1990 levels. However, if the countries were allowed to emit without any controls in place, their emissions would increase steadily achieving some high values by 2010. When compared to these uncontrolled values projected for the same commitment period, the expected reductions could be as high as 25-30 %. The Kyoto Protocol has provided several mechanisms through which this burden could be shared with developing countries.

3.10 Ozone Layer Depletion

3.10.1 Occurrence of the Layer

It was mentioned earlier that the ozone layer is responsible for the absorption of harmful UV-B radiation in the range. Ozone is found in the atmosphere as a minor constituent extending from ground level up to about 100 km, concentrated into a layer between about 15 – 50 km with a peak around 20 - 25 km, depending on the latitude. The distribution of ozone concentration with height has a latitudinal dependence, the profile being thin at low latitudes and thick at high latitudes. A low latitude profile based on rocket and balloon measurements is shown in **Fig. 3.14**. The thickness of the ozone layer is generally expressed by its column density, which is the height of a hypothetical ozone column, expressed in milli-cm, formed when the ozone layer is compressed to a uniform pressure of one atmosphere. It is sometimes referred to as the Total Density or Total Ozone, using a unit known as the **Dobson Unit** (DU). The global average thickness of the ozone layer is 300 DU.



Source :IPCC (2001)

Fig. 3.14. Height distribution of ozone concentration

The ozone column density is measured from ground level using special spectrometers developed for that purpose. A large number of observatories have been established around the world to monitor ozone level using these instruments. The ozone layer thickness is also measured from the top of the atmosphere using satellites. These measurements show that the average thickness of the ozone layer around the tropics is in the range 240 – 260 DU, while it is in excess of 350 DU in mid-latitudes, where the thickness has a seasonal variation also, being highest in the spring around 450 DU, and minimum in the autumn.

3.10.2 Ozone Depleting Substances

Chloro-fluoro-carbons (CFC)

Sherwood Rowland and Mario Molina, in 1974, predicted that the ozone layer could get depleted due to the action of a certain family of industrial chemicals, which were released to the atmosphere in large quantities annually. This family is the same as that mentioned previously as one of the industrial greenhouse gases, namely, chloro-fluoro-carbons (CFC). When a fluorine atom is substituted in place of a chlorine atom in carbon-tetrachloride, (CCl_4), depending on the number of chlorine atoms replaced, a series of compounds could be obtained - $CFCl_3$, CF_2Cl_2, CF_3Cl etc. These compounds are very stable in the troposphere.

Rowland and Molina found that these substances would slowly diffuse into the stratosphere and get exposed to UV radiation there. Though they are not chemically active while in the troposphere, it was found that UV radiation in the stratosphere could easily detach a chlorine atom from these compounds. The released chlorine atom could then initiate the following reactions with ozone, converting it to oxygen.

$$O_3 + Cl \rightarrow ClO + O_2$$
$$O + ClO \rightarrow Cl + O_2$$

There are several other reaction paths, and their overall effect is to convert ozone into oxygen. One important factor here is that the chlorine atom, which initiates the chain of reactions will end up again as a free atom. Here, chlorine atom acts as a catalyst, and it is estimated that each chlorine atom could convert several tens or even hundreds of thousands of ozone molecules into oxygen, which makes it possible for a man-made substance to make a dent in the naturally occurring ozone layer.

Bromine Compounds

Subsequently, it was found that any chemical containing either chlorine or bromine, which is relatively stable at ground level, but breaks down in the stratosphere due to the action of UV radiation, could destroy ozone. These substances include hydro-chloro-fluoro-carbons (HCFC) used in domestic air-conditioners, bromo-fluoro-carbons (BFC), also known as halons used in fire extinguishing, carbon tetrachloride, methyl chloroform, methyl bromide used as a soil fumigant in tea nurseries and many others.

In one reaction path, ozone gets converted to oxygen without the intervention of atomic oxygen. This is

$$O_3 + Cl \rightarrow ClO + O_2$$
$$O_3 + Br \rightarrow BrO + O_2$$
$$ClO + BrO \rightarrow O_2 + Cl + Br.$$

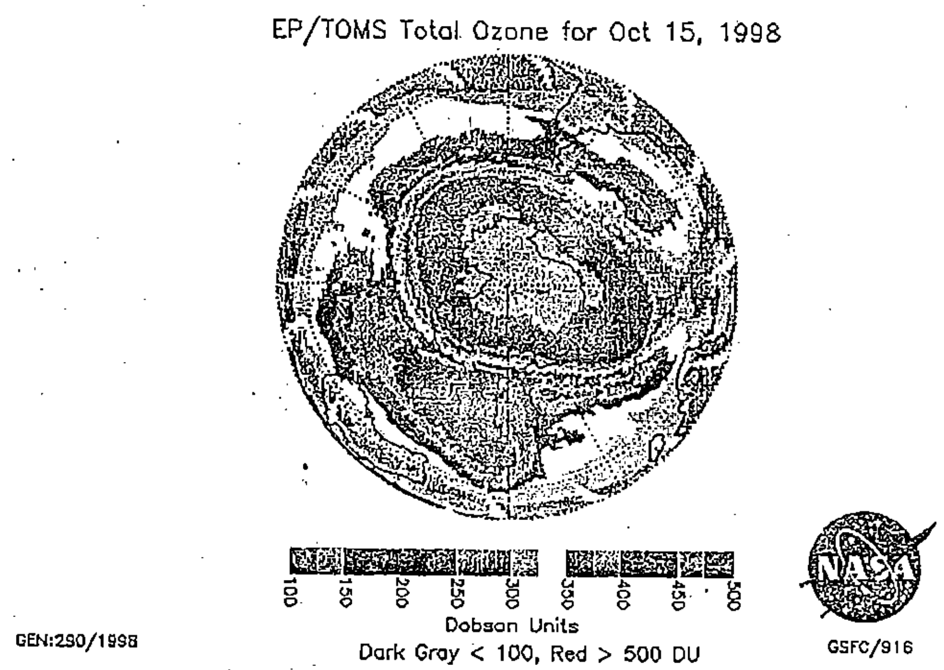
Since atomic oxygen concentration in the stratosphere is not very high, this path is considered to be more efficient than that involving atomic oxygen. Also bromine is considered to be more efficient in converting ozone into oxygen than chlorine.

3.10.3 The Depletion

The Ozone Hole

The depletion of the ozone layer was first detected over Antarctica by a team of British scientists who have been measuring the ozone content there from 1957 onwards using a Dobson Spectrometer. Up to about 1972, these measurements remained more or less constant around 300 DU, though there had been a year-to-year fluctuation. However, they noted that beginning 1972, the measurements showed a gradual decline during the spring months of September and October each year, reaching values below 200 DU by 1985.

Satellite measurements also showed a reduction in ozone column density during the months of September to November each year beginning 1979. During 1992, the reduction had been even below 150 DU, which was more than 50% from previous observations, and the depleted area covered the entire Antarctica. This region of depleted ozone was in fact referred to as the '**Ozone Hole**' in the sky. Measurements taken in subsequent years showed that in the Ozone Hole, the column density has dropped even below 100 DU. A mapping of the ozone hole over the south pole made by satellite instruments is shown in **Fig. 15.**



Source: www.unep.org/ozone/

Fig. 3.15. Ozone levels over the south pole showing the ozone hole.

Non-Polar Latitudes

Scientists also found that during winter months, ozone column density suffered a significant reduction in other parts of the globe too. An assessment of data taken during the 20-year period from 1979 to 1997 has concluded that in both hemispheres, in mid- and high latitudes in all seasons, the depletion has been large and statistically significant, while in the low latitudes, between 20 °N and 20 °S, there has been no statistically significant depletion. The actual observed trends, averaged over the longitudes, are shown in **Table 3.9**

Latitudes	Mean Depletion (% per decade)		
	Annual	Dec - May	June - Nov
50° - 65 °N	3.7	4.4	2.8
30 ° - 50 °N	2.8	3.8	1.7
20 ° N – 20 ° S	0.5	0.3	0.7
30 ° – 50 ° S	1.9	2.4	1.4
50 ° – 65 ° S	4.4	3.4	5.2

Source: WMO (1998)

Table 3.9 Depletion of the ozone layer in non-polar latitudes

It should be remembered that in tropics, even though the depletion has not been significant, the year-round total density has been the lowest, resulting in the highest flux of solar UV radiation reaching the ground.

3.10.4 Precautionary Measures

The prediction in 1974 of a possible depletion of the ozone layer caused fear among scientists and policy makers of its grave consequences, which included an increase in the incidence of skin cancer cases, particularly among the fair-complexioned people, an increase in the incidence of eye cataracts, a lowering of the human immune system efficeiency, a decrease in the crop productivity and a reduction in the life of synthetic materials. This prompted the UNEP to adopt in March 1985, an international agreement called the Vienna Convention for the Protection of the Ozone Layer, with the objective of encouraging countries to undertake studies on the ozone problem, develop alternatives to ozone depleting substances and coordinate exchange of information.

The discovery of the 'Ozone hole' in 1985 and the growing scientific evidence that man-made CFCs were indeed responsible for the ozone layer depletion, made UN Environment Programme (UNEP) to take the initiative to adopt a legally binding protocol, called the Montreal Protocol on Substances that Deplete the Ozone Layer, for the purpose of controlling the consumption of ozone depleting substances, initially applicable for the developed countries. Under this Protocol, all developed countries have already phased out the use of CFCs and other ozone depleting substances, while the developing countries have frozen their consumption, with a target set for 2010 for a complete phase out.

There has been a somewhat conflict between the Kyoto Protocol and the Montreal Protocol, in that HFC, which is the substance that was introduced under the Montreal Protocol as an alternative to CFCs, has been included as a substance to be controlled under the Kyoto Protocol, because of the global warming potential of HFC.

Summary

The atmosphere is essential for the survival of life on Earth. Its major constituent, oxygen provides life-giving support to animals and prevents harmful solar radiation from reaching the ground. Its constituent, CO_2 plays a key role in the carbon cycle, and along with many minor constituent gases, serves as a radiative gas keeping the Earth warm enough to maintain life on it. Some of them filter out harmful solar radiation and control the pollution in the atmosphere. Moisture and aerosols are important players in the energy balancing process, through absorption and scattering of solar and terrestrial radiation.

The key climate parameters, pressure, temperature and precipitation, have a wide variation with the altitude and also a spatial and temporal variation globally. The stratification of the upper atmosphere into several regions having different temperature gradients is characteristic to the Earth. The variation of pressure and temperature over the surface causes winds to blow, which have regular as well as sporadic features. The variation in climate parameters could be linked to variations in solar energy deposited on the Earth, which as a seasonal and a diurnal variation. The Earth is classified into several climate zones depending on the temperature and rainfall distribution, and also the moisture content.

Certain human activities over the past several decades have caused a threat to the future of the Earth's ability to sustain life on it. One is the enhanced global warming, arisen as a result of changing the composition of the atmosphere through emissions from uncontrolled combustion of fossil fuel and biomass. The other is the depletion of the ozone layer caused by the emission of chemical substances that could convert ozone into oxygen in the

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UNDERSTANDING THE ENVIRONMENT

BLOCK
3

HYDROSPHERE

Unit - 1

Hydrosphere

Unit 4
Hydrosphere

Contents

Introduction

The hydrosphere is the water portion of the earth, which consists of liquid water in lakes, rivers, oceans, sedimentary rocks as well as ice in sea ice. This lesson focuses on various aspects of the above components of the hydrosphere. It begins with an overview of water, which is essential for life on earth.
As you know, water is a peculiar substance. It is the only liquid commonly found on the surface of the planet and that is useful to life. **The structure of the water molecule and its specific properties** have created surprising physical and chemical conditions. This has made it biologically important thus enabling many organisms to exploit all systems of water.

All water on the planet is constantly recycled , by a system known as **the hydrologic cycle** driven by solar energy. In this cycle, water is lost to the atmosphere as vapour from the earth, and in turn is precipitated back in the form of rain, snow, frost etc. This precipitation and evaporation continues forever and there by maintain the amount of water in the air, land and ocean.

When you learn about the **composition of the hydrosphere**, you will find that our Earth contains an immense amount of water, covering 71% of its surface up to about an average depth of 3800 meters. Out of this large amount of water, only about 0.25% could be used for human consumption. Over 97% of this water is deposited in the ocean depressions.

The habitats in the **freshwater environment** frequently integrate with one another, and they are divided into two categories: the standing water bodies which comprise lakes and ponds etc. and the running water bodies that include streams and rivers.
The physical structure of the above water bodies is determined by the distribution of factors such as light, heat and currents which vary by day and season. The chemical structure results from the uneven distribution of chemicals.

The **estuaries** are partially enclosed coastal embayment, where fresh water and sea water meet and mix. Most of the characteristics of the estuaries result from the flow of fresh water and sea water. Tides add additional complexity to the circulation pattern and salinity gradients, within the estuaries.

The estuaries discharge their contents into **the Ocean,** which is a large interconnected body of water that is constantly in motion, mixed and moved by winds, waves and currents.

4.1 The structure of water and its properties

Water is probably the most important natural resource on the earth. It is a vital chemical constituent of living cells and a habitat for large number of organisms. It is worth while then, looking at some of its chemical and physical properties.

Water has the chemical formula H_2O, and it exists as a liquid, solid and as vapour. This simple molecule has many surprising properties which are biologically important, and many of them result from its molecular structure.

A molecule of water results, from the formation of two single covalent bonds between an oxygen atom and two hydrogen atoms **(Fig. 4.1a).** The oxygen atom has a greater power in attracting electrons than that of the hydrogen atoms. As a result, the oxygen atom acquires a partial negative charge, and each hydrogen atom a partial positive charge **(Fig 4.1a).** Molecules such as water, that exhibit charge separation are termed polar molecules and they have a marked affinity for each other.

The charge separation has resulted in an arrangement of a tetrahedron for the water molecule and a bond angle of 104.5 degrees. It also allows the oxygen molecule to form weak hydrogen bonds with the oppositely charged hydrogen atom of another molecule. This allows water molecules to combine together in an uninterrupted net work **(Fig 4.1b).** Each of these hydrogen bonds is individually very weak and transient and contains only about 1/16 of the energy of a normal covalent bond.

Therefore, when water changes, from the solid phase to liquid and then to vapour, hydrogen bonds strain or break. However, the cumulative effects of very large numbers of such bonds could be enormous, and are responsible for many of the important thermal properties of water about which you are going to learn now.

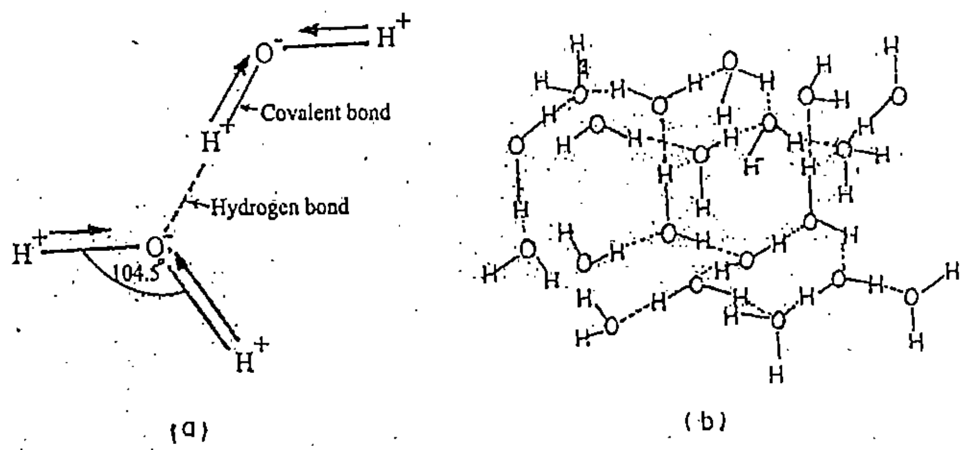


Fig. 4.1 a. The arrangement of H and O atoms in a molecule of water and its charge separation
b. Lattice of Hydrogen bonds

Thermal properties of water

High specific heat capacity

Specific heat is the amount of heat (4.2 KJ) required to raise 1 Kg of water by 1°C. Water has a high heat capacity. Therefore it can absorb large amounts of heat with only a small increase in temperature. This is because much of the energy is used in breaking the hydrogen bonds. High specific heat capacity of water enables aquatic organisms to survive even the intense solar radiation at the equator, as water provides a very constant environment. Therefore biochemical processes can operate only over a small temperature range and proceed at more constant rates are less likely to be inhibited by extremes of temperature in aquatic environments. Water also stores a lot of heat per unit volume and large volumes of water can therefore alter climate.

High latent heat of fusion

Latent heat of fusion is the amount of heat gained or lost per unit mass of a substance, changing from a solid to a liquid, or liquid to a solid phase, without an accompanying rise or fall in the temperature.
For 1 Kg of water at 0°C to freeze, a large amount of heat ((3.34×10^5 J) must be released. Conversely when ice at 0°C melts to water at 0°C, an equal amount of heat must be absorbed.
High latent heat of fusion of water helps to maintain the temperature at a critical point (at 0°C), before freezing. This prevents sudden freezing of water and melting of ice in temperate lakes, which make the transition between seasons less abrupt, enabling organisms to adjust gradually to the changing climate. In addition, it prevents the formation of ice in the tissues of organisms when the body is freezing. (Ice crystals are particularly damaging if they develop inside cells). Also this property inhibits large scale freezing of oceans.

High latent heat of evaporization

Latent heat of evaporation is the amount of energy that required when liquid water changes into vapour, or vapour into liquid water, at 100°C. As a large amount of energy is required to break the hydrogen bonds between the molecules (so that they can escape as a gas), water has an unusually high boiling point. This character of water causes most of it to exist in liquid form at atmospheric temperature.

The energy needed by water molecules to vaporize, results in the loss of energy from their surroundings. This results in a cooling effect. Sweating and panting by mammals bring down their body temperatures. The evaporation of sweat and saliva allows, loss of large amounts of body heat with a minimal loss of water from the body. For instance every gram of water that evaporates from the human body removes 586 calories of heat from the body.
This high latent heat of evaporation of water is also important in reducing the evaporation of water from lakes and seas and moderating sea surface temperatures by transferring large quantities of heat to the atmosphere through evaporation.

Cohesive properties and surface tension

Water molecules being very polar themselves, are attracted to each other and also to other polar molecules. When this attraction is between each other; it is referred to as **cohesion.**

It is because of this cohesive property, that water can form a lattice of hydrogen bonds, which allows it to remain as a liquid, at normal atmospheric temperatures. At the air/water boundary they stick together so thoroughly, resulting a high **surface tension**, that will support light weight substances if they are spread out over a larger area. For example, a floating leaf with a large surface area, or an insect like a water strider, that distributes its weight over a larger area (by spreading its long legs), can float on water. The surface tension decreases with increasing temperature, salinity and addition of organic surfactants produced by plants and animals.

Adhesive properties and capillary action

The attraction between a water molecule and a polar molecule of a different substance is called **adhesion.** Water is adhesive to any substance with which it can form hydrogen bonds. This is why certain things get "wet" when they are dipped in water, and why waxy substances composed of non polar molecules do not get wet.

The adhesion of water to substances with surface electrical charges, is responsible for capillary action (rising of water and watery substances in narrow tubes). This ability helps to pull water up through conducting tissues, in trees as tall as 100m, and water to creep up trough minute spaces in soil making it available to the roots of plants. It is also responsible for the uptake of water (imbibition) by seeds of plants, to permit germination.

Solvent properties

Water is almost a universal solvent, with the ability to dissolve more substances than any other liquid. This is because the solvent action is of two types. One depends on the polar characters of the molecules and the other on the hydrogen bonding. The polar character helps to dissolve various salts by the interaction of their ions with the charges on the water molecule. (E.g. table salt – NaCl – dissolving in water) **(Fig 4.2).**
Various non polar organic and inorganic compounds are held in solution by hydrogen bonding thus producing hydrates (E.g. water and protein molecules).

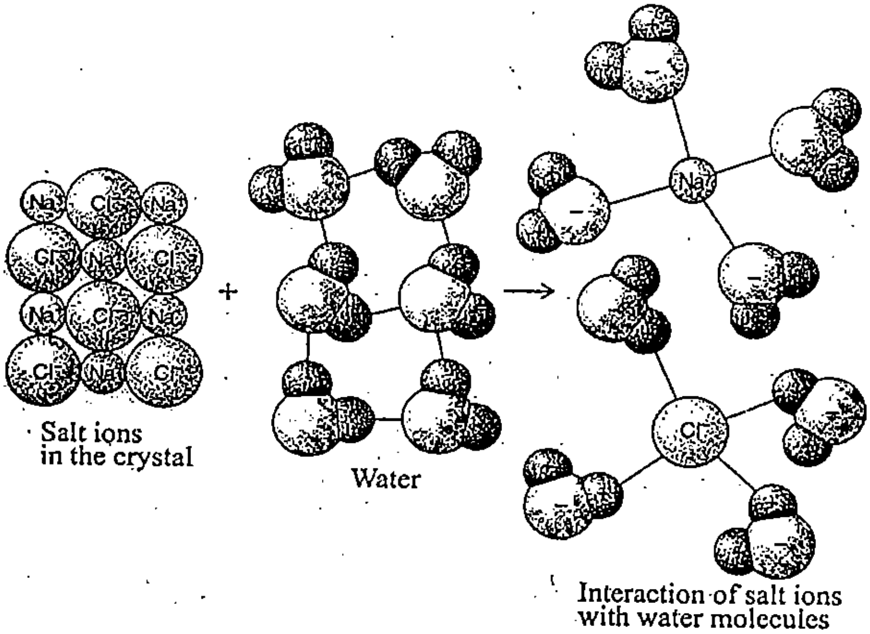


Fig 4.2 Action of charged water molecules with Na^+ and Cl^- in a salt crystal

Density and Freezing properties

The density or specific gravity of pure ice at 0°C is 0.9168. It is about 8.5 % lighter than liquid water at 0°C (0.99987) and that is why ice floats on water. Solid ice has a regular molecular structure with slight gaps between the molecules. As ice melts, the regular structure breaks up and the gaps fill up. The density therefore increases. This effect continues until 3.98°C (~4°C), so water at this temperature has the highest density. Above 4°C, the density decreases, as molecular expansion takes place.

The difference in density of hot and cold water is responsible for the great resistance to mixing of water masses. The rate of change of water density is not constant with change in temperature: the density decreases more rapidly at high temperatures **(Fig. 4.3b).** Density of water increases with increasing concentrations of dissolved salts in a linear fashion. These variations may be small, but they cannot be ignored.

Now let us consider how these density differences enable animals in water to survive during ordinary winters.
On reaching a temperature of 4°C, water sinks to the bottom **(Fig. 4.3a)**. The water at the surface cools further and forms an ice crust, which acts as an insulator, and prevents the total freezing of the water body. The animals can live under the ice cover until winter is over.

The density of water is 775 times greater than that of air at standard temperatures & pressure (0°C, 760 mm Hg). Due to this high density, water makes aquatic organisms buoyant against gravitational pull and reduces the energy required for an organism to maintain its position. In most fresh water animals the supporting tissues are reduced (e.g. most of the plankton, benefit by just floating passively in the water column, as buoyancy supports their weight). These adaptations are conspicuous even among aquatic vascular plants.

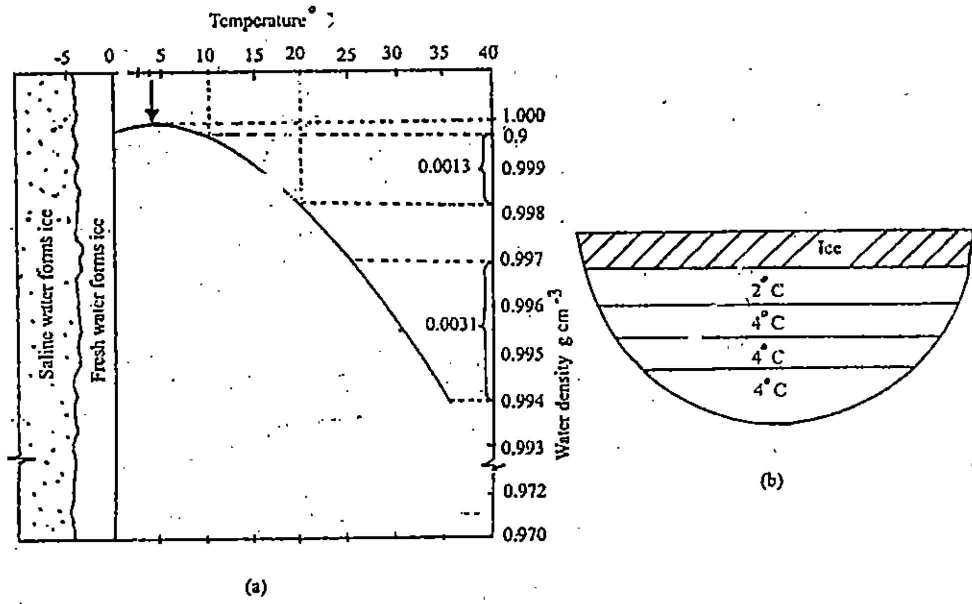


Fig. 4.3 a. Changes of density with temperature in freshwater
b. Temperature variation with depth during winter

Viscosity and other properties of water

Viscosity is a measure of a liquid's resistance to flow. It is influenced to a minor extent by the salinity of the water, but to a considerable extent by temperature. Viscosity of water is much higher at lower temperatures, and doubles when temperature is lowered from 25°C to 0°C. However, it decreases with increasing temperature.
The viscosity of water offers frictional resistance to moving organisms. Therefore organisms have to spend considerable amount of energy to overcome changes in viscosity. The sinking rates & the distribution of passive organisms (such as plankton), are influenced by density related changes in viscosity.

Protective functions of water

Water absorbs harmful infra red radiation of the sun. Large amount of the incoming solar radiation is dissipated in the evaporation of water from the ecosystems of the world. It is this dissipation of energy that moderates the climate and makes it pleasant for life to exist on earth.

Effluent from industrial plants and factories, and insecticides from agricultural lands are discharged into water, where they become diluted and sometimes even change the chemical composition. Thus, water makes them less dangerous to life. In industrial countries, oil pollution and nuclear pollution can become serious threats to aquatic life as evident from some recent accidents. Also, gases released by certain factories cause rain water to become acidic. The deleterious effects of the above are reduced by dilution with water.

Activity 4.1

State the physical properties of water that are important, in the biological processes given below.

1. Prevent formation of ice in the vascular tissues of plants and in animal tissues.------------------------------------
2. Help in cooling the bodies of animals. ---------------------------------
3. Moderate daily and seasonal temperature variations and stable body temperature of organisms.-------------------------------------
4. Causes ice to float and inhibits complete freezing of large bodies of water. ------------------------------------
5. Enhance a variety of chemical reactions.--------------------------------
6. Allow aquatic animals to survive in ordinary winters.-----------------
7. Allow water striders to skate on the surface water.-------------------------

4.2 Hydrologic Cycle (Water cycle)

The total amount of water in the world remains constant. What changes is its state and availability. Water is constantly being recycled in all its forms, (ice, liquid and vapour), by a system known as the **hydrologic cycle** which is driven by solar energy **(Fig 4.4).** It involves the continuous recycling of water between the atmosphere, land and oceans by several processes. Within the atmosphere vertical and horizontal air movements including winds transfer moisture from place to place. The streams, rivers and glaciers transfer water from land to the ocean where large scale currents transfer water within the oceans.

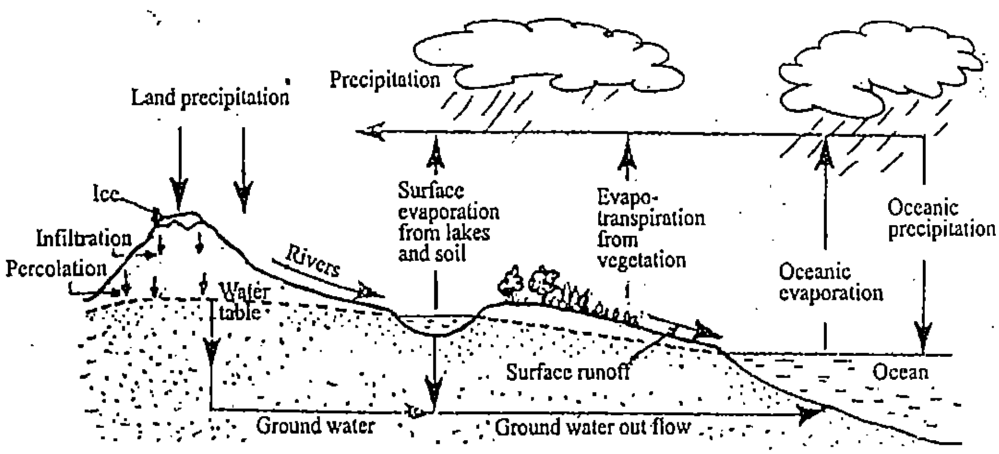


Fig 4.4 The Hydrologic cycle

The three major processes involved in the hydrologic cycle are: evaporation and evapotranspiration, precipitation and surface run off.
Atmospheric water, surface water and ground water are all part of the hydrologic cycle. Let us consider how water re-cycles within and between these realms by the above mentioned processes.

Evaporation and evapo-transpiration

The hydrologic cycle can be crudely visualized as starting with the evaporation of water from the oceans, seas, lakes and rivers. A significant quantity of water that reaches the earth (soil) is lost into the atmosphere by evaporation. Some of it is lost through plants by transpiration from their leaves. During the above process, water is transferred through the soil by capillary action, then from the soil through the roots of plants by osmosis and then to their leaves. Due to the fact that the process of evaporation and transpiration are difficult to separate, evapo-transpiration is generally used to describe the combined process. This water that undergoes evaporation and evapo-transpiration becomes a part of the atmospheric store of water vapour, which condenses and creates clouds.

Precipitation

Precipitation is process by which water returns to land. There are four major types of precipitation, namely drizzle, rain, snow and hail. With the exception of high latitude and high altitude regions, rain tends to be the most important form of precipitation. During winter in some temperate latitudes, however, snow can be more important than rain, and when it melts suddenly it releases large volumes of water. .
The rate at which, groundwater resources (aquifers) is replenished, is basically dependent upon the quantity of precipitation.

Surface run-off

Water falling on land as precipitation will either accumulate on the surface soil and eventually returns to the atmosphere by evaporation, or will infiltrate the surface layer. Subsequently this water percolates to deep levels and reaches the water table to form ground water. These waters are discharged either directly or indirectly to rivers and seas by way of seepage and springs **(Figs. 4.4).**

If the rain fall intensity is much greater than the infiltration capacity of the soil, excess water moves in surface channels as surface runoff into streams, rivers and lakes and later empties into the ocean.

Activity 4.2

The major processes of the hydrologic cycle are given below. Briefly explain the involvement of each process in the hydrological cycle.

Precipitation --

Evapo-transpiration --

Surface run off --

4.3 Composition of hydrosphere

The total volume of water in the global water cycle is estimated at about 1.384 million km^2. Depending on the salt concentration or salinity, this water could be categorized into fresh water, and salt or saline water. [Salinity is a measure of the total concentration of all salts (principally sodium and chloride). The salt concentration is usually given the symbol ‰ (Parts per thousand)]. In the sea water, the salinity varies between, 33‰ - 37‰. The mean salinity of sea water is 35‰. For fresh water the salinity is always less than 0.5‰.

At any one point in time, around 97.6% of the world's water is saline or in other words, salt water. Most of the water is found in the oceans, which clearly play an important role in the global water cycle **(Fig. 4.5).** The remainder of the salt water makes up the salt lakes. This means that only 2.5% of the volume of water in the world is actually fresh water. Some 75% of this fresh water is locked up as polar ice caps and glaciers with a further 24% located underground as groundwater. This means that less than 1% of the total freshwater is found in lakes, rivers and the soil.

Nearly 0.01% of the world water budget is present in lakes and rivers, another 0.01% present as soil moisture which is unavailable to human supply. So while there appears to be lots of water in the world, there is in reality very little, which is readily available for the maintenance of terrestrial life on Earth **(Table. 4.1).**

Storage component	Volume ($km^3 \times 10^3$)	Total percentage of water
Oceans	1 350 400	97.6
Saline lakes and inland seas	105	0.008
Ice caps and glaciers	26000	1.9
Groundwater	7000	0.5
Soil moisture	150	0.01
Lakes	125	0.009
Freshwater rivers	2	0.0001
Atmosphere	13	0.0009
Total	1384000	

All figures are approximate estimates and rounded

Table 4.1 Major natural stores of water within the global hydrologic cycle

Activity 4.3

1. The percentages of the stock of water at different natural storages in the hydrologic cycle are given in **Figure. 4.5.** Identify the storage basin with the help of table 4.1. The largest storage basin is already identified.

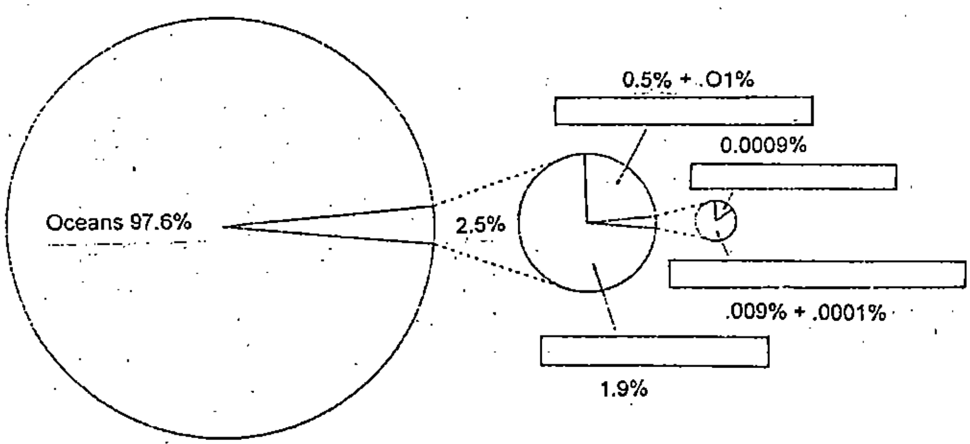


Fig. 4.5 Composition of water on Earth

2. Identify the following lakes in the world map.
   Lake Baikal in Siberia, Lake Tanganyika in Africa, Caspian Sea (largest salt lake)
   Lake superior, Huron, Michigan, Ontario, Erie of U.S.A.

Now you know that there are various types of storages in the world that that store global water. From now on, you will learn about those storages. Firstly, we shall deal with the fresh waters. After that we shall briefly review the estuaries which have brackish water, which is actually a mixture of fresh water and salt water. Finally you will learn about the oceans that contain saline water, which comprise the largest portion of the global water.

4.4 Freshwater Environment

There are a wide range of freshwater habitats. In the most basic classification of water, these habitats can be separated into surface and ground water. We do not intend to discuss ground water in this lesson, but you should remember that it accounts for a significant percentage of the world's fresh water and is fairly well distributed throughout the world. It provides a reasonably constant supply which is not likely to dry up under natural conditions, as surface sources may do, and is often of quite a high quality.

The surface waters are categorized into two types. They are:

- **lentic habitats** or standing water habitats
- **lotic habitats** running water habitats

There are no sharp boundaries between the two categories, or the types of water habitats within one category.

Lentic habitats

Lentic habitats may be natural or artificial. They may be further classified as follows, according to the nature and duration of the existence of the water body.

Natural lentic habitats

Temporary habitats such as pools and puddles, which last only for a few days, especially after rains.

Semi-permanent habitats such as ponds, ditches and villus which last for a longer period, but dry up for a time or remain in a semi-aquatic state.

Permanent habitats, such as lakes and perennial tanks.

Artificial lentic habitats
These include reservoirs, small village tanks which are man made.

Lotic habitats

Lotic habitats may be further classified as vertical and horizontal habitats.

Vertical habitats- These habitats refer to the water flowing vertically to the ground such as in waterfalls.

Horizontal habitats- These habitats refer to water moving horizontally to the ground like slow and fast streams and rivers.

Activity 4.4

Give examples of:

1. A temporary lentic habitat that last only for a few days.-----------------.
2. An artificial lentic habitat.---------------------------
3. A torrential habitat -----------------------------
4. Horizontal lotic habitat with a laminar flow -------------------------
5. First order stream---------------------------

In this lesson we shall consider a natural lake, in order to study the characteristics of lentic habitats. To begin with let us learn how various types of lakes in the world are formed.

Characteristics of Lentic Habitats

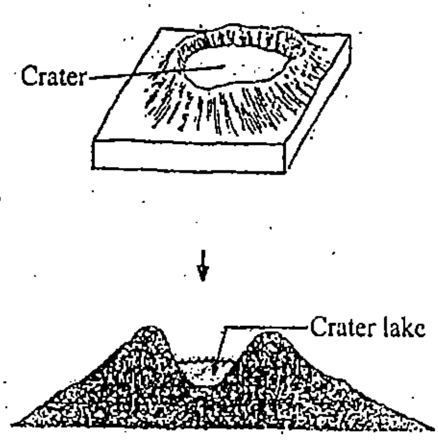
Formation of lakes

Much effort has been made to classify lake formation. In this book we have considered the classification given by Hutchinson in 1975.
According to that classification, there are three general causes that have resulted in originating lakes.
They are the **Rock basins, Barrier basins** and **organic basins**. Each of these is again divided into specific subdivisions, as given in **Table 4.2.**

Types of lakes	Description
Rock basins	These lakes are formed by a depression of the landscape. Five subdivisions of this type of lakes are considered below.
1) Lakes formed by volcanic activity	As a result of volcanic eruption, volcanic material or magma are ejected and create depressions or a cavity forming a lake **(Fig 4.6a).**
2) Lakes formed by glacial activity	During the Pleistocene era, much of the earth was covered by glaciers. When these melted and shrunk, the water that accumulated in the rock basins resulted in forming lakes.
3) Lakes formed by tectonic activity	These are the lakes formed by the movements of deeper portions of the earth crust. These movements may be a due to: • a faulting (break in continuity of strata) E.g. Lake Malawi of Africa. • up lift of a portion of the sea floor. E.g. Caspian sea. • tilting sinking or rising of the earth crust **(Fig. 4.6 b&c)**
4) Lakes formed by solution of the bedrock	Most solution lakes are formed in depressions resulting from the solution of limestone ($CaCo_3$) by slightly acidic water containing CO_2. These basins are usually circular and conically shaped.
5) Lakes formed by meteorite impacts	The impact of meteors on the surface of the earth occasionally creates depressions which later may be filled by water. E.g. chub lake in Quebec.
Barrier basins	These lakes are formed by the imposition of a barrier across a previously open channel. Three subtypes of this type of lake are considered here.
1) Lakes formed due to landslides	Sudden movements of large quantities of unconsolidated material in the form of landslides in to the floors of stream valleys can result in dams and create lakes often of very large size.

2) Lakes formed due to windblown deposits	Soil particles may be carried by the wind blocking the passage of streams giving rise to a lake.
3) Lakes formed by the action of the shore line	Sand pits parallel to the shore line may completely cut off a coastal body of water. It is also possible that material derived from erosion, block the river mouth preventing water moving to the sea, giving rise to a lake.
Organic basins	These are created by the action of living things.
Lakes of organic origin	For example, animals such as beavers construct dams with logs which cut off the pond from the main stream. Man has also entered this process by building dams and submerging areas which were once exposed.

Table 4.2 Types of lakes



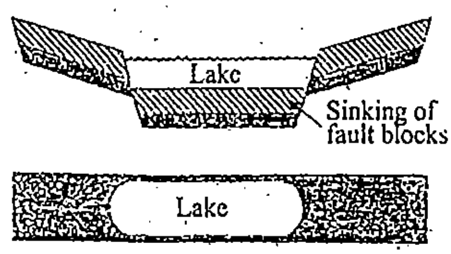


Fig. 4.6 a. Lake occupying a volcanic crater
b. Side view of a lake formed by sinking of faults

If you look at table 4.1 again, you will find that the lakes store more water than rivers. About two thirds of all the fresh surface water on earth is stored in about 250 large lakes which are not distributed evenly over the world. You will be surprised to learn that Lake Baikal in Siberia contains one fifth of the world's freshwater. It is the largest lake in the world.

Activity 4.5

Fill in the column "B" with the name of the lake that is formed as a result of the process given in the column "A"

A-Process	B-Lake
As a result of volcanic eruption a great void is left inside:	
Movements such as unwarping or faulting in the earth crust:	
blocking the river mouth by formation of a dune by windblown sand:	
Formation of a dam by beavers:	

Let us focus on some of the **physical and chemical characteristics** of water, which are common to fresh water and saline water although there are some differences. In this lesson you will study the details of the above characteristics in fresh water, and only those which are most important to brackish and saline water. Firstly we will consider the physical characteristics in lentic habitats and then compare these with that of the lotic habitats.

Physical characteristics of lentic habitats:

- Light and Colour
- Transparency / Turbidity
- Temperature and Heat
- Water movement
- Substratum

Light and Colour

When light strikes the surface of water, a certain amount of light is reflected. The portion that enters the water column is subjected to further reduction by absorption and scattering due to various suspended particles in the water.

Sun light is composed of radiation, which include all the visible colours ranging from very short ultraviolet to very long infrared (wave lengths from about 400 to 700 nm) **(Fig. 4.7)**. As these wave lengths enter clear water, the violet and red light is absorbed within the first few meters.

The green and blue components are absorbed less rapidly, and a fraction of blue light penetrates to more than 100m in transparent water. Hence waters appear blue in colour, as the smallest particles present in the water scatter blue light. In less transparent lakes, dissolved and particulate matter normally obscure back scattering of blue light. These lakes will appear green in colour. If there is a large quantity of dissolved material, especially organic matter the lake appears yellow or brown.

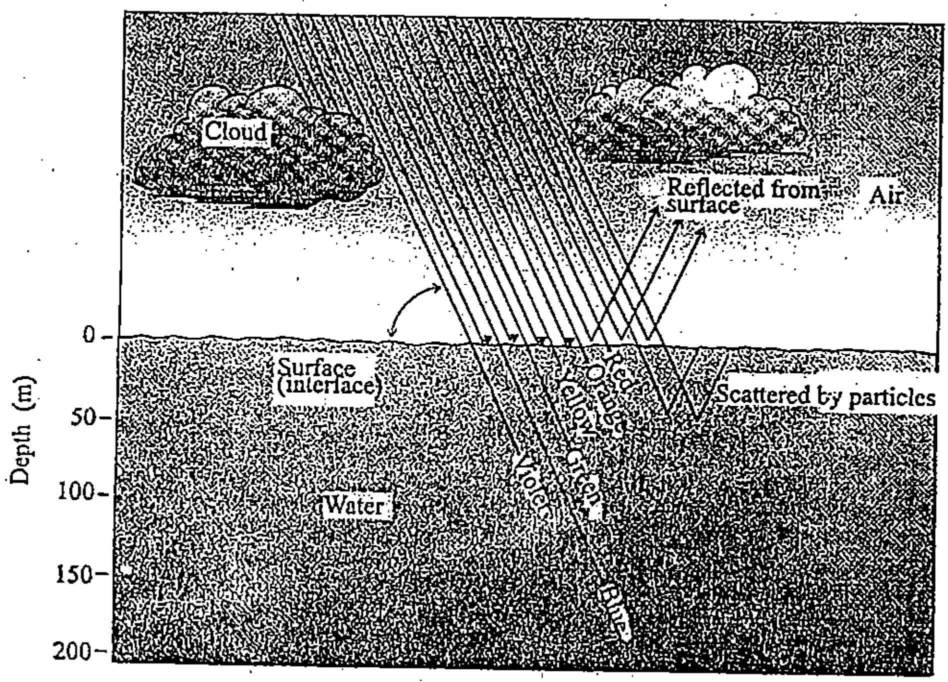


Fig. 4.7 Penetration of different wavelengths through water

From the transmitted light that passes through the water column 53% is absorbed (on cloudless days) and converted to heat, and the temperature of the water body rises. The rest of the light is available for photosynthesis and used by algae.

Lentic habitats contain regions of different light intensity. The well lit surface waters comprise the **euphotic zone**. In this region there is enough light for the net growth of plankton. The lower limit of this zone is the **compensation level**. This is the region where only 1% of the light at the surface remains. At this level the energy harnessed by photosynthesis will only equal to the respiratory requirements of plants. Below this level is called the **profundal zone**, where the plants cannot photosynthesize and grow **(Fig. 4.8)**.

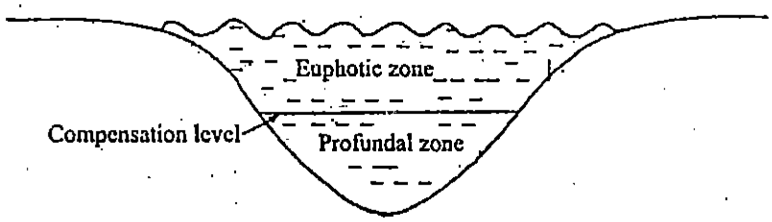


Fig. 4.8 The zonation of a lake according to light penetration

The precise depth of the compensation level varies with seasons and changes in water colour and cloudiness (turbidity). In a clear water lake the euphotic zone may extend down to 20m or more but in tropical lakes it is often about 3-5 m or even as little as 0.5m deep as there are suspending particles.

Transparency / Turbidity

Transparency is a measure of the depth to which one may see into the water. Obviously this is variable with the day condition, season and the eye sight of the observer.

Turbidity results from suspended material (organic matter, mud particles, plankton etc.) and could be measured by the turbid meters, or by a filtrator. If the particles that produce turbidity are imported to the water body, then they are called **allochthonous material.** If they are produced within the system itself, it is called **autochthonous material** (dead parts of animals and plants).

Temperature and Heat

The temperature of a water body, changes horizontally (with time) and vertically (with depth). We can recognise two patterns of temperature variations:
1. Daily temperature
2. Seasonal temperature variation

1. Daily temperature variation

Aerial temperature has a significant correlation with the surface water temperature. When aerial temperature rises surface water temperature also rises. However temperature in a water body alters much rapidly than that of air. As a result there is no abrupt difference in temperature by day and night as on land.

In **tropical** countries daily variation of temperature in surface water layers is very prominent. It fluctuates between 1 °C - 6°C and the highest temperature was recorded between 1-3pm. However, bottom strata of a deep water body does not show significant changes and remain cooler **(Fig. 4.9).**

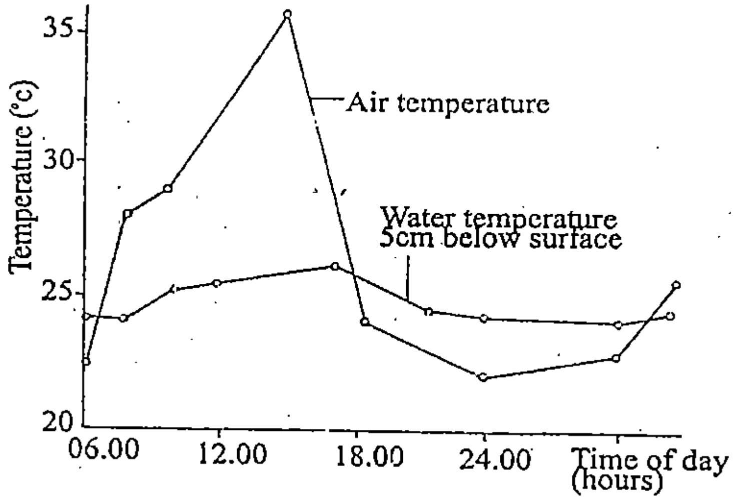


Fig. 4.9 Daily temperature change in air and water in a small tropical lake.

The water temperatures in **temperate** water bodies closely follow the air temperature as in tropical countries but this variation is not prominent. However in winter water temperature does not go below 0 °C and in spring air warms more rapidly then water.

Temperature variation with depth

As you know, a water body gets heated due to sunlight and this heat penetrates to deeper layers by conduction and mixing. Conduction is a very slow process and if there is no wind (calm day) water does not mix and it will take a long time for heat to penetrate down. In shallow waters, sediments can absorb significant quantities of solar radiation.

Globally, a situation is developed where there is a marked difference in temperature between the waters of upper layers and deep layers. This type of thermal discontinuity is known as the **thermal stratification (Fig. 4.11).** In such a situation there is also a density difference along the water column between the upper and deep layers. The warm upper layers of a thermally stratified lake are called the **epilimnion.** The cool deeper layer of a lake that is not heated by the sun and is too deep to be circulated directly by the wind is termed the **hypolimnion.** The transitional zone between the epilimnion and the hypolimnion is termed as the **metalimnion.** In this there is a region where a sharp temperature gradient is present. This region is called the **thermocline (Fig. 4.10 a&b).** The direction of the current may be different in the epilimnion and the hypolimnion, as the density of water in these two regions differs very much.

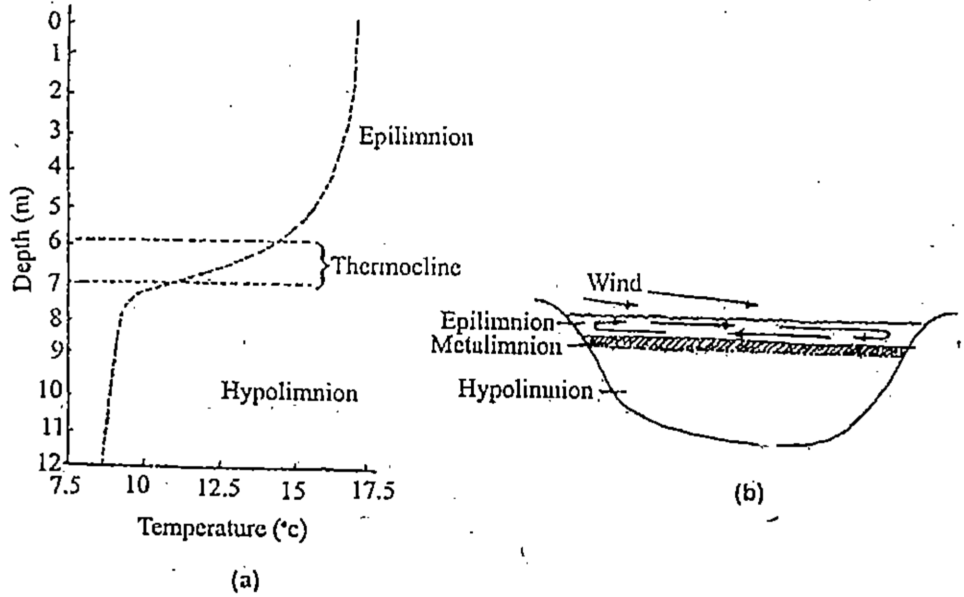


Fig. 4.10 a. Temperature variation with depth during summer in a temperate lake.
b. Different zones of a stratified lake

Thermal stratification of lakes

Thermal stratification is the most important physical event in a lake's annual cycle and dominates most aspects of the lake's structure. In shallow ponds and tanks the typical thermal stratification may not be seen. Although there is some form of temperature discontinuity. In such water bodies temporary thermocline may be present between two to three meters during the day time which disappears in the night.
In tropical deep water bodies, thermal stratification is present throughout the year and in temperate water bodies thermal stratification is seen only during summer.

Thermal classification of lakes

Lakes around the world can be classified according to the types of **thermal stratification patterns, mixing and the formation of the hypolimnion.** Several types of such lakes are given in **Table 4.3.**

Table 4.3

Type of lake	Description
Holomictic lakes.	In these lakes, circulation occurs throughout the entire water column,
Meromictic lakes	These lakes do not undergo complete circulation, and the primary water mass does not mix with the lower portion The deeper stratum of these lakes are separated from the upper stratum by a steep salinity gradient called the chemocline
Dimictic lakes	These lakes mix twice a year, once in the autumn and once in the spring. They are covered with ice in winter and may show inverse stratification.
Monomictic lakes	These lakes do not freeze as the temperature does not fall below 4°C. They are stratified in summer.
Polymictic lakes	These lakes are shallow. They mix every few days or even daily all year round.
Amictic lakes	These lakes have year round ice cover and never mix and limited to Antarctica.

2. Seasonal temperature variation

The seasonal temperature variation of a shallow body of water in a tropical country is not prominent. However, the water bodies of temperate countries show clear seasonal variations about which we will discuss in the following paragraphs.

The beginning of the annual stratification cycle in temperate lakes is considered to be in mid winter. As the temperature continues to drop, water at the top of the lake freezes and seals the water body. The presence of a layer of ice on the surface can result in the development of an inverse stratification between 0°C water at the surface and 4°C water at the hypolimnion (**Fig. 4.11**).

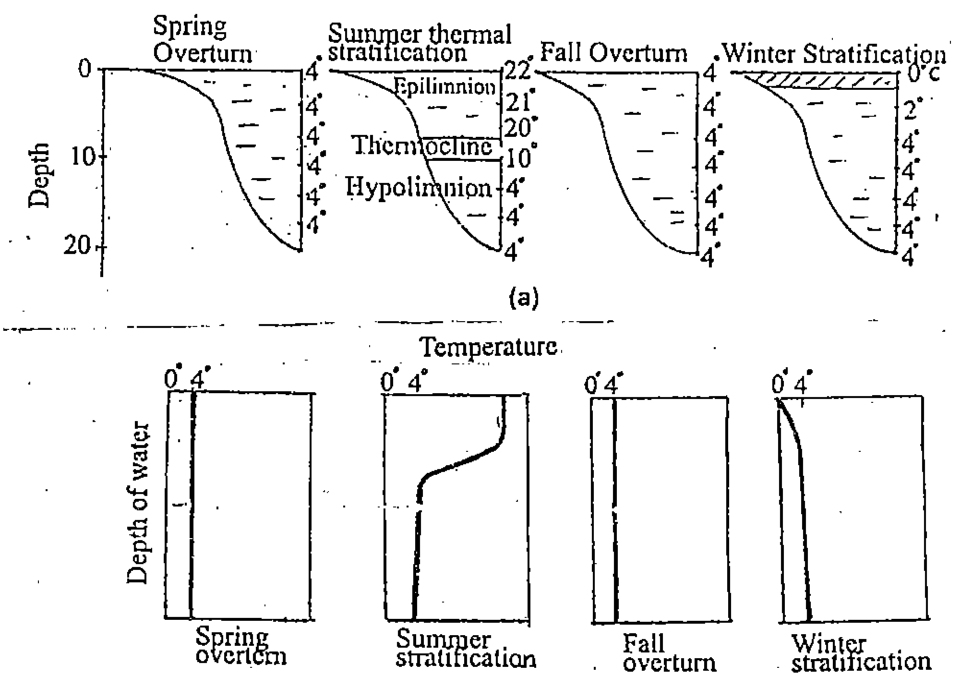


Fig. 4.11 a. Temperature variation during the four seasons in a temperate lake.
b. Graphs showing the variation of temperature with depth

As spring approaches, the upper layers receive more heat than can be returned, resulting in an increase of the surface temperature. When the surface temperature reaches 4°C and become same as that of the hypolimnion the thermal and density distribution becomes more uniform. At this stage the resistance to vertical mixing is at its lowest level and even a small wind can achieve partial or complete vertical mixing and circulation. This type of wide circulation is called the **overturn.**

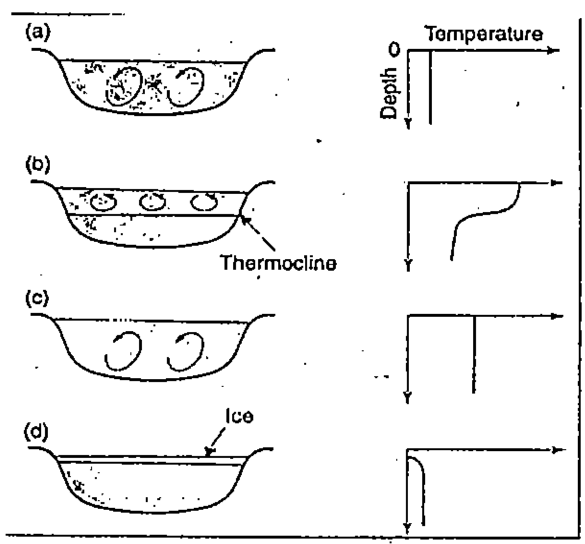
During late spring and early summer, the temperature of the surface waters increase, while colder heavier water remains near the hypolimnion. This leads to direct thermal stratification.

In late summer, and during autumn (fall) the surface temperature begins to decrease. Due to continuous heat loss, during late fall and early winter, the surface temperature drops again and reaches a point where it is essentially the same as the hypolimnion. This results in another overturn and the cycle begins again.

In this way, the complete annual cycle consists of two stratification periods (in winter and in summer) and two turnovers (in spring and in fall). You have to remember that this does not occur in tropical countries.

Activity 4.6

The diagram given below shows the stratification cycle of a temperate lake, showing zones of mixing and temperature profile with depth during different times of the year. Answer the questions given below, which are based on the diagrams a to d.



1. Which one of the diagrams show an inverse stratification?
______ ______________

2. Complete circulation is shown in ______________.
3. The situation in Autumn (fall) is shown in ______________
4. Give one difference between the temperature variation between spring and autumn
______________________.
5. According to thermal stratification of lakes to which category does this lake belong?
______________________.

Water movement

The flow of the water can be laminar or turbulent. depending on the bottom topography of the water body.

Laminar flow is the smooth slipping of water particles, past each other or an obstruction and has little drag on moving objects.

In contrast, **turbulent current** is the random, chaotic tumbling of the water particles around each other, or any object passing through the water. These tumbling motions are described as **eddies**. Turbulent currents occur in mountain streams.

There is no strong continual unidirectional current in lentic habitats as in lotic habitats. Wind is the primary force moving the water at all depths of a water body. Temperature, density differences and gravity are also important factors that cause water movements.

Substratum

Substrate is a complex aspect of the physical environment. In lentic habitats the stillness of the waters results in a heterogeneous stable bottom. The substratum is widely composed of sand, clay, mud and detritus that will tend to increase the nutrient content, thus supporting aquatic plant and animal life.

At the bottom region, due to the action of aerobic bacteria on detritus, low oxygen concentrations will be present.

Due to anaerobic conditions in the bottom anaerobic bacteria come into function thus forming gases such as hydrogen sulfide, ammonia and carbon dioxide.

Chemical characteristics

Now that you have studied the physical characteristics of fresh water bodies, let us discuss the chemical characteristics such as:

- Dissolved Oxygen
- CO_2 concentration, pH, Alkalinity and Acidity
- Water Conductivity
- Hardness
- Dissolved solids and nutrient status

Dissolved oxygen

The amount of dissolved oxygen in water is one of the most important factors of lakes and other water bodies aside from water itself. The main source of dissolved oxygen, are from the atmosphere and the photosynthetic processes of the green plants. The turbulence and mixing of water, increase in pressure and cold temperatures increase, the amount of oxygen dissolving in water.

Oxygen levels often show marked daily cycles as plants release oxygen during daylight, sometimes supersaturating (i.e. greater than 100 per cent saturation) the euphotic zone, then respire at night causing a deficit.

Oxygen is transferred to deeper waters by either diffusion or circulation of surface waters. If circulation in a water body is less, and diffusion is the only means to get oxygen to deep levels, anoxic conditions may prevail at the bottom layers of lakes. In addition, bacteria present at the bottom layers, involved in the decay of organic matter use oxygen for respiration resulting anoxic conditions.

In very clear waters, where there is low productivity (oligotrophic lakes), oxygen distribution will largely be a function of temperature, resulting in a fairly uniform orthograde distribution. Initially there is a slight increase in the dissolved oxygen with depth and after that the curve is characterized by no appreciable decrease or increase **(Fig. 4.12a)**.
There is another type of oxygen distribution which is called clinograde distribution displayed in eutrophic lakes **(Fig. 4.12b)**. In this type, the temperature and oxygen at the surface layers is higher and both decrease with depth.

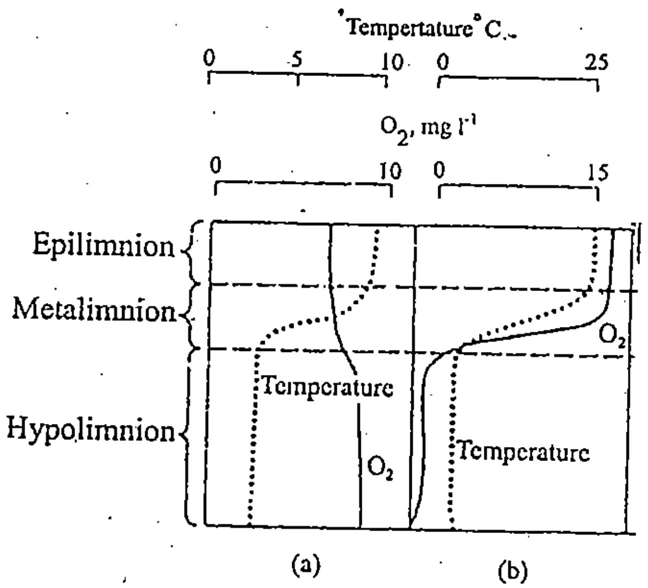


Fig. 4.12. Patterns of the distribution of oxygen
a. Orthograde distribution
b. Clinograde distribution

CO_2 concentration, pH, Alkalinity and Acidity

The pH of most natural waters falls in the range of 4.0-9.0. Many being in the range of 6.0-8.0. The majority of natural waters have a somewhat alkaline pH, because of the presence of carbonate and bicarbonate ions.
In stratified lakes pH may drop with depth, and the bottom, pH remains very low. This condition is common in deep water bodies in tropics. In temperate water bodies this condition is experienced in summer, but during the overturn the pH is constant from surface to bottom.
pH in the epilimnetic region of lakes fluctuates daily due to the fluctuations of CO_2 concentration, resulting from the dissolved atmospheric CO_2, photosynthesis and respiration by plants. In the hypolimnion, CO_2 is formed due to the decomposition of organic matter.

Surface waters become acidic towards the later part of the night and early morning due to the accumulation of CO_2, resulting from respiration coupled with inhibition of photosynthesis.

Carbon dioxide combines chemically with water to form undissociated carbonic acid (H_2CO_3), which dissociates partly to produce hydrogen and bicarbonate ions. The latter decompose further forming more hydrogen and carbonate ions.

$$CO_2 + H_2O \leftrightarrows H_2CO_3 \leftrightarrows H^+ + HCO_3^- \leftrightarrows H^+ + CO_3^{--}$$

All these reactions are reversible and the whole system reaches equilibrium, so that natural waters will contain various proportions of CO_2, HCO_3^- and CO_3^{--}. As shown in the figure 4.20, pH value inversely correlates with the dissolved CO_2 and directly with HCO_3^- concentration, and also there is no free CO_2 in water above pH 8. However the pH shift is reduced or buffered by the higher concentrations of carbonates and bicarbonate in water.

In acidic waters where pH is low the above reaction will move backwards, while under alkaline conditions it will move forward and HCO_3^- and CO_3^{--} become more common **(Fig. 4.13).** Acidity, the ability of water to neutralize alkalinity is normally determined by pH, a measure of the concentration of hydrogen ions in solution (H^+). Water is said to be alkaline when the concentration of the hydroxyl ions exceed the hydrogen ions.

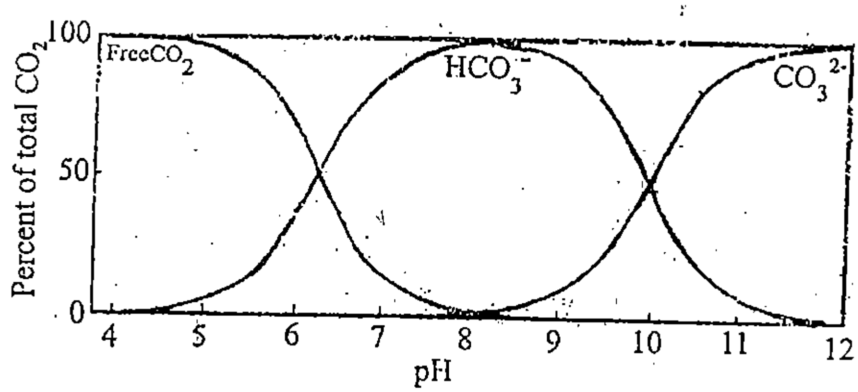


Fig. 4.13 Variation of pH with different forms of carbon dioxide in water

Water Conductivity

Conductivity is the measure of the total quantity of charged particles (ions) dissolved in the water. In freshwaters it is measured in micro-Siemens per centimeter ($\mu S/cm^{-1}$ at 25°C). Normally in lentic habitats, conductivity increases vertically.

Hardness

Hardness in water is mainly due to the presence of ions of the metals calcium, magnesium and iron. Water with bicarbonate salts of calcium, magnesium and iron [e.g. ($Ca(HCO_3)_2$)] when boiled above 70°C, carbonate salts of the metals are precipitated in kettles. Such water is known to exhibit temporary or carbonate hardness, because the carbonate salts (e.g. calcium carbonate) are largely insoluble and on heating is removed from the water and deposited as scales.

$$Ca(HCO_3)_2 \xrightarrow{\text{heat}} H_2O + CO_2 + CaCO_3$$

When calcium, magnesium and iron, are present as chloride or sulphate salts (e.g. $CaCl_2$) the hardness is called **permanent or non carbonate hardness**. Although this type of hardness also contributes to scaling, in this case the precipitate is due to the decreased solubility of these metal salts at higher temperatures and not due to the formation of new insoluble compounds.
Both temporary and permanent hardness make lathering with ordinary soap difficult. The result is the formation of scum that, float on the surface of washing water.

Dissolved solids and nutrient status

Dissolved solids include mainly nitrogen salts, phosphorus salts, magnesium and calcium salts and trace elements such as manganese, copper, zinc cobalt etc.
Nitrogen and phosphorus are very important for the growth of aquatic plants and phytoplankton.
The rate of supply of nitrogen salts into a water body is intimately connected with land use practice of the watershed. Phosphorus enters a lake via rainfall, from upstream lakes, sewage etc. Phosphorus supplies to the aquatic plants are enhanced by internal recycling within the lake itself, both from the sediments and through animal excreta. Phosphorus is lost from the ecosystem much more than nitrogen and carbon, because it reacts with mud and chemicals in water in ways that make it unavailable to plants.
The terms **oligotrophic** and **eutrophic** are widely used to describe lakes with respect to their nutrient conditions. An oligotrophic lake is one with low nutrient levels. Phytoplankton production in these lakes is low and therefore water clarity is high. Decomposition in the hypolimnion uses up oxygen more slowly than it can be replaced by mixing from the surface, so the water remains well oxygenated. An eutrophic lake has high nutrient

levels and therefore has high rates of phytoplankton production. High densities of algal cells reduce water clarity. Decomposition of detritus in the hypolimnion uses up oxygen more rapidly than it can be replaced. This may lead to anoxic conditions at the bottom.

Activity 4.7

Fill in the blanks.

1. A lake where water mixes continuously all year around is called --------------------
2. The materials that are brought into a lake from outside are referred to as ------------ --------------------
3. The upper layer of a stratified water body above the thermocline is the---------------- -
4. The trophic condition of a water body with high nutrient content is called.-------------- ----------------------
5. The region of a lake where a rapid change of temperature takes place is referred to as------------------------------.
6. The depth of a water body at which the processes of photosynthesis and respiration are equal is called ----------------------------------
7. Typical Oxygen distribution curve that could be expected in a productive lake is termed as. --
8. The visible colour of the spectrum, that penetrates deep into the water is,------------ --------------------------------
9. Scales form at the bottom of kettles when boiling----------------------------water.

The biography of lentic habitats.

Once formed, lentic waters develop and change greatly with time. Newly formed lakes are generally barren of life **(Fig.4.14a)**. Few plants and animals will have established themselves, sediment may be limited and nutrients scarce. Since they are depositing habitats allochthonous input from catchment area, together with autochthonous input tend to fill them with time **(Fig.4.14b)**. Filling with organic matter, changes the lentic habitat into a different habitat. Macrophytes found in wetlands encroach these habitats replacing the open water plants **(Fig.4.14a)**. The ultimate fate of these systems is the conversion to a fully terrestrial flora. This results in the final disappearance of most ponds and lakes **(Fig. 4.14d)**.

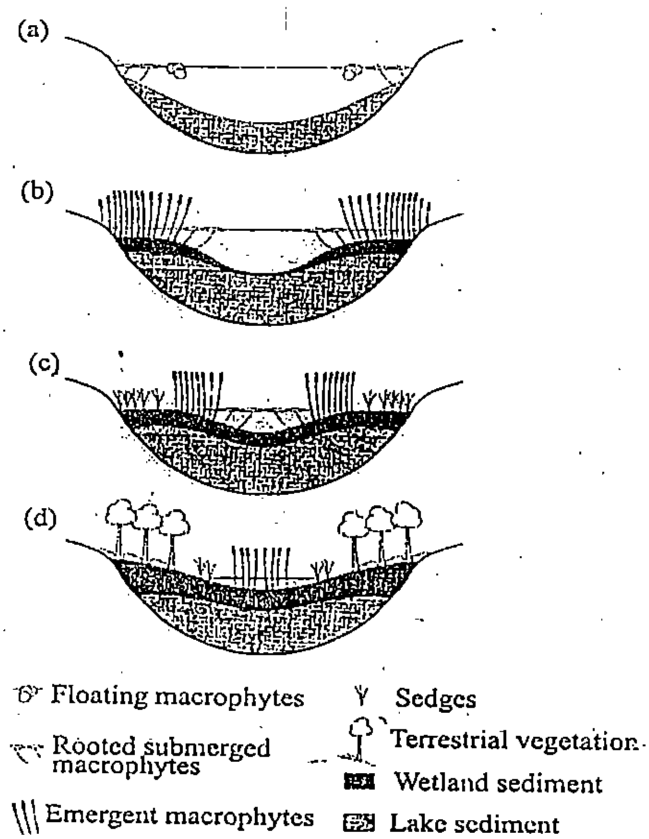
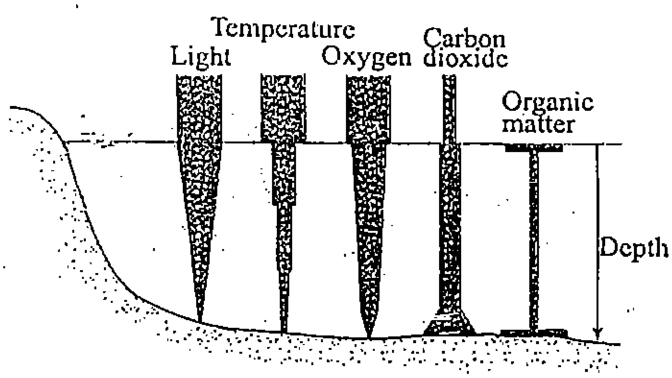


Fig. 4.14 Stages of development of terrestrial vegetation in a lake

Activity 4.8

Vertical distribution of some of the chemical and physical characteristics in a lake is shown in the diagram.
Answer the questions given below, which are based on the diagram.



ii. Explain the changes in the vertical distribution of temperature.
iii. Explain why Oxygen concentration at the bottom is lower than that of carbon dioxide?
iv. Explain the processes that influence absorption of oxygen at the epilimnion.
v. State whether this lake is stratified or not. Give reasons for your answer.

Up to now you have being studying various characteristics of lentic water habitats. Now let us study the nature of lotic water habitats, and their characteristics.

Characteristics of lotic habitats

Lotic habitats are best described by flow, erosion, deposition and channel form. They vary form raging torrents and waterfalls to rivers, whose flow is so smooth as to be almost unnoticeable. The characteristics of any river vary greatly along its length, as it changes from a tiny trickle present at the head waters, to its full size at the lower reaches **(Fig. 4.15)**.

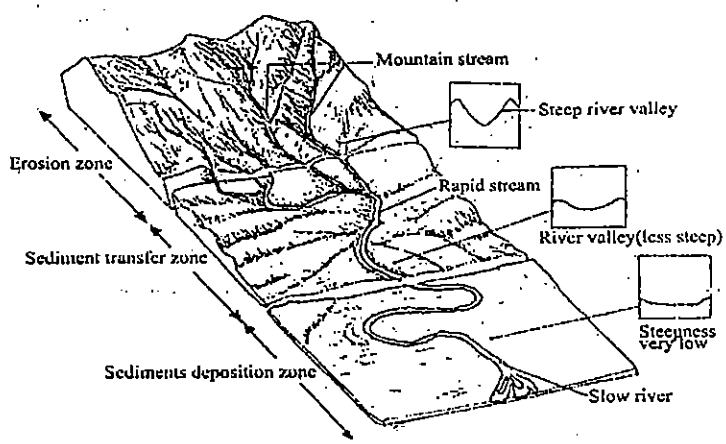
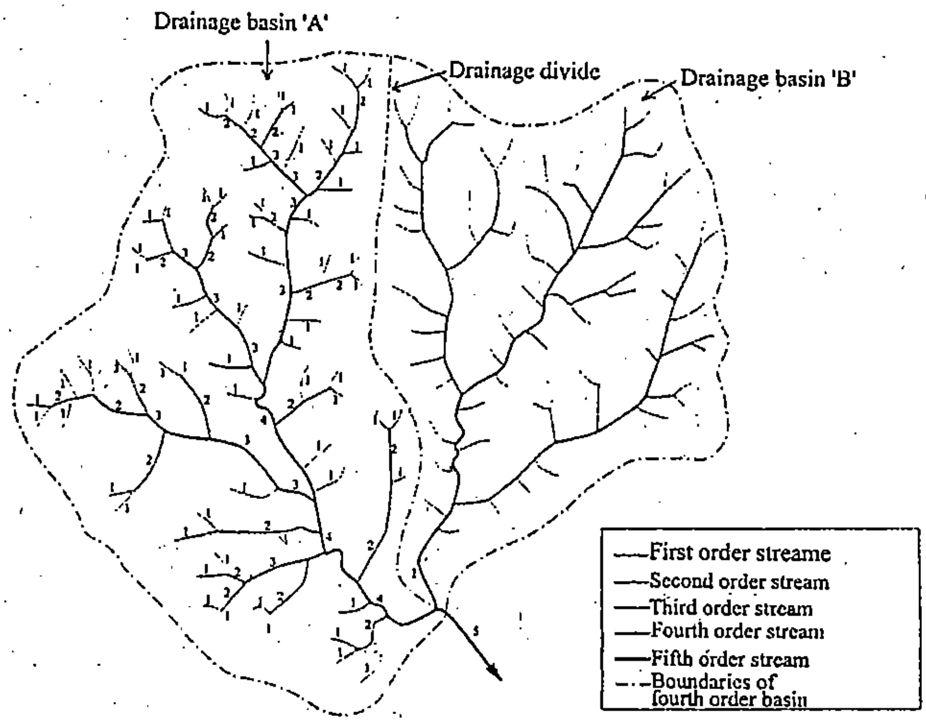


Fig. 4.15 Pathway of a river along its length

River classification
Many attempts have been made to classify types of rivers. The geo-morphological description that is widely used is the **stream order,** developed by Robert Horton in 1945.

Stream Order

Stream ordering is a method of assigning a number, to each river stretch. This number is an indication of its relative size within a drainage basin. The order is based on the hierarchy of how channels link up. **First order** streams have no tributaries, they are the first streams (e.g. trickles, brooks etc.) fed by springs, ground water and runoff. When two first-order streams join, the confluent stream is **second order**. Joining of two second-order streams make a **third order.** Notice that it needs two stream segments of equal order to join to produce a segment of a higher order. The order remains unchanged if a higher order stream joins a lower order stream (e.g. a third into a first, or a second) **(Fig. 4.16)**.



The ordering hierarchy has allowed comparisons within and between individual rivers. Also this network structure determines how flood waters collect within the channels, which in turn determines the magnitude and frequency of river flooding.

Activity 4.9

Refer Figure 4.16 and identify the order of streams in the drainage basin "B" and mark them.

Zonation along the river channel

Now let us consider various regions along a river, in order to study the characteristics of a lotic system. To make our study convenient, we will consider an idealized river system, which can be divided into three zones **(Fig 4.15)**.

- Erosion zone
- Sediment transfer zone
- Sediment deposition zone

Erosion zone

This zone comprises of drainage basins and headwater streams. The most important component governing the physical, chemical and biological structure of this zone is the flow rate. Rivers with current velocities >0.3 mS^{-1}, are classified as erosive, while rivers with velocities < 0.3 $mS^{-1}$are classified as depositing.

The environments from which water drains into a lake or river is the drainage basin or the catchment. The term 'watershed' may also be used in Europe though this tends to mean the dividing line between neighbouring catchments rather than the catchment's drainage area itself. Drainage basins are open systems through which water and sediments are transported **(Fig. 4.16).** The head water streams, found in this zone are low order streams (mainly first order).

Natural river channels are never homogenous in form, but contain a variety of structural features. Among these, torrential habitats such as water falls, cascades, and alternating patterns of riffles and pools may be found. A major feature of the riffle is that it is wider and shallower and water flows through or over stones or gravel beds **(Fig. 4.17a &b).** Stones with sunlit riffles are often covered with algae and mosses. In contrast, the pools are narrow deep sections, which usually cover several times the area of the riffles, contain different less dense biota living among a mixture of stones and fine grain sediments. A characteristic feature of pools is the accumulation of decaying terrestrial debris.

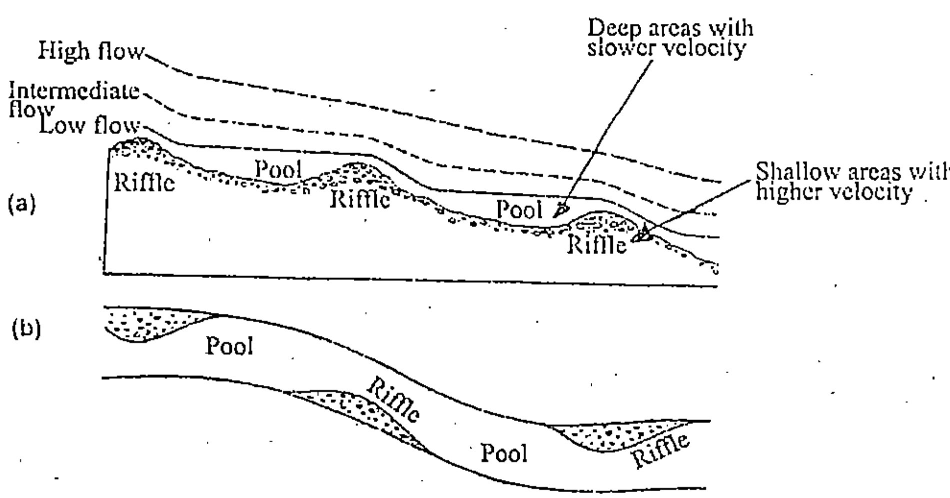


Fig. 4.17 riffle pool sequence
a. longitudinal profile
b. plan view

The streams in the erosion zone are small and often have steep gradients. Since most of them are shaded by trees, which considerably reduce solar heating the water tends to be cool. The most distinctive feature of this zone is the **shear force exerted by the fast flowing water,** which has a great erosive power. Due to this eroding nature of water, the sediments, detritus and mud do not accumulate in a flowing stream, except in sheltered areas. Instead the substratum is often composed of cobbles and boulders which are rounded and smoothed by water. Occasionally the stream may be eroded to the bedrock.

The shear force and rate of water movement tend to be quite different on top of a rock fragment, between rock fragment and beneath rock fragments. Different species can exploit these differences in microhabitats. For instance the regions near rocks where no movement of water is present are called **dead zones**. These are important micro habitats for benthic organisms.

Due to turbulence, and the aeration caused by splashing of water on rocks, the amount of oxygen incorporated into water is very high. Also it prevents thermal stratification.

Plankton if present, are found in the sheltered areas. There are fewer microorganisms accompanying detritus. Aquatic plants usually do not grow in the erosion zone of a stream, except few species which are often confined to the banks, as everything that is not attached or weighted is swept away by the current, including organisms and sediment particles.

Sediment transfer zone

This zone is the region in which the river gradient is reduced so that water and sediment are transported without net loss or gain. The rivers are much wider. Flood plain may be present. A flood plain, is a flat valley bottom of loose sediments which are deposited during river flooding **(Fig. 4.18a).**

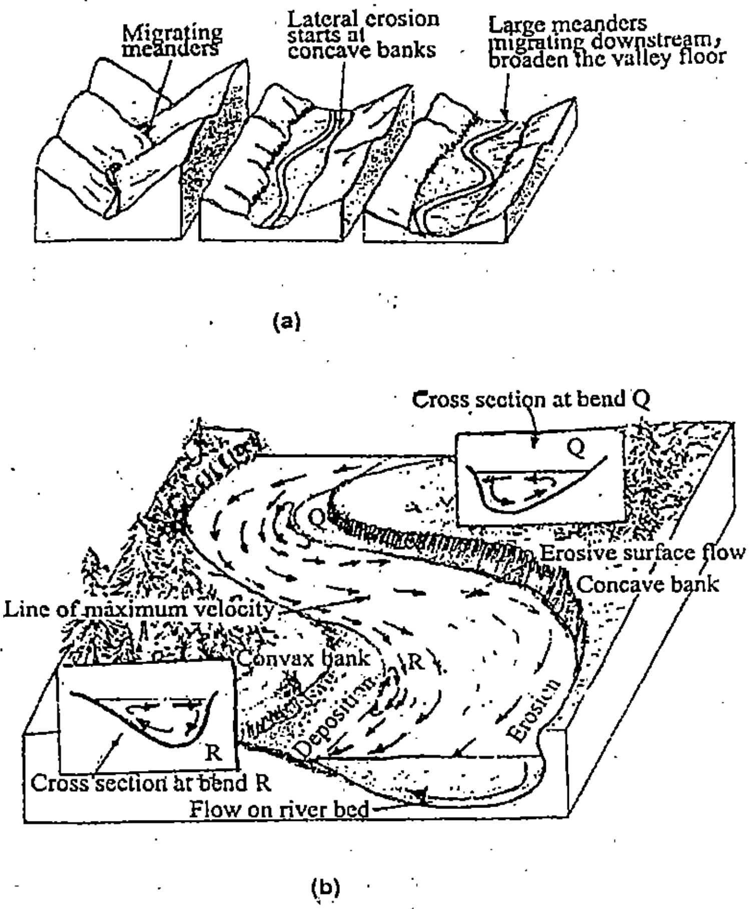


Fig. 4.18 a. Formation of flood plains in a meandering stream.
b. Erosion and depositing areas of a meander

Deep rivers generally have a small gradient and the shear stress on the bed is low. This condition produces muddy debris laden sediments and riffles are absent. The average current, as well as discharge becomes greater as the overall size increases so that river waters move faster than streams, despite their lesser slope.

Although stratification is unusual in rivers, given their turbulence, movement and relative shallowness thermal conditions do vary. Seasonal and diurnal changes occur, and thermal regimes vary along the length of a river with topography and volume.

Sediment deposition zone

The deposition zone is where the river deposits its sediments load, typically as it approaches the sea and develops a delta or an estuary. The flow tends to be laminar. The substrate is dominated by fine silt. This part of the river is usually found on flat land and has a low gradient. Flood plains are generally present, especially in rivers which flood regularly.

When the river flows further down it normally follows a sinuous course. A meander is a winding sinuous course. The gentle flow of the river normally erodes the valley from side to side nibbling away its banks, so flattening and boarding the valley floor. Highest turbulence occurs on the concave side of the stream **(Fig. 4.18b)** thus eroding into the outer banks of bends. Deposition starts at the slower and less turbulent waters on the concave side. This side becomes partially sedimented.

The debris moved by the river, its load plus suspended algae make the water in this region of the river turbid. Other nutrients, dissolved ions and pH influences rivers and streams as much as they do in lakes and ponds. Since various ions are accumulated along the path of the river the conductivity shows higher values than the other regions. This zone of the river is subjected to more pollution than upstream.

Activity 4.10

Fill in the table by giving the nature of the physical and chemical characteristics in each of the habitats given below.

Habitat	Type of current	Nature of Substrate	Temperature	Oxygen at the bottom	Turbidity
Water fall					
First order mountain stream					
Flood plain					
Riffle					
Concave side of a meandering river					

4.5 Estuaries

Estuaries are found along the coastlines of the world, but most are evident in wetter climates of temperate and tropical latitudes. In such areas, land drainage (river water) provides the necessary freshwater input at the head of the estuary, which mixes with the salt water that enters to the estuary from its mouth, to form brackish water. The salinity of brackish water is below those of adjacent open ocean waters and range between 0.5‰ to 35‰.

Types of Estuaries

Estuaries are semi enclosed coastal embayment where freshwater rivers meet the sea. Here fresh water and sea water mix, creating a unique and complex ecosystem where there are various types of habitats namely, **mangrove habitat, sea grass beds and salt marshes.**

This part of the lesson outlines the characteristics of estuaries without studying the habitats mentioned above. However, when you study the biotic component of the aquatic systems in **unit 3, of Block II**, you will come to know most of the characteristics of these habitats.

Depending on the geomorphology of an estuary, the geological history of the area, and the prevailing climatic conditions, there may be different estuarine types, each displaying somewhat different physical and chemical conditions. These may be grouped into few basic types as follows (**Fig. 4.19**).

- **Coastal plain estuary**
- **Tectonic estuary**
- **Bar built estuary or lagoon**
- **fjord**

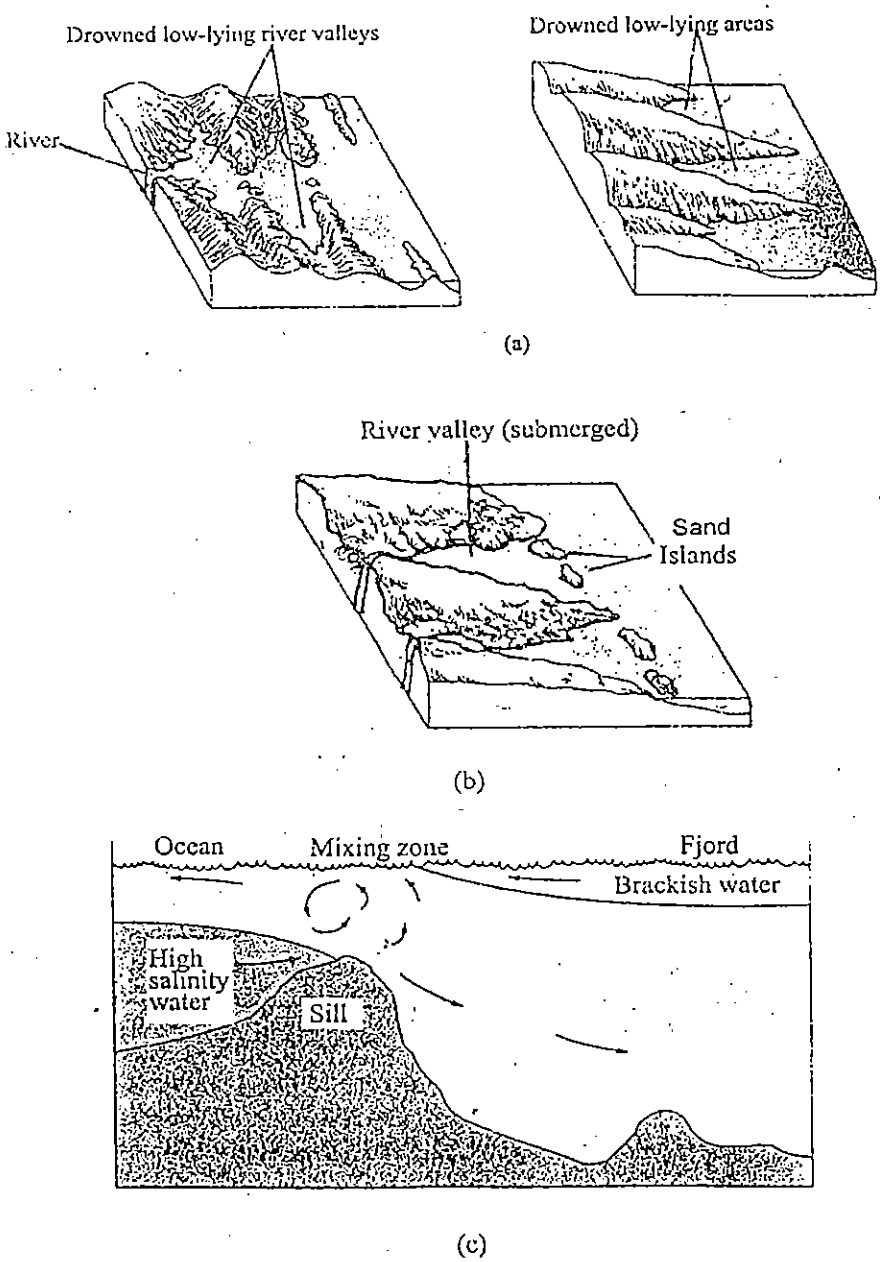


Fig. 4.29 Types of estuaries

a. Coastal plain estuary
b. Bar built estuary
c. fjord

Coastal plain estuary

At the end of the last ice age, the earth became warmer and the ice melted and water was released. As a result of this water, sea level rose as much as 400m relative to the level of the land. The rising sea level invaded low lying coastal river valleys thus forming coastal plain estuaries which are the most common type **(Fig. 4.19a).**

Tectonic estuary

In this type of estuary, the sea reinvades the land due to subsidence (large earth movements may cause a depression in the earth crust), of the land.

Bar built estuary or lagoon

The off-shore barriers such as sand pits form parallel to the coastline and partially cut off the waters behind them from the sea forming bar built estuaries. Some of these estuaries merge with coastal water bodies and form lagoons behind the sand bars which collect the freshwater discharge from the land. The water in such lagoons varies in salinity depending on the climatic conditions **(Fig. 4.19b).**

fjord

These are valleys that have been deepened by glacial action and are then invaded by the sea. They are characterized by a shallow sill at the mouth that greatly restricts water interchange between the deeper waters of the fjord and the sea. Often, these deeper waters are stagnant because of lack of circulation **(Fig. 4.19c).**

Classification of estuaries based on salinity

Estuaries may be classified in another way, depending on the basis of the relative distribution of salt water and fresh water and the degree to which they mix. In most estuaries there is a gradient in salinity from being fully saline (33‰ -37 ‰ –parts per thousand) at the mouth to fresh water (0.5‰ - 3‰) at the landward end. The position of the gradient also moves up and down the estuary with the tidal cycle.

Three fundamental types of estuaries can be identified based on salinity.

- **Positive estuary or salt wedge estuary**
- **Negative estuary**
- **Neutral estuary**

Positive estuary or salt wedge estuary

In these estuaries, salt water enters from the bottom. A substantial amount of out going fresh water tends to float on the seawater because of its low density. There is a salinity gradient from surface to bottom as well as from head to mouth. Mixing occurs where the two waters come in contact, but this does not take place completely. In such a situation a cross section of the estuary shows isohalines (lines of equal salinity), which extends upstream at the bottom **(Fig. 4.20a).** Along any vertical line in the estuary, salinity will be highest at the bottom and lowest on the surface. This type of estuaries is found in temperate countries where there is little evaporation from the surface waters.

When positive estuaries show salt wedge, they are called salt wedge estuaries. These estuaries develop where, a river discharges into a virtually tide less sea.

The position of the salt wedge is dependent on the freshwater flow. Under low flow conditions, the volume of fresh water input is small and salt water can penetrate much further inland than under high flows **(Fig. 4.20b)**. When there is very little mixing between the two water masses, the salt wedge will be very prominent.

Negative estuary

In tropical climate, where the amount of fresh water input to the estuary is small and the rate of evaporation is high, a negative estuary results. In this estuary, the incoming salt water enters at the surface and is somewhat diluted by partial mixing with the small amount of fresh water. The high evaporation rate however, causes this surface water to become hyper saline. Hyper saline water is denser than seawater, sinks to the bottom, and moves out of the estuary as a bottom current. A salinity profile of such an estuary is the reverse of the positive estuary, with highest values at the bottom and lowest at the top.

The isohalines in this estuary extends in the direction of the sea **(Fig. 4.20c)**

Neutral estuary

In these estuaries the river flows are small but tides & tidal currents are large. The water may be mixed almost completely from top to bottom. Salinity remains constant from surface to the bottom of the estuary. Rarely, fresh-water inflow equals evaporation from the surface to the estuary, which also makes a static salinity pattern **(Fig. 4.20d)**

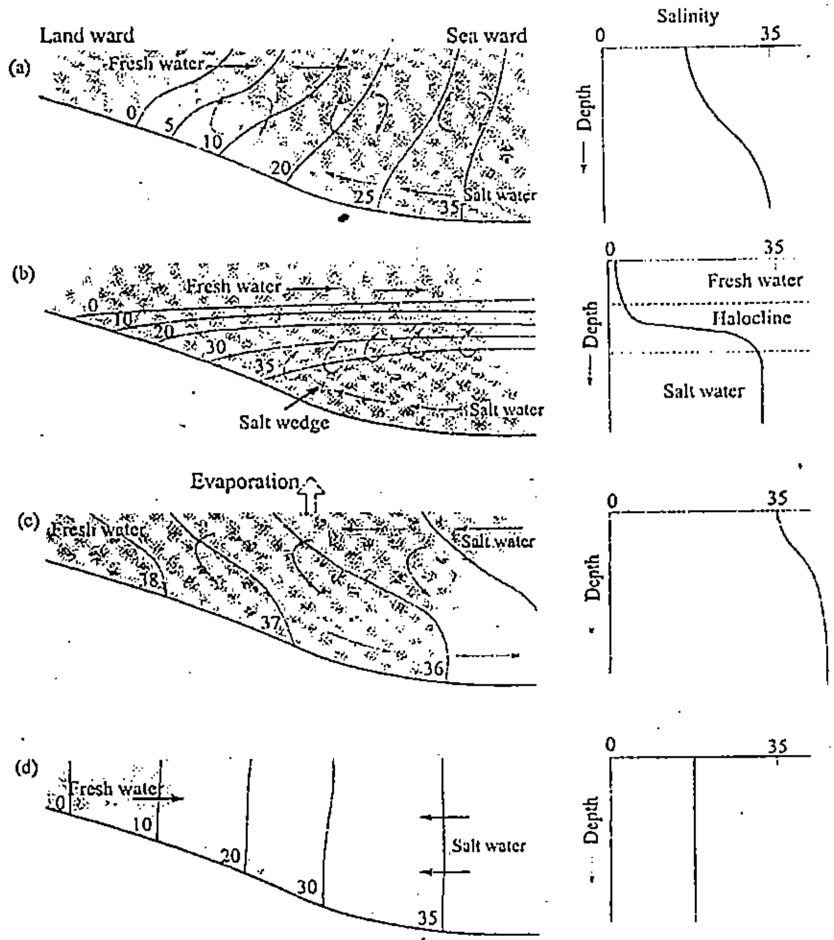


Fig. 4.20. Classification of estuaries based on salinity

a. Positive estuary, b. Salt wedge estuary
c. Negative estuary d. Neutral estuary

Activity 4.10

1. Among the four basic types of estuaries, what is the most common type?
   ____

2. Name the estuary, which is formed due to subsidence of land.__________________

3. Name the estuary that is formed due to the action of glaciers.__________________

4. Name the estuaries, (categorized according to the salinity gradient) that have isohalines of the following types:
   i. extend up stream at the bottom.____________________
   ii. extend down stream at the bottom.____________________
   iii. straight from top to bottom.____________________

Some important abiotic (physical and chemical) characteristics of estuaries

In this lesson we shall consider in detail, the characteristics that have a major effect on the estuarine system. They include tide and salinity. Other characters will be explained only briefly.

Tide

In a freshwater aquatic system the change in the water level is due to rain and evaporation. But in the seas and estuaries, it is mainly due to the tidal action, which is actually the periodic rise and fall of the water level, due to the attraction of the sun and moon. Tides are ocean surface phenomena, familiar to anyone who has spent time on seashore. Let us learn more about tides now.

When the water level rises it is called the high tide and when it lowers it is called the low tide. During high tide, sea water flows into the estuary and the salinity increases. During low tide, fresh water flows into it and the salinity decreases correspondingly. Due to these fluctuations the habitat along the margin of the estuaries is covered at high tide and exposed at low tide. This is a prominent feature in estuaries.

Tidal changes are less noticeable in the estuaries located close to the equator. But in countries like Australia there are estuaries with a tidal fluctuation of about 8m.

Tides occur mainly due to the following forces. **(Fig.4.21)**

1. Gravitational attraction of the sun on the earth resulting in solar tide.
2. Gravitational attraction of the moon on the earth resulting in lunar tide.
3. Centrifugal force generated by the rotating earth.

There are two types of tides.

- Daily tides
- Monthly tides

Daily tides

When there is one high tide and one low tide per day it is called diurnal tides (E.g. Some parts of Mexico). Some countries have two high tides and two low tides per day and this pattern is called semi diurnal tides. (e.g. Sri Lanka). The two high tides are quite similar to each other, as the two low tides. There is another daily tidal pattern, which has two high tides and two low tides that occur each day, but successive high tides are quite different from each other (e.g. West coast of North America) **(Fig. 4.22)**

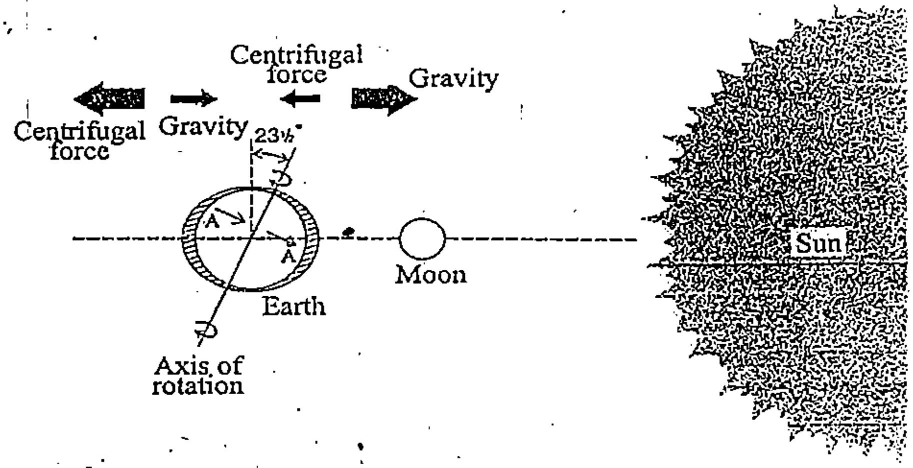


Fig. 4.21. Different forces that result tides

Monthly tides

Two types of tides can occur within one month and they are the:

Spring tides

Neap tides

Spring tides

These tides occur during full and new moon days, when the sun, moon and earth are in alignment. On these days the solar tide has an additive effect on the lunar tide. This creates extra-high, high tides and very low, low tides **(Fig. 4.23a)**

Neap tides

These tides occur in the first quarter and the third quarter of the lunar month. On these days the sun and moon are at right angles to each other, the solar tide partially cancels the lunar tide, to produce moderate tides known as neap tides. The water level has only small changes from high tide to low tide as it creates extra-low high tides and very high, high tides **(Fig. 4.23b)**

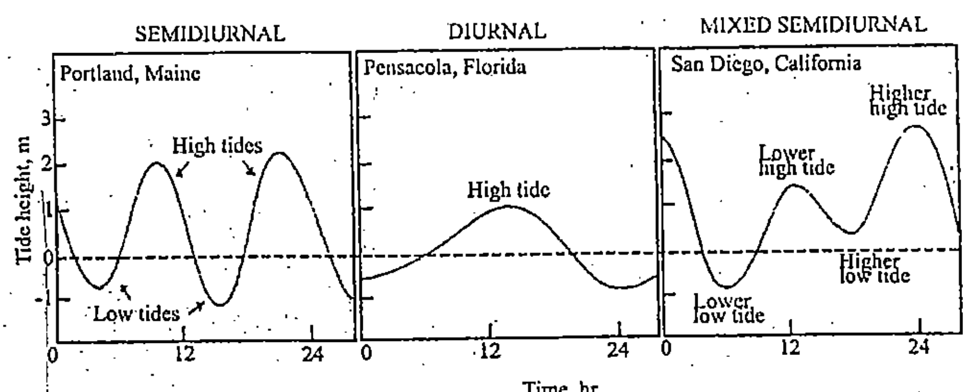


Fig. 4.22 a. Three common types of daily tides

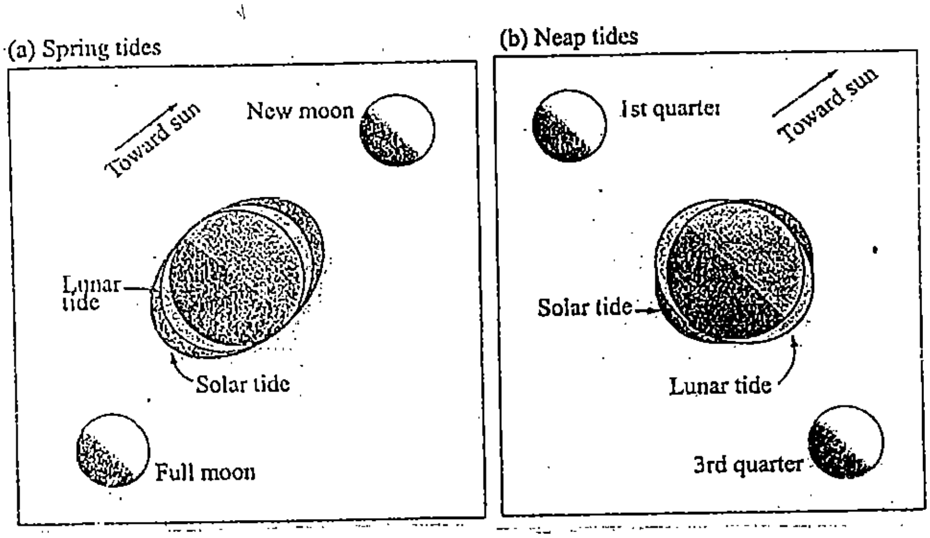


Fig.4.23 Two types of monthly tides
a. Spring tide
b. Neap tide

Salinity
The dominant feature of the estuarine environment is the fluctuation in salinity. Normally a salinity gradient exists at sometime in an estuary. We have already discussed the pattern of salinity distribution in different types of estuaries. There are, however other factors that alter salinity patterns. They are the tides, Coriolis effect and the seasonal changes. Let us look at each of these factors in detail.

Effect of tides on salinity
If there is a significant tidal range, during high tide the salt water moves further up the estuary displacing the isohalines upstream, Low tides by contrast, displace the isohalines down stream. As a result a certain area of the estuary is subjected to a salinity regime that changes with each tide.

Force of Coriolis effect on salinity
The rotation of the earth deflects the flowing water. In the Northern Hemisphere this effect deflects out flowing fresh water in north-south oriented estuaries to the right, as one looks down the estuary toward the sea. The opposite is true, in the Southern hemisphere. As a result, two points (each on opposite side) in an estuary may have different salinities.

Effect of seasonal changes on salinity
Seasonal changes in salinity in the estuary are usually the result of seasonal changes in evaporation, fresh water flow or both. In areas where freshwater discharge is reduced or absent higher salinities may be found further upstream (salt wedge). With the onset of increased freshwater flow, the salinity gradients are moved toward the mouth. High evaporation makes the water more saline.

Extreme shallowness
Next to salinity, the most important environmental feature of estuaries is the shallow depth. Due to shallowness, the light penetrates to the bottom. Especially during low tide, when most part of the edges or boundaries expose, the dark mud, which get heated by the sun. This increases the decaying process of detritus by numerous bacteria found in estuaries. This activity increases the organic food in the estuaries.

Substrate
Most estuaries have soft, muddy substrates. These are derived from sediments carried into the estuary by both seawater and fresh water. Currents and the size of the particle control the deposition of particles. Both sea water and fresh water drop their coarse sediments first, the former at the mouth of the estuary and the latter in the upper reaches of the river itself, so the area of mixing (middle region of estuary) is dominated by fine silt (mud) forming a muddy bottom. The interstitial water between the mud particles, changes its salinity slower than in the overlying water, so that the animals dwelling within them are subjected to less drastic salinity changes.

Temperature, and currents
Water temperatures in estuaries are more variable than in the nearby coastal waters. Rivers in the temperate regions are colder in winter and warmer in summer than adjacent seawater. When this fresh water enters the estuary and mix with the seawater they alter the temperature.

Currents in estuaries are caused primarily by tidal action and river flow. The highest velocities occur in the middle of the channel where the frictional resistance from the bottom and side banks is lowest.

Turbidity
Because of the great number of particles suspended in the water of estuaries the turbidity is high. This results in decrease in the penetration of light.

Oxygen
The regular flux of fresh and saltwater in to the estuary, coupled with the shallowness, turbulence, and wind mixing, usually means there is an ample supply of oxygen in the water column. However, oxygen is severely depleted in the substrate, as the high organic content and high bacterial populations of the sediments exert a large oxygen demand on the interstitial water.

Nutrients

Since estuaries are energy rich, nutrient rich ecosystems they are the most biologically productive ecosystems on earth. The tidal action concentrates the nutrients that are brought to the estuary from the riverside and the seaside. It takes much longer, for a nutrient particle to traverse an estuary than to pass through a similar length of even the most slowly flowing river. Thus the estuary acts as a nutrient trap.

Activity 4.11

Fill in the blanks

1. Daily tides with two high tides and two low tides per day are called ________________.
2. Spring tides occur in ________________.
3. When neap tides occur the sun and moon are a t________________ to each other.
4. The tidal fluctuations are highest during ________________ tides. In the Southern hemisphere, the flow of water in the estuaries deflected to the ____________ side of the estuary.
5. Oxygen concentration in the substrate of the estuaries is low due to ____________ ______________.

4.6 The Oceans

The oceans are not equally distributed over the earth. Oceans cover more than 80% of the Southern hemisphere but only 61% of the Northern Hemisphere, where most of the earth's land mass occurs **(Fig. 4.24).**

The oceans have been separated for convenience into three major divisions: Pacific, Atlantic and Indian. They are connected to each other, on their southern end by Antarctic Ocean. There is the Arctic Ocean too, which is usually considered to be a marginal sea, connected to the Atlantic.

The Pacific Ocean is the deepest and largest basin, occupying more than one-third of Earth's surface. The Atlantic is a relatively narrow basin connecting the Arctic and Antarctic Oceans. It is also relatively shallow, and has wide continental margins.

The Indian Ocean lies primarily in the Southern Hemisphere. It is the smallest of the three major ocean basins. Three of the world's largest rivers (Ganges, Brahmaputra, and Indus) discharge into the northern Indian Ocean. This region has an abundance of both fresh water and sediment from the discharge of these rivers. Thus, the northern Indian Ocean is the most affected by nearby lands.

The northern Indian Ocean also has two major sources of warm saline water. They are the red sea and the Arabian Gulf , which produce warm saline subsurface waters that can be traced for hundreds of kilometers below the surface of the Indian ocean.

Projecting from, or partially cut off from, these larger oceans are smaller marginal seas, such as the Mediterranean, Caribbean, Baltic, Bering, China sea and Okhotsk **(Fig. 4.24).**

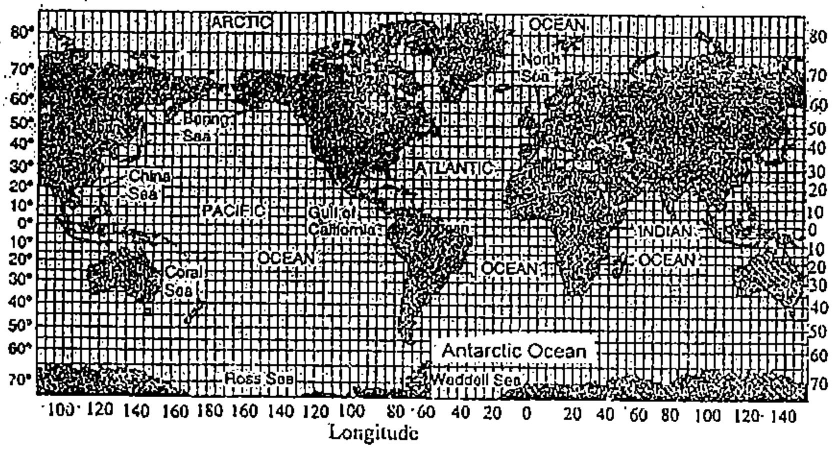


Fig. 4.24 Major oceans and the seas of the world

Features of the ocean basin

Two boundaries can be recognised, in the ocean basin.

- Continental margin
- Deep ocean basin

Continental margin

The principal boundary between any continent and ocean basin is the submerged area called the continental margin, which consists of a **continental shelf, a continental slope** and a **continental rise (Fig. 4.25).** Continental shelf is the submerged upper part of the continental margin. It is close to the land and mostly shallow. It ranges in width from less than 1 Km to more than 1300 Km off the North shore of Alaska and Siberia. Continental shelf is the basic source of nutrients for ocean plants. Most biologically productive waters in the ocean overlie the continental shelf. Continental slopes are the edges of continental blocks. It is the zone of steeply sloped sea floor, leading from the continental shelf towards the ocean bottom. It extends down to depths of 2-3 Km. at the base of the continental slope, the steepness disappears and the bottom begins to slope gently again. The average angle of inclination is about 4.

Continental rise separates the continental slope from the ocean bottom. This area is usually divided into an upper and lower part, with the slope being different in the two.

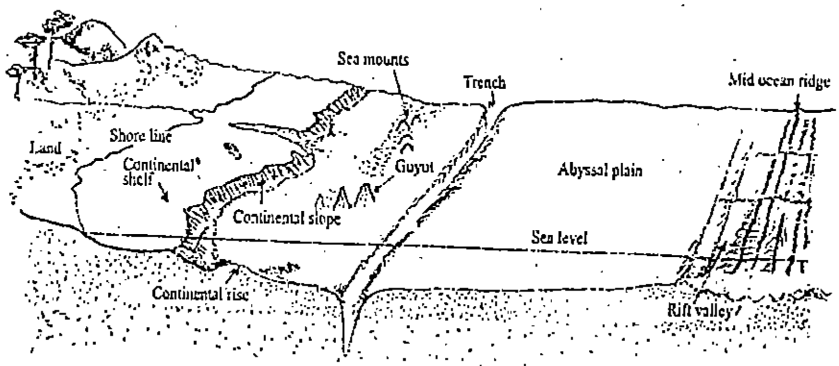


Fig. 4.25 Diagrammatic cross section of the ocean basin

Deep ocean basin
The main feature of the deep ocean basin is the **abyssal plain.** It is among the flattest parts of earth's surface (**Fig. 4.25**). Most abyssal plains appear to be covered with thick deposits of sediments that likely come from nearby lands. These deposits include lithogenous particles (rock origins) carried from the continental shelf by ocean currents and winds, biogenous particles (of biological origin).

There are some characteristics features in the abyssal plain. These include, **Mid ocean ridges, Rift valley, Sea mounts, Trenches, and Volcanoes**

The long underwater mountain ranges are called mid-ocean ridges. They are the most distinctive world wide oceanic features. One ridge rises the entire length of the Atlantic Ocean. It continues through the Indian Ocean along the east side of the Pacific Ocean. Small earthquakes occur frequently on the crest of mid ocean ridges and most points on the ridges are far below the sea level. Most prominent feature of this ridge is, its steep sided central valley called rift valley.

A rift valley is 1Km to 2Km deeper and 25-50 Km wide. It is bordered by rugged mountains whose tallest peaks come to within 2 Km of the sea surface.

Thousands of mountains are scattered across the ocean basin called seamounts. Many seamounts were formed as volcanoes or volcanic peaks, which are especially common along the mid-ocean ridges. Volcanic peaks that have been flattened because of wave erosion before subsiding beneath the ocean surface are called table mounts or guyots. Their flattened tops are more than 200m below sea level. As you see each guyot is an old seamount whose top was once close to sea level.

The deepest places in the oceans are called trenches (Fig. 4.25). They are relatively narrow cannels and 3 to 4 km deeper than the surrounding floor. Most trenches occur in the Pacific, especially the western Pacific. There are also trenches in the South Atlantic (south sandwich trench) and in the Indian Ocean (Java trenches). Trenches are associated with active volcanoes and earthquakes.

Volcanoes and volcanic islands are common in the ocean. They usually stand 1 km or more above the surrounding Ocean floor. Most volcanic eruptions occur quietly and usually go unnoticed on mid ocean ridges. The second type of oceanic eruption occurs in shallow waters and can be quite violent.

Oceanic depth zones

Now we consider the ocean's vertical structure. There are three principal depth zones: surface zone, pycnocline, and deep zone (Fig 4.26).

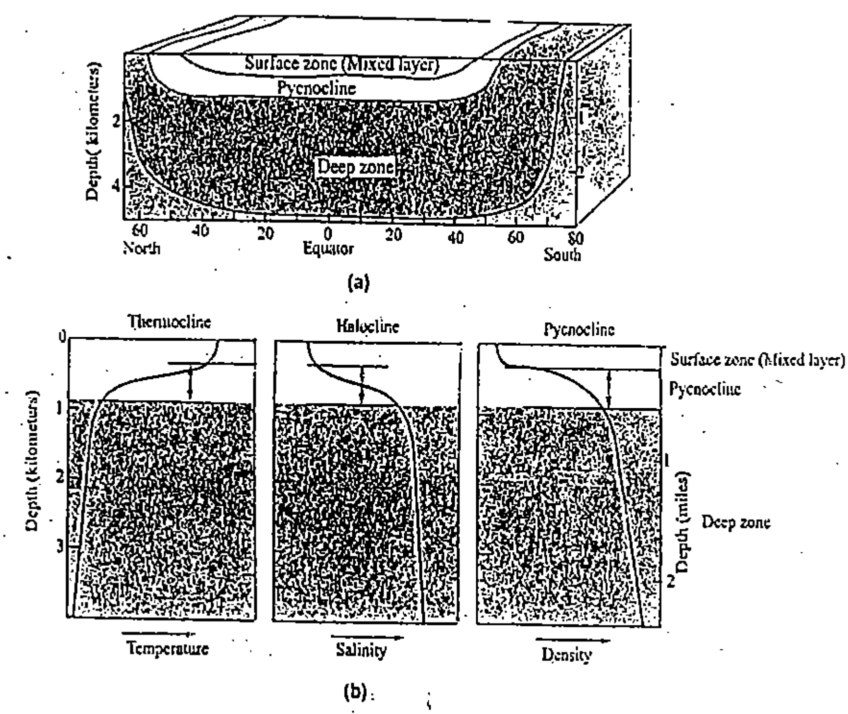


Fig 4.26 a Oceanic Depth Zones

b. Marked variation in temperature, salinity and density with depth

The **surface zone** is 100 to 500 meters thick and contains about 2 percent of the ocean volume. It is intimately linked with the overlying atmosphere. For instance, water temperatures and salinities in the surface zone change seasonally because of variation in precipitation, evaporation, cooling and heating. This zone contains the warmest and least dense waters in the oceans. Average surface-water temperature is 17.5°C **(Fig 4.26a).**

Near-surface waters are well mixed by winds, waves and cooling or heating of the surface. For this reason, the surface zone is also called the **mixed layer,** because the waters there move vertically very easily. These vertical motions are mainly wind-driven.
The **pycnocline** is where water density changes markedly with depth. The exact depth of the pycnocline is controlled by those factors which influences the density of seawater, namely temperature and salinity.

Where the seawater density is controlled primarily by changes in temperature, the pycnocline coincides with a zone of marked temperature change; called a **thermocline (Fig. 4.26b).**The zone where seawater density is controlled by marked changes in salinity is referred to as a **halocline.** Because temperature changes are more important in the open ocean, where salinity changes little, the depth of the pycnocline, is controlled by a thermocline. In coastal ocean areas where salinity changes dominate and temperature changes are less important, halocline controls the depth of pycnocline.

Below the pycnocline is the **deep zone,** which contains about 80 percent of the ocean's volume. Except in the high latitudes **(Fig. 4.26b),** the deep zone is separated from the atmosphere. This isolation of the deep zone prevents interactions with the atmosphere and warming of the deep ocean water by solar heating. Thus, deep zone retains its low water temperature -3.5°C- characteristic of the surface waters in the Polar Regions. Since the temperature and salinity of deep-ocean waters are unaffected by surface processes, temperature and salinity are conservative properties.

Activity 12

Select the correct word from the list and fill in the blanks.
[Continental shelf, Abyssal plain, trenches, Pycnocline, Halocline]

1. the zone where seawater density is controlled by marked changes in salinity is referred to as ____________________.
2. The vertical movements of waters in the surface zone and seasonal changes in their temperature or salinity do not penetrate the ____________________ as this zone has a great stability.
3. A large portion of the deep ocean consists of flat, sediment covered areas called ____________________.
4. The deepest areas of the ocean are called ____________________.
5. The relatively smooth underwater extension of the edge of the continental shelf and the deep ocean basin.

Currents in the Ocean

Ocean waves and currents are large-scale water movements, which occur everywhere in the ocean. They are formed due to various causes, mainly the prevailing wind system. Let us learn about the wind system first.

Wind patterns
The uneven heating of the atmosphere generates the prevailing wind systems that drive the ocean current systems. The maximum rate of the atmospheric heating occurs at the equator and the minimum rate at the poles. Because of this distribution pattern, warm, humid low-density air rises above the equator and cold, dry dense air descends over the poles.

These winds are cool at high altitudes, sink at around 30° north and south, forming a belt of high pressure. After, the air in these zones sinks, it spreads along the earth's surface, causing the surface winds.

Part of the air moves towards the equator as the **trade winds**. The rest of the air flows generally toward the poles as **westerlies (Fig. 4.27)**. The direction of the currents changes according to these winds. Hence the current moves in clockwise direction into northern hemisphere and anticlockwise direction into southern hemisphere.

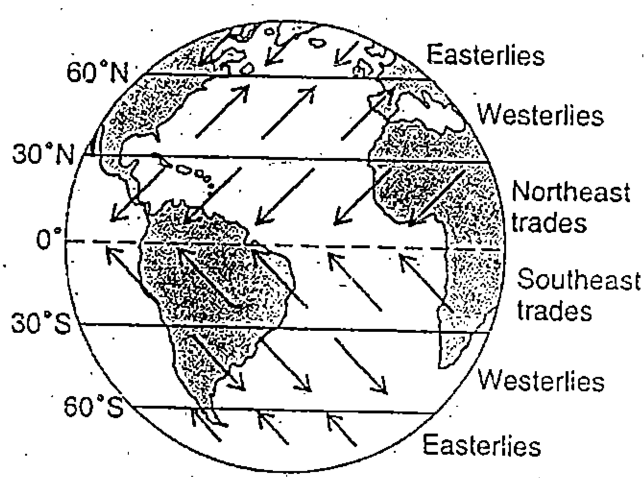


Fig. 4.27 Prevailing surface winds

Circulation of Ocean waters
Within the open ocean there are two separate water circulation systems; surface water circulation driven by the winds and deep water circulation, driven by the continuous sinking of surface waters, at the high latitudes. The latter is a slow circulation present through out the deep ocean.

Waves

Most of the waves are generated by winds blowing across the ocean surface. They may also be generated by earth quakes, volcanic explosions, and underwater landslides. Where waves are forming, the ocean surface is chaotic. Waves range in size from ripples only a few centimeters in height to storm waves, which may tower as high as 30m. Ripples form first and then grow into larger waves as winds continue to put energy into the water surface. Longer waves travel faster than short ones. Most ocean waves are less than 6 meters high. Waves are altered when they enter shallow water. They change direction by refraction, moving most slowly in shallow water and fastest in deep water. As the waters get shallower, waves eventually become unstable and break, forming breakers.

Currents

The direction of a current is determined not only the wind system but also the land masses. There is another force that affect the direction and that is the Coriolis force (remember the effect of this force in the estuaries). This force causes the ocean current to deflect into roughly circular gyres that move clockwise in the northern hemisphere and counter clockwise in the southern hemisphere

There are two entirely different ocean current systems. They are the **horizontal and vertical currents**. We are more familiar with the horizontal currents (which are known as) surface currents. These are primarily driven by winds. Vertical currents, referred to as subsurface currents, are driven by chilled waters sinking in the polar and sub polar oceans are less familiar.

Horizontal Currents

Trade winds and westerlies create these currents. They are surface currents and are divided into 3 types.

Boundary currents

The Equatorial currents

Antarctic circumpolar currents

Boundary currents

These currents flow close to and parallel to the continental margins. They are more changeable than open ocean currents. Mainly, because the many irregularities in continental coastline, cause the direction of flow, of boundary currents to change frequently..The two principal types of boundary currents are named according to which side of an ocean basin they appear on. These are the western boundary current and the Eastern boundary current.

Western boundary currents

The western boundary currents are very swift and its speed ranges from 100-1000 km/day. Speeds are highest at the surface. They are narrow (less than 100km) but deep (2km). They are the strongest surface currents. They are especially conspicuous in the northern hemisphere. (E.g. Gulf Stream in the Atlantic) (**Fig 4.28a**).

Eastern boundary currents

Eastern boundary currents occur in the return flow toward the equator (**Fig 4.28a**). They are relatively shallow (about 500m) but are broader (about 1000km) and less deep. They move much more slowly. In the southern hemisphere the western boundary currents (Brazil current) and the eastern boundary currents are not as strongly developed.

The Equatorial current system

The circulation of the upper layer of the ocean in the tropical and subtropical regions is dominated by the north and south equatorial currents driven primarily by the trade winds.

Among the outstanding branches of the equatorial current systems are the equatorial countercurrents (**Fig 4.28a**). The locations of these countercurrents are closely related to the location of the doldrums. Near the equator there are belts where the air is rising because it is warm and contains a lot of moisture and climate is characterized by high rain fall, much cloudiness, light, variable winds and low atmospheric pressure, such a belt is called doldrums.

The equatorial undercurrent is another extremely narrow eastward setting current branch in the equatorial current system. It is centred on the equator. Most often it does not reach the sea surface.

Monsoon currents

During February- March, the northeast monsoon drives a small clockwise gyre circulation in the Arabian Sea, a clockwise gyre in the Bay of Bengal, and westerly flows throughout the remainder of the northern Indian Ocean. During August -September, the southwest monsoon creates a clockwise gyre circulation in the northern Indian ocean and Arabian Sea with an anticlockwise gyre in the Bay of Bengal. The result is a reversal of the Somali Current between the two seasons and the presence of an Equatorial Counter Current in the Northeast monsoon period (**Fig. 4.28b**).

The Antarctic circumpolar current

Currents around Antarctica flow mainly from west to east. The great eastward flow is called the Antarctic circumpolar current. In contrast to the wind driven currents in tropical and subtropical regions this current is deep. In some parts of the ocean it reaches the bottom at 3000-5000m depth. Therefore its course is strongly affected by bottom topographical features, like submarine ridges. The surface speeds are rather small but as a result of the great depth, the water volume transported exceeds that of any other oceanic current systems.

We hope now you have a clear understanding about the surface current system in the ocean. Next we will focus our attention to vertical current system in the oceans.

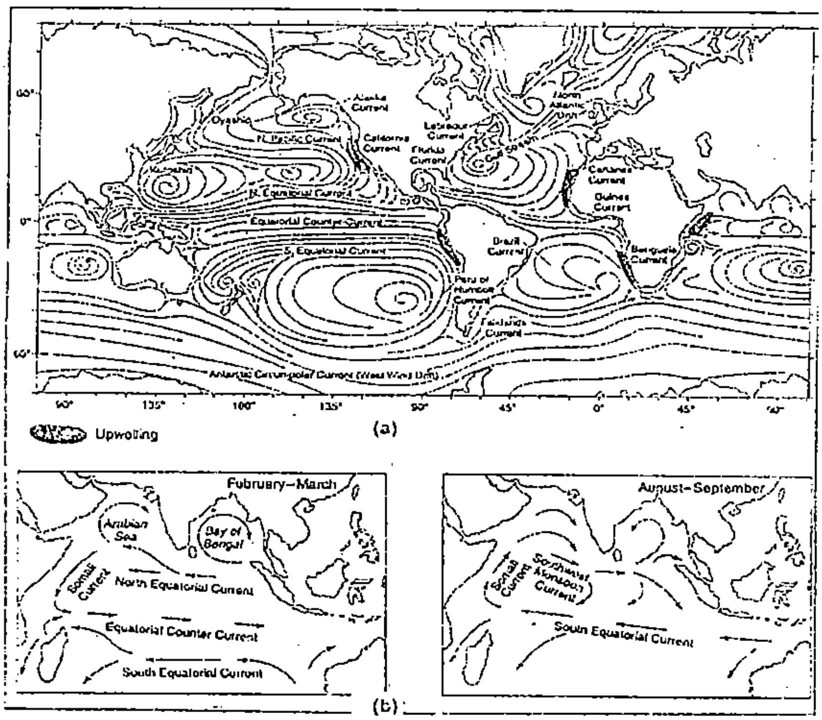


Fig. 4.28 a. Surface ocean circulation patterns
b. Monsoonal cutrrents

Vertical Currents

Vertical water movements are produced by sinking and upwelling processes. Such processes tend to break down the vertical stratification established by the thermocline

Sinking

Sea water sinks when the density increases. The physical processes that increase sea water density are strictly surface features. This, dense sea water, which is from the surface and is usually highly oxygenated, transports dissolved oxygen to deep areas of the ocean basins. The chief areas of sinking are located in the colder latitudes where sea surface temperatures are low.

Upwelling

Subsurface water is carried to the photic zone by several processes which are collectively termed upwelling. These waters are rich in nutrients. Upwelling is especially conspicuous on the eastern margin of ocean basins, for example, the wind driven surface currents along the continental margins flow toward the equator. At the same time, the Coriolis effect tends to push these surface waters offshore. This water is then replaced by deeper water transported vertically along sloping surfaces to the surface **(Fig. 4.29).** Similarly along the equator the two equatorial currents flowing east are deflected away to the right north of the equator and left to the south of the equator. To replace this water, subsurface water up wells to the surface. Such wind induced upwelling is limited to the upper 200 m of the water column. near coasts (near capes or other irregularities in the coast line) and equator (E.g. Peru and Chile).

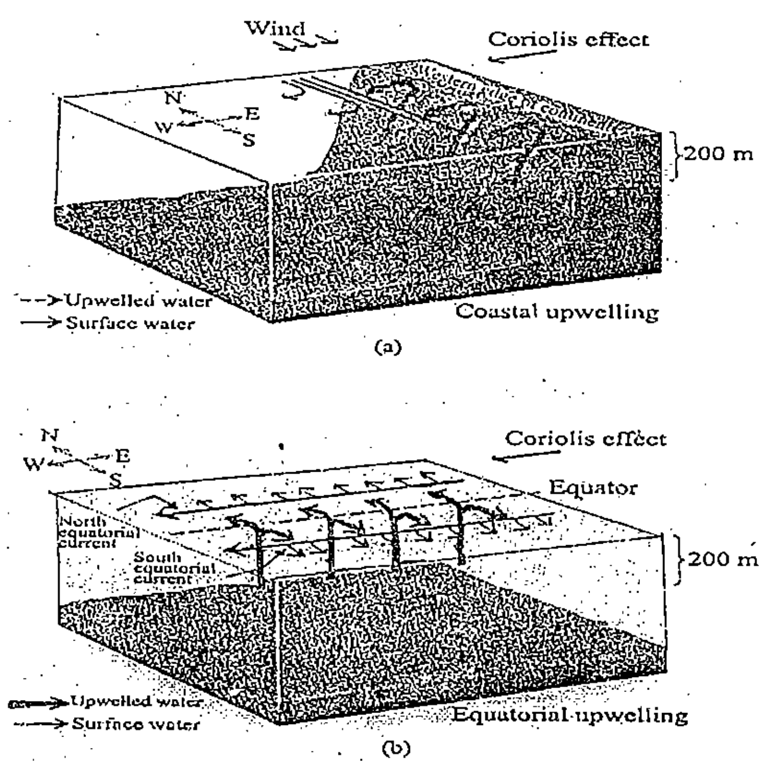


Fig. 4.29a) Coastal upwelling in the Northern Hemisphere
b) Equatorial upwelling

Sea ice

Sea ice forms where sea water cools below its initial freezing point. Each winter, sea ice covers the entire Artic Ocean and completely surrounds Antarctica. It also forms in bays and along the coast of Alaska, the western coast of Canada, and the Atlantic coast of North America as far south as Virginia. In spring much of the ice melts but large areas in the polar oceans remain ice-covered throughout the year. The annual expansion of ice – covered ocean areas and the retreat of the ice in local summer affect climate worldwide. In this section, we discuss how and why sea ice forms.

When seawater chills to its temperature of initial freezing, clouds of tiny needle like particles form, making the water surface slightly turbid. The ocean surface dulls and no longer reflects the sky. As ice particles grow, they form hexagonal spicules 1 to 2 centimetres long. The needles and spicules of newly formed ice are called *frazil ice*. When the surface is stirred by winds and waves, the ice forms a soupy-looking layer known as *grease ice*.

As sea ice continues to form, ice crystals eventually form a blanket over the water surface. When the surface is calm, an elastic layer of ice forms. It is only a few centimeters thick. Waves and especially winds break these ice sheets into large pieces called **floes**.

Since salt is excluded from ice, the remaining water becomes more saline and its freezing point is lowered. Some brine pockets remain trapped in the ice. Salinities of newly formed ice are typically 7 to 14, but this value depends on temperature. The more slowly the ice forms, the easier it is for brines to escape, resulting in lower ice salinities. Conversely, at very low temperatures, ice forms rapidly, and the salty brines cannot easily escape. This quick freeze results in higher ice salinities. Salinity of sea ice is always lower than that of the surrounding waters. As sea ice ages the brines are expelled. Thus, multiyear ice typically has salinities around zero at the top and around four at the bottom.

Snow accumulates on top of, and freezes to, the ice surface. Thus, sea ice grows from both top and bottom. First –year ice is flat and usually snow-covered. During a single winter, new sea ice can reach a thickness of two metres. Where sea ice never completely melts, multiyear ice continues to grow, and older ice has a rough hilly surface. In the central Arctic, multiyear ice reaches thicknesses of three to four meters. Ice melts during the summer, down to about two metres in the central Artic. The fresh water released by melting ice forms a thin layer of low-salinity surface water.

Currents and winds move large pieces of sea ice together, forming mounds called **hummocks** or **pressure ridges** that are the ice pack's most conspicuous features. At these pressure ridges, the ice is deformed and thickened, up to 20 metres thick. These ridges can extend many tens of metres below the ice and are hazards to submarines moving under the ice.

When floes move apart, they expose open waters in areas called **leads.** Leads range from few centimetres to many hundreds of metres wide and can extend for many kilometres. Ships moving through sea ice utilize leads where possible to avoid having to break ice. Mammals stay near the leads and near holes in the ice. This permits them to catch fish and other food in the underlying waters.

Activity 13

Select the correct statement and mark them with a tick (√)

1. As waves enter shallow water and begin to encounter the bottom effects, they slow their forward motion and the wave length decreases.
2. The Choriolis effect is the result of the rotation of the Earth on its axis.
3. The western boundary currents are very slow.
4. The direction of the boundary currents are not affected by the continents.
5. Due to vertical currents, nutrients at the bottom layers are brought to the surface.
6. Sea ice contains a high percentage of salt.

Summary

Water is a common, yet very remarkable, substance on the earth's surface. While abundant in its liquid form, large quantities of water also exist as a gas in the atmosphere and as a solid in the form of ice and snow. The asymmetrical shape of a water molecule creates an electrical charge separation that initiates hydrogen bonding which in turn, affects water's thermal properties such as, heat capacity, latent heat of fusion and vaporization, and other basic properties such as surface tension, density-temperature relationships, solvent capability, viscosity etc.

All water on the planet is constantly recycled, by a system known as the hydrological cycle driven by the solar energy. In this cycle, water is lost to the atmosphere as vapour from the earth, which is in tern precipitated back in the form of rain snow frost etc. This precipitation and evaporation continues forever and thereby maintain a balance between the two. There is an immense amount of water on the earth's surface. Out of this, large quantity, only a little percentage; (about 0.25) could be used for human consumption. Over 97% of water is deposited in the ocean depression.

Lakes are not uniformly distributed over the surface of the earth. They are formed as a result of a number of different processes both natural and artificial. They can be divided into three main categories namely rock basins, barrier basins and organic basins.
Freshwater habitats are categorized as lentic or standing water and lotic or running water.

In lentic habitats there is no unidirectional current and the water movement is mainly due to wind. The surface waters are clearly lit throughout the day unless it is in a shady area.
The surface waters are clearly lit and temperature, fluctuate widely, daily and annually. Thermal stratification is present and a sharp temperature gradient zone or thermocline may be formed. (if it is deep).
The substratum is widely composed of sand, clay or mud. The stillness of the waters results in a very stable bottom.
Accumulation of organic matter tends to increase the nutrient content, thus supporting plant and animal life.
If the mineral nutrient or soluble nitrogen compounds are high a rich plankton growth may appear especially in early spring and summer.
At the bottom region, due to the action of aerobic bacteria on detritus, low oxygen concentrations may be present. Due to the action of anaerobic bacteria gases such as hydrogen sulfide, ammonia and carbon dioxide may be formed.

Two clearly different zones can be identified in a river. The upper region or the head waters of a stream consist of clear cool water flowing over a substrate composed of cobbles and gravel. The most distinctive feature of this zone is the shear force exerted by the fast flowing current, which affect most of the physical and chemical characteristics of the zone. The next region is the sediment transferring zone, which is comparatively wider and has an average flow rate. The lowest region is the sediment depositing zone, which is larger, deeper and muddier with a laminar flow which encourages deposition of sediments.
Estuaries are partially enclosed coastal embayment where fresh water and seawater meet. Based on geology and geomorphology, there are four basic estuarine types: coastal plain, tectonic, bar build estuary and fjord. Based on salinity gradients, there are three groups of estuaries. They are the Positive, negative and neutral estuaries.
The dominant feature of the estuarine environment is the fluctuation in salinity, which is mainly affected by, tidal regime, Coriolis effect, and seasonal changes in evaporation, freshwater flow or both. Most estuaries have soft, muddy substrates, with large amounts of particulate organic matter, which serve as food for estuarine organisms. Oxygen is usually in ample supply in the water column but is severely depleted in the substrate.
There are three major oceans in the world and they are connected to each other. These oceans are connected to the continents by shallow extensions of the continents called continental shelves. Most of the ocean basins are flat

abyssal plains, but these plains are cut by deep trenches in some places, and have volcanoes sea mounts etc. in the other areas.

The open ocean has a three –layer structure: surface zone, pycnocline and deep zone. The surface zone responds quickly to changes in the overlying atmosphere. The pycnocline inhibits exchanges between atmosphere and deep zone. The deep zone is exposed to the atmosphere only in high latitudes, which causes its waters to be cold.

The oceanic water is constantly in motion. It is mixed and moved by winds, waves and currents, sinking water masses, and upwelling.
Sea ice is a major feature of the ocean. Its freezing in winter and melting in summer dominates surface waters in the polar oceans. Sea ice also influences the deep- ocean circulation. The coldest and densest water masses form in the polar oceans. Freezing sea ice expels salt, and this excess salt in the remaining liquid water increases the density of water masses, which is especially important near Antarctica.

Objectives

Now you should be able to:

- understand the special properties of water in relation to its molecular structure
- Giving examples, explain how some of the physical properties of water have become biologically important
- Describe the main processes, that drive the hydrological cycle
- Give an account on the composition of the hydrosphere.
- List the major types of freshwater habitats in the world
- Give a concise essay on formation of lakes
- Describe the temperature variation in tropical and temperate lakes
- Describe the physical and chemical characteristics of lentic habitats
- Write an account on the stratification pattern in the lakes of the world
- Explain how the streams are classified according to stream order
- Explain the nature of the different zones of a river that originates from a high mountain.
- Give a concise account on the type of estuaries classified, based on the geology and geomorphology
- Explain how estuaries are classified according to salinity gradient
- Describe the important abiotic characteristics of estuaries
- Describe the features of the ocean basin
- Describe the variation between the oceanic depth zones
- Describe the formation of waves and the patterns of current in the world ocean
- Write an account on Sea ice