Fundamentals of Soil Science

Dean Student Welfare

Alluvial soil = 75 mha
Black soil = 74 mha
Red soil =
Laterite soil = 25 mha
Desert soil : 29 mha
forest area : 75 mha.
Salt effected soil = 10 mha.

IRS-IA = 1988
" -IB = 1991
" -IC = 1995
" -Id = 1997
" -IE = 1993
P_2 = 1994
P_3 = 1996
P_4 = 1999
P_6 = 2003,

Cartosat-1 = 2005
" -2 = 2007
" 2A = 2008

Indian mini satellite = 2008,

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Chapter 1

Introduction

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Soil and land, though related, are two different entities. Land is two-dimensional entity representing geographical area and landscape, while soil is a three-dimensional body with length, breadth and depth and is hidden below the land surface. It is largely hidden from the outside world until it is lost and goes out of the site. It is recognized by digging a pit and exposing its profile. The profile of a soil with well marked horizons called A, B and C, tells the history of its formation and bears the imprint of many physical, chemical and biological processes which have led it to the present form. The soil entity dictates much of what societies can do. Soils are formed by weathering processes of rocks which in the words of Hans Jenny can be depicted by the equation:

$$S = f(cl, r, p, o, t, \ldots)$$

where, S = Soil, cl = Climate (rainfall, temperature, snow, etc.), r = Relief or topography, p = Parent material (rocks, minerals and geological formation), o = Organisms (plants and animals), and t = Time or age.

Some of these factors of soil formation are passive but the organisms (biosphere) and man are most active in the transformation of soils and in modifying its properties. These are the effects of human activities or the human-induced changes in the modern times, which are largely responsible for the existing diversities in soil properties. Human beings, through their negligence, ignorance or greed, can accelerate soil degradation and through scientific management, improve its quality. A soil scientist has to understand both the human-made and the natural soil processes and develop strategies for appropriate remedial action to meet the needs of the society or the current managers of the land.

'Soil Science' provides the eyes and ears to the society in translating what the idealistic dreamers envision as the utopian society into the reality of what nature will allow. Soil scientists must know soil, understand its potential and limitations to answer the following type of questions:

- What is the carrying capacity of a given soil?
- What is its productivity?
- How should we realize the potential of a soil and how should we remove its constraints?
- How can one prevent soil degradation and restore its productivity?

The relationship of soil, environment and society is intimate and depends on soil quality and its management. Those who develop the concepts, the science and the facts about this thin skin that interspaces the lithosphere and atmosphere and help manage it, often control the fate of societal visionaries. The laws of physics, chemistry and biology are inviolable and the manner in which a society utilizes them to manage the soil, determine the returns on investments in soil management and its use.

Soil is an interface of the organic and inorganic chemistry of the terrestrial world, combining nitrogen and carbon from the atmosphere with the various elements of mineral lithosphere via the organisms anchored in the soil, intercepting energy of the sun and moisture from the hydrosphere and atmosphere. The soil fractionates the hydrologic cycle in the terrestrial area into runoff, percolation and evapotranspirational components. It is a source as well as a sink to adsorb, desorb, fix or release mineral elements and gases and grow plants and decompose crop residues incorporated into it. It is a living factory where millions of tiny organisms are ceaselessly working day and night, transforming the organic matter and participating in carbon and nitrogen cycles and many mineral elemental cycles. Without life, there is no soil and without soil, there is no life on the planet Earth. It is no wonder that the astronauts landing on the moon searched for evidence of soil, life and water there and brought with them soil samples to make detailed studies for an evidence of life.

To understand and rationalize the use and management of soil, the soil scientists, especially pedologists, have been developing methods of soil survey and classification of soils. A widely used system is the 'Soil Taxonomy' developed by the United States Department of Agriculture. Indian soil scientists, using the same system, have mapped and classified their soils. India is virtually a museum of soils of the world, as almost all the Soil Orders classified anywhere in the world, are found in India. The National Bureau of Soil Survey and Land Use Planning is a repository of soil information in the country. The Bureau has delineated 20 agro-ecological regions and 60 agro-ecological sub-regions by integrating the information on soils, physiography, climate and effective plant growing period. This is a valuable information for land use planning, soil management and use.

Ever since the dawn of civilization, man's interest in soil has been to produce food, fibre, fuel and timber, but in modern times besides the growing demands of society for these products of the soil, the use of land for industry, civic purposes, roads, airports, habitations, etc. is also increasing and the problems of soil degradation are assuming greater dimensions. Prime lands are going out of agricultural use and marginal lands are coming under the emerging needs of the society, including agriculture. Thus, the demand on soil scientist's expertise is becoming too wide and too difficult. He/She is expected to respond to man's ever increasing, constantly changing needs and respond to most challenging tasks. Thus, a shift in paradigm of soil science and information technology research and education has become necessary to meet the demands of the time and of the situation. Advances in science, such as biotechnology, space technology, computer sciences and modelling techniques have their impact on soil science and soil management technology and a major shift in education becomes inevitable. Since soil is a pivot of natural resources, its management has to be fully integrated with the eco-friendly techniques. It is no surprise that many research and educational organizations are making soil science as an integral part of natural resource education and research systems.

Soil and water are integral parts of the system for plant growth and no study of soil is complete without consideration of water regime and its interaction with the soil and plant as one system. It also draws heavily on knowledge of soil physics and water management. Likewise for understanding the dynamics of plant nutrients in the soil, knowledge of soil chemistry, soil fertility, soil biology, soil biochemistry and soil microbiology is essential.

The chemical and physical properties of soil, its mineral composition and stock of nutrient resources for plant growth and the changes therein call for a regular monitoring of the availability, rate of depletion and addition to the soil system. Ever since the dawn of civilization, man has been exploiting the native fertility of soil but with the advancement of science and compulsion of growing needs of agricultural products from the soil, the use of chemical fertilizers has emerged as the most important input for increasing productivity in agriculture. No country, whether developed or developing, can hope to produce more and more food from less and less land without resorting to the enhanced use of fertilizers. However, their imbalanced application or use without organic manures/ green manures and overmining of soil resources of nutrients is leading to declining yields and increasing deficiency of many nutrients. Moreover, poor management of irrigation, neglect of rainwater management and soil conservation practices is accelerating soil degradation, salinization, alkalinization and groundwater pollution and in nutshell, deterioration of soil quality. A soil scientist needs to have intimate knowledge of soil degrading processes and should develop strategies for integrated use of all the inputs such as plant nutrients, water, energy and crop management technology for optimizing productivity and minimizing environmental degradation.

India is home to 18% of the world's human population and 15% of the animal population, but owns only 2.0% of the world's geographical area, out of which 51% is the net cultivated area. This meagre resource has to support the ever-growing population, which has grown from 361 millions in 1951 to over 1000 millions in 2000, and is projected to become about 1400 millions in the next 25 years. The land to man ratio will be reduced to 0.10 ha by 2025 from the 0.14 ha in the year 2000. The scope for extending agriculture to new areas is extremely limited and the only option left for us to meet the challenge of growing needs of food, fibre, fuel and timber is through increase in productivity of the land already under the plough. But, this has to be done without detriment to environment.

The country receives on an average 1130 mm of rainfall every year but it is unequally distributed, both in time and space and most of it comes in a few months of the monsoon season. It is the erratic distribution of rainfall, which leads to frequent droughts and floods and makes Indian agriculture a gamble in monsoons. No doubt one-third of the cultivated area is irrigated but its productivity is hardly one-third of the potential attainable with the present day technology and inputs. This gap needs to be bridged. Two-thirds of the country's cultivated area is unirrigated and follows rainfed farming. These soils are not only thirsty but also hungry. Farmers' poverty makes the soils poor and they, in turn, make the farmers poorer. This is a vicious circle. Whether it is irrigated or unirrigated area, the future food security lies in increasing its productivity. Soil and water (irrigation as well as rain water) management and scientific knowledge-based precision farming offer great scope for realising the potential of these favourable as well as harsh environments. Diversification of agriculture, and alternative land use systems are the needs of the times. Increasing use efficiency of agricultural inputs, including water for improving the economy of the system along with prevention of degradation of environment, are important challenges of the day.

Area irrigated 1/3
productivity 1/3

Soil is neither an inexhaustible store of plant nutrients nor a renewable natural resource, capable of withstanding onslaught of the exploitative agriculture and soil degrading processes. It is only through scientific management that its productivity, quality and sustainability can be maintained. Arresting the fall in organic matter is the most important weapon to fight soil degradation and to ensure sustenance of soil quality and agricultural productivity. Addition of nutrients such as N, P, K, Ca, S, Mg, Cu, Mn, Fe, Zn, B and Mo becomes essential from external sources to maintain input-output balance and to ensure their ready availability for meeting the needs of growing crops. Harnessing microorganisms for augmenting nutrient supply and availability will make integration of nutrient management system more economical and sustainable.

Long-term experiments under intensive cropping systems in different agro-ecological regions have provided ample evidence of rapid exhaustion of soil and nutrient deficiencies which are increasingly becoming limiting factors for crop production. Indian soils are mostly deficient in N and the application of nitrogenous fertilizers is essential for crop production in all soils, but the lower use-efficiency of nitrogenous fertilizers and high cost of fertilizers dictate the conjunctive use of nitrogenous fertilizers, organic manures and biological N fixing technology should be practised for improving their use-efficiency and economy. Soil organic matter is the key to N economy and soil quality. The world is becoming more and more conscious of depleting soil organic matter stock and its consequences for both crop production and environment quality.

Increasing industrialization, urbanization and negligent use of agricultural chemicals are multiplying the problems of pollution and deterioration of quality of groundwater. Soil, no doubt functions as a buffer and a sink for these pollutants, but it can do so only within certain limits and an understanding of these limitations is the key to soil health and environment improvement.

Water is the critical factor for crop production. But the cost involved in developing sources of surface irrigation, rapidly falling groundwater-table due to over-exploitation of underground water resources and increasing salinization and alkalization in irrigated areas are becoming the chronic problems of irrigated farming. Increasing water-use efficiency and prevention of soil degrading effect of poor water management should receive high priority for research and education in soil science.

Soil is a limited and non-renewable resource but a pivot for agriculture, food security, nutritional security, environmental safety and quality of life. A dynamic system of soil management requires a knowledge of all the sciences, bearing on the nature of problems that the soil scientists are called upon to handle and the technologies that can be put to use. Soil and society are undoubtedly two inseparable entities on this planet Earth. 'Soil scientist moves from the field to the laboratory to diagnose and develop a technology for solving the soil-related problems and gets back from the laboratory to the field to evaluate his strategy'; it is the secret to success. The chapters included in this textbook endeavour to address all these issues in a coherent manner.

Chapter 2

Weathering and Soil Formation

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1. Introduction

Soil is a surface-covering on most of the earth's land area, an aggregation of unconsolidated mineral and organic particles produced by the combined action of wind, water and organic decay. Soils are formed from hard rocks, loose and unconsolidated transported inorganic materials and accumulated organic residues. Even the loose mineral materials from which soils are formed, are originally constituted by the weathering of rock masses to stones, gravels, sands, silts, clays and soluble salts. Organic soils are developed mostly from plant parts that have accumulated in stagnant waters where decomposition is slow due to lack of oxygen (anaerobic conditions).

2. Soil and Regolith

The unconsolidated material overlying the rocks is known as regolith (Figure 1). Its thickness may vary from a few centimetres on sloping hills of the Himalayas to tens of metres in the Indo-Gangetic alluvial plains. It might have accumulated *in situ* from the weathering of underlying rock (as in the Deccan plateau) or might have been transported from elsewhere by water (in alluvial plains and coastal areas), wind (in the Thar desert) or ice (in higher reaches of the Himalayas) and deposited on the bedrock or on other materials covering the bedrock. Thus, the composition of regolith can vary widely.

The characteristics of upper 1-2 metres of regolith are markedly different from those of the underlying materials. It generally has higher content of organic matter and has been subjected to more weathering than the lower portions. Consequently, it can be differentiated into characteristic layers called 'horizons'. This upper and biochemically weathered portion of the regolith is called 'soil'.

Soil formation comprises two different processes. The first is referred to as 'weathering' which changes the hard consolidated rock mass into unconsolidated loose materials. The second is referred to as 'soil formation' and covers changes occurring within the loose materials as time passes. In nature, the change from a solid rock mass to loose soil materials and other changes within the soil profile occur simultaneously. The term soil formation is used to mean both the formation of unconsolidated materials by the weathering processes, and the soil profile development which encompasses the changes involved in the development of horizons in a soil.

3. Weathering

Weathering refers to the chemical and physical disintegration and decomposition of the rocks, and the minerals contained in them. It is basically a combination of transformation and synthesis. As a result of weathering, the rocks are broken down physically into the smaller fragments and eventually into individual mineral grains. Simultaneously, the rock fragments and minerals undergo chemical changes to form new minerals by either minor modifica-

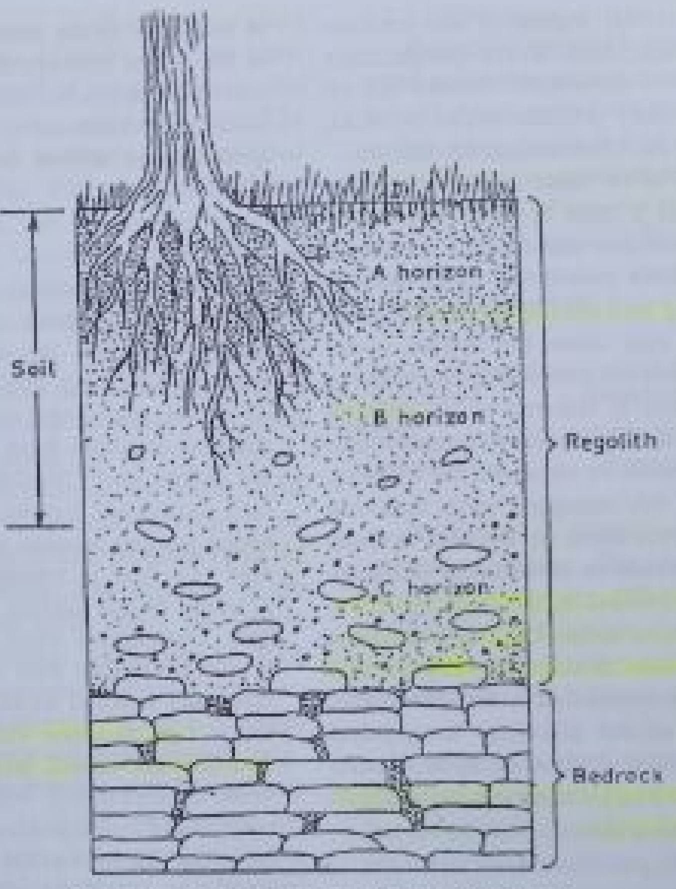

Figure 1. Regolith in relation to soil and bedrock

tions (simple transformations) or complete chemical and structural changes. The chemical changes are usually accompanied by a reduction in particle size and release of soluble constituents which may be lost through drainage water or may be recombined to form new (secondary) minerals. Weathering of the initial materials precedes soil formation in hard rocks and accompanies it in soft rocks and soil materials. It is a continuing reaction during soil development to the point where no more reactants are available. Thus, weathering results in mechanical (or physical) and chemical breakdown of rocks. The former is often designated as 'disintegration'. However, both these processes operate simultaneously. In drier (Thar desert) and/or cooler (upper Himalayas) climates, physical breakdown is more pronounced. In humid tropical climates, chemical breakdown predominates resulting in the formation of deep to very deep soils.

3.1. Physical Weathering

Physical weathering is a mechanical process, causing disintegration of consolidated massive rocks into smaller pieces. Under very cold or very hot and dry conditions, such as in the central Himalayas or the Thar desert, physical weathering is prominent and is operated through various agents, viz. temperature, water, ice, wind and living organisms. The role of these agents are discussed below.

3.1.1. Temperature

Rocks may be classified into two types, viz. monomineralic (made up of one mineral), and polymineralic (made up of more than one mineral). In nature, polymineralic rocks are common. As a result of diurnal temperature changes, the rocks get heated during the day and cooled during the night. Different minerals in the rocks have different coefficients of expansion and thus heating and cooling of rocks results in differential expansion and contraction of minerals. With every temperature change, therefore, differential stresses are set up which eventually produce cracks in rocks, thus facilitating their mechanical breakdown.

Because of slow heat conductance, the outer surface of a rock is often warmer or cooler than the inner parts which are more protected. With the passage of time, this differential heating and cooling produce, lateral stresses which may cause the surface layers to peel away from the underlying parent mass. This phenomenon is referred to as 'exfoliation'. Some rocks may have entrapped water also. Water expands by about 9% on freezing. The freezing of this entrapped water thus may exert a tremendous force on the rock surface, if space is not available to dissipate this pressure. The force exerted by the freezing of water is equivalent to about 1465 tonnes (mega grams) per square metre (Mg/m^2) or 150 tonnes/ft^2 — a pressure that widens the cracks in huge boulders and dislodges the mineral grains from smaller fragments. Role of temperature thus can be overwhelming in the physical weathering of rocks.

3.1.2. Water

The rainwater falls on the land and travels towards the low lying areas such as rivers, lakes and oceans, continuously detaching, and shifting, sorting and reworking the sediments that it carries. When loaded with sediments, flowing water has a tremendous cutting power, resulting in the formation of gorges, ravines and valleys. Of all the agents of physical weathering, the effect of water perhaps is more pronounced and widespread. The rounding of sand grains on ocean beaches in the Eastern and Western Ghats is indicative of the abrasion that accompanies the water movement.

3.1.3. Ice

The moving ice is an erosive detachment and transporting agency of tremendous capacity. Snow received at higher elevations or polar regions accumulates and starts moving in the form of glaciers. During their movement, glaciers cause great deal of cutting and crushing of the bedrocks. Although glaciers are not so extensive in the present day environments; in the recent geological past, they had transported and deposited parent materials over millions of hectares on this planet. At present, glaciers are active in upper parts of the Himalayas.

3.1.4. Wind

Wind, an important agent of transportation of suspended particles, also exerts an abrasive action. Generally, the effect of wind is more pronounced during arid climates. Dust storms of almost continental extent have occurred in the past. Particles from the bare land surface are picked up by the blowing winds, particularly during dry months. Poorly aggregated or single grained deposits are prone to wind erosion. As dust is transported and deposited, abrasion of one particle against the other occurs. Dust-ladden winds also act as an abrasive agent against such obstacles as rock outcrops and exposed rocks, which come in their way. The rounded rock remnants in the Aravalis are caused largely by wind action. Wind in combination with ocean waves causes mechanical disintegration along the sea coasts, as in the Eastern and the Western Ghats.

3.1.5. Plants and Animals

Some plants, like mosses and lichens, grow on the exposed rock. They accumulate dust, which further encourages plant growth, and a thin film of highly organic material is formed. Sometimes, roots of higher plants exert a prying effect on rocks (as the root girth increases with plant growth), which results in some disintegration. Burrowing by rodents, movement of animals, and human activities (cultivation, quarrying, land levelling, construction of roads, buildings, railway lines, etc.) also result in physical weathering. Such influences, however, are of relatively limited importance in producing parent material when compared to the drastic physical effects of water, ice, wind, and temperature changes.

3.2. Chemical Weathering

Physical disintegration is accompanied by chemical decomposition which produces changes in the nature and composition of rocks and minerals. Chemical weathering takes place mainly at the surface of rocks with the alteration or disappearance of some minerals and formation of (new) secondary minerals. The process of chemical weathering is controlled by various agents, viz. water, temperature, and different organisms. The presence of water facilitates chemical weathering. The rate of chemical reaction increases with increase in the amounts of dissolved carbon dioxide and other minerals in water, and with increase in temperature. Chemical weathering is minimal in the desert areas, due to scarcity of water and in the cold regions, due to low temperatures. Due to availability of sufficient water and the presence of favourable temperature, chemical weathering is well pronounced in the humid tropical climate. The presence of organic and inorganic acids, which are formed as a result of microbial breakdown of plant residues, also accelerates chemical weathering. These agents commonly act in a synergistic manner to convert primary minerals (e.g., feldspars, micas, amphiboles, etc.) into secondary minerals (e.g., kaolinite, vermiculite, hydrous oxides, etc.) with release of water soluble ions. The principal processes of chemical weathering are described below. For a clear understanding of individual processes, these have been described separately, but in nature they mostly occur simultaneously.

3.2.1. Solution

Water is a universal solvent. Its solubilizing action is enhanced when it contains dissolved carbon dioxide, organic and inorganic acids or salts in it. Most of the minerals are affected by the solubilizing action of water, though by varying degrees. Some minerals such as halite (NaCl) dissolve readily in water, whereas the solubility of some silicates such as quartz in water is very low. Solution helps in the continuous removal of weathered materials but the total removal by simple solubilizing action is very limited. In the arid climates, due to paucity of water, even water-soluble minerals remain in rocks and sediments, whereas these are completely washed away in the semi-arid and humid regions.

3.2.2. Hydration

Hydration means chemical combination of water molecules with a mineral to form a new mineral. Many anhydrous minerals undergo hydration when they come in contact with water. Hydration reactions occur primarily on the surface and edges of mineral grains, but may pervade the entire structure in simple salts. Some examples of hydration reactions are given in Equations (1) and (2):

$$\underset{\text{(Anhydrite)}}{CaSO_4} + \tfrac{1}{2}H_2O \rightarrow \underset{\text{(Hemihydrate)}}{CaSO_4.0.5H_2O} \qquad \ldots(1)$$

$$\underset{\text{(Anhydrite)}}{CaSO_4} + 2H_2O \rightarrow \underset{\text{(Gypsum)}}{CaSO_4.2H_2O} \qquad \ldots(2)$$

Hydration is always accompanied by an increase in volume. The characteristics of hydrated minerals are different from those of their anhydrated counterparts. The hydrated minerals are usually soft and are readily weatherable. The absorbed water provides a bridge or entry way for the hydronium (H_3O^+) ions or protons (H^+) to attack the structure. Slaking of certain rocks is mainly due to hydration of their mineral constituents. Under the hot desiccating conditions, dehydration (reverse of hydration) can also occur.

3.2.3. Hydrolysis

Hydrolysis is one of the most important processes in chemical weathering and results in complete disintegration or drastic modification (in structure and composition) of weatherable primary minerals. Hydrolysis involves the partial dissociation of water into H^+ and OH^- ions. The pure water undergoes very limited dissociation, but in the presence of dissolved carbon dioxide, minerals and organic acids in it, the concentration of H^+ ion increases, resulting in an accelerated hydrolytic action of water. Hydrolysis is a double decomposition process and a hydroxide of some kind is usually formed. Water thus acts like a weak acid on silicate minerals, as depicted in Equations (3) and (4):

$$H_2O \rightarrow H^+ + OH^- \qquad \ldots(3)$$

$$\underset{\text{(Orthoclase)}}{KAlSi_3O_8} + H^+ \rightarrow HAlSi_3O_8 + K^+ \qquad \ldots(4)$$

Another example of hydrolysis reaction is the replacement of interlayer potassium in micas by protons or hydronium ions to produce vermiculite:

$$\underset{\text{(Muscovite)}}{K(Si_3Al)Al_2O_{10}(OH)_2} + H^+ \rightarrow \underset{\text{(Vermiculite)}}{(K,H)(Si_3Al)Al_2O_{10}(OH)_2} + K^+ \quad \ldots(5)$$

9 E

The products of hydrolysis are wholly or partially removed by the percolating water, depending on the climatic conditions and permeability of the residual materials. They may also recombine with other constituents to form clays. In a way, hydrolysis reactions may be considered as the forerunners of clay formation.

3.2.4. Oxidation

Oxidation is an important chemical reaction occurring in well-aerated rock and soil materials where oxygen supply is high and biological demand is low. It is particularly important in rocks and minerals that contain iron and manganese — elements that are easily oxidized. In most of the primary minerals, iron is present in the ferrous (Fe^{2+}) form. On oxidation, it undergoes the following reaction [Equation (6)]:

$$Fe^{2+} - e^- \rightarrow Fe^{3+} \quad \ldots(6)$$

Oxidation of iron is a disintegrative weathering process in minerals containing ferrous as part of their crystal structure. Reduction in size and increase in electrical charge on oxidation of Fe^{2+} to Fe^{3+} create electrical and structural imbalances in these minerals. Rocks containing ferromagnesian silicates such as pyroxenes, hornblende, biotite, glauconite and chlorite, are susceptible to oxidation. In other cases, ferrous iron may be released from the mineral and is almost simultaneously oxidised to the ferric form. An example of this is hydration of the mineral olivine and the release of ferrous ions from it, which may be immediately oxidised to ferric form which has very low solubility. When ions such as Fe^{2+} are removed or are oxidised within the minerals, the rigidity of the mineral structure is weakened and the mechanical breakdown becomes easier. This provides a favourable environment for further chemical reactions. Similarly, minerals containing the manganous form of manganese and sulphide groups are also susceptible to oxidation.

3.2.5. Reduction

Reduction occurs where a material is water-saturated (such as below the water-table level), oxygen supply is low, and the biological oxygen demand is high. The net effect of these conditions is the reduction of a metal, say iron, to its highly mobile form, ferrous (Fe^{2+}) in this case. In this form, it may be removed from the system if there is a downward and/or outward movement of groundwater. If ferrous iron persists in the system, it tends to form sulphides and other ferrous compounds. These impart characteristic green and blue colours to many reduced soil materials. Under certain conditions, there could be formation of lepidocrocite (γ-FeOOH), resulting in the characteristic orange and yellow mottles.

3.2.6. Carbonation

Carbonic acid (H_2CO_3), although a weak acid of carbon dioxide, is very important in the chemical weathering of rocks and minerals as it makes minerals more soluble. The atmosphere contains only 0.03% carbon dioxide, but the rain water may contain as high as 0.45% carbon dioxide. The decomposition of organic matter and the respirations of roots and other soil organisms (macro and micro) liberates carbon dioxide in large amounts. The CO_2 concentration in soil is thus much higher (0.25%) than atmosphere. Carbonation tends to produce carbonates and bicarbonates as shown in Equations (7) and (8):

$$H_2O + CO_2 \rightarrow H_2CO_3 \quad \ldots(7)$$

$$\underset{\text{(Calcite)}}{CaCO_3} + \underset{\text{(Carbonic acid)}}{H_2CO_3} \rightarrow \underset{\text{(Calcium bicarbonate)}}{Ca(HCO_3)_2} \quad \ldots(8)$$

The solubility of calcium bicarbonate is considerably higher than that of calcite.

3.2.7. Integrated Weathering Processes

Different types of chemical reactions described above occur simultaneously and are interdependent. For example, hydrolysis of a given primary mineral may release ferrous ion that is quickly oxidised to ferric form, which, in turn, is hydrated to give a hydrous oxide of iron. Hydrolysis also may release soluble cations, silicic acid, and aluminium or iron compounds. These substances can recombine to form secondary silicate minerals such as silicate clays.

3.3. Factors Affecting Weathering

Different minerals weather at different rates. Three major factors which affect weathering of rocks and minerals are: climatic conditions, physical and chemical characteristics of the rocks, and stability of minerals.

3.3.1. Climatic Conditions

The climatic conditions influence the rate and nature of weathering profoundly. Under arid conditions, the physical weathering predominates. The size of particles decreases with relatively little change in the chemical composition of a mineral. The original primary minerals are prominent, whereas the content of secondary minerals is low. Physical changes due to temperature fluctuations and wind action are accompanied by only limited chemical changes. Consequently, the soils of arid regions are remarkably like the parent materials from which they are formed. Similarly, in extremely cold climates, the rocks and minerals undergo mechanical disintegration with little modification in chemical composition. In humid regions, the forces of weathering are more varied. Vigorous chemical changes accompany physical disintegration. New minerals such as silicate clays and oxyhydroxides and oxides of iron and aluminium are predominant. Chemical alteration is accelerated and intensified by the action of organic acids formed by the decomposition of large quantities of organic matter produced by abundant plant growth.

In the humid tropical regions, as in southern and north-eastern India, the year-round high temperatures and the luxuriant plant growth provide optimum conditions for intensive weathering. In these regions, the primary silicate minerals succumb to weathering and only the highly-weathered silicate clays persist. Resistant products of chemical weathering, such as oxides of iron and aluminium, tend to dominate soils of the humid tropical regions.

3.3.2. Physical Characteristics

The physical characteristics that influence weathering include particle size, hardness, and nature and degree of cementation. Rocks comprising minerals with large crystals, disintegrate easily than those with fine crystals because of pronounced expansion and contraction due to changes in temperature. However, once the rocks get disintegrated into smaller fragments, the finer crystals undergo relatively rapid chemical changes than the larger ones because the fine-grained materials provide larger surface area for chemical attack.

Rate of weathering also depends on the hardness and cementation. For example, a dense quartzite or a sandstone cemented strongly by a slowly weatherable mineral can resist mechanical breakdown and presents only a small total surface area for chemical activity. On the other hand, porous rocks, such as volcanic ash, coarse limestone or sandstone, are readily broken down into smaller particles and are easily decomposed.

3.3.3. Chemical and Structural Characteristics

Chemical composition and structural characteristics of minerals also influence the ease of their removal or breakdown. Some minerals such as gypsum ($CaSO_4.2H_2O$) or calcite ($CaCO_3$) can be solubilized in water saturated with carbon dioxide and can be easily removed from the parent material. Some minerals such as ferromagnesian silicates like olivine and biotite, contain readily oxidizable iron (ferrous), and their component ions are not very tightly packed in the mineral crystal structures, and therefore can be easily weathered. In contrast, the relatively tightly packed nature of crystal units and lack of oxidizable iron in muscovite impart considerable resistance to weathering.

3.3.4. Stability of Minerals

Minerals can be arranged in the order of stability or weatherability. In view of the differences in surface area and consequent reactivity, it is desirable to separate mineral par-

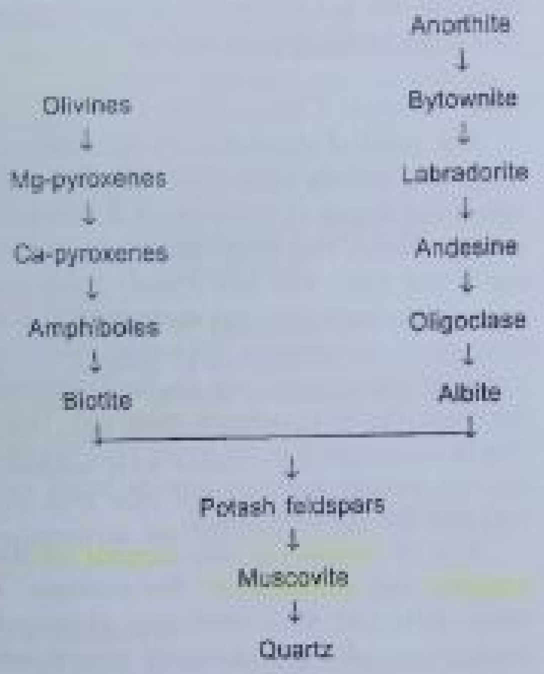

Figure 2. Stability sequence of different minerals

ticles into two classes when we discuss their weatherability: clay size and sand-silt size. The weatherability of the common primary minerals is illustrated by the 'stability series' (Figure 2) proposed by Goldich (1938). On the left hand or the 'basic' branch of this series, there is an increasing silica-tetrahedral linkage with increasing stability from top to bottom. The least stable mineral (olivine) has independent tetrahedron and the silicon tetrahedra are held together by forming bonds with easily hydrolyzable magnesium or oxidizable iron. On the other hand, in quartz, one of the most stable minerals, there is a three-dimensional network of silicon tetrahedra. All the oxygen atoms are shared with the adjacent tetrahedra. Also, there is a decrease in the content of easily hydrolyzable bases from the least to the most stable mineral. On the right hand or the feldspar branch, there is a decreasing distortion of the lattice from calcic to potassic feldspars. In anorthite, 50% of the silicon in tetrahedral sites is replaced by aluminium, whereas in potash feldspars, this replacement is only 25%. Divalent calcium (in anorthite) does not fit well into the framework structure of feldspars. On the other hand, the large monovalent potassium fits snugly in the feldspar structure. Thus, orthoclase (albite) is more stable than anorthite.

northite →

The minerals at the top of both series were formed during the early stages of cooling and crystallization of magma at high temperatures, whereas quartz was the last mineral to crystallize from cooling of magma. Thus, higher the temperature of crystallization of a mineral from the magma, the more unstable it is. Obviously, this higher instability is related to the greater disequilibrium with the environment in the pedosphere (lithosphere-atmosphere interface) as compared with temperature of formation of these minerals from magma.

Besides the stability of primary minerals, the persistence of clay-size minerals in soils is also important. The sequence of weathering is largely controlled by the intensity and capacity factors as a function of time. Weathering produces minerals which may be assigned with a particular index. This index known as 'weathering index' (WI) may help to array the clay-size minerals according to weathering sequence (Jackson *et al.*, 1964). Higher WI shows increase in the resistance to weathering (Table 1). During weathering of a primary mineral, each particle is considered as a closed system. The easily-weatherable minerals in order of increasing weathering indices are: gypsum<

Table 1. Weathering indices of some clay-size minerals in soils

G,C. OD BA QM
VBM, PK, KA HA

Typical clay-size minerals	Weathering index
Gypsum	1
Calcite, dolomite, aragonite, apatite	2
Olivine-hornblende, pyroxenes, diopsides	3
Biotite, glauconite, nontronite	4
Albite, microcline	5
Quartz, cristobalite	6
10Å Mica, muscovite, sericite	7
Vermiculite, 2:1:1 minerals, where brucite layer is Fe/Mg	8
Montmorillonite, beidellite	9
Pedogenic chlorite (2:1:1 mineral)	9
Kaolinite, halloysite	10
Gibbsite, boehmite	11
Allophane	11
Hematite	12
Anatase	13

Source: Bear (1964)

calcite < olivine < biotite. Interestingly, the secondary chlorites (2:1:1 mineral) have more weathering stability (WI 8) than the ideal chlorite (WI 4). The other form of chlorite (2:1:1 mineral) with aluminous hydroxy-interlayering has been grouped under minerals showing weathering index of 9. The kaolinites are the most resistant phyllosilicate minerals (WI 10).

Gibbsite has for a longtime, been considered as index mineral showing advanced stage of weathering. Although gibbsite has been reported to be formed from kaolinite, this transformation appears improbable since it involves a desilication process which can happen only above pH 9 (Millot, 1970). Interestingly the soils showing the presence of gibbsite are all acidic and oxidic. Recent evidences show that gibbsite present in the Ultisols of the Shillong plateau are formed in an alkaline pedo-environment (Bhattacharyya *et al.*, 2000). The model understanding of the formation of gibbsite indicates two important points, viz., (i) gibbsite is present as a remnant of an earlier alkaline pedochemical environment, and (ii) its formation even in the presence of a considerable amount of 2:1 minerals discounts the hypothesis of an anti-gibbsite effect (Jackson, 1963; 1964). The presence of gibbsite in these soils should not, therefore, be considered as a conclusive proof of extreme weathering conditions of soils (Macias Vasquez, 1981; Jenkins, 1985; Lowe, 1986). This fact assumes importance since Jackson's weathering index assigns gibbsite as a mineral with very high weathering index (WI 11). [illegible] Iron and titanium containing minerals have been found to be most resistant to weathering as may be judged by their weathering indices (WI 12 and 13, respectively, Table 1).

There are different ways to evaluate the weathering stage of a soil. The concept of molar ratio (viz., silica:sesquioxides; silica:iron oxides; silica:aluminium oxides; silica:bases) has been found to be effective. Other quantitative evaluation, namely 'weathering mean' (m) has been useful since it involves the knowledge of quantified value and the weathering index (WI) of each mineral (Table 1). The weathering mean (m) is given by

$$m = \frac{\Sigma(p * WI)}{\Sigma p}$$

where,

p = Percentage of the mineral, and

WI = Weathering index of that mineral.

Weathering mean is a sensitive measure of degree of weathering and it is more sensitive to minerals with weathering index 4 to 9 (Table 1). This mean value increases as a function of time and intensity of soil weathering. If p (% mineral) is plotted against the weathering index (WI), a distribution curve is obtained (Figure 3). This curve gradually moves to the right (in Figure 3) when the soil represents advanced stage of chemical weathering.

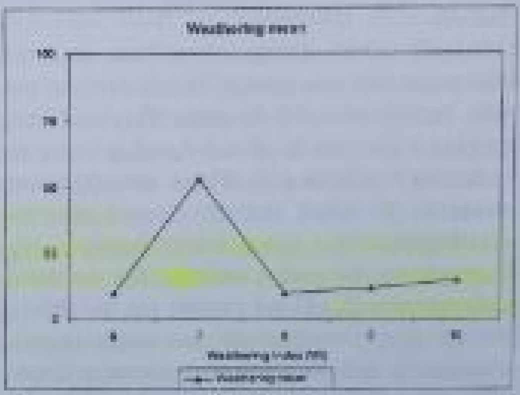

Figure 3. A schematic diagram of frequency curves of a soil showing 10 % quartz (WI 6), 50% mica (WI 7),10% vermiculite (WI 8), 12% montmorillonite (WI 9) and 15% kaolinite (WI 10) The minerals are in clay fractions (<2µm)

4. Soil Formation

At any specific location on the surface of earth, at least five factors act simultaneously to produce soil. These are:

- Parent material
- Climate
- Relief
- Biosphere, and
- Time or age.

These factors are not of equal significance in development of different soils. Although some of them may be more effective in determining the nature of soils under a particular set of conditions, all of them are inter-related and complement one another. Jenny (1941) expressed the relationship of these five factors to the soil properties by Equation (9):

$$\boxed{S} = f(cl, r, o, p, t, \ldots\ldots) \qquad \ldots(9)$$

where,

S = Any soil property, e.g. organic matter content of surface horizons, pH, soil texture, etc.
f = Function of or dependent upon
cl = Climate
r = Relief or topography
o = Organisms (flora and fauna)
p = Parent material, and
t = Time or age.

Thus, any soil property is a function of the collective effects of all these five soil forming factors.

Joffe (1949) divided these factors of soil formation into two groups, viz. active and passive factors of soil formation. Passive factors represent the source of soil forming mass and conditions affecting it. These are: (i) parent material, (ii) relief, and (iii) time. Active factors represent the agents which supply energy that acts on the parent material for the development of soils. These factors are the driving forces that promote the processes causing changes in soil during the course of soil genesis. Climate and organisms are the active factors of soil formation. Soils are often defined in terms of these factors as 'dynamic natural bodies having properties derived from the combined effect of climate and biological activities, as modified by topography acting on parent material over a period of time'.

4.1. Parent Materials

The parent materials on which soils are developed can be divided into two broad groups, viz. sedentary (formed in place) and transported. The transported materials can be sub-divided according to the agencies of transportation and deposition as shown in Figure 4. Different types of parent materials are described in the following sections.

4.1.1. Residual Parent Materials

When the soils develop at a place from the underlying rocks, they are said to have been formed from the residual parent materials. Such materials develop *in situ* from the underlying rock. Typically, they have undergone prolonged and often intense weathering. In a warm, humid climate, as in some parts of eastern states, north-eastern states, Kerala, Tamil Nadu and Karnataka, the parent materials are likely to be thoroughly oxidized and well leached. Red and yellowish brown colours imparted by haematite (α-Fe_2O_3) and goethite (α-FeOOH) minerals, are characteristics of intense weathering in hot, humid climates. In cool and especially dry climates, weathering is less drastic and the oxidation and hydration of iron is hardly noticeable. On the other hand, such parent materials may contain high contents of calcium and magnesium due to limited leaching. Residual parent materials are encountered

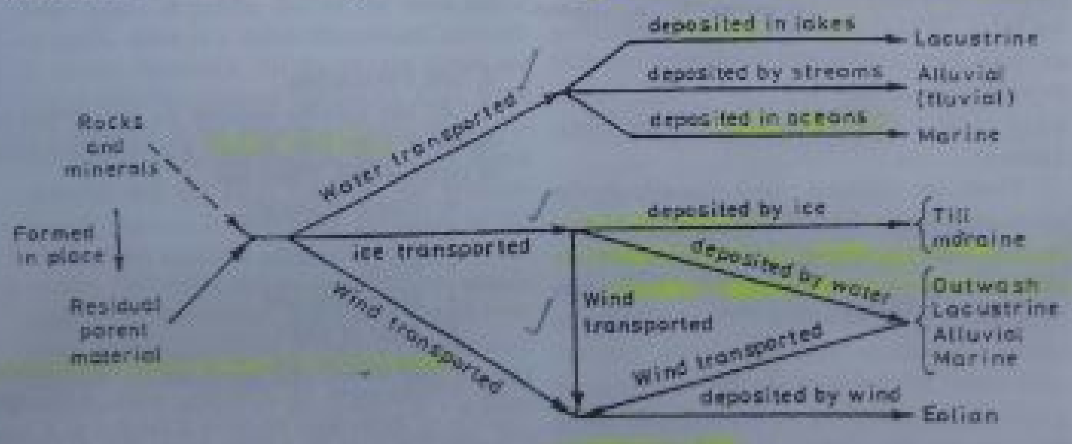

Figure 4. Different types of parent materials

mainly in the Deccan plateau and some parts of central India and are of relatively limited significance in the Aravalis and the Himalayan region.

4.1.2. *Colluvial Parent Materials*

Rocks or soil debris at the foot of a slope that have moved there due to gravity, are called colluvial materials. A colluvial material exists to some extent at the base of all slopes, but it is especially noticeable in the mountainous or hilly topography where rock slides, slips and avalanches are common. It is made up of fragments of rock detached from the higher parts and is carried down the slopes mostly by gravity. The 'frost action' has also much to do with the development of such deposits. A parent material developed from colluvial accumulation is usually coarse and stony, because physical rather than chemical weathering is dominant. Soils developed from the colluvial materials are not of significant agricultural importance generally, because of their small surface area, inaccessibility, and unfavourable physical and chemical characteristics. However, some useful timber, horticulture, plantation and grazing lands in mountainous regions have colluvial materials.

4.1.3. *Parent Materials Transported by Water*

Sediments that are deposited by the flowing water such as streams, rivers, etc. are called 'alluvial materials'. These deposits include floodplains, alluvial terraces, alluvial fans, and deltas. These are described in the following paragraphs.

Floodplains — During rainy seasons, streams and rivers overflow their banks and flood the surrounding area. The parts of a valley and alluvial plains, which are inundated during floods, are referred to as 'floodplains'. Sediments carried by the swollen streams are deposited during the flood, with the coarser materials being laid down near the river channel and the finer materials farther away. Such deposits are found to some extent along every stream. Depending on topography and lithology of the area, and gradient of the river or stream, the floodplains may be very narrow or wide. For example, at some places, the floodplain along the Brahmaputra and the Ganges are tens of kilometres wide. The soils derived from these sediments are generally rich in nutrients, but they may require drainage and protection from the overflow during the rainy seasons. The floodplain soils are young and highly stratified.

Alluvial Terraces — The sedimentary deposits that were laid down by the rivers but are not subject to flooding now are called 'alluvial terraces'. These terraces are generally located at a higher elevation than the floodplains and are separated by escarpments of different heights. Most of the Indo-Gangetic Plains are alluvial terraces.

plain

Alluvial Fans — Sediments deposited by streams of flowing water when they enter plains from the hilly areas are known as 'alluvial fans'. Because of the presence of steep slopes, the flow velocity of streams in hilly and mountainous regions is high. When these streams leave a narrow valley in an upland area and suddenly descend a relatively flat and broader valley below, the sediments are deposited in the shape of a fan. Generally, these deposits are coarse textured, gravelly and stony, somewhat porous and well-drained. Alluvial fans occur in the transition zones between the lower Himalayas and the Indo-Gangetic alluvial plains.

Delta Deposits — Most of the coarser sediments suspended in the stream water are deposited as alluvial fans, alluvial terraces and floodplains. By the time these streams reach the oceans and seas, the flow velocity is considerably reduced and the finer particles are deposited near the mouth of the river as 'delta deposits'. These deposits are fine-textured and are subject to flooding, i.e. they are clayey and swampy in nature. Delta deposits of the Ganges are well known.

4.1.4. *Marine Deposits*

Marine deposits are formed by the deposition of sediments carried by streams into the bodies of static water. The stream-carried sediments are eventually deposited in oceans, lakes, seas, and gulfs. The coarser fragments are deposited near the shore and the finer particles away from the shore. These underwater sediments are built-up over long periods of time. In some areas, these marine deposits get raised above the sea level due to changes in the elevation of the earth's crust. These deposits are variable in texture. Some are sandy, others are clayey. Since the marine sediments have been subjected to soil-forming processes for relatively short periods of time, the properties of the soils that develop on them are predominantly determined by the nature of the marine parent materials. Although not of widespread occurrence, marine deposits are found in some coastal areas.

4.1.5. Lacustrine Deposits

The sediments deposited in fresh water lakes are known as 'lacustrine deposits'. Lakes are inland bodies of static water. As a result of deposition of sediments, these lakes that were once filled with water may become filled with mineral sediments. The source of these sediments is erosion of lake banks and/or particles suspended in streams which get discharged into the lakes. The lacustrine deposits are relatively rare. These may also be formed as a result of glacial activity. When an ice front comes to a standstill at a point where there is no ready escape for the water, ponding begins and ultimately very large lakes are formed. The deposits formed in these glacial lakes range from coarse delta materials and beach deposits near the shore, to larger areas of fine silt and clay deposited from the deeper and relatively still waters at the centre of the lake. Areas of inherently fertile soils have developed from these materials as the lakes dried.

4.1.6. Glacial Deposits

In polar regions, snow continues to accumulate and the pressure of its weight changes the snow to ice. After centuries of such a build-up, the ice starts moving in the form of glaciers. The glaciers transport materials embedded in them and also remove the rock protuberances which come in their way. Materials deposited by the glaciers are called glacial deposits. The materials deposited directly by the glacial ice are called glacial till. These are heterogeneous mixtures of the debris of wide diversity, which vary from rocks and boulders to clay. Glacial till is found mostly as irregular deposits called moraines, of which there are various kinds. Terminal moraines are hilly ridges that characterize the southernmost extensions of the various ice lobes in the northern hemisphere when the margin was stationary for long enough period to permit an accumulation of debris. Terminal moraines occupy little land area. Relatively thinner and more level deposits called ground moraine are by far the most widely distributed glacial deposits which are formed when the ice retreated to the north. An outstanding feature of glacial till is its variability. Consequently, the soils derived from such material are highly heterogeneous.

4.1.7. Glacial Outwash Sediments

[illegible]

As the glaciers melt, torrents of water gush [illegible] constantly from the ice lobes of glaciers, especially during the summer. The vast loads of sediments carried by such streams are either dumped immediately or carried to other areas before deposition. An outwash plain is formed by streams flowing from the ice which are heavily loaded with glacial sediments. Since the sediment is sorted by the flowing water, sands and gravels are common. Outwash deposits are found in valleys and on plains where the glacial waters were able to flow away freely.

4.1.8. Aeolian Deposits

Sand dunes and depositions from the materials transported by wind are called 'aeolian deposits'. These occur in the Thar desert covering western parts of India and the adjoining areas. Except for places where they have been stabilized by vegetation, sand dunes are of limited agricultural value and may become menace to agriculture if they are moving. During the glacial period, conditions were ideal for wind erosion. The weather was cold and windy, and the vegetative cover in areas to the south of the glaciers in the North America was sparse or non-existent. During winters, winds picked up fine alluvial materials, deposited during previous summers by the ice-fed streams and moved them southward. Fine particles from glacial till and even the residual material were similarly transported. These wind-blown materials, comprised primarily silt or silt with some fine sand and clay, are called 'loess'. Aeolian deposits other than sand dunes and loess include volcanic ash. Soils derived from the volcanic ash occur in some parts of the USA, Japan and New Zealand. The volcanic soils are light and porous and are generally of lower agricultural value than the soils developed from the loess.

4.2. Parent Materials and Soil Formation

The parent materials influence soil formation by their different rates of weathering, the levels of nutrients they contain, and their particle size distribution. During the initial stages of soil development, the parent materials exert a greater influence in determining the properties of soil. Thus, in slightly weathered soils such as those in floodplains, the parent material is dominant in determining soil properties. Soil developed on weakly cemented sandstones are sandy. Soils developed on shales are shallow and fine-textured. Glacial till, loess or limestone deposits undergo very limited changes, except accumulation of organic matter, development of some structure and loss of soluble salts. Even moderately to well weathered soils may retain rocks in glacial moraines or preserve nature of clays in deltas or sands in river terraces.

The nature of parent material influences the soil characteristics to a great extent. For example, soil texture is highly influenced by the parent materials. In turn, soil texture influences the downward movement of water, and thereby affecting the translocation of fine soil particles and plant nutrients. Clay illuviation and the identification of Bt (textural B) horizon have been considered as the most important signatures of highly-weathered soils such as Alfisols and/or Ultisols unlike Oxisols (Soil Survey Staff, 1999). Ideally, the term laterite is equivalent to Oxisols in the USDA Soil Classification System (Buol and Eswaran, 2000). The primary requirement for the formation of an Oxisol is the development of an oxic horizon with low CEC, low ECEC (effective cation exchange capacity) and low amount of weatherable minerals (< 10%). The states of Kerala, and Maharashtra (in the Western Ghats) and the north-east region fulfill all the requirements conducive for the formation of Oxisols. Yet, these soils have not reached the stage of Oxisols (Bhattacharyya *et al.*, 1993; Krishnan *et al.*, 1996; Sen *et al.*, 1999; Velayutham and Bhattacharyya, 2000). Studies on these soils as well as a few selected Oxisols from abroad (Beinroth, 1982; Jones *et al.*, 1982; Fox, 1982; Macedo and Bryant, 1987; Bourman *et al.*, 1996; Muggler, 1998) indicate an adequate amount of weatherable minerals in these soils. This suggests that the transformation of Alfisols and Ultisols to Oxisols with time, as often envisaged in the model of soil genesis, is difficult to reconcile (Chandran *et al.*, 2005).

The chemical and mineralogical compositions of parent materials also influence weathering directly and can simultaneously affect the natural vegetation. For example, the presence of limestone in the parent material delays the development of acidity, a process that humid climates encourage. In addition, the leaves of trees growing on limestone deposits are relatively high in calcium and other base-forming metallic cations. As these high-base-content leaves are incorporated into the soil and are decomposed, they further delay the process of acidification.

The parent material also influences the quantity and the type of clay minerals present in the soil profile. The parent material itself may contain different types of clay minerals. The nature of parent material also influences the kind of clays that can be formed. The nature of the clay minerals present, in turn, markedly affects the kind of soil that develops. The quality of parent material in influencing formation of different type of soils has been illustrated by various case studies.

4.2.1. Formation of Ferruginous Red Soil (Alfisols)

Myth about the highly-weathered red ferruginous materials that were hitherto referred to as laterites on the basaltic Deccan plateau has been resolved in the light of quality of basalts as parent material. In soil science parlance, these weathered materials are of 60 million years. The persistence of these non-kaolinitic and/or non-oxidic red soils has been possible due to the presence of base-rich zeolites of amygdoloidal basalt. The quality of parent material thus helps in the persistence of these red soils in the humid tropical climate. It is also an evidence of the role of parent material in preventing the loss of soil productivity even in an intense leaching environment to preserve the natural soil resources in the Western Ghats (Bhattacharyya *et al.*, 1999).

4.2.2. Formation of Black Soils (Vertisols)

Although formation of smectitie minerals in soils has been a phenomenon of humid climate, the soils rich in these minerals are common in the semi-arid to arid climates. Since smectite minerals are ephemeral in high rainfall areas, the formation and persistence of black soils (Vertisols) which are rich in these minerals, are difficult to reconcile. The quality of basalts in the Western Ghats of Maharashtra permits the persistence of black soils even under very high rainfall (> 3,500 mm).

The amygdoloidal zeolites in the basalts provided sufficient bases to prevent the complete transformation of smectite minerals. The presence of smectites and zeolites made the formation of black soils possible in micodepressions even in a tropical humid climate (Bhattacharyya *et al.*, 1993).

4.2.3. Formation of Mollisols

In contrast to generally observed non-acidic and less-weathered Mollisols in temperate semi-arid and humid climate (Fenton, 1983; Fanning and Fanning,1989), acidic and fairly-weathered Mollisols on the Deccan basalt are observed in the hills of central (Satpura Range, Madhya Pradesh) and western India (Western Ghats, Maharashtra) under forest in the tropical humid climate condition. The presence of zeolites as a part of parent material inhibits the transformation of smectites. Retention of adequate amount of smectite and continuous supply of bases from zeolites make the formation of Mollisols possible. The formation and persistence of these Mollisols in association with acidic Alfisols under sparse forest in basaltic landscape of millions of years in central and western India demonstrate the primary importance of the quality of parent material rich in expanding clay minerals and bases on their formation. This unique example thus expands the basic knowledge on the formation of Mollisols even in the humid tropical part of the world (Bhattacharyya *et al.*, 2006).

4.3. Climate and Soil Formation

Climate is a dominant factor in soil formation, mainly because of the effects of precipitation and temperature. Some direct effects of climate on soil formation include:

(i) Retention or accumulation of lime (carbonates) at shallow depths in areas having low rainfall. It is because calcium carbonate and bicarbonate (from dissolving carbon dioxide, minerals, and lime) are not leached out due to limited amount of water moving through the soil. Such soils are usually alkaline.

(ii) Formation of acidic soils in the humid areas is due to intense weathering and pronounced leaching of basic cations (calcium, sodium, magnesium, and potassium).

(iii) Erosion of soils on sloping lands constantly removes developing soil layers.

(iv) Deposition of soil materials downslope buries the developing soils.

(v) Weathering, leaching and erosion are more intense and of longer duration in the warm and humid regions where the soil does not freeze. The reverse is true in cold climates, as in the central Himalayas.

Water plays a key role in the soil formation. A soil is said to be developed when it has detectable layers (horizons), such as of accumulated clays, organic colloids, carbonates, or soluble salts that have been moved downward by water. The extent of colloid movement and the depth of their deposition are determined partly by the amount and pattern of precipitation, which produce the leaching action.

Climate influences the formation of soils through its influence on different soil forming processes, namely clay illuviation, formation of CaCO$_3$ and development of soil sodicity. Vertisols (commonly referred to as black soils) are common in humid, sub-humid (moist), sub-humid (dry), semi-arid (moist), semi-arid (dry) and arid (hot) climates in India. The characteristic shrink-swell behaviour of these soils is governed by high amount of clay (< 2 µm) dominated by smectitic minerals formed from weathering rocks (mainly by basalts) during humid climate. Formation of Vertisols thus occurred in a much wetter climate and are still preserved in the non-leaching clay environment. These Vertisols show different types of properties due to the change in climate (from humid to semi-arid and arid) and bear signatures of more than one genetic episode. Consequently, they qualify for polygenetic soils (Pal *et al.*, 2001). A view of the polygenesis in Vertisols of a climo-sequence from humid to arid the evolution of soils is presented in Figure 5. This evolution sequence remains as an evidence of climatically-induced natural soil formation process (Pal *et al.*, 2003a;b).

Black Soil,

Climate influences the soil formation indirectly also through its action on vegetation. Semi-arid climates support scattered shrubs and grasses. Arid climates supply only enough moisture for sparse, short grasses or shrubs, which may not be dense enough to protect the soil against wind and water erosions. Many arid soils show very limited profile development and contain low amounts of organic matter.

The soils formed under the year-round hot and humid climate are very deep, reddish in colour (due to the presence of oxidized iron as haematite), contain well-decomposed organic matter, and are low in essential elements because of intense leaching.

4.4. Biota and Soil Formation

The activities of living plants and animals, and the decomposition of their organic wastes and residues (the living environment, the biota) markedly influence the soil development. Differences in soils that have resulted primarily from the variability in vegetation are specially noticeable in the transition zone where trees and grasses meet. Under the humid forest vegetation, soils that develop may have many horizons, but are leached (washed, eluviated) in the surface layers, and have slowly decomposing organic-matter layers on the surface. In contrast, some grassland soils near the transition zone of forests have surface horizons rich in well-decomposed organic matter, frequently extending down to a depth of 30 cm or more into the mineral soil.

Burrowing animals— rodents, earthworms, ants, and termites— are very important in the soil formation, if present in large numbers. Soils that are habitat of many burrowing animals have fewer but deeper horizons because of the constant mixing within the profile, which nullifies the organic colloid and clay movements downward.

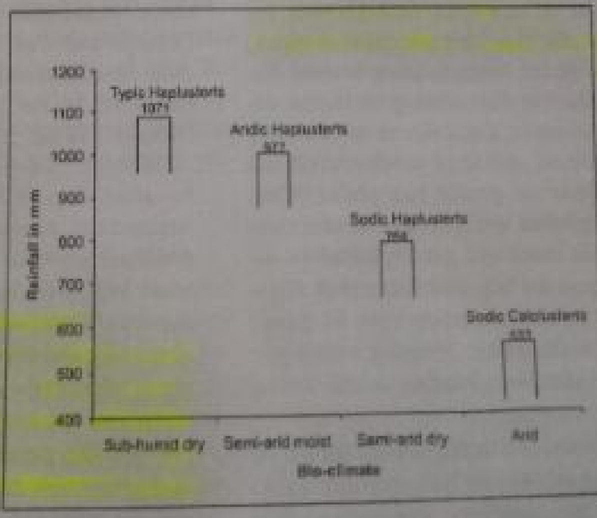

Figure 5. Pedogenic evolution of Vertisols in a climosequence

Microorganisms help the soil development by decomposing organic matter slowly and forming weak acids that dissolve minerals faster than water. Some of the first plants to grow on weathering rocks are crustlike lichens, which are beneficial (symbiotic) combination of algae and fungi.

4.5. Topography and Soil Formation

The configuration of land surface is known as 'topography' or relief. Topography influences soil formation primarily through its effects on modifying water and temperature relations. Soils within the same general climatic area developed from the similar parent materials and on steep hillisides typically have thin horizons because limited amount of water moves down through the profile as a result of rapid surface runoff and rapid erosion of surface. Similar materials on gently sloping hillisides have more water passing vertically through them than do materials on steeper slopes. The soils on gentle slopes generally are deeper, have more luxuriant vegetation, and the organic matter level is higher than in soils on similar materials on steep slopes.

The materials lying in landlocked depressions receive runoff waters from the surrounding higher areas. Such conditions favour better vegetation growth but exhibit slower decomposition of dead plants because of oxygen deficiency in the waterlogged (saturated) soils. This results in soils with a large amount of organic matter. If the area remains wet for many months of a year, organic (peat or muck) soils develop. If the accumulating waters dissolve salts from the surrounding soils, the depression may become a salt marsh with unique salt-tolerant plants, or it may develop toxic salt conditions where no plants can grow. When soils in a watershed are strongly acidic, iron may leach from them and get deposited in depressions to form the bog iron (limonite). Alkaline soils on the sloping topography in humid regions may result in lime leaching which gets eroded into depressions, leading to the formation of a marl.

Fe leached due to acid
⇓
Bog soil.

alkaline condition
lime leached
⇓
Marl soil formation.

In the northern hemisphere, soils on south- and west-facing slopes receive more direct rays of the sun and are, therefore, warmer and drier than north- and east-facing slopes. In the arid climates, these drier south- and west-facing slopes are often less productive than the soils on north- and east-facing slopes. The opposite slopes are affected in the southern hemisphere. In cold, wet areas, these warmer sites may be highly productive. Higher temperature on south- and west-slopes results in greater loss of water by evaporation; the net result in regions where water is limited, is formation of soils with thinner horizons and less vegetative cover than the soils on north and east slopes.

Topography has been reported to influence the soil properties and hence soil types. Interestingly, micro-topographic variations on a landscape have immense control over the soil formation. In the Western Ghats of Maharashtra, micro-topographic variations have influenced the formation of black soils even in the humid tropical climate with very high mean annual rainfall (Bhattacharyya *et al*., 1993). Recent studies on soil formation in the micro-high (MH) and micro-low (ML) in two important food production zones of India, namely the Indo-Gangetic Plains (IGP) and the black soil region (BSR) have shown the formation of soils with different characteristics even within a short distance in the landscape. In the BSR, formation of sodic Vertisols in MH alongside non-sodic Vertisols in ML positions is a unique phenomenon. The different types of soil formation has been possible due to micro-topographic variations which modify distribution of water across the landscape and facilitate deeper penetration of rainwater in ML position (Vaidya and Pal, 2002). In the IGP, on the other hand, non-sodic and moderately sodic soils (Alfisols) in the MH and highly sodic soils (Alfisols) in the ML position are formed. The main soil-forming processes have been clay illuviation, deposition of calcium carbonate and concomitant development of sodicity. The ML positions are repeatedly flooded with surface water in IGP during the brief and high intensity showers. These soils are thus subject to cycles of wetting and drying, leading to (i) steady supply of alkalis (Ca and Na ions) by hydrolysis of feldspars, (ii) precipitation of $CaCO_3$ at high pH, and (iii) development of highly sodic soils throughout the depth. Such events of flooding and the associated effects are minimal in the MH position, resulting in the formation of non-sodic and/or moderately sodic soils (Pal *et al.*, 2003b).

4.6. Time and Soil Formation

The length of time required for a soil to develop the distinct layers called the 'genetic horizons' depends on many interrelated factors of climate, nature of parent material, organisms, and topography. Horizons tend to develop rapidly under the warm, humid and forested conditions when there is adequate water to move colloids. Acid sandy loams lying on the sloping topography appear to be the soils most conducive for a rapid soil profile development.

Under ideal conditions, a recognizable soil profile may develop within 200 years. Under less favourable circumstances, the time may be extended to several thousand years. Soil development proceeds at a rate determined by the combined effects of time and intensities of climate and biota (organisms), further modified by the effect of land relief (topography) on which the soil is situated and the kind of parent material from which it is developing.

Different surfaces of earth's lands have been exposed for different lengths of time. Some plateau soils have been exposed for hundreds of thousands of years. Glacial till surfaces are more recent but may still be a few hundred thousand years old. More recently, rivers have flooded and covered floodplains and valley bottoms with recent deposits; these land surfaces may be only a few years or decades old, and the soil development has just started.

The soil on different-aged surfaces have been forming for different lengths of time. Recent deposits have shown little soil development, whereas land surfaces exposed for thousands of years may have well-developed profiles that are quite different.

5. Interaction of Soil Forming Factors

The formation of soil is a diverse and complex process, with effects from the five major factors working in combination. For instance, soil with good drainage (topography or parent material) and mild temperature and high rainfall (climate) will probably support ample plant life (biota) because favourable drainage provides an aerated location for the plant roots. The plants, in turn, decompose, producing carbon dioxide (CO$_2$), which combines with water from rainfall to form carbonic acid (H$^+$ and HCO$_3^-$). The resulting acidity increases the solubility of parent materials; sodium, potassium, calcium and magnesium are dissolved and leached in the draining soil water, and the soil becomes more acidic. High rainfall also translocates some clays and organic colloids deeper into the B horizons.

The same climate (or clayey parent material) but different topography (a low lying valley' or depression instead of a gentle slope) might produce a waterlogged soil (which has poor drainage). The poor drainage results in stagnant water and non-aerated soil, often with the resultant poor plant growth and decreased rates of organic matter decomposition. The accumulating drainage waters also lead to the accumulation of dissolved salts. The horizons of these soils, developed from common factors but with differing topography, will become very different.

The regions of high rainfall and good drainage typically develop acid soils having upper horizons (A) of organic matter accumulation, and removal of colloids from the upper horizons and their accumulation in the deeper horizons (B). The soluble materials move deeper or completely below the rooting depth. If the soils have been developing for a long time (dozens of centuries), they tend to have high clay content, well differentiated horizons, acidity in wet climates, and salt accumulation in some soils of arid regions. The wet and cool forest soils become strongly acidic, accumulate slowly decomposing organic matter on their top, and exhibit extensive colloid translocation. The wet and warm climates promote a faster organic matter decomposition. In the semi-arid areas, vegetation is sparse, and carbonates often accumulate at shallow depths to form whitish carbonate zones. Accumulation of soluble salts is also common.

A parent material may not be highly altered during its exposure to the factors of soil formation because of many other conditions that retard the development of a soil profile. Some of these conditions are:

(i) Low rainfall (slow weathering, washing of little soluble material from soil) as in the Thar desert

(ii) Low relative humidity (little growth of micro-organisms such as algae, fungi, and lichens) as in the Thar desert

(iii) High lime or carbonate content of parent material such as limestone (keeps soil material less mobile)

(iv) Parent material that are mostly quartz sands, with low amounts of silt and clay (slow weathering, few colloids to move)

(v) High clay content (poor aeration, slow water movement)

(vi) Resistant parent rock materials, such as quartzite (slow weathering)

(vii) Very steep slopes (erosion removes soil as fast as the upper horizon develops; low water intake reduces leaching)

(viii) High water table (slight leaching, low weathering rate)

(ix) Cold temperature (slowing of all chemical processes and microbial activity)

(x) Constant accumulations of soil material by deposition (continuous accretion of new material on which soil development must begin afresh) as in floodplains

(xi) Severe wind or water erosion of soil material (exposes new material to begin afresh to develop a profile)

(xii) Mixing by animals (burrowing) and humans (tillage, digging) minimizes the net downward colloid movement, and

(xiii) Presence of substances toxic to plants, such as excess salts, heavy metals, or high concentration of herbicides and pesticides.

Conditions opposite to the above favour more rapid rates of soil development. For example, under nearly level topography (not eroding) in an area of high rainfall and warm humid climate, easily weatherable and porous parent materials would rapidly develop into a well differentiated soil profile.

6. Pedogenic or Soil Forming Processes

Coming into being of a well-formed soil from regolith or newly deposited parent material is the outcome of several processes which are collectively called pedogenic or soil forming processes. Though physical and chemical weathering discussed earlier continues to operate even during pedogenesis, there are several other processes involving addition, leaching, redistribution, neo-synthesis or reorganisation which lead to the development of a soil with distinct profile. Example of addition is the build-up of humus or salts; that of redistribution is movement of clay or lime from the upper part of a soil profile to lower depths; that of leaching is the removal from entire soil profile of soluble salts; and that of neo-synthesis is formation of organo-mineral/clay-humus or free sesqui-oxides. The development of soil structure is a reorganisation phenomenon. Some of these processes such as redistribution of lime or build-up of humus take place at a faster rate and reach a near equilibrium state in a few hundred to few thousand years, others like laterisation require a few thousands to millions of years to reach maturity.

An examination of a soil pit or a fresh roadcut indicates distinctive horizontal layers. These layers will not be found if a similar cut were to be made in unconsolidated materials recently deposited by a volcanic eruption or flowing water during floods. Obviously, significant changes are made as soils develop from the relatively unconsolidated materials. A study of soil formation (genesis) gives an indication of as to how these changes occur and why they can lead to the formation of so many different kinds of soils. Soil genesis is brought about by a series of processes as mentioned above, the most significant of which are:

(a) Weathering and organic matter breakdown, by which some constituents are modified or dissolved and others are synthesized,

(b) Translocation of inorganic and organic materials up and down the soil profile, the materials being moved mainly by water but also by soil or organisms, and

(c) Accumulation of soil materials in layers (horizons) approximately parallel to the soil surface.

The role of these major processes can be seen by following the changes that take place as soils form from relatively uniform parent material. When plants begin to grow and their residues are deposited on the surface of the parent materials, soil formation has truly begun.

Table 2. Some important pedogenic processes

Structure development	Grouping of individual particles (clay, silt and sand) together with humus and free sesquioxides into aggregates or peds of fairly distinctive size and shape
Humification	Transformation of raw organic matter into humus
Translocation of lime	Removal in solution of lime from the upper part of the profile and its partial or total accumulation in the lower part. The process leads to the formation of a *kankar* or calcic horizon
Leaching	Removal in solution of a constituent from soil, e.g. soluble salts
Salinization	Accumulation of soluble salts in soil
Clay migration or lessivage	Removal of clay, particularly of fine clay in suspension from the upper part of soil profile and its accumulation in the lower part
Braunification/Rubification/ Ferruginisation	Release of iron from primary minerals and their dispersal as coatings on soil particles or as complexes with organic matter/clay or as discrete aggregates to impart a brown to red colour to the soil
Laterization	Removal of silica from soil and accumulation of sesquioxides (goethite, gibbsite, etc.) with or without the formation of iron stone and concretions
Podzolization	Removal of iron and aluminium, often as complexes with humus, from the upper part and its deposition at some depth
Regur fomation	Formation of intensely dark colour complex of smectitic clay and humus. It is a dominant process in black cotton soils
Gleization	The reduction of iron under anaerobic (waterlogged) conditions with production of bluish to greenish grey colour with or without mottles or ferro-manganese concretions

Common soil forming processes are listed in Table 2. Several of these processes operate simultaneously though some may start acting sequentially. An example of the latter is translocation of clay within a profile which occurs after leaching of soluble salts. However, in the long-run, one or two of these processes dominate and lead to the development of a soil with a distinctive profile dictated by factors of soil formation described earlier.

7. References

Bear, F.E. (Ed) (1964) *Chemistry of the Soil*, van Norshtand Reinhold, New York, 515 p.

Beinroth , F.H. (1982) Some highly weathered soils of Puerto Rico, 1. Morphology, formation and classification. *Geoderma* **27**, 1-27.

Bhattacharyya, T., Pal, D.K. and Deshpande, S.B. (1993) Genesis and transformation of minerals in the formation of red (Alfisols) and black (Inceptisols and Vertisols) soils on Deccan basalt in the Western Ghats, India. *Journal of Soil Science* **44**, 159-171.

Bhattacharyya, T., Pal, D.K. and Srivastava, P. (1999) Role of zeolites in persistance of high altitude ferrigenous Alfisols of the Western Ghats, India. *Geoderma* **90**, 263-276.

Bhattacharyya, T., Pal, D.K. and Srivastava, P. (2000) Formation of gibbsite in presence of 2:1 minerals: An example from Ultisols of North-East India. *Clay Minerals* **35**, 827-840.

Bhattacharyya, T., Pal, D.K., Lal, S., Chandran, P. and Ray, S.K. (2006) Formation and persistence of Mollisols on zeolitic Deccan basalt of humid tropical India. *Geoderma* **136**, 609-620.

Buol, S.W. and Eswaran, H. (2000) Oxisols. *Advances in Agronomy* **68**, 151-195.

Buurman, P. and Velthorst, E.J. (Eds) (1996) *Manual of Soil and Water Analysis*, Back Huys Publishers, Leiden.

Chandran, P., Ray, S.K., Bhattacharyya, T., Srivastava, P., Krishnan, P. and Pal, D.K. (2005) Lateritic soils of Kerala, India: Their mineralogy, genesis, and taxonomy. *Australian Journal of Soil Science* **43**, 839-852.

Fanning, D.S. and Fanning, M.C.B. (1989) *Soil: Morphology, Genesis and Classification*. John Wiley and Sons, USA. 395 p.

Soil Classification

J.L. SEHGAL

1. Introduction

Classification is the grouping of objects in some orderly and logical manner. It is based on the properties of objects for the purpose of their identification and study. They are termed as differentiating characteristics as they differentiate and serve to separate one class from the others; for instance, soils are classified as sandy, loamy or clayey soils on the basis of their characteristics. For classifying the individuals of a large and widely varying population, such as soils, it is useful to group individuals into classes, and further into higher classes. This kind of grouping is called a multi-categoric or hierarchical system of classification. The individual soils are grouped into classes of lower category (e.g., *soil series*), which are further grouped into classes of higher categories (e.g., *soil orders*). The lower categories are defined by a large number of differentiating characteristics and higher categories by a few differentiating characteristics. Within each class, there is a central core or nucleus to which the individual members are related in varying degrees. It is called the central concept or an idealized individual which typifies the class.

1.1. Purposes of Soil Classification

Like the flora and fauna, soils are classified in a systematic manner, to remember their properties and understand their relationships. The purpose of a classification is to:

- Organise knowledge leading to economy of thoughts
- Recognize properties of the objects
- Learn new relationships and principles in the population being classified
- Establish groups or subdivisions (classes) of the objects under study in a manner useful for practical and applied purposes in:
  - predicting their behaviour,
  - identifying their potential uses,
  - estimating their productivity, and
  - transferring agro-technology from research farms to cultivators' fields.

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2. Evolution of Soil Classification Systems

2.1. Early Systems

The early systems of soil classification were quite simple and practical; their aim was utilitarian; for instance:

- *Economic Classification*: It is the grouping of soils based on their productivity for the purpose of taxation.
- *Physical Classification*: It is the grouping of soils based on their texture — a property of soil closely associated with soil productivity and management, for instance, loamy, sandy and clayey soils.
- *Chemical Classification*: Grouping of soils based on the composition of soil having a bearing on their chemical characteristics, e.g. acidic, alkaline, calcareous, gypsiferrous soils, etc.
- *Geological Classification*: Grouping of soils based on the nature of underlying par-

ent rock/parent material, e.g., basalt, limestone, sandstone, etc. or transported material like alluvium, aeolian material.

- *Physiographic Classification*: Grouping of soils based on the characteristics of landscape, for instance, levee, basin, terrace, mountain, valley, upland and lowland soils, etc.

Since the above systems were based on a single character, their utility was limited. Hence, the need for a more comprehensive system was felt.

2.2. Dokuchaiev's Genetic System

In the later part of the 19th century, Dokuchaiev, working in Central Russian Upland, observed that a rather uniform loess parent material extended for hundreds of kilometres with increasing temperature gradient from north to south and an increasing rainfall and moisture gradient from east to west. These differences in climatic conditions were associated with important vegetation patterns, varying from forest to steppe (prairie), which left their imprint on the parent material, producing distinct soil differences. Such observations led Dokuchaiev– the founder of the modern pedology– to establish the concept of soil as an independent natural body and develop a series of publications on soil genesis and classification. The Russian approach tends to emphasize on soil genesis, and hence is termed as *Genetic System* of soil classification.

Dokuchaiev (1900) divided soils into three categories; *Normal*, *Transitional* and *Abnormal*. These categories were later termed as *Zonal*, *Intrazonal* and *Azonal soils*, respectively.

Zonality Concept — The soils that have fully developed soil profiles, and are in equilibrium with the environmental conditions, such as climate and vegetation, are termed as *Zonal Soils*, for instance, Sierozem, Chestnut, Podzol and Laterites. The soils formed in regions, where time has been a limiting factor to produce fully-developed horizons are termed as *Azonal Soils*, for instance, Alluvial soils and Regosols. Still others, occurring within the zonal areas and having characteristics that are determined largely by the local conditions, like topography, parent material, etc. are termed as *Intrazonal Soils*, for instance, Calcimorphic and Hydromorphic Soils.

2.3. Marbut's Morpho-Genetic System

Although the Dokuchaiev's (Russian) approach was based on sound principles of soil genesis, it had an inherent weakness of laying undue emphasis on climate and vegetation rather than on the intrinsic properties of the soils. Marbut in the USA was greatly influenced by the Dokuchaiev's approach. He accepted the concept of the Russian *Soil Type*, but gave it the name of Great Soil Groups. Marbut (1927) was the first to advocate classification of soils on the basis of their intrinsic properties rather than on soil-forming factors, thus reducing emphasis on geology or parent rock. Marbut evolved his scheme in successive steps and published it in the *Atlas of American Agriculture*. It was based on the iron-alumina and lime contents. At the highest category level, he divided *Zonal Soils* into two classes: *Pedalfers* and *Pedocals*, the former showing accumulation of iron and aluminium oxides and the latter of calcium as calcium carbonate. The Pedalfers were presumed to occur in the areas of high rainfall having real surplus of water for leaching. The Pedocals were supposed to occur in areas of high evaporation having real deficit of water.

2.4. Baldwin and Associates Genetic System

The major limitation of Marbut's system was that it was based, in part, on assumptions concerning soil genesis. As such, many of the soil series, recognized in the USA, could not find a place in the system. Hence, the Morphogenetic System of Marbut (1935) was revised and elaborated by Baldwin *et al.* (1938) and Kellogg and Thorp (1949) (Table 1). The system marked the beginning of a comprehensive approach. The salient features of this system are:

- A return to the zonality concept of Russian School,
- The pedocal-pedalfer concept was de-emphasized, and

Table 1. Genetic system of soil classification

Order	Suborder	Great Soil Groups
Zonal soils (6)	1. Soils of the cold zone	Tundra soils
	2. Light-colour soils of arid regions	Sierozem soils Brown soils Reddish Brown soils Desert soils Red Desert soils
LLL, CDS	3. Dark-colour soils of the semi-arid subhumid, and humid grasslands	Chestnut soils Reddish Chestnut soils Chernozem soils Prairie or Brunizem soils Reddish Prairie soils
	4. Soils of the forest grassland transition	Degraded Chernozem soils Noncalcic Brown soils
	5. Light-colour podzolized soils of the timbered regions	Podzol soils Gray Wooded, or Gray Podzolic soils Brown Podzolic soils Soil-brun acide Gray-brown Podzolic soils Red-yellow Podzolic soils
	6. Lateritic soils of forested warm-temperate and tropical regions	Reddish-brown Lateritic soils Yellowish-brown Lateritic soils Laterite soils
Intra-zonal soils (3)	1. Halomorphic (saline and alkali) soils of imperfectly drained arid regions and littoral deposits	Solonchak or Saline soils Solonetz soils (partly-leached Solonchak) Soloth soils
Lithoral deposit,	2. Hydromorphic soils of marshes, swamps seep areas, and flats	Humic-gley soils Alpine Meadow soils Bog and Half-bog soils Low Humid-Gley soils Planosols Groundwater Podzol soils Groundwater Laterite soils
Lithosole	3. Calcimorphic soils	Brown Forest soils Rendzina soils
Azonal soils (3)	No suborder	LAR Lithosols Alluvial soils Regosols

Source: Baldwin et al. (1938), as modified by Thorp and Smith (1949)

- More emphasis was laid on soil as a three-dimensional body with its own characteristics.

= Group-[illegible]

A new category, viz. *Soil Family*, was introduced between Great Soil Group and Soil Series; but neither soil families nor the higher categories were defined in terms of soil properties.

Serious problems arose when some soil series did not fit in any of the existing Great Soil Groups. Therefore, as a stop-gap strategy, the system was revised in which three new Great

Soil Groups were introduced and three others were merged with the existing Great Soil Groups. The soils were grouped in three Orders, viz. *Zonal*, *Intrazonal* and *Azonal*, following the Russians zonality concept, as under:

Zonal Soils — The soils whose characteristics are determined primarily by the environment, especially climate and vegetation (Figure 1).

Intrazonal Soils — These soils occur within a zone, but reflect the influence of some local conditions, such as topography and/or parent material.

Azonal Soils — The soils that have poorly developed profiles because of time as a limiting factor, e.g. young soils without horizon differentiation. Baldwin classification.

The three orders were further subdivided into nine suborders on the basis of specific climatic and vegetative regions. Each suborder, in turn, was divided into Great Soil Groups, which are an expression of more specific conditions. The Great Soil Groups were further subdivided into numerous Soil Families, Series and Soil Types.

2.5. *Limitations of the Genetic Systems*

The major limitations of the Genetic Systems are:

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Limitations

(i) The two highest categories are defined in genetic terms and not on the basis of properties of the soils.

(ii) The concepts and definitions of the highest category, i.e. the Order, in terms of soil properties, are not clear.

(iii) The Great Soil Group concepts and definitions are based on environmental factors, rather than on the soil properties. Hence, their definitions are comparative and qualitative.

(iv) Many of the soils are defined in terms of properties that were obvious under virgin soil conditions and are destroyed during cultivation and hence the classification of such arable soils becomes ambiguous.

(v) There are tendencies to attempt definition of units in the lower categories in terms of too few properties thought to be important for one interpretation.

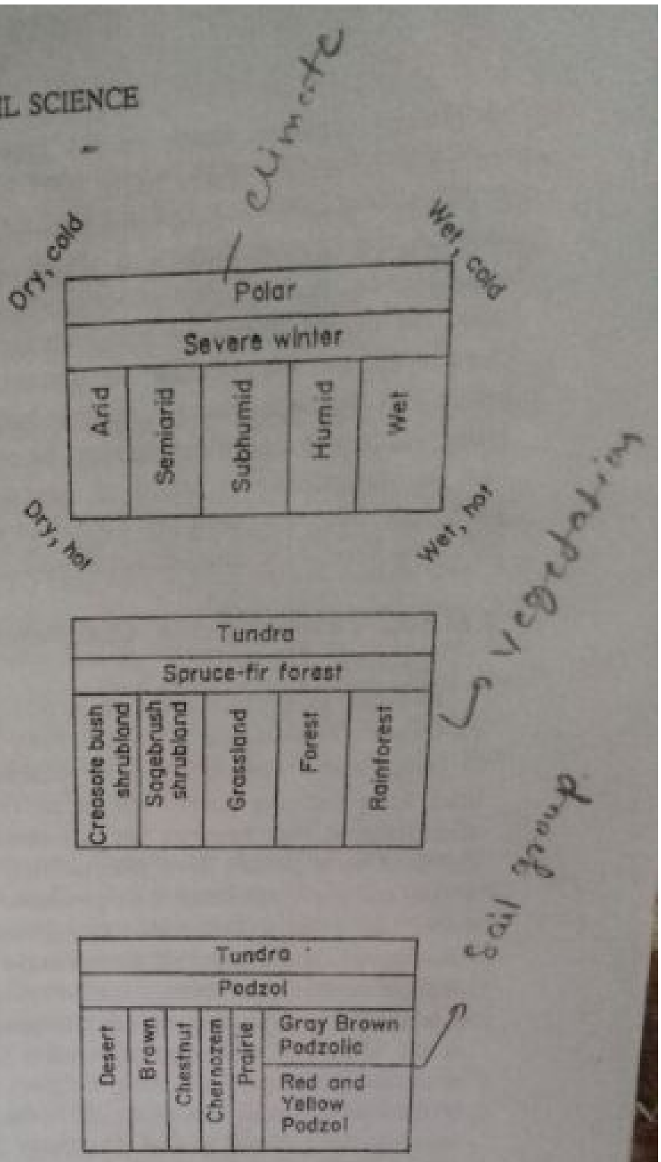


Figure 1. A schematic relationship of climate, vegetation and soils (top: climate types; middle: vegetation types; bottom: soil groups)

Source: Sehgal, 1996

(vi) The nomenclature in the highest categories laid too much emphasis on colour or vegetation rather than on the salient properties of the soils.

(vii) The nomenclature was evolved from several languages, and with the mixtures of nouns and adjectives; it was difficult to name the intergrades.

According to Kellogg (1963), most systems of soil classification, as used in 1950s, including that of the USA, had one or more serious limitations because most of these over-emphasized the virgin soils. The conclusion is that different schools of thought inevitably needed some compromise in their divergent opinions to meet the important need to classify arable soils (of which at least the upper horizons had been mixed); the alluvial soils (which have no well-developed soil horizons) did not find a suitable place in the system having a strong genetic bias. Hence, a desirable system should be based on combinations of soil characteristics known to be significant to genesis and behaviour, but not directly on the either. In other words, the classification must be one that can be interpreted in terms of genesis and behaviour, but the genesis and behaviour should be a step behind the classification itself.

2.6. *Soil Taxonomy : A Comprehensive System*

In order to overcome the shortcomings of the Genetic System, the US Soil Survey Staff, in cooperation with many other institutions, have been working since 1951 to arrive at a classification that narrows the differences in different view points. Each year, starting from 1953, an approximation was produced and tested. In 1960, a Comprehensive System of Soil Classification, popularly known as the "*7th Approximation*" (Soil Survey Staff, 1960), was published. The system was put into official use in the USA and was adopted in many other countries, including India, Iraq, Belgium, The Netherlands, etc. Ultimately in 1975, the system was brought out as Soil Taxonomy (Soil Survey Staff, 1975); its second edition appeared in 1999. The Soil Taxonomy was designed to serve the needs of soil survey.

2.6.1. Salient Features

The Comprehensive System is a morphogenetic system in which morphology of soil, that is an outcome of soil genesis, serves as a guide (Smith, 1963). It is based on the properties of soils as they exist today. Although one of its objectives is to group soils similar in genesis, the specific criteria used to place soils in different groups are those of soil properties. The system has an edge over the earlier systems in the following respects:

- Unlike the Genetic Systems, the Comprehensive System is based on measurable soil properties that exist today.
- It considers all such properties which affect soil genesis or are the outcome of soil genesis.
- The common definition of a class of taxonomic system is type or orthotype.
- The nomenclature, using coined words, is derived mainly from *Greek* and *Latin* languages. Although it appears difficult, once understood, it is the most logical nomenclature and helps in relating the place of taxon in the system and in making interpretations.
- A new category, viz. *Subgroup*, has been introduced to define the central concepts of Great Groups and their intergrades in order to express and recognise more clearly that soils are in continuum and show gradual change in many properties.
- Unlike the Genetic System, it is an orderly scheme without prejudices, but facilitates easy recognition of the objects.

2.6.2. Diagnostic Horizons

Diagnostic horizons are understood to reflect genetic horizons widely occurring in soils, which fairly well describe and define soil classes. Thus, a *diagnostic horizon* is defined as one, formed through pedogenic processes and having distinct properties or features that can be described in terms of measurable soil properties. In fact, a set of pedogenic processes operate in a soil for a significant period to leave their mark in the form of distinct horizons or features, which can be measured. The diagnostic horizons are largely used not only for identifying soils but also in classifying them at various categoric levels, especially Great Groups. A number of diagnostic horizons have been defined in Soil Taxonomy. Based on their locations in soil profiles, these horizons are of two types, viz. surface and subsurface.

The diagnostic surface horizons are called ***epipedons*** (Greek *epi*, over, upon and *pedon*, soil). The epipedons are simply the upper most soil horizons and include the upper part of the soil darkened by organic matter. They are not synonymous with A-horizon. Nine epipedons, viz. folistic, histic, melanic, mollic, anthropic, umbric, ochric, plaggen and grossarenic are recognised, but generally speaking, three of these, viz. *mollic*, *ochric* and *umbric*, are of importance in India.

epipedm = 9. Mam, PG, F. U. O, (9)

Endopedm = Gianga, PDF, SSSSC, CO, TS, K,K(19)

The diagnostic subsurface horizons are called ***endopedons*** (Greek *endodermis*, subsurface or deep-seated and *pedon*, soil). The endopedon includes the lower part of the soil where soil materials accumulate. Nineteen endopedons, viz. argillic, natric, agric, spodic, sombric, cambic, kandic, oxic, sulphuric, salic, placic, albic, glossic, calcic, gypsic, duripan, fragipan, petrocalcic and petrogypsic are recognised. Of these, eight, viz. argillic, natric, cambic, kandic, oxic, salic, calcic and gypsic are commonly observed in India. The major features of the diagnostic epipedons and endopedons are briefly given as under:

2.6.3. Diganostic Surface Horizons (Epipedons)

The nine epipeons are described below:

Folistic Epipedon – A surface layer that is never saturated with water for more than 30 days (cumulative) in normal years (and is not artificially drained) and either consists of organic soil material (containing 75% or more, by volume, of *Sphagnum* fibres) or has a bulk density (moist) of less than 0.1 Mg/m^3 or is 15 cm or more thick; *or* is an Ap horizon (up to 25 cm) that has an organic carbon content (by weight) of 16% or more if the mineral fraction contains 60% or more clay or 8% or more with no clay. Most folistic epipedons comprise organic soil materials that remain saturated for less than 1 month. (30 days).

O.C By Value 75%

By weight 16%

Histic Epipedon – A thin organic horizon (if virgin) or horizon with very high organic matter content depending on clay content, that remains saturated with water for 30 days or more (cumulative) during some season of the year, unless artificially drained. It is thinner than 30 cm if drained, or 45 cm if not drained. Note the differences in Folistic and Histic epipedons, especially in terms of saturation with water.

Melanic Epipedon – A thick black horizon at or near, but within 30 cm of the soil surface and having andic soil properties. It contains 6% or more organic carbon as weighted average, and 4% or more organic carbon in all layers. The deep dark colour is due to the accummulation of organic matter resulting from root residues, supplied by graminace-eous vegetation, result in colour value (moist) and chroma of two or less throughout.

Mollic Epipedon – A thick, dark colour, soft mineral horizon with high (> 50%) base saturation and strong structure. It contains 1% or more organic matter (when mixed to a depth of 17.5 cm) with colour values darker than 3.0 (moist) and 5.0 (dry) and chroma less than 3.5 (moist). Soil structure cannot be massive and hard, very hard or extremely hard when dry. Base saturation is over 50% and the epipedon is not naturally dry in all parts for more than nine months in a year.

Anthropic Epipedon – A surface horizon, like the mollic, but formed under long continued system of farming that involves large additions of organic matter (compost). It contains 1500 mg or more of citric acid (1%) soluble P_2O_5 per kg soil, which distinguishes it from mollic and umbric epipedons.

Umbric Epipedon – A surface horizon like the mollic, but is low (< 50%) in base saturation (dominantly saturated with H$^+$) with high C : N ratio and is not naturally dry for more than three months in a year.

Ochric Epipedon – A surface horizon that is light in colour [colour values > 3.5 (moist) or > 5.5 (dry)]. It contains less than 1% organic matter, or is hard or very hard and massive when dry; or remains dry for more than three months in a year.

Dry = ocolur = ochric

Plaggen Epipedon – A thick (> 50 cm) man-made surface horizon, produced by a long and continued manuring with sod. It can be easily identified by artefacts, such as bits of brick, pottery, etc. It also shows spade marks.

Phospho
↓
plagg

Grossarenic Epipedon – A sandy (loamy fine sand or coarser) horizon, 100 cm or more thick over an argillic horizon.

2.6.4. *Diagnostic Subsurface Horizons (Endopedons)*

The characteristics of 19 endopedons are discribed below:

Argillic Horizon – A silicate-clay-enriched horizon formed by illuviation of clay. The fine clay is carried downward by percolating water and is deposited as clay skins or cutans on ped faces and on the walls of pores. In general, a B-horizon that has at least 1.2-times as much clay content as some horizon above, or 3% more clay content if the eluvial (bleached) layer has < 15% clay, or 8% more clay if eluvial layer has > 40% clay. It should be at least 1/10 of all overlying horizons or more than 15 cm thick, whichever is thinner.

Natric Horizon[a] – A high sodium-clay-enriched horizon with columnar or prismatic structure. The horizon meets all the requirements of an argillic but in addition, has 15% or more of the exchange complex saturated with Na^+, or has more exchangeable Mg^{2+} plus Na^+ than Ca^{2+} plus exchange acidity at pH 8.2. 8. 8.5,

Agric Horizon – An illuvial horizon of clay, silt and humus formed directly under the plough layer due to long and continued cultivation.

Spodic Horizon – A humus- and/or sesquioxides-enriched subsurface horizon with or without iron. There are many specific limits dealing with Al, Fe, organic matter and clay ratio, depending on whether the overlying horizon is virgin or cultivated. It is generally formed in cold humid regions. A spodic horizon must have 85% or more of spodic materials in a layer 2.5 cm or more thick that is not a part of any Ap-horizon. Such a horizon is rarely observed in India. The so-called Podsols, as reported to occur in the Himalayas, do not meet the requirements of spodic horizon as laid down in Soil Taxonomy.

Sombric Horizon – A free-draining horizon located not under an albic horizon; has colours (darkness) and base status like an umbric epipedon, and has been formed due to illuviation of humus and not of aluminium or sodium.

al = 8cm

under

Some = humous,

Cambic Horizon – A colour or structural B-horizon formed due to alteration by the physical movement or chemical weathering. The pedogenic processes have altered the material enough to form soil structure (if the texture is suitable, i.e. very fine sand, loamy very fine sand or finer) to liberate free iron oxides to form silicate clays, and to obliterate most evidences of the original rock structure in more than 50% of the volume. The alteration is not enough to completely destroy volcanic glass, allophane, glass, feldspars and micas. Similarly, illuviation of oxides, humus or clay should not be enough that will qualify it for *argillic* or *spodic* horizon. The horizon is extremely variable in mineralogy because of its pedogenic youthfulness, occurs under widely differing environment and may develop in the presence or absence of fluctuating groundwater.

Kandic Horizon – A subsurface horizon of low activity clays with or without clay-skins. It has CEC of < 16 cmol (p^+) kg^{-1} soil at pH 7, and effective CEC (ECEC) of < 12 cmol (p^+) kg^{-1} soil. It shows a clay content increase at its upper boundary of > 1.2 × clay within a vertical distance of < 15 cm, that is, abrupt or clear textural boundary. It shows no stratification.

Oxic Horizon – A horizon enriched with Fe- and Al-oxides with dominance of 1:1 type clay minerals and from where silica has leached. It is atleast 30 cm thick and is sandy loam or finer in texture. It has dominance of low charge clays (1:1) type, e.g. kaolinite with CEC of 16 or less cmol (p^+) kg^{-1} at pH 7 and apparent ECEC of 12 or less cmol (p^+) kg^{-1} clay. The increase in clay content is gradual than in Kandic horizon. It contains less than 10% weatherable minerals in the sand fraction.

Sulphuric Horizon – A mineral or organic soil horizon that has a pH of < 3.5, is toxic to plant roots, and has yellow mottles of jarosite.

Salic Horizon – A horizon with secondary accumulation of water soluble salts (NaCl, Na_2SO_4, etc.) at some depth in the soil profile. The horizon is atleast 15 cm thick with secondary soluble salt enrichment of over 2%, and

$<\frac{1}{3}$, $\frac{1}{3}$-$\frac{2}{3}$, $\frac{2}{3}<$,

the product of thickness and salt percentage is 60 or more. The salt content, in terms of electrical conductivity (EC) in a saturated paste should be 30 dS/m or more, and the product of EC (dS/m) and thickness (cm) is equal to 900 or more.

Albic Horizon – A bleached E-horizon of podzols and planosols. It has typical colour values of > 5 (dry) or > 4 (moist).

Glossic Horizon (Gr. *Glossa*, tongue) – Glossic horizon is a horizon which shows albic horizon characteristics gradually intruding into an argillic, a kandic or a natric horizon. It may be 5 cm or more thick and consists of an eluvial part which constitutes 15 to 85% (by volume) of the glossic horizon and an illuvial part (i.e. partially-altered or degraded argillic, kandic or natric horizon).

Calcic Horizon – A horizon with secondary Ca- and/or Mg-carbonate-enriched materials. It is 15 cm or more thick, has 15% or more of secondary accumulation of carbonates and contains at least 5% more carbonates than any underlying horizon.

Gypsic Horizon – Calcium and/or magnesium sulphate-enriched horizon. It is more than 15 cm thick, and contains at least 5% more calcium sulphate than the underlying horizon/layer.

Petrocalcic Horizon – An indurated calcic horizon that has hardness of 3 or more (Moh's scale) and whose one-half or more of dry fragments break down in acid, but not in water.

Petrogypsic Horizon – A strongly cemented gypsic horizon whose dry fragments do not slake in water.

Placic Horizon – A thin (2-10 mm thick), slowly-permeable, dark reddish brown to black colour iron or manganese pan that lies within 50 cm of the surface.

These are not the complete definitions, but are highly abbreviated versions. Many of these are still being discussed and revised. The basic meaning attached to each term (given above) remains the same.

2.6.5. *Diagnostic Organic Materials*

Different diagnostic organic materials are described below:

Fibric Soil Material (formerly peat) – The fibrous material in an unrubbed condition, constituting over 2/3 of the mass, and yields almost clear solution when extracted with sodium pyrophosphate.

Hemic Soil Material (formerly mucky peat or peaty muck) – The fibrous material in an unrubbed condition constituting 1/3 to 2/3 of the mass, in an intermediate stage of decomposition.

Sapric Soil Material (formerly muck) – The identifiable fibrous material in an unrubbed condition, constituting less than 1/3 of the mass; it produces sodium pyrophosphate extracts with colours lower in value and higher in chroma than 10YR 7/3.

↳ Sompric soil,

Humilluvic Material – Illuvial humus that accumulates after prolonged cultivation of some acid organic soils.

Limnic Materials – Organic or inorganic materials deposited in water by the action of aquatic organisms, or derived from underwater and floating organisms. For instance, marl, diatomaceous earth, and sedimentary peat (coprogenous earth).

2.6.6. *Diagnostic Characteristics for Both Mineral and Organic Soils*

Aquic Conditions – Soils with aquic (Latin *aqua*, water) conditions are those that are continuously or periodically saturated with water and undergo reduction. For details, see under soil moisture regimes.

Cryoturbation (Frost Churning) – It is the mixing of soil matrix within a pedon that results in irregular or broken horizons, accumulation of organic matter over the permafrost table, silt capping on rock fragments and oriented rock fragments.

Densic Contact (L. *densus*, thick) – A thick contact between soil and densic material(s) below. It has no cracks.

Not cemented

Densic Materials – These are relatively unaltered materials that do not meet the requirements for any other named diagnostic horizons that have a non-cemented, rupture-resistance class. It has bulk density that does not allow root penetration, except through cracks. These are dense earthy materials, such as mine spoils, volcanic mud flows till. Being non-cemented, these differ from paralithic materials which are cemented.

cemented.

Bot

Gelic Materials – These are mineral or organic soil materials that show evidence of cryoturbation (frost churning), which is manifested by irregular and broken horizons, accumulation of organic matter generally on top of the permafrost and silt-enriched layers. The structure is platy, blocky and/or granular.

2.6.7.Other Diagnostic Soil Characteristics

Several other layers or horizons and macrofeatures are recognized and named in Soil Taxonomy (Soil Survey Staff, 1999). These features are usually used as diagnostic criteria at a somewhat lower level in the classification system than the subsurface diagnostic horizon (endopedons) and epipedons.

HAT

Abrupt Textural Change – It refers to a boundary of a considerable increase in the clay content within a short vertical distance between a surface (ochric epipedon or albic horizon) and an underlying subsoil (argillic horizon). It can be established using the following criteria: If the surface material has < 20% clay, clay content doubles within 7.5 cm; if the surface (ochric epipedon) horizon has 20% or more clay, an absolute increase of at least 20% clay content is required within 7.5 cm.

Durinodes – These are weakly-cemented to indurated nodules, cemented by SiO_2.

horizon.

Duripan – It is a subsurface horizon at least half cemented by SiO_2. The air-dry peds do not slake in water or HCl, but are destroyed by hot KOH after acid washing.

layer,

Fragipan – It is a subsoil layer(s) of high bulk density. It is brittle when moist, and very hard when dry. It does not soften on wetting, but can be broken in the hands. The air dry fragments slake in water.

Low Chroma Mottles – These are moist soil colours of two or less chroma; 4 or more value often represent gley conditions.

Vg vegetable vg.

Permafrost – It is a layer where soil temperature is always < 0 °C. It may either be very hard or loose.

PG

Plinthite (Greek, *Plinthose*, brick) – It is a humus-poor, sesquioxide-rich horizon, which hardens irreversibly to ironstone hardpans or aggregates with repeated wetting and drying. The red, indurating portions of the layer, are usually mottled with yellowish, greyish or white materials.

Soft Powdery Lime – It is authigenic lime translocated within the soil, normally present as coating on the ped surfaces.

b → albic

Tonguing – It is used when albic horizon material (of at least 5 cm deep and 5 mm wide) penetrates into an underlying argillic or natric horizon (*see* glossic horizon).

2.6.8.Diagnostic Contacts (to "Non-Soil" Material)

Lithic Contact – A boundary between soil and continuous coherent, underlying material that has hardness of > 3 on the Moh's scale and through which roots cannot penetrate.

PH

Paralithic (Lithic-like) Contact – A boundary between soil and continuous non-coherent, underlying material that has hardness of > 3 on the Moh's scale. The roots can penetrate, at irregular and infrequent intervals, to 10 cm or more.

Petroferric Contact – A boundary between soil and an indurated layer of iron cemented material.

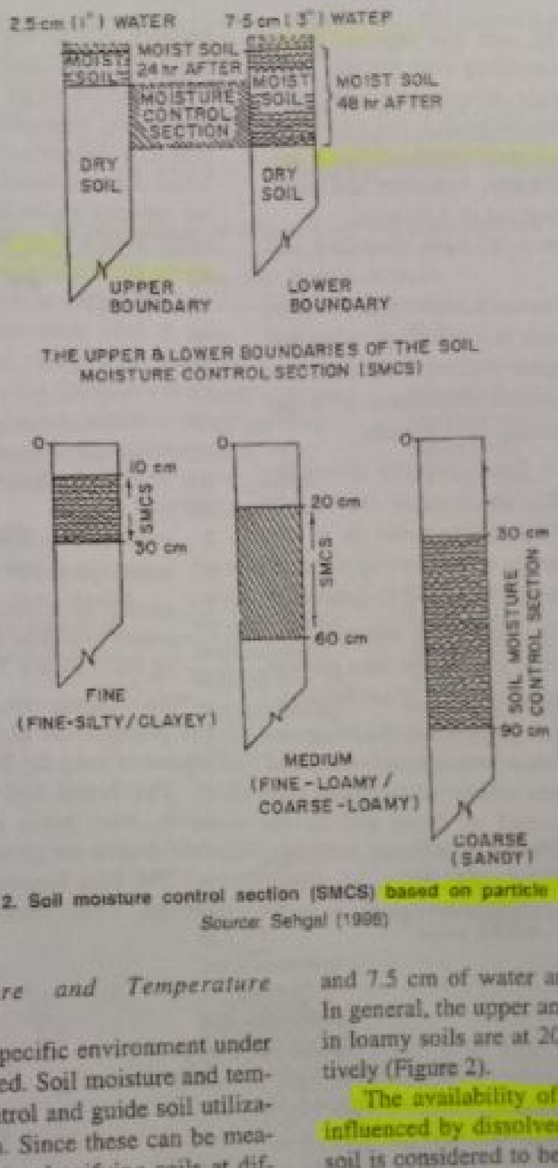


Figure 2. Soil moisture control section (SMCS) based on particle size class
Source: Sehgal (1996)

2.6.9. Soil Moisture and Temperature Regimes

Each soil has a specific environment under which it has developed. Soil moisture and temperature regimes control and guide soil utilization for plant growth. Since these can be measured, they are used in classifying soils at different categoric levels.

Soil Moisture Regimes (SMR) – Soil moisture regime (*SMR*) refers to the presence or absense of water in a soil at different times of a year. Soil is considered moist when it is at moisture tension of less than 1500 kPa (15-bar) and dry when the tension is 1500 kPa (15-bar) or more within the *Soil Moisture Control Section* (SMCS). The limits of SMCS are determined by the soil depth to which the soil at wilting point is moistened when 2.5 cm and 7.5 cm of water are added at the surface. In general, the upper and lower limits of SMCS in loamy soils are at 20 cm and 60 cm, respectively (Figure 2).

The availability of water to plants is also influenced by dissolved salts. Consequently, a soil is considered to be dry when it is at moisture tension of 1500 kPa (15-bar) or more, or when salt concentration reaches the level that limits moisture availability to plants. Under such conditions, the soil is considered to be physiologically dry.

The SMR is a partial function of climate, soil and landform. It is defined in terms of the number of days the SMCS remains moist with moisture tensions between 33 kPa (1/3-bar) and 1500 kPa (15-bar). The SMCS, considered important for moisture supply to crops, is defined as the depth limits of soil that regulate the moisture supply, and this depends on its texture. Besides controlling crop growth, SMR is important not only in understanding pedogenesis and nutrient availability, but also in the classification of soil at different categoric levels, such as Soil Family, Suborder and occasionally at Order level (as in Aridisols).

In nature, there exist three dominant soil moisture regimes:

(i) ***Saturated*** – Taxonomically characterized as AQUIC, which is a condition when soil pores are completely filled with water, resulting in anaerobic conditions. It is not conducive to normal crop growth.

(ii) ***Non-leaching*** – Taxonomically characterized as ARIDIC, where water moves into SMCS for a very short period in a year and gets completely withdrawn by high potential evapotranspirative (PET) demand.

(iii) ***Leaching*** – Taxonomically characterized as UDIC, where water moves into the soil almost throughout the year (if not frozen).

In between leaching and non-leaching environments, there exist two partially-leaching soil moisture regimes which, according to Soil Taxonomy, are termed as *Ustic* and *Xeric*. The former is indicative of summer monsoon rains (as in India) and the latter, of winter rains (as in the Mediterranean region). In these regimes, farmer can grow one crop a year (in summer or in winter).

A knowledge about the prevailing bioclimate, quantified by precipitation (P) and potential evapotranspiration (PET), and soilscape characteristics, in terms of available water capacity (AWC), are helpful in characterizing SMRs. The model runs as the SMCS starts moistening when P exceeds 0.5 PET and the supply of moisture to SMCS continues till P falls short of 0.5 PET. Thereafter, the water need is met from storage (if stored water is available) and continues till about 80% of the available stored moisture is exhausted by evapotranspiration. Thus, the number of days the SMCS remains moist is worked out from the total sum of:

a. moist period (PET < P or P > 0.5 PET);
b. humid period (P > PET);
c. period to evaporate more than 80% of stored available moisture after P falls short of PET (Figure 3).

As a thumb rule, any month with an average precipitation of more than 50 mm is considered to keep the SMCS partially moist.

The broad soil moisture regimes, viz. *aquic*, *udic*, *ustic*, *xeric*, *aridic*, *torric*, and their criteria are given in Table 2.

The Soil Moisture Regime Map of India (Sehgal, 1996) shows that India is dominantly

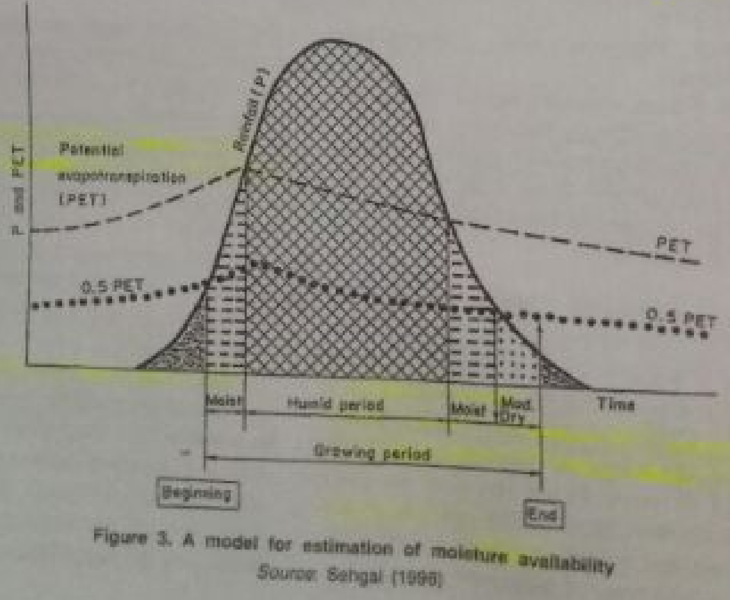


Figure 3. A model for estimation of moisture availability
Source: Sehgal (1996)

represented by *ustic* soil moisture regime covering almost three-fourth (72%), followed by *aridic* (16%) and *udic* (12%). It suggests potential need for additional irrigation water, in areas representing *aridic* and *ustic* SMRs for supporting one to two crops in a year which is most crucial for increasing agricultural production.

Soil Temperature Regimes (STR) – These are the ranges in temperature classes within which biological activity of different degrees prevails.

Table 2. Class criteria of different soil moisture regimes

Gleying/ Aquic	Moist soils throughout Udic	Soils with limited moisture		Soils with negligible moisture Aridic
		Ustic	Xeric	
Saturated for some time in a a year. Reduced soils because of lack of dissolved oxygen Persquic Moisture regime in a close, land locked depression with perennial streams	Not dry any part for as long as 90 cumulative days (to exclude from Ustic) Not dry in all parts for for as long as 45 consecutive days following summer solstice* (to exclude from Xeric)	If MAST > 22 °C or MSST and MWST differ by < 5 °C < 5 °C (tropical condition), then Dry in some or all parts of 90 cumulative days. But the SMCS is moist in some parts for 180 days (cumulative) or continuously moist in some parts for atleast 90 consecutive days. If MAST > 22 °C and MSST and MWST differ by ≥ 5 °C. Dry in some or all parts for ≥ 90 (cumulative) days. But is not dry in all parts for > 6 months (the Aridic and Torric regimes). It is not dry in all parts for 45 consecutive days within 4 months following summer solstice. It is not moist in all parts for 45 consecutive days or more in 4 months following winter solstice (Xeric)	Associated with Mediterranean climate MAST < 22 °C, MSST, MWST differ by > 5 °C, the SMCS is: Dry in all parts for 45 or more consecutive days within 4 months following summer solstice in 6 or more years out of 10 years. 6/10 Moist in all parts for ≥45 consecutive days following winter solstice	Dry throughout for more than 180 days (cumulative), the soil temperature is > 5 °C Never moist in some or all parts for as long as 90 consecutive days when the soil temperature is above 8 °C. No moisture in all parts for as long as 45 consecutive days during 4 months following the winter solstice (if MSST and MWST differ by > 5 °C and MAST is < 22 °C (i.e. to exclude from Xeric)

90+180+90
3+6+3

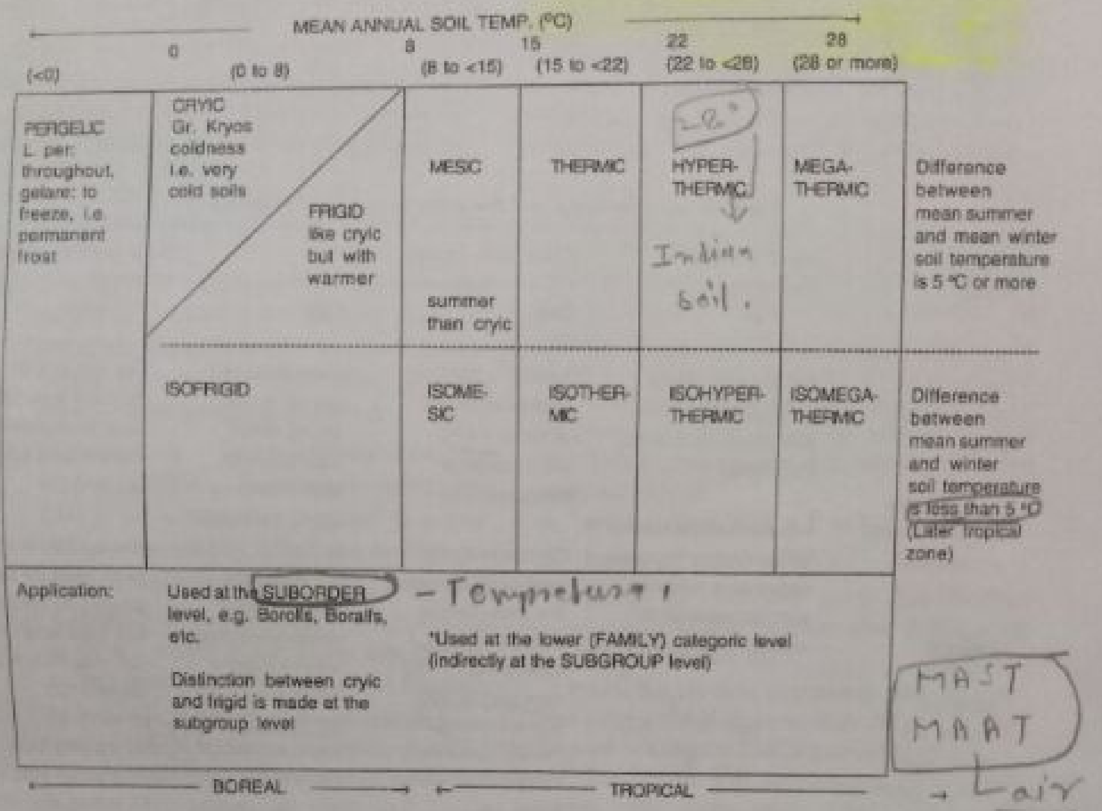
Tropical MAST - MWST difference
< 5°C,

90+180
3×6×

* from 22nd June onwards; * from 23rd December onwards

day [illegible] more
moist is short

MAST = Mean Annual Soil Temperature
MSST = Mean Summer Soil Temperature
MWST = Mean Winter Soil Temperature
SMCS = Soil Moisture Control Section



(<0)	0 (0 to 8)	8 (8 to <15)	15 (15 to <22)	22 (22 to <28)	28 (28 or more)	
PERGELIC L. per: throughout, gelare: to freeze, i.e. permanent frost	CRYIC Gr. Kryos coldness i.e. very cold soils FRIGID like cryic but with warmer summer than cryic	MESIC	THERMIC	~28° HYPER-THERMIC Indian soil.	MEGA-THERMIC	Difference between mean summer and mean winter soil temperature is 5 °C or more
	ISOFRIGID	ISOMESIC	ISOTHERMIC	ISOHYPERTHERMIC	ISOMEGATHERMIC	Difference between mean summer and winter soil temperature is less than 5 °C (Later Tropical zone)
Application:	Used at the SUBORDER level, e.g. Borolls, Boralfs, etc. Distinction between cryic and frigid is made at the subgroup level	-Tempreture , *Used at the lower (FAMILY) categoric level (indirectly at the SUBGROUP level)				

Figure 4. Class criteria of different soil temperature regimes
Source: Soil Survey Staff (1975)

There is no biotic activity at temperatures below freezing point (0 °C). Between 0 and 5 °C, root growth and seed germination of most plants are impossible; there is no respiration by plants and thus soil is called inactive. Like lower limit of soil temperature (around 5 °C) that acts as thermal pan to root growth and germination, there exists an upper limit of soil temperature (35 °C or more) above which root growth and germination of most of the mesophytic plants are severely restricted. Therefore, the temperature range from 5 to 35 °C is important in determining the degree of biological and chemical activities and physical processes going on in soils and hence are used in soil classification systems.

Soil temperature regimes play an important role in classifying soils at the Family and Suborder levels. The nomenclature of the different soil temperature regimes and the criteria used are given in a schematic diagram (Figure 4).

In all there are six soil temperature regimes, viz. *pergelic*, *cryic/frigid*, *mesic*, *thermic*, *hyperthermic* and *megathermic*. The prefix *iso* is used if the difference between the mean summer and mean winter temperature is less than 5 °C to separate tropical areas. Further the diagram shows that a Mean Annual Soil Temperature (MAST) of 8 °C is used as a limit to separate *frigid* (cold) soils from *mesic* (cool to warm) soils. Similarly, a MAST of 22 °C is used to separate *thermic* (warm to hot) from *hyperthermic* (hot) soils and a MAST of 28 °C (introduced for Indian Soils) is used as the limit to separate *hyperthermic* (hot) from *megathermic* (very hot) soils.

The MAST can be computed from the mean annual air temperature (MAAT) by adding:

- 3.5 °C to the mean air temperature for areas situated between 8° and 16° N latitude;
- 2 °C for areas situated between 16° and 26° N latitude and;

8 ___3.5°C___ 16 ___2°C___ 26 ___1.5___ 33 ___1°C___

°C (N latitude)

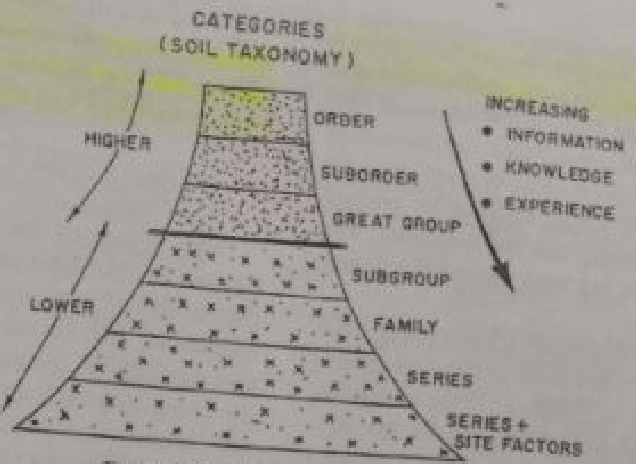


Figure 5. The hierarchy of Soil Taxonomy
Source: Sehgal (1996)

- 1.5 °C for areas situated between 26° and 33° N latitude.

Similarly, for areas situated above 33° N latitude, a temperature of 1°C is added to the MAAT for determining the mean annual soil temperature.

2.6.10. Definition of Soil Taxa

In the definitions of the taxa, differentiating characteristics selected are properties of the soils, including soil temperature and moisture regimes. The genesis of soil is neither directly deployed nor is in the forefront, except as a guide to relevance and weighing of soil properties. The definitions are quantitative and precise rather than comparative and are written in "operational" terms.

2.6.11. Nomenclature (Soil Taxonomy)

Most of the names used in the genetic system are biased in favour of colour (brown/ red soils) and vegetation (prairie/ forest soils) and show no similarity or relationship with each other. They also do not show the place of taxon in the system. In view of these limitations, the nomenclature used in Soil Taxonomy is altogether different and is based on coined words from Greek or Latin languages. The basic principles followed in coining the names, according to Heller (1963), are that the name should :

- be most easily remembered,
- suggest some properties of the object,
- suggest the place of a taxon in the system,
- be as short as possible,
- be as euphonic as possible, and
- fit readily in as many language as possible.

EPPSEL

PPE)

The names of the classification units are combinations of syllables. These names may sound strange at first, but, with experience, one can appreciate and use these with advantage. Each part of the name conveys a concept of soil character or genesis. For example Aridisol (L. *aridus* means dry; *solum*, soil) – the soils of dry places; Vertisol (L. *verto* means turn; *solum*, soil) – the soils which churn or invert.

Structure –The system has six categories of classification (Figure 5) from the highest to the lowest levels of generalization. These are grouped under two broad categories, viz. higher and lower.

Higher Categories

1. *Order* – The highest category in the system.
2. *Suborder* – Comparable to great soil groups of the genetic system
3. *Great Group* – A basic category, based on diagnostic subsurface horizons

Lower Categories

4. *Subgroup* – A new category designed to define the central concepts of Great Groups
5. *Family* – A practical category for making predictions for land use plans.
6. *Series* – The lowest and the most specific category.

The category **Subgroup** has been added to define the central concepts of Great Groups. The Subgroups are intergrades between two Great Groups (and extragrades transitional to a not-soil population) in order to emphasize the fact that soils are a continuum with gradual changes in morphology and soil properties. The category Soil *Type* has been dropped in favour of surface *Phase*.

The nature and kind of differentiating characteristics used in the six categories (mentioned above) are given in Table 3. These categories may be compared with those used for the classification of plants. Just as *Acacia nilotica* identifies a specific kind of plant, the *Ghabdan clay loam* or *Tigris silt loam* identifies a specific kind of soil. This similarity between the soil and plant classification schemes can be followed to the highest category, for instance *Phylum* for plant and *Order* for soil.

2.6.12. *Application of Differentiating Criteria*

The names of the Orders, the formative element in the Order name, used as an identifier at lower categoric levels, derivation or source of the formative element and the mnemonicon or memory device for each Order along with their approximate equivalents in the Genetic and FAO Systems are given in Table 4. The names (of the Orders) are coined words and have a common ending, 'sol' (L.

Table 3. Differentiating characteristics of different categories in soil taxonomy

Category (Number of Taxa)	Differentiating characteristics and description
Order (12)	There are 12 Orders 'A VAGAMI HOUSE*' phrase suggested to facilitate naming of all the Orders. These are based largely on morphology, as produced by soil-forming processes, and indicated by the presence or absence of major diagnostic horizons.
Suborder (63)	There are 63 Suborders within 12 Orders. These emphasize genetic homogeneity, wetness, climatic environment, parent material and vegetational effects. The differentiate used vary, but most tend to emphasize wetness and moisture regime.
Great Group (240+ approximately) G.P.	There are 240 Great Groups within 63 Suborders of 12 Orders. The major emphasis is on the diagnostic horizons (except in Entisols which have no such horizon) and presence or absence of diagnostic layers, base status, soil temperature and moisture regimes.
Subgroup (1000+)	There are more than 1,000 Subgroups. The Typic is used to define the central concept of a great group; the others are used to indicate intergraded to great groups, suborders and orders, and the extragraded to 'not soil'.
Family	The soil properties that are most important for plant growth (like texture, mineralogical class (dominant of solum), soil temperature class (based on MAST at 50 cm depth) and pH are used to differentiate families. They meet the need for making practical prediction for land-use planning.
Series (Approximately 200+ in India; and 12,000 in the USA)	It is the lowest category in the system. The series is a collection of soil individuals, essentially uniform in differentiating characteristics (like colour, texture, structure, consistence, pH and EC) and in arrangement of horizons. It is the series which is most useful for making land-use plans of a small area. The series are named after the geographic name of the place where it was first recognized or where they have wide extent of distribution, e.g. Jodhan loam, Tigris silt loam. The textural name, along with the series name, suggest the surface phase.

*A-Aridisols, V-Vertisols, A-Alfisols, G-Gelisols, A-Andisols, M-Mollisols, I-Inceptisols, H-Histosols, O-Oxisols, U-Ultisols, S-Spodosols, E-Entisols.

Table 4. Formative elements of soil orders and their derivation

Name of Order	Formative element* in name of Order	Derivation of formative element	Mnemonicon or Memory Device	Soils equivalent in other systems
Entisol	ent	Nonsense syllable	Recent	Azonal, some low-humic Gley soils
Vertisol	ert	L.** *verto*, turn	Invert Vert	Grumosols, Black Cotton soils (Regur)
Inceptisol	ept	L. *inceptum*, beginning	Inception	Ando, Sol Brun Acide, Forest and Humic Gley soils
Aridisol	id	L. *aridus*, dry	Arid	Desert, Reddish Desert, Sierozem Solonchak some brown and reddish brown soils
Mollisol Bruni-mol	oll	L. *mollis*, soft	Mollify more	Chesnut, Chernozem, Brunizem (Prairie) Rendzinas, some Brown Forest and associated Solonetz
Spodosol	od	Gk. *spodos*, wood ash	Podzol; odd	Podzols, Brown Podzolic and Ground-water Podzols
Alfisol	alf	Nonsense syllable	Pedalfer	Grey-brown Podzolic, Grey Wooded, and Non-calcic Brown soils, Degraded Chernozems and associated Planosols — Alfi.
Ultisols	ult	L. *ultimus*, last	Ultimate	Red-yellow Podzolic, Grey wooded, and associated Planosols and Half-bog soils
Oxisol	ox	French *oxide*, oxide	Oxide	Laterite soils, Latosols
Histosol	ist	Gk. *histos*, tissue	Histology	Bog soils
Andisol	and	Jap. *and*,	Ando	Volcanic ash soils
Gelisol	el	Gk, *gel*, ice	Frost churning	Frozen Tundra soils, showing cryoturbation

* The element (ent, ert, id, etc.) starts with the first vowel of the Order name and ends with the last consonant preceding the connecting vowel, e.g. oll, ist, etc.

** L stands for Latin and Gk. for Greek

Source: Soil Survey Staff (1975)

solum, soil) with the connecting vowel 'o' for Greek and 'i' for Latin and other roots. A formative element is abstracted from the name of each Order, e.g. *oll* from Mollisol; the element starts with the first vowel (o) and ends with the last consonant (l) preceding the connecting vowel (i). The formative element is used as an ending for the names of all Suborders, Great Groups, Subgroups and Families within an Order, such that with a little experience one can make several statements about soil properties simply by analysing the name of the soil.

Each Suborder name consists of two syllables. The first is suggestive of a property of the class (i.e. Suborder), and second, the name of the Order as reflected by the formative element (e.g. *oll* for Mollisol). For instance, the names: *Udoll, Xerert, Fluvent* are first unique to the Suborders and secondly, suggest the property of the Suborder (*ud* from udic*, *xer* from xeric*, *fluv* from fluviatile) and the name of the Order (*oll* from *Mollisol*, *ert* from *Vertisol*, and *ent* from *Entisol*).

Likewise, the names of Great Groups are coined by prefixing one additional prefix (formative element) to the appropriate Suborder name, for instance, *Hapl*-udoll, *Calci*-xerert, and *Usti*-fluvent. If a Great Group is distinguished from others within the same Suborder by a particular diagnostic horizon or property, the name of that property is used as the root of a prefix to form the other Great Group(s), e.g. *Argiudoll, Haploxerert, Udifluvent*, etc. The Great Group names also suggest the Suborder and Order names to which each one belongs.

Subgroup names consist of the name of the appropriate Great Group modified by one more adjectives, e.g. *Typic Hapludoll, Typic Ustifluvent*. The adjective *Typic* is used to typify the central concept of the Great Group (*Hapludoll or Ustifluvent*).

The nomenclature, as it relates to the different categories in the classification system, is illustrated as follows:

present

Mo*ll*isol	Order
Ust*oll*	Suborder
Argiust*oll*	Great Group
Typic Argiust*oll*	Subgroup

②

The three letters '*oll*' identify each of the lower categories as being in the *Mollisol* Order. Likewise, the Suborder name Ustoll is included as part of the Great Group and Subgroup names. From the Subgroup name (*Typic Argiustoll*), the names of the Great Group (*Argiustoll*), Suborder (*Ustoll*) and Order (*Mollisol*), to which the soil belongs, can easily be known.

2.7. *Salient Characteristic and Differentiating Properties of Different Soil Orders*

In Soil Taxonomy, there are twelve Soil Orders, including the newly-proposed Andisols and Gelisols. They replace the Zonal, Intrazonal and Azonal Orders of the 1938 (revised in 1949) system and are roughly equivalent to the Great Groups of that classification with regard to the level of abstraction. The names of the Orders and their equivalent Great Groups in the 1949 (revised) system are shown in Table 4. These represent broadly the geographic areas that almost coincide with the bioclimatic boundaries (Figure 6).

2.7.1. Key to Soil Orders

Key to Soil Taxonomy – To avoid confusion, one must use the key schematically represented in Figure 6 to check differentiating characteristics for keying out soils. For instance, the presence of a mollic epipedon may not always suggest that the soil is a Mollisol, as some soils having a mollic epipedon may key out as another soil Order. Similarly, all soils having an argillic horizon may or may not qualify for Alfisols or Ultisols. Hence, the use of Key becomes imperative.

ENTISOLS (recent)

Concept

These are very recently developed mineral soils with no diagnostic horizon other than an ochric or anthropic epipedon. The main feature of Entisols is a slight degree of soil formation, because of either limiting time or exceedingly unfavourable conditions.

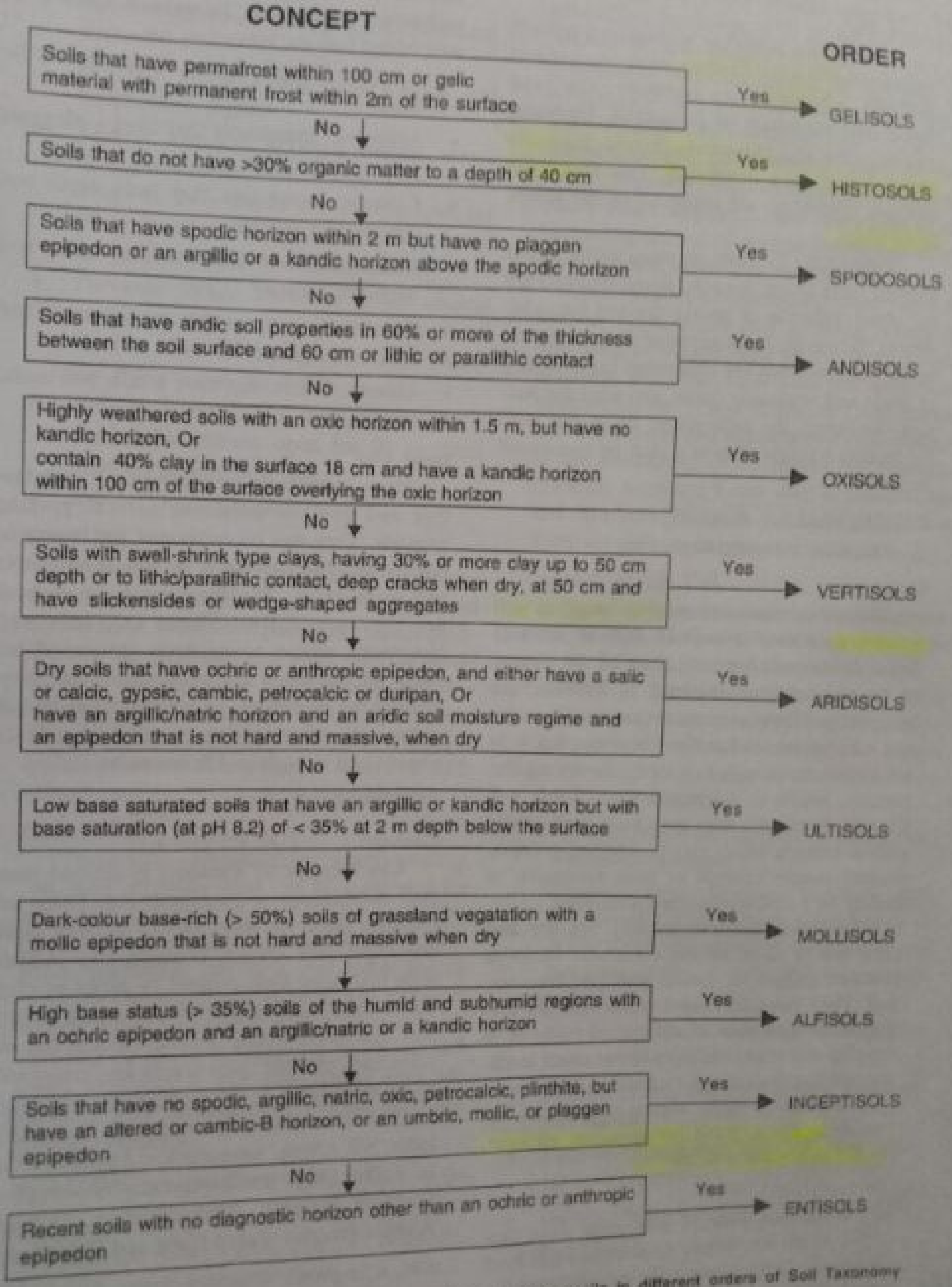
Entisols, by virtue of plant growth and root penetration, show some mineral weathering and surface accumulation of organic matter, but the original crystalline, metamorphic or sedimentary features of their parent materials are little altered by soil formation. Entisols are thus as variable as their parent materials, which range from recent alluvium, through sand dunes, to a variety of rocks. Their topographical setting is also variable. Some are found on young geomorphic surfaces, such as floodplains and on steep slopes where erosion removes soil material as it is formed. They are formed under a variety of climatic conditions.

Classification A.P. AO. f.

Based on the fluvial nature, extreme texture and remnants of old artifacts, soil moisture and temperature regimes, Entisols have been subdivided into the following five Suborders showing 2 syllables; first is characteristic of suborder and the second, of order name:

- *Aquents* : Entisols developed under aquic or hydromorphic conditions
- *Fluvents* : Entisols showing fluvial nature
- *Psamments*: Entisols suggesting extreme sandy nature
- *Arrents*: Entisols showing fragments of diagnostic horizon within one metre
- *Orthents*: Other common well-drained non-fluviatile Entisols

It is beyond the scope of this chapter to discuss further about the suborders and their lower categories (Great Group, Family, etc.)



Soils that have permafrost within 100 cm or gelic material with permanent frost within 2m of the surface

Soils that do not have >30% organic matter to a depth of 40 cm

Soils that have spodic horizon within 2 m but have no plaggen epipedon or an argillic or a kandic horizon above the spodic horizon

Soils that have andic soil properties in 60% or more of the thickness between the soil surface and 60 cm or lithic or paralithic contact

Highly weathered soils with an oxic horizon within 1.5 m, but have no kandic horizon, Or
contain 40% clay in the surface 18 cm and have a kandic horizon within 100 cm of the surface overlying the oxic horizon

Soils with swell-shrink type clays, having 30% or more clay up to 50 cm depth or to lithic/paralithic contact, deep cracks when dry, at 50 cm and have slickensides or wedge-shaped aggregates

Dry soils that have ochric or anthropic epipedon, and either have a salic or calcic, gypsic, cambic, petrocalcic or duripan, Or
have an argillic/natric horizon and an aridic soil moisture regime and an epipedon that is not hard and massive, when dry

Low base saturated soils that have an argillic or kandic horizon but with base saturation (at pH 8.2) of < 35% at 2 m depth below the surface

Dark-colour base-rich (> 50%) soils of grassland vegetation with a mollic epipedon that is not hard and massive when dry

High base status (> 35%) soils of the humid and subhumid regions with an ochric epipedon and an argillic/natric or a kandic horizon

Soils that have no spodic, argillic, natric, oxic, petrocalcic, plinthite, but have an altered or cambic-B horizon, or an umbric, mollic, or plaggen epipedon

Recent soils with no diagnostic horizon other than an ochric or anthropic epipedon

Figure 6. A flow diagram giving simplified key for classifying soils in different orders of Soil Taxonomy

Land Use

In the rocky humid or subhumid mountain regions of Himachal Pradesh, medium-textured Entisols (Orthents) are common. On sand dunes and sand-bars of Rajasthan, Haryana, central Punjab, extremely sandy Entisols (Psamments) are observed. The fluviatile Entisols (Fluvents) are observed in the recent flood plains.

Some of the Entisols on steep rocky slopes support trees along streams. Some Entisols are too stony, infertile or poorly drained for cultivation. However, large areas of Entisols in alluvial bottom lands are cultivated for a variety of grain and vegetable crops, and used for pasture; however, the inter-dunal areas are used for raising a single crop in a year on conserved moisture.

INCEPTISOLS

(L. *Inceptum*, meaning beginning)

Concept ATP. OV.

The soils, representing early stage in soil formation, which is beyond that of Entisols (least developed) but still short of the degree of development, as observed in Alfisols, are Inceptisols. They may have some accumulation of clay in a subsurface horizon, but it is not sufficient enough to qualify for an argillic horizon, which is diagnostic for Alfisols and Ultisols. Similarly, they may have dark colour surface horizon which may be limiting in depth, organic matter content or base saturation to qualify for a mollic epipedon (Mollisols).

The Inceptisols are usually not dry and have one or more of the diagnostic horizons (cambic, umbric, or mollic with low base status). They have developed rather recently owing to the alteration of the parent material to develop soil structure, but without much leaching and accumulation of material in the subsoil.

These soils occur almost throughout the world. In India, these are observed all over, except in hot and arid regions.

Classification

In the past, most of these soils were classified as Brown Forest, Andisol and Sol Brun Acide. In India, these soils are observed extensively in many states. Based on the criteria of wetness, temperature and moisture regime, and the presence of diagnostic epipedon, viz. anthropic or plaggen, Inceptisols have been subdivided into six Suborders, viz.

- *Aquepts:* Inceptisol, that have aquic condition
- *Anthrepts*: Inceptisols that have a plaggen/anthropic epipedon
- *Cryepts*: Inceptisols that have cryic soil temperature regime
- *Ustepts*: Others that have an ustic soil moisture regime
- *Xerepts:* Inceptisols that have a xeric soil moisture regime
- *Udepts*: Others that have a udic soil moisture regime

Land Use

Many Inceptisols are agriculturally productive and provide excellent natural grazing grounds. They can be cultivated for improved pasture and for growing a variety of vegetable and grain crops. The Inceptisols, in general, are very productive. There are, however, Inceptisols whose productivity is limited because of imperfect drainage and/or sodicity. Most of the Inceptisols (e.g. Ustepts) of Punjab under assured irrigation conditions produce wheat yield as high as 4.5 tonnes/ha.

VERTISOLS (L. *Verto*, meaning invert)

Concept TAX. UC,

The concept of Vertisols is derived from their morphological manifestations. These are uniform, thick (at least 50 cm) tropical black and other dark colour, cracking-clay mineral soils that have high content (> 30%) of clay. These soils swell on wetting and shrink on drying. The swell-shrink process induces the development of wide, deep cracks associated with gilgai microrelief or intersecting slickensides.

These soils are predominantly observed in India, NE Africa, Australia and Southern USA. In India, they are dominantly observed in Madhya Pradesh, Maharashthra, Gujarat, Andhra Pradesh, Tamil Nadu and Rajasthan.

Classification

Based primarily on the soil moisture and termperature regime, and the period the cracks remain closed or open (up to the surface in a year), the Vertisols have been subdivided into six Suborders, viz.

- *Aquerts*: Veritsols having aquic conditions within 50 cm
- *Cryerts*: Vertisols that have a cryic soil temperature regime
- *Xererts*: Vertisols having a thermic, mesic or frigid soil temperature regime and if not irrigated, have cracks that remain open for 60 or more consecutive days and closed for > 60 consecutive days in winter
- *Torrerts*: Other Vertisols under natural conditions, have cracks that remain closed for less than 60 consecutive days (when soil temperature is > 8 °C)
- *Usterts*: Vertisols, if not irrigated, have cracks that remain open for 90 or more cumulative days in a year.
- *Uderts*: Other Vertisols

Land Use

Because of their shrink-swell character and development of deep, wide cracks and slickensides, these soils pose many tillage problems during cultivation, or when used for building foundations, laying drainage and gas pipelines and installing irrigation networks. Tilting of trees, fences, poles and sinking of floors are typical characteristics observed wherever such soils occur.

Constraints

These soils offer excellent natural grazing, and, with irrigation, they can be cultivated for two crops in a year. Under rainfed conditions, these soils can support one crop of cotton, sorghum, pigeonpea, etc. in a year either during monsoon (*kharif*) period or on stored moisture in *rabi* (winter) season. Paddy is grown on Vertisols in lower topographic positions and where irrigation is available. These soils are basically very productive, provided, they are managed properly.

MOLLISOLS (L. *mollis* meaning soft)

ABY, RU, AU.

Concept

These are soils of grassland (tall or short) vegetation under subhumid (semi-arid, moist) to humid environment. They have a dark colour, well developed, base-rich, well-structured (granular or crumb) surface horizon, that is rich in organic matter (mollic epipedon). The subsurface horizons, rich in illuviated clay (argillic or Bt), calcareous (calcic or Bk) or gypsiferous materials (gypsic or By) may be present, but are not definitive of the Order. Their characteristic features result from decomposition of organic material, especially fine roots of grassy vegetation, producing some stable dark compounds, and reworking of soil burrowing activity of diverse populations of soil invertebrates, especially earthworms.

Mollisols are predominantly observed in the central USA, from east to west across the heartland of Russia, and in the central South America. In India, such soils are dominantly observed in the 'Tarai' region of Uttar Pradesh and Uttrakhand and occasionally to commonly in the Himalayas (Kulu region) of Himachal Pradesh, northern Bihar and in Maharashtra and Madhya Pradesh in association with Vertisols (Bhattacharyya *et al*., 2006)

Classification

Based largely on moisture and temperature regimes, Mollisols have been subdivided into the following seven Suborders :

- *Albolls*: Mollisols that have an albic (bleached) horizon over an argillic or natric horizon and which has chroma of 2 or less. They also show aquic conditions for some time in a year.
- *Aquolls*: Mollisols that have an aquic (wet) soil moisture regime
- *Rendolls*: Other Mollisols that have 40% or more of $CaCO_3$-content either within or below the mollic epipedon, have a udic soil moisture and/or a cryic soil temperature regime
- *Cryolls*: Mollisols that have a cryic soil temperature regime
- *Xerolls*: Others that have a xeric temperature regime
- *Ustolls*: Others that have an ustic temperature regime
- *Udolls*: Others that have an udic temperature regime

Land Use

The Mollisols are inherently the best agricultural soils of the world. In India, these soils produce optimum yields, both under irrigated and unirrigated conditions with minimum inputs. The soils in the drier regions are used mostly for grazing. In subhumid regions, they are widely cultivated for a variety of crops, such as wheat, maize, sorghum, etc. They can also produce a variety of vegetables. In the Tarai region (Uttar Pradesh and Uttrakhand) of India, because of summer monsoon rains, these are cultivated for maize and sorghum.

ARIDISOLS (L. *Aridus*, meaning dry)

Concept

CCA
SAS

These are mineral soils of dry places (arid and semi-arid) and of areas having high groundwater table. The soils remain dry for most part of the year, and salts accumulate at the surface and/or in the solum, resulting in the development of a salic, gypsic or calcic horizon.

The Aridisols widely occur in the arid climatic environment, especially in South Asia, north and north-east Africa, Australia, south-west part of south America, south-west and northern USA, South Africa and through east to west of Russia. In India, these soils occur in western and north-west India (Rajasthan, Gujarat, Haryana, etc.) in pockets.

Classification

These soils were formerly designated as Desert, Reddish Desert, Sierozem, Reddish Brown Solonetz and Solonchak. Aridisols have been classified into the seven Suborders, largely based on the presence of diagnostic subsurface horizons, viz. cambic, argillic, calcic, gypsic, salic, etc. and on soil temperature regime, viz. cryic.

- *Cryids:* Aridisols that have a cryic soil temperature regime
- *Salid:* Aridisols that have a salic horizon within 1 m of the soil surface
- *Durids:* Aridisols that have a duripan within 1 m of the soil surface
- *Gypsids:* Others that have a gypsic or petrogypsic horizon that starts within 1 m of the surface
- *Argids:* Aridisols with an average argillic (or natric) horizon that starts within 1 m of the soil surface
- *Calcids:* Aridisols that have a calcic or petrocalcic horizon that starts within 1 m of soil surface
- *Cambids:* Other common type of Aridisols

Land Use

These soils do not support a crop without irrigation. These are best left for sparse native grazing. However, if irrigated for cultivation, they pose the risk of salinization or developing sink-holes in gypsiferous landscapes. A network of such sink-holes formed due to slow dissolution of gypsum, leads to formation of big drainage channels underneath and the land sinks leaving behind uneven surfaces for cultivation and/or irrigation. However, in non-gypseferous landscapes of Haryana and Punjab (India), they are used for growing one crop (usually pearl millet) in a year by following water conservation practices during summer monsoon period and growing a crop on stored moisture. The Aridisols have limitations of xerophytic climatic conditions and brackish groundwater which comes to surface due to high evaporative demand, rendering the soils saline and unsuitable for plant growth, except for salt-loving vegetation.

ALFISOLS (from *Pedalfer* of Marbut)

Concept

AUX, BU

Alfisols are base-rich, timbered, mineral soils of subhumid and humid regions. These are characterized by a light-colour surface horizon (ochric epipedon) over a clay-enriched argillic subsurface horizon that is rich in exchangeable cations with base saturation of more than 35%. Alfisols are less weathered than the Ultisols.

These soils are widely distributed in the USA, South America, Northern Europe, south-east and north-west Africa, Brazil and south-east Asia. In India, the Red Soils (Rhod-, Pale-, Hapl- ustalfs) are predominantly observed in different agro-ecological regions covering parts of the states of Andhra Pradesh, Assam, Bihar, Arunachal Pradesh, Himachal Pradesh, Maharashtra, Karnataka, Tamil Nadu, Orissa and Madhya Pradesh.

Classification

In the past, these soils were classified as Gray Brown Podzols, Planosols and some Half

Bog Soils. Based on soil moisture and temperature regimes, Alfisols have been subdivided into five Suborders, viz.:

- *Aqualfs*: Alfisols that have aquic conditions
- *Cryalfs*: Alfisols that have a cryic or isofrigid temeperature regime
- *Ustalfs*: Other Alfisols that have an ustic moisture regime
- *Xeralfs*: Other Alfisols that have a xeric moisture regime
- *Udalfs*: Other Alfisols that have a udic moisture regime

Land Use

Because of the high base saturation, favourable textures and location in semi-arid (moist) to humid regions, Alfisols are naturally fertile and productive. They are used for forestry, grazing, and once cleared, they are suitable for growing a variety of fruit, vegetable and grain crops. In India, they are used for growing crops, like wheat, maize, sorghum, rice and horticultural crops, especially mango, litchi, cashew and jackfruit. The sodium-rich Alfisols need ameliorative measures, such as application of gypsum (as an amendment).

Alkali soil
use Amendment

SPODOSOLS

(Gr. *Spodos*, meaning wood-ash)

Concept A.O.H.F.

These are mineral soils with accumulation of sesquioxides and humus in the subsurface horizons. These develop under cool, humid climate and coarse-textured siliceous parent material which favours free leaching conditions. These are sparsely formed in hot humid tropical, and in warm humid regions, where the parent material is sandy (quartz-rich) and groundwater fluctuates. The diagnostic feature of these soils is an illuvial horizon, enriched with free sesquioxides (Bs) and humus (Bh) underlying a bleached, wood-ash-colour eluvial E-horizon.

These soils are largely distributed in north-west Europe, northern and north-east USA, and Canada. In India, Spodosols are seldom reported.

Classification

In the earlier systems of soil classifications, these soils were classified as Podzol, Brown Podzolic, and Hydromorphic Soils (groundwater Podzols) by recognizing the bleached, ash-colour eluvial, E-horizon.

In the Soil Taxonomy, emphasis has been laid on the subsoil diagnostic horizon, i.e. spodic B-horizon rather than the characteristic of the eluvial E-horizon. The Spodosols have been classified into the following four Suborders based on mixed criteria, such as soil moisture regime and humus content.

- *Aquods*: Spodosols that have aquic conditions as manifested by redoximorphic features
- *Cryods*: Spodosols that have a cryic soil temperature regime
- *Humods*: Other Spodosols having high (> 6%) organic carbon within the spodic horizon
- *Orthods*: Other common types of Spodosols

ULTISOLS (L. *Ultimum*, meaning last) (UUAXH).

Concept Ultisol – Advance stage of

The Ultisols are comparable with Alfisols, except for having low (< 35%) base saturation on the exchange complex, which is due to their advanced stage of weathering. These are base-poor, timbered, mineral soils of humid regions developed under high rainfall and forest vegetation. They are characterized by a clay-enriched subsurface argillic or kandic horizon.

The soils are most extensive in the warm, humid and subtropical regions of the world. These are most predominant in south-east USA, South America, south-east Asia, central Africa and India. In India, they are observed in Kerala, Tamil Nadu, Orissa, Assam and some north-eastern states.

Classification

Based on the criteria of moisture regime and humus content, the Ultisols have been subdivided into the following five Suborders.

- *Aquults*: Ultisols having aquic conditions
- *Humults*: Ultisols having high organic carbon (12 kg/m^2 or more) within 1 m of the soil surface
- *Udults*: Soils having udic soil moisture regime

- *Ustults*: Other Ultisols having an ustic soil moisture regime
- *Xerults*: Other Ultisols that have a xeric moisture regime

Land Use ultisols : forestery.

Because of the low-fertility and low-base status, these soils pose limitations for agricultural use. Generally, these soils are used for forestry, but may produce good agricultural and plantation crops, when they are adequately limed and fertilized. In the tropical regions, such soils are cultivated for pineapple, sugarcane, coffee, cocoa, coconut, rubber, etc. These are also cultivated for vegetables, but only upon adequate fertilization.

OXISOLS (Fr. *Oxide*, oxide)

Concept PUT, AU

These are strongly and deeply-weathered mineral soils of the humid tropics that are poor in fertility. These are characterized by a uniform profile having negligible amounts of weatherable minerals, and are dominated by the kaolinitic and sequioxidic-rich deep subsurface (oxic) horizon. In some cases, deeply-weathered mottled horizons (plinthite) may also develop. Most Oxisols have brick-red colour; but some are also yellow or grey in colour.

Oxisols are dominantly observed in the tropical world, on gently sloping very old landscapes in central Africa, north-east South America and south-east Asia. In tropical India, such soils are not reported yet. Their occurrence in other areas (desert savannas) is indicative of a climatic change.

Classification

In the past, these soils were classified as true Laterites, Latosols and Groundwater Laterites. Based on the considerations and the criteria of soil moisture regimes, the Oxisols have been subdivided into the following five Suborders:

- *Aquox*: Oxisols that have aquic features
- *Torrox*: Oxisols that have an aridic moisture regime
- *Ustox* : Oxisols that have an ustic or xeric moisture regime
- *Udox*: All other Oxisols and those having a udic moisture regime
- *Perox*: Other Oxisols that have a perudic moisture regime

Land Use

Oxisols are chemically-degraded soils and need careful management for agricultural use. These are mainly used for grazing and forestry; some are used for growing crops such as coffee, rubber, cocoa, sugarcane and tropical fruit crops (pineapple, coconut, jackfruit, etc.)

Shifting agriculture is a common practice adopted in such areas. This practice was satisfactory when the shifting cycle was of >20 years, but with the increasing population pressure and demand for food, the cycle has been reduced to 3 to 5 years, resulting in serious problems of soil erosion by water and nutrient depletion.

HISTOSOLS (Greek *histos*, meaning tissue)

Concept

The central concept of Histosols is that of soils formed on organic soil materials. As a thumb rule, a soil without permafrost is classified as a Histosol if half or more of the upper 80 cm is organic. A soil is also called as a Histosol, if the organic materials rest on rock. These are organic-rich soils with thick, peaty horizon (histic epipedon) that develop in lower topographic positions in a permanent water-saturated environment, where plant species (bog, swamp and marsh) are limited in diversity. The main process in their formation is the accumulation of peat, and that happens when organic matter produced by native vegetation is more than its decomposition in the soil. The rate of peat accumulation for woody peat is 0.5 to 1mm per year; however, the rate is much lower in swamps than herbaceous vegetation (marsh, grasses). The amount of organic matter required for a soil to qualify as Histosol depends on the clay content. The minimum amount of organic matter required is 20%, if the clay-content is nil (Figure 7).

The Histosols are of local importance and occur in some selected, nothern parts of Europe, northern USA, Canada and north-west Asia. In India, such soils are expected to occur in pockets of Kerala, Tripura and Andaman and Nicobar Islands.

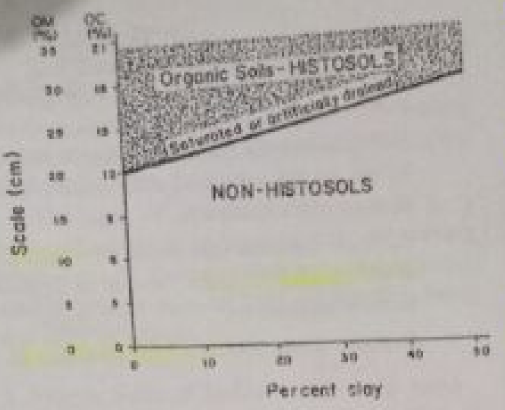


Figure 7. Distinction between Histosols and Mineral Soils based on organic matter content (Organic soils that are never saturated for more than a few days and containing 20% organic carbon are also included in Histosols)
Source: Sehgal (1996)

Histosolc

Classification HF,SF

In the past, these soils were termed as Peat, Muck, and Bog soils. Based largely on the degree of decomposition of the humus, the Histosols have been subdivided into four sub-orders, viz. Hemists, Fibrists, Saprists, and Folists.

Land Use FB

Most of the Histosols, especially Fibrists, flat areas are best located in depressions and areas are best left under natural vegetation. The Folists are freely drained. They represent ustic, xeric or torric soil moisture regime and are used for growing vegetables. The Hemists show moderately decomposed organic matter and are used for natural vegetative cover and as wood-land/ rangelands. For their management, it is necessary to drain these soils for plant growth and to facilitate operation of farm machinery. However, when drained, they oxidise and subside, making the drainage difficult.

fol

Hemi half midrate

ANDISOLS (from Japanese, black soil)

A1- [illegible]

Concept

Andisols were first recognised in Japan and were named as Volcanic ash material. The central concept is of soils developed on volcanic ash. They are typically of dark colour, low bulk density soils that do not have an albic horizon, but must have andic properties [loose or friable colour mollic or umbric surface horizon overlying a brown cambic B-horizon and one or both of the following: low bulk density (< 0.9 Mg/m^3) or 60% or more vitric volcanic ash within 60 cm of the mineral soil surface].

The typical characteristic of these soils is the high content of allophane which gives them a very low bulk density and fluffiness, especially in the B-horizon. The low bulk density and high fertility of these soils make them easy to cultivate.

Normally, Andisols can be observed on all topographic positions and under all soil moisture and temperature regimes. However, these are identified on uplands with undulating topography in Indonesia. Since the Andisols are dominantly developed on volcanic ash, or materials derived from it, such as volcanic alluvium, volcanic loess, they are dominantly observed around the Ring of Fix in the Pacific, including Japan, New Zealand, Indonesia and Philippines. They are also observed in the 'Rift Valley' of Africa, especially Kenya, Western Coast of USA and South America. In India, such soils are expected in the Islands of Andaman and Nicobar where volcanic activity is observed often.

Classification

They are keyed out after Histosols in the Soil Taxonomy. The Andisols have been recently defined in *Soil Taxonomy* (Soil Survey Staff, 1999) in the form of a newly introduced soil Order, viz. Andisols. These soils were earlier grouped under the Order Inceptisols, Sub-order Andepts.

Based generally on the soil moisture and temperatue regimes, the Andisols ~~are divided~~ into eight suborders, viz.: AGTX, UUVU

- *Aquands*: Andisols that have either a histic epipedon or aquic conditions
- *Cryands*: Other Andisols that have a cryic soil temperature regime
- *Torrands*: Andisols that have an aridic moisture regime
- *Xerands*: Other Andisols that have a xeric moisture regime
- *Vitrands*: Other Andisols that have a duripan or a petrocalcic horizon

- *Ustands*: Andisols that have an ustic moisture regime
- *Udants*: Andisols that have a udic moisture regime
- *Vitrands*: Other Andisols that have a duripan or a petrocalcic horizon

Land Use

Andisols that occur on steep slopes, have the limitation of mechanized farming. However, Andisols in favourable climatic conditions, are thickly populated and intensively cultivated. It is advantageous to grow crops that show little response to the application of phosphorus, for instance sweet potatoes (*Ipomoea batata*). The major limitation in the use of Andisols is their high phosphorus fixation capacity, which renders P unavailable. Andisols are generally stable and resist water erosion because of high infiltration and permeability rates. When dry, they are susceptible to wind erosion.

GELISOLS

(from *gelic*, meaning frost churning)

The Order Gelisols is the most recently introduced one to accommodate soils with permafrost in the extreme northern hemisphere.

MGUME

Concept

The Gelisols are conceptually the soils with gelic materials (mineral or organic soil materials that show evidence of cryoturbation, i.e. frost churning) underlain by permafrost. Diagnostic horizons may or may not be present in Gelisols as thawing and freezing play an important role in their evolution. Permafrost influences their pedogenesis by acting as a barrier to the downward movement of the soil solution. Cryoturbation (commonly known as frost churning or mixing) is an important process in Gelisols development.

Gelisols are dominantly observed in the areas of permafrost, which include northern USA, Canada, Greenland, Russia, Mongolia and other areas in the extreme northern hemisphere. In India, such soils have not been observed or reported so far. They may be observed in higher Himalayas with permafrost conditions (ice caps in extreme north of Jammu and Kashmir and Sikkim).

Classification HOT

The Gelisols are subdivided in the following three suborders:

- *Histels*: Gelisols that have organic soil materials or are saturated with water for 30 or more days in a year.
- *Turbels*: Gelisols that show cryoturbation in the form of irregular, broken or distorted horizon boundaries. ADI
- *Orthels*: Other common type of Gelisols.

Land Use

The Gelisols are not cultivated and left to support natural vegetation. The Orthels of anhydrous condition, support little or no vegetation.

2.8. Advantages and Applications

ejp

The recently-developed system is the most elaborate system marked by great care and precision with which the classification units have been defined. For the following three reasons, the students of soil science may find this system very useful.

- The primary basis for identifying different classes in the system is the properties of soils as they exist in the field.
- The nomenclature (with Latin and Greek origins) employed gives a definite connotation of the major soil characteristics.
- The system has an in-built mechanism to permit addition of new soil groups, suborder, or even new order without being disturbed, e.g. new Orders, viz. Gelisols, Andisols, have been included in the system recently.

Other advantages of this system are:

- It permits classification of soils rather than soil forming processes.
- It focuses attention on the soil rather than on the related sciences, such as geology and climatology.
- It permits the classification of soils of unknown genesis.
- It permits greater uniformity of classification as applied by different soil scientists located far apart from each other.

2.8.1. *Applications*

Soil Taxonomy is based on widely-accepted principles and criteria and is being used by many countries, including India. The system has emerged as the most logical soil classification scheme today.

The goal of a classification system is to help soil survey in correlating soils and making predictions by assessing their potential and constraints and in agro-technology transfer, using soil taxa as wheels.

3. Major Soils of India

India, situated between the latitudes of 08°04' and 37°06' N and longitudes of 68°07' to 97°25' E, has a geographical area of 329 Mha. Physiographically, it can be divided into the three broad regions, viz. Peninsula (a triangular Plateau in the Deccan and south of the Vindhyas), Extra-Peninsula (mountain region of the Himalayas), and the Indo-Gangetic Plain separating the two above-mentioned regions (Figure 8).

PR

Geologically, a great part of the Peninsula is occupied by the Archean rocks comprising gneiss, schists and other rocks of diverse nature. Red soils (*Alfisols*) generally predominate in this region. Next in order of age are the *Cuddapah* and *Vindhyan* rocks, followed by the coal-bearing Gondwana formations supporting rocks of the Mesozoic and Tertiary groups. These are mainly distributed over the north-central and NE-central parts where red soils (Alfisols, Inceptisols and Entisols) have

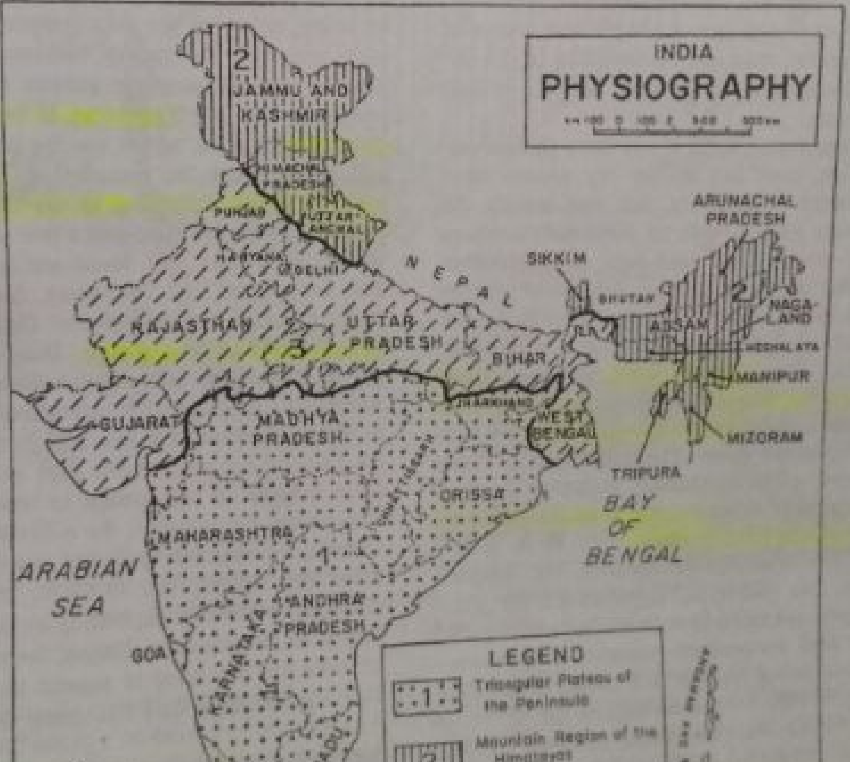


Figure 8. Physiography of India

Source: Sehgal (1996)

developed. The western and central parts are covered by lava flows of the Deccan trap where basaltic rocks predominate. Here, black cotton soils of different thickness (Vertisols, Inceptisols) predominate.

The extra-peninsula, on the other hand, shows the development of marine sediments of all ages, especially in north of the Himalayas. The major rock formations are Tertiary-old sedimentary (sandstone, limestone, etc.) and igneous (granites) (at places metamorphosed to gneisses and schists). Here, non-calcic brown soils (Inceptisols, Alfisols, Entisols, Mollisols) predominate.

The vast Indo-Gangetic and other plains of the Pleistocene origin are composed of alluvium of the great river systems flowing in this region. These are the alluvial soils. However, depending on the age of alluvia and degree of development, they can be classified in the Orders, Inceptisols, Entisols or Alfisols in Soil Taxonomy.

Climatically, India shows three distinct seasons, viz. cool and mainly dry winters from November to February, hot and mainly dry summers from March to June and monsoon rainy season from mid-June to September. Rainfall is received during June to September (from the south-west monsoon) and during December to February (from the north-eastern winds). The Western Ghats and the eastern Himalayan ranges receive the maximum rainfall because of their alignment across the summer monsoon winds. The Indo-Gangetic Plains receive moderate rainfall; Rajasthan receives little rainfall as the Aravallies lie along the path of moisture-bearing winds.

The other climatic element, viz. temperature, has particular significance for plant growth, especially in northern India during winter; high temperature promotes evapotranspiration during summer and causes crop-affecting aridity. Peak temperatures of 42 °C to 47 °C are observed during summers (May to June), which fall sharply with the onset of monsoon rains (in July to September). Sub-zero temperatures are observed during winters in the extreme northern regions where germination of seed becomes a problem.

The bio-climate is the climate prevailing a few metres above and below the soil surface where biological activity prevails. The different climatic elements of relevance to plant growth are expressed in the bioclimatic map (Figure 9). The soil temperature and moisture regimes as discussed earlier, play an important role in classifying soils and crop planning, as each plant species requires, specific soil temperature and moisture conditions for its optimal growth. India is dominantly represented by thermic and hyper-thermic temperatures and Ustic, Aridic and Udic moisture regimes.

India with a wide variety of landforms, geological formations and climatic conditions, exhibits a large variety of soils; the variety is so diverse that barring a few soil orders (Andisols, Spodosols,Oxisols), India represents all the major soils of the world.

The grouping of these soils can be achieved by using either of the two systems, viz. Genetic and Soil Taxonomy. Whereas the Genetic System is based on genetic factors and processes, the Soil Taxonomy is based on the properties of soils, which are the reflection of soil genesis and are measurable. The major soils of India, according to the Genetic approach, can be classified into a few soil groups, viz. alluvial, black, red, forest and desert soils. But, in terms of Soil Taxonomy, they key out in nine out of the twelve Soil Orders and a large number of great groups. Since the objective of this section is to apprise students of the major soils of India, it is considered desirable to do so in terms of a few soil groups as per the Genetic approach rather than explaining a large number of soil groups in terms of Soil Taxonomy. For instance, the Alluvial Soils (in Genetic System) can be keyed out in three or more soil orders, viz. Entisols, Inceptisols, Alfisols, or Ultisols, depending on the age and degree of development. Hence, the major soils of India are discussed in genetic terms giving their equivalents in Soil Taxonomy (Figure 10).

GS

TP

T = 9

3.1. Alluvial Soils

The name alluvial is given to the soils that have developed in alluvium, irrespective of their place of occurrence and degree of development. They are one of the important groups of soils for agricultural production. They are extensively distributed in the states of Punjab, Haryana, Uttar Pradesh, Uttrakhand, Bihar,

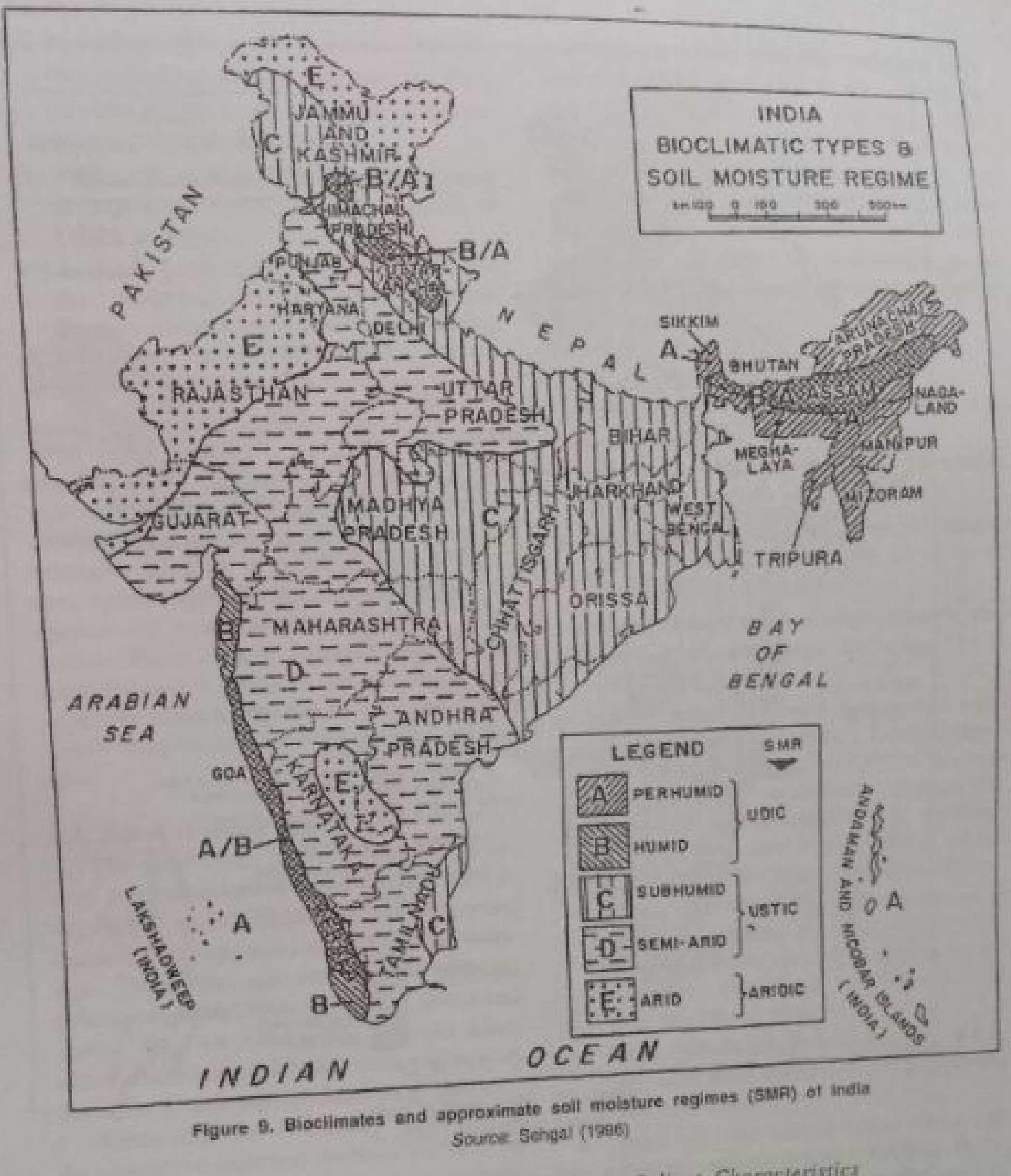


Figure 9. Bioclimates and approximate soil moisture regimes (SMR) of India

Source: Sehgal (1986)

West Bengal, Assam and coastal regions of India and occupy an estimated area of 75 Mha in the Indo-Gangetic Plains and Brahmaputra Valley alone.

The parent material of these soils (alluvium) is of recent origin and has been derived from the deposition of erosion products brought and laid down by various river systems; coastal alluvium is, however, laid down by the sea currents.

3.1.1. Salient Characteristics

The salient characteristics of alluvial soils are:

- The Alluvial soils, in general, are varible in texture, depending on the source of parent material and their place of deposition. They are coarser near the source and become finer in proximity to the sea (delta area).
- They are fluvial in nature (stratified) which is reflected in their texture and irregular distribution of organic matter with depth.

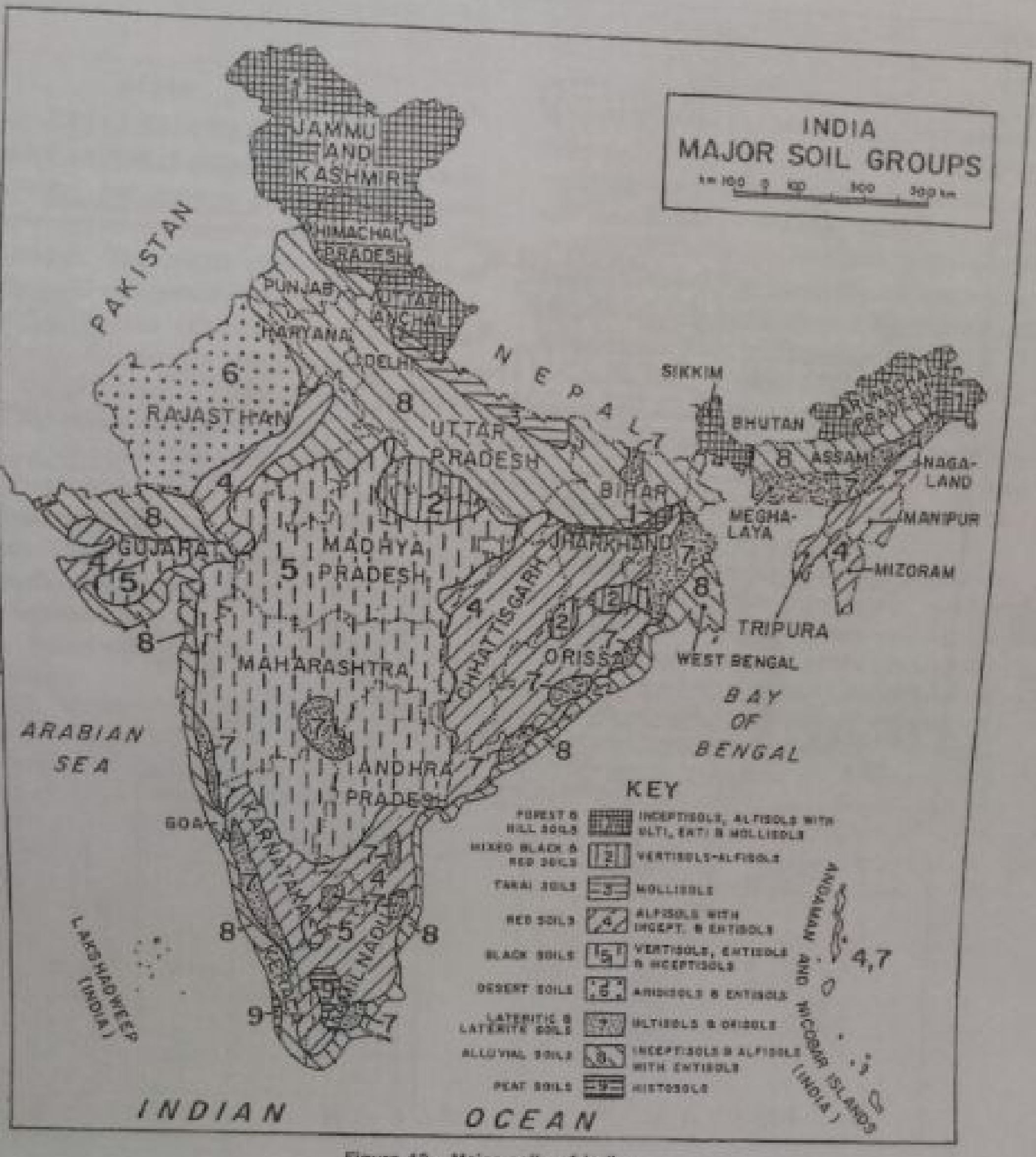


Figure 10. Major soils of India
Source: NBSS&LUP (1985)

Al = US

- They are either alkaline (in aridic and/or ustic regime) or acidic (in udic regime).
- They exhibit different profile developments from least-developed A-C to well-developed A-B-C profile, depending on the prevailing climate and age of the alluvium.
- They are inherently rich in plant nutrients. In general, they are fairly sufficient in phosphorus and potassium, but are deficient in nitrogen and organic matter contents.

In the Genetic System, these soils are classified as 'Alluvial soils' belonging to the Order 'Azonal'. In terms of Soil Taxonomy, such soils may key out in any one of the following Orders:

(i) Entisols (with A-C profiles), and depending on their nature and morphology, they may fit in with Fluvents, Psamments, Arrents, Orthents or Aquents (at Suborder level).

(ii) Inceptisols (with A-(B)-C profiles) and further depending on their moisture and temperature regimes, as : Aquepts, Cryepts, Ustepts, Xerepts or Udepts.

(iii) Alfisols [with A-Bt-C profiles] and further as Aqualfs, Cryalfs, Ustalfs, Xeralfs or Udalfs as above.

(iv) Aridisols [with A-(B)-C profiles] and further as Cryids, Salids, Durids, Gypsids, Argids, Calcids or Cambids, depending largely on the kind of diagnostic horizon and temperature regime.

3.1.2. Land Use

The major constraints of some of these soils are stratification that restricts leaching and drainage, extreme sandy nature that promotes excessive leaching of water and plant nutrients, hydromorphic condition that promotes reduction and results in poor aeration for plant growth. These soils, if managed well, can be fruitfully used for most of the agricultural and vegetable crops. They can be rendered saline where evaporation exceeds rainfall (aridic zone).

3.2. Black (Cotton) Soils

The name black is given to soils that are very dark in colour and turn extremely hard on drying and sticky and plastic on wetting, and hence are very difficult to cultivate and manage. These soils are comparable with the Grumosols of the USA. In view of thier black colour, they are comparable with the Chernozems of Russia and Prairie soils of the USA, but differ in their physicochemical properties.

These soils are dominantly distributed in the central, western and southern states of India (Figure 10). According to a recent study, different members of black soils occupy an estimated area of 74 Mha.

Typical Black (Cotton) soils, with characteristic swell-shrink nature, have developed on basaltic material (rock or alluvium) under semi-arid to sub-humid climatic conditions. The natural vegetation of the area is tropical dry deciduous and tropical thorny forests with fairly dense growth of grasses. The weathering product of the parent rocks (basalt and other metamorphic, containing lime and soda-lime feldspars) are basic and rich in clay, which is dominantly of smectitic type with high co-efficient of expansion and contraction, and therefore set up a steady churning process in the pedon. Churning causes vertical mixing in deep soils and leads to the development of deep (> 50 cm) and wide (> 1 cm) cracks, gilgai micro-relief and/or closely intersecting slickensides.

3.2.1. Salient Characteristics

The salient characteristics of these soils are as follows:

- These are highly clayey with clay content ranging from 30% to 80%.
- Being calcareous, they have pH ranging from 7.8 to 8.7, which may go up to 9.5 under sodic conditions.
- Being rich in smectitic-clay minerals, they have high exchange capacity [30-60 cmol(p^{+})kg^{-1} soil].
- These have high water and nutrient holding capacity. The moisture holding capacity, although is high (150-250 mm/m), large part is of it is not available for plant growth, because it is held tenaciously by the dominant smectitic clay. It also reduces permeability and hence poses drainage problem.
- These do not exhibit any eluviation and illuviation process, because of churning. Moreover, under calcareous conditions, the clay stays flocculated and doesn't move with percolating water.
- These have high bulk density (1.5-1.8 Mg/m^3) because of the swelling and shrinking nature of the soil.
- These are very dark in colour which may be due to clay-humus complexes and/or the presence of titaniferous magnetite mineral.
- These are highly sticky and plastic and thus pose many problems in both agricultural and non-agricultural sectors.

3.2.2. Classification

These soils, as per the Genetic system, are classified in the Orders: Intrazonal (Grumosols, '*regur*') and Azonal (Regosols, Alluvial soils). According to Soil Taxonomy, the deep, black soils are keyed out in the Order: Vertisols and Suborders Usterts, Torrerts based on the moisture regime. The shallow soils, however, key out in the Orders: Inceptisols and Entisols, depending on the degree of profile developement.

3.2.3. Land Use

The major contraints in their land use are: narrowing of workable moisture range, low infiltration rate and poor drainage, poor moisture and nutrient availability for plant growth, poor in some available plant nutrients, especially nitrogen, phosphorus, sulphur, their calcareous nature adversely affects the availability of micronutrients and due to their swell-shrink nature, these are unsuitable for laying foundations, construction of buildings, laying of pipelines and electric communication poles, etc.

The soils are inherently very fertile and, under rainfed conditions, they are used for growing cotton, sorghum, millet, soybean, pigeon pea, etc. Under irrigated conditions, they can be used for a variety of other crops, such as sugarcane, wheat and citrus plantations.

3.3. Red Soils

The name Red is given to soils rich in sesquioxides that have developed on rocks of Archean origin (granite, gneiss) and on welldrianed, stable and higher land forms under hot, semi-arid to humid subtropical climatic conditions. Under such conditions, the weathering is moderately intense and leads to enhanced decalcification. Some weathering products are leached out leaving behind the less mobile elements, like silica, iron and alumina. The iron and aluminium under oxidised conditions, form sesquioxides (Fe-oxides and Al-oxides), imparting red colour to these soils.

These soils are predominantly observed in the southern parts of the Indian Peninsula, comprising the states of Andhra Pradesh, Tamil Nadu, Karnataka, Maharashtra, Orissa and Goa and in N-E States. In Andhra Pradesh, the Red and the Black soils occur under similar bioclimatic conditions, but on different parent materials and landforms. The Red soils develop on igneous (acidic) rocks and occupy higher topographic positions, whereas the Black soils develop on basalt (basic) rock or on alluvium derived from basalt, and occupy lower positions on the landscape. The soils grade from shallow, gravelly and light-colour (of the uplands) to much fertile, deep, dark reddish brown in the plains and valleys.

Red igneous).

3.3.1. Characteristics

Hot FC,

Due to FeO, Fe_2O_3

The main characteristics of these soils are:

- These are red to yellowish in colour which is the result of coating of ferric oxides on soil ped sufaces, rather than to a very high proportion of iron. It is red when ferric oxide occurs as haematite or anhydrous FeO and yellow when it occurs in the hydrated form (called limonite). In general, they have hues of 7.5YR (or redder) in upper topographic positions and darker hues in the valleys.
- These soils are highly variable in texture, ranging from loamy sand to clayey; but in general they vary from loam to clay loam in texture.
- They are shallow and poor gravelly in the uplands to very deep, fertile in the plains and valleys.
- They are well to excessively drained, depending upon their topographic position and texture.
- They are, in general, neutral to acidic in nature, depending upon the content of iron oxides.
- The SiO_2/sesquioxide ratio (in clays) varies from 2.5 to 3.0; the amounts of iron and aluminium are generally high (30-40%).
- The cation exchange capacity and base saturation of these soils are relatively lower than those of the Black or Alluvial soils. Some Red soils, occurring in association with Lateritic soils, have low exchangeable bases [~20 cmol(p$^+$)kg^{-1} soil]. The CEC (clay) ranges from 25 to 40 cmol(p$^+$)kg^{-1} soil.
- These soils are generally deficient in nitrogen, phosphorus and potassium. They are also poor in organic matter and lime contents.
- These show common presence of kaolinite (1:1 type) clay mineral.

Laterite
→ Presence of wet, dry.
→ S-Ratio < 2
→ B.S.-Ratio < 40%
→ CEC < 16%
→ oxisols

Lateritic
not present
S-Ratio > 2
B.S. > 40%
Alfisols and ultisols

AU = Red

3.3.2. *Classification*

In the Genetic scheme, they qualify as Red loam, Reddish and Yellowish-brown soils. According to Soil Taxonomy they key out as: Alfisols and Ultisols (well developed soils with high and low base saturation, respectively), Entisols (very shallow and crusty soils), or as Inceptisols (moderately developed soils).

3.3.3. *Land Use*

The soils pose limitations of soil depth (on hills and hill slopes), poor water and nutrient holding capacity, surface crusting and hardening, excessive drainage and runoff, poor natural soil fertility (N, P, Ca, Zn, S). Under good management practices, these soils can be profitably used for a variety of agricultural, horticultural and plantation crops (depending on the moisture regime), such as millets, rice (both direct seeded and transplanted), groundnut, maize, soybean, pigeonpea, green gram, jute, tea, cashew, cocoa, grapes, banana, papaya, mango, etc.

3.4. *Laterite and Lateritic Soils*

The term **'Laterite'** was originally used by Buchanan in 1807 for the highly ferruginous, vesicular and unstratified material observed in the Malabar Hills of south India. The laterites are typically formed in tropical climate experiencing alternate wet and dry seasons. With monsoon type of climatic conditions acting on the basic parent rock, the siliceous matter is leached almost completely during weathering and the sesquioxides are left behind. On drying, these are converted into irreversible iron oxide and aluminium oxide. The soils thus formed are rich in sesquioxides, devoid of bases and primary silicate minerals, hard or capable of hardening like bricks when exposed to drying after wetting. It is a compact to vesicular rock-like material composed of a mixture of hydrated oxides of iron and aluminium with small amounts of manganese oxides and titania. They are generally observed on hill-tops and Plateau landforms of Orissa, Kerala, Tamil Nadu, etc.

The Lateritic soils are formed under almost comparable climatic conditions as described in the preceding paragraph, but do not require alternate wet and dry conditions and the groundwater level may not be very near the surface. Such soils are widely distributed in the states of Maharashtra, Andhra, Karnataka, Tamil Nadu and North-East region, and occupy about 25 Mha of the total geographical area of India.

3.4.1. *Characteristics*

The Laterites have following characteristics:

- They have hues of 5 YR or redder with maximum intensity in the B-horizon. RB-5
- These are deeply-weathered soils with high clay-content.
- These soils lose bases (Ca, Mg, Na, K) and silica due to pronounced leaching with relative accumulation of sesquioxides and the soils are rendered acidic with silica: sesquioxides ratio of < 2. Laterite low
- These are predominant in kaolin clay minerals.
- These are low in CEC (clay), which is <16 cmol(p$^+$) kg^{-1}, and base saturation (< 40%).

The Lateritic soils (Ultisols, Alfisols) remain dry for 4 to 5 months in a year, and have SiO_2/R_2O_3 and SiO_2/Al_2O_3 ratios of > 2. Kaolin is the dominant clay mineral with occasional presence of gibbsite. These are relatively high in base saturation (> 40%).

3.4.2. *Classification*

The true Laterites (Oxisols in Soil Taxonomy) are not yet observed in India. Most of the laterite soils key out in the order of Ultisols (Chandran *et al.*, 2005) and Suborders, Aquults, Udults, Ustults, depending upon the moisture regime.

3.4.3. *Land Use*

The major limitations posed by these soils are deficiency of P, K, Ca, Zn, B, etc. and high acidity and toxicity of aluminium and manganese. Liming of these soils is not practical because of non-availability of lime due to high cost of transportation. The experience of liming of temperate soils cannot be applicable in tropical area having low-activity clays. Evidences suggest that calcium is more important as a nutrient than for neutralization of acidity

The Laterites of lower topographic positions are used for growing rice, banana, coconut and arecanut, and of higher topographic positions, for cocoa, cashew, tea, coffee, rubber, etc. Shifting agriculture is mainly practised in these areas, but the shifting cycle should be of 20 years or more.

3.5. *Desert (Arid) Soils*

The name Desert/arid is given to the soils supporting negligible vegetation, except xerophytic plants, unless irrigated. Such areas may be observed in cold or hot temperature regime. A large tract of hot arid region, with a growing period of < 60 days in a year is situated in the north-west India (Rajasthan, Gujarat, Haryana and Punjab). It covers an area of about 29 Mha, and poses desertic conditions of geologically recent origin. It differs from other deserts that are cold (in the extreme north, Leh and Ladakh) and tropical (in the south in Karnataka and Andhra Pradesh). The soils of hot arid belt are comparable with those of Alluvial soils, but have an aridic moisture regime. The aeolian action moves and carries the sandy material and deposits it in the direction of wind in the form of a thick mantle of sand at the surface. The sandy material, under arid climatic conditions, results in poor profile development.

Recent origin

3.5.1. Salient Characteristics

The salient characteristics of these soils are:

- These are sandy to loamy fine sand in texture with clay content varying from 3.5% to less than 10%.
- These are pale brown to yellowish brown in colour and are either single grained or have weak subangular blocky structure.
- These are generally poor in nutrient and water-holding capacity
- These are slightly to moderately alkaline in reaction (pH 7.8-9.2) because of their calcareous nature. The calcitic nodules in varying amounts (5-50%), are observed in dispersed form or as a layer; their depth of accumulation varies depending on the rainfall and topographic position.
- These are gypsiferrous in extreme arid areas (such as Bikaner, Jaisalmer) that may form a gypsic horizon. Such gypsiferrous soils need special attention to avoid formation of sink holes, if irrigated.
- These are poor in soil fertility status, especially N, P, K, S, Zn.

3.5.2. Classification

In the Genetic Scheme, the calcareous desertic soils are classified as Pedocal Sierozems or Calcareous Sierozems. The sandy soils of the Thar desert with 300-500 mm rainfall are classified as Grey Brown (Desert) and the more drier, as Desert soils. In general, these sandy soils can also be classified as Regosols within the Order Azonal.

According to Soil Taxonomy, these soils qualify for the Orders: Aridisols (Suborders: Cambids, Calcids, Salids) and Entisols (Suborders: Psamments — the sandy members and Orthents).

3.5.3. Land Use

The major constraint of these soils is less water which restricts their use for raising agricultural crops. However, the Psamments (sandy) member, situated in the interdunal valleys receiving additional rain water as runoff from the surrounding areas, are used for growing a crop (millet or pulse) during monsoon period. The Orthents are used for raising a crop on conserved moisture of monsoon rains. The Gypsids may best be used as pasture land to avoid dissolution of gypsum to form sink holes. If irrigated, all members (except gypsiferous) can be profitably used for growing two crops in a year.

3.6. *Forest and Hill Soils*

This name is implied for soils developed under any forest cover. In India, the total area under different forest species (tropical, deciduous, coniferous, tropical evergreen) is estimated to be 75 Mha and is observed dominantly in the states of Himachal Pradesh, Jammu and Kashmir, Uttar Pradesh, Uttrakhand, Bihar, Madhya Pradesh, Maharashtra, Kerala and North-East region. While the climatic conditions and altitude control the kind of forest species, the kind of forest and topography control the kind of soils and their degree of profile development. The

Podsolic
pH – 4.5-6.0
O.M = 3.-3.5%
CEC = 10-15 cmol(P+) kg-1
clay = 20-30%.

Base saturation Brown forest soil
pH – 6.0-7.0 (cal go to 8.2)
O.M : 2-3%
CEC = 15-20 cmol(P+) kg-1,
B.S = 70-90%.

major soils observed in different forest areas are: Brown Forest and Podsolic (in northern Himalayas) and Red and Lateritic (in the Deccan Plateau). The Himalayan soils have developed on sand-limestone, conglomerates, granite, gneisses and schists under cool/cold (sub) humid climate (acidic environment), and the Deccan Plateau soils are formed on igneous and metamorphic rocks (basalt, granite, gneisses) under (sub) tropical climatic conditions (slightly acidic, neutral or basic environments), and hence differ in their properties.

3.6.1. Podsolic Soils

The soils, found under coniferous vegetation in the presence of acid humus and low base status, show some characteristics associated with Podsols, but because of the unfavourable (non-siliceous) parent material and absence of breaking down of soil minerals in the unsaturated organic acids, the process of podsolization is restricted up to the mobilization of sesquioxides and hence the true podsols in northern Himalayas are not formed. The Podsolic soils have the following characteristics:

- These soils are moderately to strongly acidic in reaction (pH 4.5-6.0).
- These soils are high in organic matter content (3.0-3.5%) and, in general, low (< 50%) in base status.
- These soils are variable in exchange capacity [10-15 cmol(p$^+$)kg^{-1}]. The clay content varies from 20% to 30%.
- These soils are deficient in phosphorus since it gets precipitated as iron phosphate and aluminium phosphate.

The soils developed under pinus vegetation and cool (sub) humid conditions qualify largely for Alfisols, but those under (per) humid conditions, for Ultisols. The soils on slopes are Inceptisols.

These soils face a major problem of erosion by water. They are used for growing a variety of crops, especially rice, maize, soybean, etc. on terraces and tea and other fruit plants on slopes.

3.6.2 Brown Forest Soils

The other soils, developed on sedimentary rocks and/or alluvium under subhumid to humid climate and mixed vegetation are Non-Calcic Brown or Brown Forest soils. They, in general, possess the following characteristics:

- These are neutral to slightly acidic in reaction (pH 6.0-7.0); the calcareous members under subhumid conditions may have a pH around 8.2.
- These are moderate to high in organic matter content (2-3%) which decreases regularly with depth; they also show biological acitivity.
- These are moderate in cation exchange capacity [15-20 cmol(p$^+$)kg^{-1} soil] and the exchange complex is almost saturated with bases (70-90%).

The soils key out in the Orders: Inceptisols and Mollisols on stable landforms and as Entisol on eroded surfaces or steep slopes. The aspect and topography play an important role in their evolution and classification. While the north-facing slopes largely support Alfisols and Inceptisols, the south-facing slopes support Mollisols and Inceptisols.

NA
SM,
SI.

The soils have great potential for growing agricultural crops such as rice, maize and fruit plants, such as apple, almond, pear, apricot, etc.

3.7. Salt Affected Soils

The soils, occurring in the arid and semi-arid regions, are Intrazonal as they are interspersed with the other Zonal dominant soils of the region. According to the estimates of the Central Soil Salinity Research Institute (CSSRI), Karnal, such soils occuppy 10 Mha, of which a major fraction (say 7 Mha) is sodic and occurs in the Indo-Gangetic Plain, followed by the Deccan (Peninsula) Plateau supporting Black soils; the rest (about 20%), in the arid and coastal regions, are saline.

The sodic soils of the Indo-Gangetic Plain, occupy relatively lower topographic positions where products of weathering accumulate during the monsoon rains by surface runoff; evaporates and the soil solution becomes concentrated resulting in increased sodium adsorption ratio (SAR) and hence, increased ESP and pH. The displaced calcium is precipitated at high pH and temperature, as calcium carbonate. The process, repeated over years, results in the formation of sodic soils (Pal *et al.*, 2003). The saline soils of the coastal region result from the rise of brackish underground water due to capillary action under excessive evaporation; the water evaporates leaving the salts at or near the surface depending upon the equilibrium established between leaching and capillary action. The salts precipitate as white efforescence and may qualify for a salic horiozn.

The soils are classified as Solanchaks (saline) and Solonetz (sodic). In terms of Soil Taxonomy, they fit in the Orders of Aridisols, Inceptisols, Alfisols and Vertisols, depending upon the profile development.

The sodic soils pose serious problems of high sodium on the exchange complex, poor physical conditions, especially soil structure and drainage, nutrient and water availability and micronutrient deficiency. Another problem is of receding groundwater in the central sectors and the south-west sectors of Punjab, Haryana and some parts of Rajasthan. The rise in groundwater causes salinization of soils and the farmers are obliged to switch over to growing rice and eucalyptus in areas whch till recently were used for raising cotton and citrus plantation.

Despite many limitations, the sodic soils, once ameliorated by applying gypsum , are used successfully for growing rice, followed by wheat. The amount of gypsum, required for their amelioration would depend on sodium on the exchange complex (ESP). In saline soils, having high and brackish groundwater, the objective is to reduce the soluble salts by leaching with fresh water and introduction of drainage network in order to bring the salts within the safe limit for growing crops. The presence of gypsum in such soils eliminates dispersion of clay during the amelioration process.

4. References

Baldwin, M., Kellogg, C.E. and Thorp, J. (1938) Soil classification. In: *Soils and Men: Yearbook of Agriculture*, United States Department of Agriculture, Washington, D.C., USA, pp. 979-1001.

Bhattacharyya, T., Pal, D.K., Lal, S., Chandran, P. and Ray, S.K.(2006) Formation and persistence of Mollisols on zeolitic Deccan basalt of humid tropical India. *Geoderma* **146**, 609-620.

Chandran, P., Ray, S.K., Bhattacharyya, T., Srivastava, P., Krishnan, P. and Pal, D.K. (2005) Lateritic soils of Kerala, India: Their mineralogy, genesis and taxonomy. *Australian Journal of Soil Research* **43**, 839-852.

Dokuchaiev, V.V. (1900) cited by Glinka, K.D. 1927. *Dokuchaiev's Ideas in the Development of Pedology and Cognate Sciences*. Academy of Sciences, USSR, Pedology I, Leningrad.

Heller, J. (1963) The art of nomenclature of what's in a name. *Proceeding of the Soil Science Society of America* **27**, 216-220.

Higgins, G.M. and Kassam, A.H. (1981) The FAO-agro-ecological zone approach to determination of land potential. *Pedologie* **31**, 147-168.

Kellogg, C.E. (1963) Why a new system of classification? *Soil Science* **96**, 1-5.

Marbut, C.F. (1927) A new scheme for Soil Classification. *Proceedings and Paper of the First International Congress of Soil Science.*

Marbut, C.F. (1935). Soils of the United States. In: *Atlas of American Agriculture*, United States Department of Agriculture, Washington, D.C. USA.

NBSS&LUP (1985) *Soil Map of India* (1:7 million scale), National Bureau of Soil Survey and Land Use Planning Publication, Nagpur.

Pal, D.K., Srivastava, P., Durge, S.L. and Bhattacharyya, T. (2003) Role of microtopography in the formation of sodic soils in the semi-arid part of the Indo-Gangetic Plains,India. *Catena* **51**, 3-31.

Sehgal, J. (1996) *Pedology Concepts and Application*, Kalyani Publishers, Ludhiana.

Sehgal, J. Hall, C.F. and Bhargawa, G.P. (1975) An appraisal of problems in classifying saline sodic soils of the Indo-Gangetic Plain in N-W India. *Geoderma* **14**, 75-91.

Smith, G.D. (1963) Objectives and basic assumptions of the new soil classification system. *Soil Science* **96**, 6-16.

Smith,G.D. (1986) *The Guy Smith Interviews: Rationale for Concepts in Soil Taxonomy*, SMSS Technical Monograph No.11, SMSS, SCS, USDA, 259p.

Soil Survey Staff (1960) *Soil Classification. A Comprehensive System – 7th Approximation*. Soil Conservation Service, United States Department of Agriculture, Washington, DC, USA.

Singal Survey ① Soil fertility. ② crop suitability
③ Soil conservation ④ Soil erosion assessment
⑤ Irrigation suitability ⑥ land settlement
⑦ Trafficability ⑧ revenue, ⑨ Taxon,

ITI.S, R.F
E, C.C.

Chapter 4

Soil Survey and Mapping

P.K. SHARMA and DIPAK SARKAR

1. Introduction

Soil science today embraces the knowledge about soils accumulated mostly during the past 60 to 90 years. Thousands of different kinds of soils exist in the world. Their properties differ from place to place because of the interaction of various soil forming factors and processes in their evolution. Soil scientists strive to recognize and understand the soils in nature. The soil in this text is a natural body comprised of solids (minerals and organic matter), liquid and gases that occur on the land surface, occupies space and is characterized by one or both of the following : horizons or layers that are distinguishable from the initial material as a result of additions, losses, transfers and transformations of energy and matter or the ability to support rooted plants in a natural environment.

The upper limit of soil is a boundary between soil and air, shallow water, live plants or plant materials that have not begun to decompose. Areas are not considered to have soil if the surface is permanently covered by water too deep (typically more than 2.5 m) to the growth of rooted plants. The lower boundary that separates soil from non-soil matter underneath is difficult to define.

Good cultivable lands are not only limited but differ also in their production potential. The soil variability within a village, district or state influences the use of soils for different purposes. In order to make optimum use of our limited soil resources, we need detailed information about their characteristics, types, and distribution on landscape. This information is collected through soil survey and mapping.

Soil surveys began in 1899. These surveys were simple at that time and were intended to answer practical agronomic questions about soil differences and limitations which were important for improving and expanding crop production. The soil surveys expanded in details and concept with advances in the scientific knowledge and demand for more useful information.

2. Objectives of Soil Survey

The broad general objectives of a soil survey could be grouped as fundamental and applied:

- Fundamentally, a soil survey helps in gathering information about the properties, genesis, classification and nomenclature of soils.
- The applied aspect in a soil survey includes interpretation of soil data for use in agriculture, forestry, recreational purposes, urban, industrial and pasture development, etc.

Both the fundamental as well as applied objectives are pursued with equal emphasis during a soil survey. Soil surveys are planned in such a way so as to provide information about soils to different users.

3. Uses of Soil Survey

Soil surveys are of great significance to any nation as they provide inventory of soil resources. Major uses of soil surveys are indicated below:

(i) They provide information for the development of land use plans for both arable and non-arable lands and for predicting the long-term effects of a particular land use on environment.

FAAF

PED

(ii) They help in predicting the adaptability of identified soils to various uses and also their behaviour and productivity under defined sets of management practices.

(iii) Soil resource inventory helps in recognising the areas having constraints like salinity, alkalinity, acidity, erosion, waterlogging, flooding, and also in taking suitable measures for their management.

(iv) Soil information generated through soil surveys is useful for land settlement, tax appraisal, locating and designing highways, airports and other engineering structures.

In short, soil surveys provide information about the soils of a country and form the basis for land use planning.

4. Characteristics of Soil Survey

Soil surveys are basically of two types:

(i) single purpose soil surveys, and

(ii) standard soil surveys.

① soil fertility ② crop suitability, ③ soil conservation

4.1. Single Purpose Soil Surveys

These are designed for a specific objective with limited applications such as soil fertility—① appraisal, crop suitability, soil conservation, soil erosion assessment, irrigation suitability, land settlement, trafficability, revenue and taxation.

4.2. Standard Soil Surveys <SLMT>.

These involve comprehensive data collection about soils and lands in such a manner that these could be used for a variety of purposes and encompass most of the single purpose soil surveys.

A standard soil survey involves a series of interlinked activities, some of which are mentioned below.

- Study of important characteristics of soils and the associated external land features such as landform, natural vegetation, and slope.
- Laboratory analyses of soils to support and supplement the field observations.
- Correlation and classification of soils according to the standard system of classification.
- Mapping of soils, that is drawing and fixing soil boundaries of different kinds of soils on a standard base map.
- Transfer of agrotechnology through soil taxa which serves as a wheel for such transfers.

5. Practical Difficulties Encountered in Soil Survey

A soil surveyor's task is difficult and he/she is encountered with several practical constraints, as mentioned below.

- The distribution of different soils may be so complex in a field or may occupy so small areas that their delineation on a map, at any practical scale, becomes difficult.
- It is difficult to follow the geographic boundaries of soils because of the vegetative cover and their hidden nature.
- Soil survey and mapping are expensive and many developing nations may not afford them unless they serve the practical needs.
- The inaccessibility of certain areas because of transportation problems may restrict the number of sampling points.

6. Base Maps

Base - Best map

BB

Irrespective of the type of soil surveys, a fundamental requirement of all mapping activities is a suitable base map. These base maps need to be complete in details of features and accurate in their location to enable the surveyor to delineate soil boundaries more correctly and conveniently. Depending on the intensity of mapping, following types of base maps are used for soil surveys in India and in many other countries.

6.1. Cadastral Maps

Cadastral maps on the scale of 1:2640 (24" = 1 mile) to 1:7920 (8" = 1 mile) or 1:15,840 (4" = 1 mile) in plain areas and 1:1200 (52.8" = 1 mile) in hilly areas are used for detailed map-

1:1200 hilly area, 1:2640 - 1:7920, or 1:15840 for plain area.

ping. Cadastral maps show field boundaries and field or revenue survey number; however, they lack topographical details (contours, elevations, etc.).

One advantage of using cadastral base is that the soil survey information or interpretations can be communicated to individual farmers by reference to the field survey number. The cadastral maps can be procured from the village revenue official (*Patwari*) or concerned district revenue officer (*Tehsildar*).

6.2. Topographical Maps

Topographical maps are published on the scale of 1:25,000, 1:50,000 and 1:250,000. These are used as base maps for various types of soil surveys in India. Earlier, the scale of these maps used to be four inch to a mile (1:253,440), one inch to a mile (1:63,360) or even two inch to a mile (1:31,680).

Topographical maps show physical features and contain topographical details also in the form of contours and elevations above the mean sea level for Survey of India benchmarks. These maps have reliable planimetric accuracy facilitating measurement of distances and easy preparation of soil map. In India, topographical maps are prepared and published by the Survey of India, Dehradun, and can be obtained from its head office at Hathibarkala, Dehradun or its regional offices.

6.3. Aerial Photographs

Aerial photographs are the pictures taken by camera fitted in an aircraft and flying over the terrain at a predetermined height, depending on the scale of aerial photography and focal length of camera. Aerial photographs give a bird's eye view of large areas. Typically, successive photographs contain 50 to 65% overlap which is essential for stereoscopic viewing and analysis of stereo pairs. The aerial photographs ranging in scale from 1:8000 to 1:60,000 are used in different types of soil surveys.

In soil surveys, we mostly use panchromatic black and white air photos taken with a black and white film. In these photographs, there are many shades of grey colour. The black and white air photos can indicate lot of information about land forms, vegetation, human interference as well as soils. Natural colour and infrared films are also used for aerial photography, especially for forest areas to discriminate forest types/species.

Relief can be perceived by stereoscopic study of aerial photographs. Relief features help in identifying various kinds of landforms which are related to different types of soils. Many landforms, e.g. terraces, flood plains, sand dunes, coastal plains, plateaus, palaeochannels, hills, valleys and mountains can be recognised on the photographs from their shapes, relative heights and slopes. Differences in tone or colour may also reflect soil differences.

When we look at an aerial photograph, we can see various objects of different sizes and shapes. Some of these objects may be readily identifiable while others may not be, depending on the interpreter's perception and experience. Individual objects like trees, houses, roads, foot paths, field boundaries, lakes, river courses, etc. are imaged clearly depending on the scale of photographs. These 'landmarks' serve as effective 'reference points' or 'local control points' that facilitate a soil surveyor in orientation and navigation during the field work and in demarcating soil boundaries of high 'local accuracy'. Base maps for publication can be prepared from the aerial photographs economically and in a reasonable time.

Both speed and accuracy of soil mapping are increased on using aerial photographs as the mapping base. However, field work cannot be eliminated, the soils must be examined to a depth of 1 to 2 metres or to the depth of solid rocks as the soils are hidden below the surface beyond the present reach of most remote sensors (cameras or scanners).

Aerial photographs have some limitations also in soil survey and mapping. Elevations are not shown on the photographs. Scale is not precisely uniform and the differences of scale between adjoining photographs create some difficulties in matching and transferring of soil boundaries. Distances and directions cannot be measured as accurately as on topographical maps. Nevertheless, the advantages of aerial photographs generally outweigh the limitations

U.V. : < 8.8 mm
Infrared = 0.7 - 3 um
thermal infrared = 3 - 5 um - 8.15 um.
microwave 0.1 - 3.0 cm.

and make them a valuable tool for soil mapping.

6.4. Remote Sensing

Remote sensing is the science of obtaining information about objects or phenomenon in the environment through the use of sensing devices located at a distance without any physical contact between the object and the sensing device. Electromagnetic energy is the means by which information is transmitted from an object to the sensor. The interaction of electromagnetic waves with natural surfaces and atmosphere is highly dependent on the frequency of waves. Waves in different spectral bands (e.g. microwave, infrared and visible regions) tend to excite different interaction mechanisms such as electronic, molecular or conductive.

The Sun's energy commonly referred to as electromagnetic spectrum (EMS), is an electromagnetic radiation characterized by wavelength or frequency. When the radiation is incident on matter, it can be reflected, scattered, absorbed or transmitted. All the processes are strongly dependent on the wavelength of the incident radiation as well as the atomic and molecular structure of the material and its condition. The interaction between the radiation and the object of interest (e.g. soil, land, water, vegetation, crop, snow, building, etc.) conveys information on the nature of the object. This is possible by measuring electromagnetic radiations which are reflected/scattered or emitted by an object using sensing devices. Different objects behave differently and return different amount and kind of radiant energy, depending on the surface conditions, reflection coefficient, emittance, and roughness.

IRS-1A = 1988
" IB = 1991
IC = 1995
" IE = 1993
P2 = 1994
P3 = 1996
ID = 1997
P4 = 1999
P6 = 2003
Cartosat-1 2005
Cartosat-2 2007
IMS-1 = 2008

Earth orbiting satellites equipped with sensors, including cameras provide analogue (imagery) and digital data. Sensors used for remote sensing can be broadly classified as those operating in the Optical-Infrared (OIR) region and those operating in the Microwave region. These can be further sub-divided into 'passive' and 'active' sensors. The sensors which sense the reflected/scattered/emitted radiation from the Earth's surface are *passive sensors* and those which illuminate the targets with their own energy and then measure the scattered or reflected radiations from the targets are called *active sensors*. The active and passive sensors could be classified as imaging sensors and non-imaging sensors. Imaging sensors provide data which has one to one correspondence with several locations in a given area on the ground surface, whereas non-imaging sensors provide data which is the average value over a given area. In other words, an imaging sensor provides a picture, whereas a non-imaging sensor provides a numerical value. The non-photographic imaging sensors can perceive the part of EMS from ultraviolet (wavelength less than 0.38 micrometre) through microwave to the upper wavelength of 100 cm. Remote sensing technology makes use of visible (0.4-0.7 μm), infrared (0.7-3.0 μm), thermal infrared (3-5 μm and 8-14 μm) and microwave (0.1-3.0 cm) regions of EMS to collect information about various objects on the Earth's surface.

Satellite data from Indian and foreign satellites are being used for soil mapping by many national and state agencies in India. Before the launch of first Earth Resource Technology Satellite (ERST-1) in 1972 by the USA (later renamed as Landsat-1), aerial photographs were used as a remote sensing tool for soil mapping. India has well conceived the earth resource satellite programme. Till date (September, 2008), Indian Space Research Organization (ISRO) has launched fourteen Indian Remote Sensing Satellites (IRS) starting with IRS-1A in March 1988, followed by IRS-1B in August 1991, IRS-1E in September 1993 (Satellite could not be placed in orbit), IRS-P2 in October 1994, IRS-1C in December 1995, IRS-P3 in March 1996, IRS-1D in September 1997, IRS-P4 in May 1999, Technology Experiment Satellite (TES) in October 2001, RESOURCESAT-1 (IRS-P6) in October, 2003, Cartosat-1 in May, 2005, Cartosat-2 in January, 2007, Cartosat-2A and Indian Mini Satellite (IMS-1) on 28th April, 2008. The Cartosats 2 & 2A together will help in providing more frequent revisits.

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Remote sensing satellites are polar-orbiting and encircle the earth passing over or near to the north or south poles. These are launched in the near polar, sun synchronous orbits, i.e.

satellite always crosses the Equator precisely at the same local time. Indian remote sensing satellites (IRS-1D & P6) launched at an altitude of 817 km with an inclination of 98.7° in the descending mode cross the Equator at 10:30 AM ± 5 minutes. The Indian remote sensing satellite (IRS-P6) takes 101.35 minutes to complete one revolution around the earth and thus completes about 14 orbits a day. The entire earth is covered in 341 orbits during a 24-day cycle. Earlier, the planned life of satellites was three years, but it is five to seven years presently.

A large number of visible and infrared imaging sensors have been flown into space to study the earth. The imaging systems are divided into three general categories: (a) framing camera systems, e.g. Landsat RBV (Return Beam Vidicon) and Shuttle Large Format Camera, (b) scanning systems, e.g. Landsat Multi Spectral Scanner (MSS) and Thematic Mapper (TM), and (c) Pushbroom Imagers as in IRS (Indian Satellite) and SPOT (French satellite) cameras. The Pushbroom Imagers delete the Scanning Mechanism and use a Linear Array of detectors to cover all the pixels in the across-track dimension. The ground segment sensed at any instant by individual detector element of Charge Coupled Devices (CCDs) is called the ground resolution element (GRE) or ground resolution cell and its representation on the computer screen is called as PIXEL (Picture-element).

The digital image consists of tiny equal areas or picture elements (Pixels) arranged in regular lines and columns. Size of a pixel depends on the Instantaneous Field of View (IFOV) of sensor, e.g. 23.5 m × 23.5 m in the case of IRS LISS-III (Linear Imaging Self Scanner-3) and 5.8 m × 5.8 m in IRS LISS-IV. The brightness of each pixel has a numerical value ranging from zero (black) and some higher number of white, i.e. 127 or 255 for 7 bit and 8 bit radiometric resolution, respectively. Radiometric resolution is a measure of the capability of the sensor to differentiate the smallest change in the spectral reflectance/emittance between various targets/objects.

The relative intensities of light reflected from objects and features on the Earth's surface, has its distinctive band pattern or "spectral signature". In reference to these properties, it is possible to identify and classify objects by measuring the strength of light reflected from the surface. The radiation from the scene or a particular area that a remote sensor detects is measured as radiance not reflectance.

The spatial and spectral resolutions are important sensor parameters from the point of view of resource surveys. The *Spatial resolution* refers to the size of the smallest object or area that can be discriminated by the sensors. Instantaneous Field of View (IFOV) of the sensor defines the ground resolution of multispectral scanner (e.g. 23.5 m for IRS-P6 LISS-III sensor); on the other hand, *Spectral resolution* is a measure of the discreetness of spectral bands and the sensitivity of the sensor to distinguish between grey levels. The narrow spectral band (e.g. 0.52 – 0.59 μm) in which the image is taken, allows us to observe certain features more prominently. Six.

RESOURCESAT-1 (IRS-P6), the tenth satellite in IRS series launched in October, 2003 helped not only in continuing the remote sensing data services made available by the IRS-1C and IRS-1D, but enhanced also the remote sensing services by providing imagery with improved spatial resolution and additional spectral bands. Resourcesat-1 carries the following three imaging sensors operating on the 'Pushbroom Scanning' concept using linear arrays of Charge Couple Devices (CCDs):

(i) A high resolution Linear Imaging Self Scanner (LISS-IV) operating in Mono and Multispectral mode. In multispectral mode, data is collected in three spectral bands, two in visible (0.52-0.59, 0.62-0.68 μm) and one in near infrared (0.77-0.86 μm) region (VNIR) with 5.8 metre spatial resolution LISS-IV camera is steerable up to ± 26 degree across the track to obtain stereoscopic imagery and achieve five-day revisit capability. In multispectral mode, sensor provides data corresponding to preselected 4096 contiguous pixels (out of full 12k pixels) using electronic scanning scheme corresponding to 23.9 km swath. On the other hand, in mono mode, the data

um resolution → IRS. LISS] — 1 : 50000 - at district level,
↳ LAND SAT - TM - 30 M

Coarse resolution — LANDSAT - MSS - 79 M] → Reconnaissance Reginal planing.
↳ AWiFS

High resolution [IRS - LISS - IV] → Detailed and intensive soil survey
IRS - P-6 & Resources sat - 6

of full 12k pixels of any one selected band corresponding to a swath of 70 km can be transmitted.

(ii) A medium resolution LISS-III camera operating in four spectral bands, three in visible and near infrared (VNIR) regions (as in LISS-IV sensor above) and one in short wave infrared (SWIR) region (1.55 – 1.70 μm). It has a 24-day revisit cycle with 23.5 metre spatial resolution and four CCD arrays (one for each spectral band) with 6,000 detector elements to record the data. Each detector element records the data for each pixel of 23.5 m resulting in a swath of 141 km (23.5 m × 6,000 = 1,41,000 m or 141 km).

metre

(iii) An advanced wide field sensor (AWiFS) operating in four spectral bands, three in VNIR and one in SWIR region (as in LISS-III camera) with 56 metre spatial resolution, 5-day revisit cycle and a swath of 740 km.

RESOURCESAT-1 also carried a solid state recorder with a capacity of 120 giga bytes to store the images taken by its cameras which could be read out later to the ground stations.

Imaging from space has two important advantages; firstly, a large area (thousands of square kilometres) can be examined from a single point in the orbit (IRS-1D/P6-LISS-III scene covers an area of 141 km×141 km); secondly, any area can be repeatedly examined on a regular basis for monitoring.

Remote sensing technology provides, accurate, timely and cost-effective information on natural resources. The medium resolution IRS LISS-III (23.5 m) or Landsat TM (30 m) data (digital and analogue), serves as an excellent base for 1:50,000 scale mapping at the district level. The coarse resolution AWiFs (56m) or Landsat MSS (79 m) data can be used as a base map for reconnaissance soil surveys for regional level planning. The high resolution (5.8 m) multispectral data from LISS-IV sensor on board RESOURCESAT-1 (IRS-P6) is useful for detailed and intensive soil surveys. The stereoscopic imagery from IRS-P6 LISS-IV sensor can help in delineating various land forms, micro-depressions, sloping lands, erosion types and drainage pattern and thus facilitates the delineation of boundaries between various soil types.

The availability of high resolution data from WorldView-1 (0.55 m panchromatic-PAN); Quick Bird 1 & 2 (0.61 m PAN, 2.5 m Multi-spectral-Mx); Orb View 3 & 4 and IKONOS 1 & 2 (1.0 m, PAN; 4.0 m, Mx) of USA and Indian satellites, namely Cartosat-1 (2.5 m PAN) and Cartosat 2 & 2A (around 1.0 m – PAN) enables the generation of cartographic quality products as an alternative to aerial photographs. The high resolution single band (0.5 – 0.75 μm) Panchromatic data from Cartosat 1, 2 & 2A as well as Multispectral and Panchromatic data from Quick Bird 1 & 2, Orb View 3 & 4 and IKONOS 1 & 2 is an asset for cartographic applications at the village level and highly intensive soil surveys.

1 = 2.5
2. [illegible]

The high resolution black and white imagery (single band panchromatic from Cartosat 1, 2 & 2A) and False Colour Composites (multispectral data of IRS-P6 LISS-IV sensor) on 1:12,500 or larger scale serve as an excellent base map for large scale mapping as individual fields, field roads, individual trees and well locations, are clearly imaged which serve as effective 'reference points' or local control points during the field work and in demarcating soil boundaries of high local accuracy. The satellite data have the same limitations for soil surveys as in aerial photography, but no difference in scale.

Data from IRS satellites is received, processed and distributed to various users in India by the National Remote Sensing Centre (NRSC), Department of Space (DOS), Hyderabad. The NRSC Data Reception Station at Shadnagar (near Hyderabad) receives data from IRS satellites as well as from foreign satellites (NOAA of USA, ERS 1&2 of European Space Agency). The IRS data is also available to the international users under an agreement between the Antrix Corporation of the Department of Space and the Space Imaging, USA, through a network of ground stations. The National Remote Sensing Centre and Antrix Corporation supply Panchromatic (black & white) and Multispectral (Mx), digital and analog data to various users in India and abroad.

7. Examination and Description of Soils

During field operations, soils are studied by examining the soil material removed by an auger or sampling tube. The soils can also be studied by digging minipits up to 0.50 m depth and going below to a depth of 1.5-2.0 metre through augering. The samples of soils are studied for their colour, texture, consistence, mottling, concretions, pH, and effervescence.

Important properties of a pedon are not clearly observable in the small sample extracted with auger due to mixing, and detailed study of the entire pedon is needed. The study of pedon is based on the examination of a soil profile.

7.1. Soil Profile

A soil profile is a vertical section of soil up to a depth of 1.5-2.0 metres in deep soils (Figure 1). In the case of shallow soils, the vertical cut is made up to bedrock or up to water table in the case of waterlogged soils. Width of profile ranges from 1 metre to several metres. Road cuts or brick kiln or querry sites are several metres wide which can be examined for clear perception of soil properties of a pedon. The description on the entire pedon or a sample within it, should record the kinds of layers, their depth and thickness and properties of each layer. Horizons or layers are studied in both horizontal and vertical dimensions.

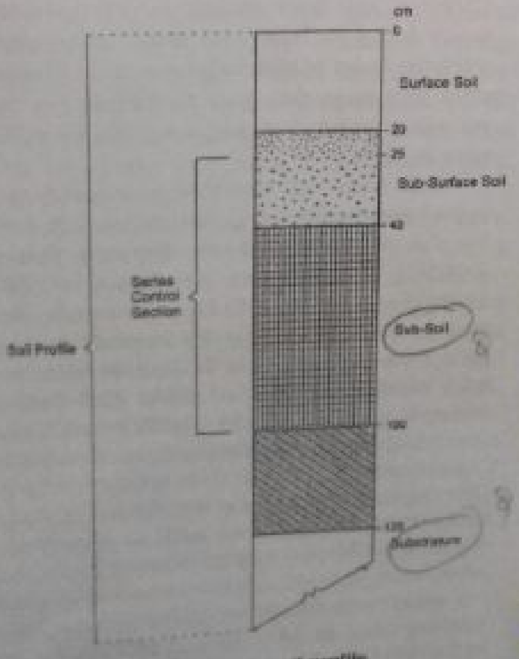

Figure 1. A soil profile

7.2. Soil Horizon

A soil horizon is a layer approximately parallel to the surface of soil, identifiable from adjacent layers by a distinctive set of properties produced by the soil forming processes over a period of time. In soils, where genetic horizons are not developed as in flood plain or sand dune soils and the properties are mostly inherited from the parent material, the term 'layer' instead of 'horizon' is used. The solum (plural sola) of a soil consists of a set of horizons that are related through the same cycle of pedogenic processes. The horizons above the parent material are collectively regarded as solum.

For selecting a profile site, following points should be considered:

- It should be typical of the soil texa (soil series or families) within the mapping unit,
- It should be away from a tree, an irrigation channel/ditch/river, human settlements and road, as these prevent/obliterate the normal development of a soil
- If possible, a virgin area should be preferred.

8. Map Units

A soil map delineates the areas occupied by different kinds of soils, each of which has a unique set of inter-related properties, characteristic of material from which it was formed, its environment and history. A map unit is a collection of soil components or miscellaneous areas or both. Each map unit differs in some respect from all others in a survey area and is uniquely identified on a soil map. Map units consist of one or more components. An individual component of a map unit represents the collection of polypedons or parts of polypedons that are members of the taxon or a kind of miscellaneous area (inclusion). Soil boundaries are shown on maps by lines. The area enclosed by a boundary is called a "soil delineation".

Map unit :- ① Consociation ② Complexy
③ Association ④ undifferentiate gro

A map unit consists of all soil delineations which are identified by a unique symbol, colour or name. There are always several map units on a map and an organised list of map unit symbols, colours and names is called map legend.

8.1. Taxonomic Units versus Map Units

The examination of soils in the field and laboratory provides the basis for placing them into taxonomic and mapping units.

Taxonomic units define the specific ranges of soil properties in relationship to the total range of properties measured in the soil. The map units and their individual delineations define areas on a landscape. Almost every map unit has more than one taxonomic units (soil series or phases of soil series, soil families or other texa).

A *soil series* is a taxonomic unit which represents a group of soils having horizons similar in characteristics and arrangement in a soil profile developed from a particular type of parent material, but they may differ in the surface texture. It is given a geographical name of either the locality where it is well developed or the place where it is recognized first.

Soil phase is a subdivision of a soil series or any other taxa in respect of surface soil texture, degree of erosion, stoniness/rockiness and salinity/sodicity. All these aspects have practical significance in agricultural and engineering interpretations.

8.2. Inclusions within Map Units

In all soil surveys, virtually every delineation of a map unit includes areas of soil components or miscellaneous areas (inclusions) that are not identified in the name of the map unit. Many areas of these components are too small to be delineated separately.

Inclusions reduce the homogeneity of map units and may affect interpretations.

8.3. Kinds of Map Units

Four kinds of map units are used in soil surveys to show the relationships of different soils on the landscape. These are: consociations, complexes, associations, and undifferentiated groups.

8.3.1. Consociation

In a consociation, delineated areas are dominated by a single soil taxon. As a rule, at least one-half of the pedons in each delineation of a soil consociation are of the same soil components that provide the name for the map unit.

8.3.2. Complexes and Associations

Regular
Bet
diss-

Complexes and associations consist of two or more dissimilar components occurring in a regularly repeating pattern. In each delineation of either a complex or an association, each major component is normally present, though their proportions may vary appreciably from one delineation to another.

8.3.3. Undifferentiated Groups

Undifferentiated groups consist of two or more taxa components that are not consistently associated geographically and therefore do not always occur together in the same map delineation. These taxa are included as the same named map unit because use and management are the same or very similar for common uses.

8.4. Naming of Map Units

Map units or groups of map units can be given any name derived from any discipline, one may call a map unit "alluvial soils", 'forest soils', "mountain soils". To be more specific one has to refer *Soil Taxonomy* (Soil Survey Staff, 1975; 1999) or *Keys to Soil Taxonomy* (Soil Survey Staff, 1990; 2003) for map unit names, for example the name "Ustept" could be used to identify a suborder of the order Inceptisol. The objectives of a survey determine the kind of map units and the taxonomic level used to identify components of map units. For example, the name Ludhiana sandy loam, 0-1% slope indicates that the soils of Ludhiana series (a Ustept) are dominant in that mapping unit.

The names of taxa of higher categories that is, suborder (Ustept), great group (Haplustept), subgroup (Typic Haplustept), family (Course loamy Typic Haplustept) are also used in map unit names, especially on small scale maps.

9. Kinds of Soil Surveys

All soil surveys are conducted by examining, describing and classifying soils in the field and delineating their areas on the maps. Soil Survey Manual (Soil Survey Staff, 1951; 1993) and Soil Survey Manual of All India Soil and Land Use Survey (AISLUS, 1971) recognize the following three types of surveys, depending on the objectives, methods, types of base material, intensity of field survey and type of map units:

(1) Detailed surveys

(2) Reconnaissance surveys

(3) Detailed - Reconnaissance surveys

To these major kinds, *semi-detailed* and *exploratory or rapid reconnaissance* surveys were added later, these surveys lead to small scale soil maps needed for macro-level planning.

The relationship between scale of soil map and frequency of observations in different kinds of surveys is described in Table 1 and the details of each type of survey are as follows:

9.1. Detailed Survey

20 mapping un

In the detailed soil map, the boundaries between mapping units are plotted on base maps (cadastral) or aerial photographs or satellite imagery (high resolution panchromatic data) from observations made throughout their course.

Cadastral maps on 1:2640, 1:7,920 or 1:15,840 scale and aerial photographs on

Table 1. Relationship of scale of soil map and frequency of field observations

Kind of survey	Scale of base map	Area represented by 1 cm^2 on map (ha)	Distance between field observations	Frequency of observation	Mapping unit	Field procedure and accuracy of soil boundaries
Rapid Reconnaissance/ extensive (for selection of area for more intensive study)	1:1,000,000 or smaller	10,000	10 km	1 in 10,000 ha	Phases of soil great groups	Soil identification is done by traversing representative areas to determine soil pattern and composition of map units. Boundaries are plotted mostly by interpretation of remotely sensed data.
RR Reconnaissance (for regional planning)	1:250,000 1:100,000	625 100	2.5 km 1.0 km	1 in 625 ha 1 in 100 ha	Phases of association of soil series families/great groups	Soil identification is done by traversing sample strips, boundaries plotted by interpretation of remotely sensed data and verified by random observations.
SD Semi-detailed (for district planning)	1:50,000	25	500 m	1 in 25 ha	Phases of soil series/association of soil series/families	Boundaries in each delineation of sample areas are identified by actual traversing
Detailed (for village/watershed planning)						
• Low intensity	1:10,000	1	100 m	1 per ha	Phases of soil series	Almost all boundaries are checked by actual traversing; remote sensing data are used as an aid in boundary delineation
• High intensity	1:5,000	0.25	50 m	4 per ha	Phases of soil series	All boundaries are checked by actual traversing

1:8,000 to 1:15,000 scale and satellite data on 1:12,500 scale are used, depending on the intensity of survey and the agricultural development needs of an area. Mapping units of this survey are narrowly defined phases of soil series (*refer* Table 1 for more details).

In detailed soil surveys, the maximum distance between route or traverses is 250 m or even closer (50-100 m), depending upon the scale of the map and the complexity of the soil pattern. Although soil boundaries are not actually traversed, they are plotted based on the observations made throughout their courses.

Detailed soil surveys furnish information needed for the proper assessment of soil properties, terrain features, erosional aspects and related factors. Such surveys are time consuming, expensive and are recommended only for priority areas such as pilot projects, agricultural research stations, agricultural farms, micro-watersheds and for areas earmarked for urban development.

9.2. Reconnaissance Survey

This type of survey is undertaken to prepare soil resource inventory of large areas. They furnish information to precede detailed or semi-detailed soil surveys. On a reconnaissance soil map, boundaries between the mapping units are plotted from soil observations made through augar sampling at an interval of 2.5 km to 1 km. The representative profiles of various soils are studied at an interval of 3-6 km or even shorter, depending on the soil heterogeneity or variability. In reconnaissance soil survey, maps are prepared relatively at smaller scale, ranging from 1:250,000 to 1:100,000 depending on the purpose. However, field mapping is frequently done at 1:250,000 or 1:50,000 scale in order to use topographical maps, aerial photographs or satellite imagery for the delineation of land types and planning of field traverses. At reconnaissance level, association of soil series, families and great groups are mapped. The variations in the land type, land use, tone and texture, as observed in aerial photographs or satellite imagery, are used as the basis for boundary delineation.

9.2.1. Rapid Reconnaissance Survey

In a rapid reconnaissance survey, field mapping is done at 1:1,000,000 or still smaller scale using the satellite imagery. Soils are mapped by traversing representative areas. The mapping units are phases of great groups.

9.3. Detailed-Reconnaissance Survey

The detailed reconnaissance survey is not really a separate kind but in this type of survey, a part of the area is covered by detailed survey and the remaining by reconnaissance survey.

It is carried out in individual areas best suited to each of these types. A reconnaissance survey is carried out over an area having limited potentialities for intensive development, in the usual way classifying and grouping the soils into soil series, or association of soil series as necessary, followed by mapping. The remaining areas which show potentialities for priority attention are covered by detailed or high intensity surveys in the standard pattern.

9.4. Semi-Detailed Survey

This survey comprises a highly detailed study of some selected sample strips cutting across many physiographic units and soils. Once the correlation between soils and physiographic units is developed, the rest of the area is checked at random and soil boundaries based on physiographic units are delineated. This kind of survey provides sufficient information about various kinds of soils including problematic or degraded soils. Scale of base maps (aerial photographs or satellite imagery) used is 1:50,000. Mapping unit is the association of soil series or families. The final maps are prepared on 1:50,000 scale.

10. Orders of Soil Survey

Every soil survey is undertaken with a purpose or a set of purposes in mind. Some soil surveys are made to serve users who need precise information about the soil resources of an area, a few hectares or less in size. These surveys require refined distinctions among small, homogeneous areas of soils. Some other surveys are made for users who need a broad perspective of heterogeneous, but distinctive areas of thousands of hectares.

Different intensities of field study and different degrees of details in mapping produce a wide range of soil surveys. The *USDA Soil Survey Manual* (Soil Survey Staff, 1993) recognizes five orders of soil survey. The order of a soil survey is a consequence of field procedures, the minimum size of delineation, and the kinds of map units that are used. The brief details of each type of survey are given in Table 2 (Key for identifying orders of soil survey).

IBLS EBCF

OSRPCS: EBLP

— planning

11. Soil Survey and Mapping Procedure

The soil survey methods have undergone large changes during the recent past with the availability of satellite imagery, apart from topographical maps and aerial photographs.

The soil survey and mapping involve the following steps:

(a) Preliminary reconnaissance of the area to investigate the major soils and their pattern of occurrence.

(b) Procurement of required base maps. Aerial photographs, satellite imagery, topographical and other maps are useful references and are used as the mapping base.

(c) Preparation of mapping legend based on the preliminary field studies. The mapping legend is modified as the survey progresses.

(d) Stereoscopic study of aerial photographs and interpretation of satellite imagery or digital data for the identification and delineation of land forms (hills, valleys, terraces, flood plains, coastal plains and sand dunes) and their sub-divisions based on the differences in tone, relief and vegetation. The differences in tone or colour serve as the basis for drawing tentative boundaries and for predicting types of soils. The predictions are verified in the field, but preliminary interpretation can increase the quality of mapping.

(e) The soils are examined at some standard interval along the traverse to locate important differences in soil properties. Interval between traverses and soil observation points cannot be specified. It depends on the scale of mapping and variability of soil properties (Tables 1 and 2 provide some guidelines).

(f) *Plotting of Soil Boundaries*: Some soil boundaries are sharply defined, others are plotted as lines mid-way in zones of gradual transition from one soil to another. The predicted boundaries are checked throughout their course in high intensity surveys and in other cases, the boundaries are plotted mostly by interpretation of remotely-sensed data and verified with some observations.

(g) Classification of soils and naming of map units based on the morphological, physical and chemical properties of soils forming the map unit.

(h) Preparation of final legend and finalization of soil map.

12. Examination and Description of Soil Profile

Different layers or horizons are marked on the basis of characteristics that can be visually observed or measured in the field, such as colour, texture, structure, consistence and presence of carbonates. Each horizon is described for the following characteristics:

(i) Horizon symbol
(ii) Depth of each horizon/layer (cm)
(iii) Soil colour under wet, moist or dry conditions
(iv) Mottling
(v) Soil texture
(vi) Soil structure
(vii) Soil consistence under dry, moist and wet conditions
(viii) Cutans (ped coatings), pressure faces, and slickensides
(ix) Nodules/concretions and cementation
(x) Content of carbonates, and soluble salts
(xi) Roots
(xii) Nature of the boundary with the horizon below
(xiii) pH
(xiv) Hard pans
(xv) Pores
(xvi) Lithorelics
(xvii) Artefacts (pottery pieces)
(xviii) Features of biological origin

Table 2. Key for identifying kinds of soil surveys

Level of data needed	Field procedures	Land area represented by the minimum size delination (ha)[1]	Typical components of map units	Kinds of map units[2]	Appropriate scales for field mapping and publication
1st order- Very intensive (i.e. experimental plots individual building sites)	The soils in each delineation are identified by transecting or traversing. Soil boundaries are observed throughout their lengths. Remotely-sensed data are used as an aid in boundary delineation.	1 or less	Phases of soil series; miscellaneous areas	Mostly consociations; some complexes	1:15,840 or larger
2nd order - Intensive (i.e. general agriculture, urban planning)	The soils in each delineation are identified by transecting or traversing. Soil boundaries are plotted by observation and interpretation of remotely-sensed data. Boundaries are verified at closely spaced intervals.	0.6 to 4	Phases of soil series; misc. areas; few named at a level above the series	Consociations and complexes; some associations and undifferentiated groups	1:12,000 to 1:31,680
3rd order - Extensive (i.e. range land, forest land, community planning)	The soils are identified by transecting representative areas and by some transects. Boundaries are plotted mostly by interpretation of remotely-sensed data and verified with some observations.	1.6 to 16	Phases of soil series or taxa above the series; misc. areas	Mostly associations or complexes; some consociations and undifferentiated groups	1:20,000 to 1:50,000 or 1:63,360
4th order - Extensive (i.e. regional land use planning)	The soils are identified by transecting representative areas to determine soil patterns and composition of map units. Soil boundaries plotted by interpretation of remotely-sensed data.	16 to 252	Phases of levels above the series; misc. areas	Mostly associations; some consociations, complexes and undifferentiated groups	1:50,000 to 1:250,000
5th order - Very extensive (i.e. selection of areas for more intensive study)	The soil patterns and composition of map units are determined by mapping representative areas and applying the information to like areas by interpretation of remotely-sensed data. Soils are verified by occasional onsite investigation or by traversing.	252 to 4,000	Phases of levels above the series; misc. areas	Associations; some consociations, and undifferentiated groups	1:250,000 to 1:1,000,000 or smaller

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1. This is about the smallest delineation allowable for readable soil maps. In practice, the minimum size delineations are generally larger than the minimum size shown.
2. Where applicable, all kinds of map units (consociations, complexes, associations, undifferentiated groups) can be used in any order of soil survey.

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The description of profile and site characteristics which are used as criteria for placing different soils in various texa in Soil Taxonomy are discussed in *Soil Survey Manual* (Soil Survey Staff, 1993); NBSS & LUP *Field Manual* (Sehgal *et al.*, 1987); and NBSS&LUP *Technical Bulletin* (Sehgal, 1992).

12.1. Horizon Designations

The capital letters O, L, A, E, B, C, R and W represent the master horizons and layers of soils. The letters are the base symbols to which other characters are added to complete the designations. Most horizons and layers are given a single capital letter symbol; some require two.

12.1.1. O-horizons

O-horizons are layers dominated by the organic material. Some of these layers have been saturated with water for long periods or were once saturated but are now artificially drained; others have never been saturated.

12.1.2. L-Horizons or Layers

Limnic horizons or layers include both organic and mineral limnic materials that were either (i) deposited in water by precipitation or through the actions of aquatic organisms, such as algae and diatoms, or (ii) derived from underwater and floating aquatic plants and were subsequently modified by the aquatic animals.

12.1.3. A-Horizons

A-horizons constitute mineral horizons that are formed at the surface or below an O-horizon. These exhibit obliteration of all or much of the original rock structure and show one or more of the following : (a) an accumulation of humified organic matter intimately mixed with the mineral fraction and not dominated by properties characteristic of E- or B-horizons (defined below) or (b) properties resulting from cultivation, pasturing or similar kinds of disturbances.

12.1.4. E-Horizons

E-horizons are mineral horizons in which the main feature is the loss of silicate clay, iron, aluminum or some combinations of these, leaving a concentration of sand and silt particles. These horizons exhibit obliteration of all or much of the original rock structure.

12.1.5. B-Horizons

These horizons are formed below an A, E or O-horizon and are dominated by obliteration of all or much of the original rock structure and show one or more of the following characteristics:

(i) Illuvial concentration of silicate clay, iron, aluminum, humus, carbonates, gypsum or silica, alone or in combination,

(ii) Evidence of removal or addition of carbonates,

(iii) Residual concentration of oxides,

(iv) Coatings of sesquioxides that make the horizon conspicuously lower in colour value, higher in chroma, or redder in hue than overlying and underlying horizons, without apparent illuviation of iron,

(v) Alteration that forms silicate clay or liberates oxides or both and that forms granular, blocky or prismatic structure, if volume changes accompany changes in moisture content,

(vi) Brittleness, and

(vii) Strong gleying.

12.1.6. C-Horizons

C-horizons are layers, excluding strongly cemented and harder bedrock, that are little affected by the pedogenic processes and lack properties of O-, A-, E- or B-horizons. Most are mineral layers. The material of C-layers may be either similar or dissimilar to that from which the solum has presumably been formed. The C-horizon many have been modified even if there is no evidence of pedogenesis.

12.1.7. R-Layers

Strongly cemented to indurated bedrock, granite, basalt, quartzite and indurated limestone or sandstone are the examples of bedrock that are designated as R-layers. A R-layer is sufficiently coherent when moist to make digging with spade impractical, although the layer may be chipped or scraped.

12.1.8. W-layers (Water)

This symbol indicate water layers within or beneath the soil. The later is designated as Wf if it is permanently frozen and as W if it is not permanently frozen. The W (or Wf) designation is not used for shallow water, ice or snow above the soil surface.

For more details one may refer to *Keys to Soil Taxonomy* (Soil Survey Staff, 2003).

12.2. Transitional or Combination Horizons

These are horizons dominated by properties of one master horizon but possess subordinate properties of another horizon also. Two capital letter symbols like AB, EB, BA, BE, BC or CB are used to represent these horizons.

The master horizon symbol which is given first, designates the kind of horizon whose properties dominate the transitional horizon, i.e.

AB/EB - Transition to B, more like A or E than B.
BA - Transition to A, more like B than A.
BE - Transition to E, more like B than E.
BC - Transition to C, more like B than C.
CB - Transition to C, more like C than B.

Horizons in which distinct parts have recognizable properties of the two kinds of master horizons are indicated by the capital letters. The two capital letters are separated by a virgule (/), such as E/B, B/E or B/C.

12.3. Subordinate Distinctions within Master Horizons and Layers

Lower case letters used as suffixes to designate the specific kind of master horizons and layers are given in Table 3.

For additional details, *USDA Soil Survey Manual* (Soil Survey Staff, 1993) and *Keys to Soil Taxonomy* (Ninth Edition, 2003) may be consulted.

Table 3. Lower case letters used as suffixes to designate specific kind of master horizons

Letter(s)	Description
a	Highly decomposed organic material
b	Burried genetic horizon
c	Concretions or nodules
co	Coprogenous earth
d	Physical root restriction
di	Diatomaceous earth
e	Organic material of intermediate decomposition
f	Frozen soil and water
ff	Dry permafrost
g	Strong gleying
h	Illuvial accumulation of organic matter
i	Slightly decomposed organic material
j	Accumulation of jarosite (potassium or iron sulphate mineral)
jj	Evidence of cryoturbation
k	Accumulation of carbonates
m	Cementation or induration
ma	Marl
n	Accumulation of sodium
o	Residual accumulation of sesquioxides
p	Tillage or other disturbances
q	Accumulation of silica
r	Weathered or soft bedrock
s	Illuvial accumulation of sesquioxides and organic matter
ss	Presence of slicken sides
t	Accumulation of silicate clay
v	Plinthite
w	Development of colour or structure
x	Fragipan character
y	Accumulation of gypsum
z	Accumulation of salts more soluble than gypsum

13. Geoinformatics

The deep understanding of earth has been limited due to lack of information. Initially, the mapping was undertaken manually and various resource maps generated were overlaid to study the earth resources in an integrated form. Later, this art was taken over by the efficient computers. The voluminous data provided by the satellites in analog and digital format reinforced the use of computers.

Geoinformatics is a fast emerging science encompassing the modern tools of Remote Sensing, Geographic Information System (GIS), Global Positioning System (GPS) and simulation models. The combination of these technologies provides a cost-effective means of acquiring high resolution real time data through remote sensing, georeferencing the ground truth data with GPS, data management, and analysis through GIS and utilisation of the information for a specific purpose. It is a new discipline that integrates elements of various disciplines dealing with geographic data. The key element that differentiates geoinformatics from other areas of information technology is that all input data is geocoded, i.e. has an address in the 3-D space and is linked to some locality on the surface of earth. Thus, geoinformatics is nothing but application of information technology for the study and management of earth resources (soil, land and forest)

13.1. Geographical Information System (GIS)

Until recently, soil maps (usually in paper form) generated through field surveys were used for a wide variety of applications and decision making, e.g. soil conservation and land use planning, selection of sites for water harvesting structures or industrial parks, alignment of roads and drains, etc.

The "spatial" information in map form with proper scale, legend, symbols and colours, enhances the understanding of "Information Inter-relationships" and thus contributes to a more appropriate location-specific developmental strategy. However, it is not easy to analyse information contained in various types of maps (in paper form), viz. soils, landuse, slope and water table depth for the same area and arrive at a logical decision like alignment of a drain for the disposal of rainwater or reclamation of waterlogged area. In recent years, the demand for storage, analysis and display of complex and voluminous data for resources (e.g. soil, water, forest) has led to the use of computers for data handling and creation of sophisticated information and decision support systems. Geographical Information System (GIS) is a new technology which is becoming an essential tool for the analysis of great diversity of data in a short time.

The GIS is computer-based systems that deal with virtually any type of information about features that can be referenced by geographic location. GIS is an integrated database management system in which large volume of georeferenced spatial data derived from field surveys, i.e. soil surveys, aerial surveys, and space remote sensing, in addition to the existing maps and reports, are efficiently stored, organized, manipulated, retrieved, analysed and displayed/presented as per user's requirement. This involves bringing together diverse information from a variety of sources on a common platform. It requires effective matching of similar entities and demand information consistency across the data sets. Any data that can be mapped has both locational and non-locational characteristics. For example, a feature may exist at an X, Y location and possess an attribute Z. The attributes can be both qualitative (soil units, namely soil series, soil family or their associations) or quantitative (e.g. elevation, slope, status of N or Zn in soil).

Many of these data have some type of explicit or implicit geographical reference associated with them. An explicit geographical reference might be a Cartesian or latitude/longitude co-ordinate and an implicit reference might be the name of a town or village or a physical feature, such as a mountain or river. In many cases, it is possible to derive an explicit reference from an implicit reference. The geographical reference has proved to be an effective means of linkage of data sets and this principle perhaps more than any other is the reason for the success of GIS.

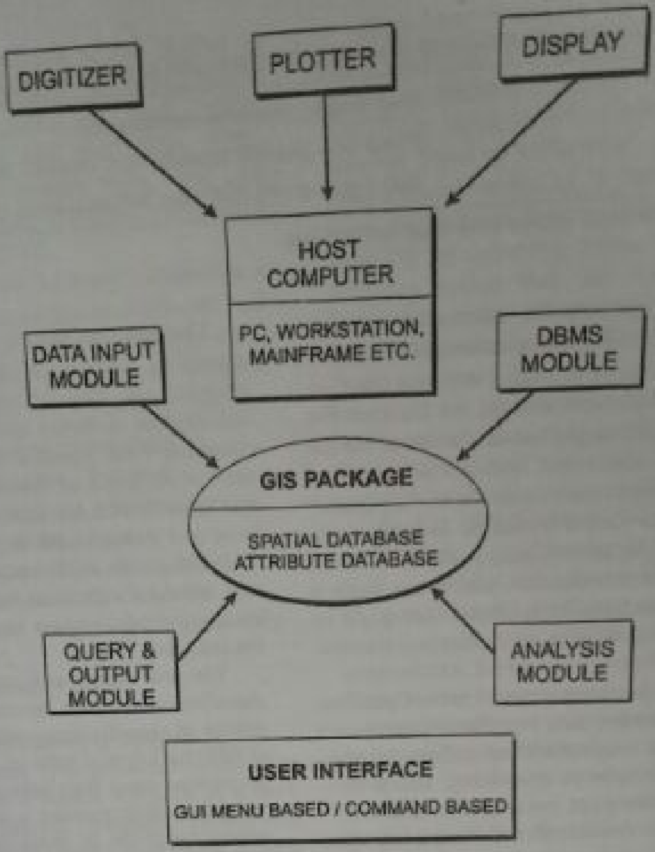
As a technology, GIS has evolved through the following three broad application domains:

(i) **As an information database:** A means of coordinating and accessing geographic data.

(ii) **As an analytical tool:** A means of specifying logical and mathematical relationships among map layers (modelling) to yield derivative maps.

(iii) **As a decision support system:** A means for deciding how to act upon the analysis produced and simulation of after effect.

A GIS database usually has a spatial component for the storage and processing of data. Hence, it has the potential to store the data and create map like products. It also offers the potential for performing multiple analyses or evaluation of scenarios such as model simulations. Any GIS comprises three basic components, viz. computer hardware, application software and a proper organizational context (Figure 2).



GUI - Graphical User Interface
DBMS - Database Management System

Figure 2. Components of geographical information system (GIS)

The hardware components include several specialized peripherals, such as digitizer or scanner for converting the resource maps into digital form, a plotter for graphical representation of the maps generated, and a visual colour graphics display unit on which the spatial data editing and display can be performed by the user. In addition to a central processing unit, and standard computer peripherals, computer software is generally designed to undertake particular tasks or solve problems. To use GIS, one has to learn how to apply various tools available in a GIS software as per specific needs.

The salient features of GIS are:

- Large volume of data from a variety of sources and formats can be integrated efficiently.
- Analysis can be carried out speedily and relevant spatial (map form) and non-spatial (attributes) information can be derived accurately with high consistency.
- Provides graphical representation and easily understandable answers to complex questions in rational use of natural resources, infrastructural development, conservation planning and decision-making in a cost-effective manner.
- Spatial modelling can be carried out conveniently through multilayer integration.
- Easy interfacing with remotely sensed data, and socio-economic data.
- Enables alternate scenarios to select the best option.

Table 4. Examples of attributes* for point, line and area features in a GIS

Point feature	Well (depth, chemical constituents)
Line feature	Power line (service capacity, age, insulator type)
Area feature	Soil mapping unit (soil type, texture, colour, permeability)

*Attributes have been shown within the parentheses

The GIS permits the automated mapping or display of the locations of features, and provides capability for recording and analyzing descriptive characteristics about the features. For example, a GIS might contain a "map" of the locations of roads and also a "database" of descriptors about each road. These "attributes" could include such information as road width, pavement type, speed limit, number of traffic lanes, date of construction, and so on. Table 4 lists other examples of attributes that might be associated with a given point, line, or area feature.

In GIS, the data are stored in multiple files. Each file contains data in a co-ordinate system (latitude and longitudes) that defines the position for each data point or entry. Characteristics of the data point are stored as "attributes". The data are associated with their geographic features. For example, water quality data would be associated with a sampling site represented by a point. Data of crop yields might be associated with field or experimental plots, represented on a map by polygons. A database of individual files is developed and the combined files may contain characteristics or attributes such as stream locations, topography, water or soil chemical characteristics, management practices, ownership details, point sources (wells), and any other data that can be collected and have meaning for the analysis.

There are two kinds of databases, viz. spatial and non-spatial that are integrated in the GIS.

- The spatial data is in the form of maps which could be topography, soil types, geology, and forests, and is stored as layers in the digital form in computer.
- The non-spatial data pertains to attribute information in the form of statistics, tables, and lists, which could be on demography, rainfall, wildlife census, etc.

The data sets need to be standardized with respect to formats, exchange, networks and software, etc. The database in a standard format enables different technologies and applications to work together. It is better to involve users at all levels of developments of database standards so that the problems related to each level and theme could be properly taken care of and the users could understand the information attached with each layer in the form of standard codes when these standards are implemented.

The soil survey and mapping help in the classification of soils according to a standard system of classification, delineating the extent of different kinds of soils on a map and making predictions about their behaviour. The soil survey information answers a wide range of soil-related questions in terms of their use, management and environmental protection. Soil survey data make up growing number of geographical information systems and models that deal with regional planning, erosion prediction, crop yields and even modelling the global change. The soil survey data has to be preprocessed and standardized to be put into a form that can be integrated with other data. This involves digitization of soil boundaries and other spatial features into a computer database as points, lines and geographically defined areas called 'polygons'. The analog (paper) maps are digitized using a digitizer or scanner. The soil database attached to different map units (polygons) includes soil profile description, i.e. horizon/layer-wise morphological, physical and chemical properties in the form of SOIL.DAT.

In the standardized database attached to digital soil map, the soil layer is linked to two tables, SOIL.DAT and SOIL.LUT. The SOIL.DAT contains profile description for each soil class. It is linked to soil layer attribute table via SOIL-CODE. SOIL.LUT describes codification scheme of soil classes up to series level. It is linked to the layer attribute table via three key fields, viz. SOIL-CODE, ASS 1 (CODE) and ASS 2 (CODE). SOIL-CODE is a 16-digit number code, with each two digits representing different levels up to soil series (ISRO, 2000).

The Nabha series in Punjab classified as (Fine silty, mixed, hyperthermic family of Typic Haplustepts as per Soil Taxonomy (for details refer Chapter 2) has been denoted by SOIL-CODE: 06-04-05-18-10-07-05-02, in which:

06 represents the order **'Inceptisol'**;
04 represents sub-order, **'Ustepts'**;
05 represents great group, **'Haplustepts'**;
18 represents sub-group **'Typic'**;
10 represents textural class, **'Fine Silty'**;
07 represents mineral class, **'Mixed'**;
05 represents soil temperature regime, **'Hyperthermic'**;
02 represents series, **'Nabha'**.

13.2. *Global Positioning System (GPS)*

The GPS has revolutionized positioning concepts though it started primarily as a navigation system. The US Navigation System with Time and Ranging Global Positioning System (NAVSTAR GPS) is satellite-based radio navigation system providing precise three-dimensional position, navigation and time information to suitably equipped users. GPS receivers passively receive signals but they do not transmit. There are atleast 24 operational GPS satellites at all times. This is the optimum size of a constellation. The satellites are operated by the US Air Force, with a orbital period of 12 hours. Ground stations are used to precisely track each satellite in orbit. Each GPS satellite has atomic clock on board and transmits data that indicates its location and time. All the GPS satellites synchronize operations so that these repeated signals are transmitted at the same instance. The distance to the GPS satellites can be determined by estimating length of time it takes for their signals to reach the receiver. When the receiver estimates distance to at least four GPS satellites, it can calculate its position in three dimensions (latitude, longitude and altitude).

Most hand held GPS units have about 5-10 metre locational accuracy. To obtain better accuracy, a technique called Differential GPS (DGPS) is used. DGPS requires an additional receiver fixed at a "known" location nearby. Observations made by a stationary receiver (base station) are used to correct the position recorded by rovering units, resulting in an accuracy greater than one metre. Earlier, in order to limit the accuracy for a civilian user, the US Government in 1990 deliberately introduced an error in the orbit data and clock accuracy for L1 signals. This part of GPS operation is called "Selective Availability" and has been discontinued since May, 2000 due to wider civilian applications of GPS.

Global Navigation Satellite System (GLONASS) being developed by Russia is a rival to the US Global Positioning System and Europe's Planned Galileo System. India is a participant in both GLONASS and Galileo programmes. Europe is developing the 30 Satellite Galileo network as a navigational system, independent of any military control. Russia and India are jointly developing a new generation of global positioning satellite (GLONASS-K) for wider navigational applications which will function parallel with the American GPS. Russia intends to operationalize GLONASS by 2010 by completing the cluster of 24 satellites. The GLONASS will provide commercial services to surface transport, shipping and airlines. As per ISRO Chairman, by launching seven satellites, the Indian Regional Navigation Satellite System (IRNSS) is expected to be operational by 2012. The IRNSS will be used for surveying, telecommunication, transport, identification of disaster areas and public safety among others. India is planning to develop its own version of GPS.

Besides the navigational applications, the GPS can be effectively used in location of precise control points, conducting geophysical and cadastral surveys and updating the existing base maps. The wells can be precisely mapped and georeferenced samples of water, soil and plants can be collected. This helps in revisiting the earlier surveyed spots for re-sampling and monitoring. A similar approach can be adopted for mapping the extent of spread of disease in crops and its monitoring.

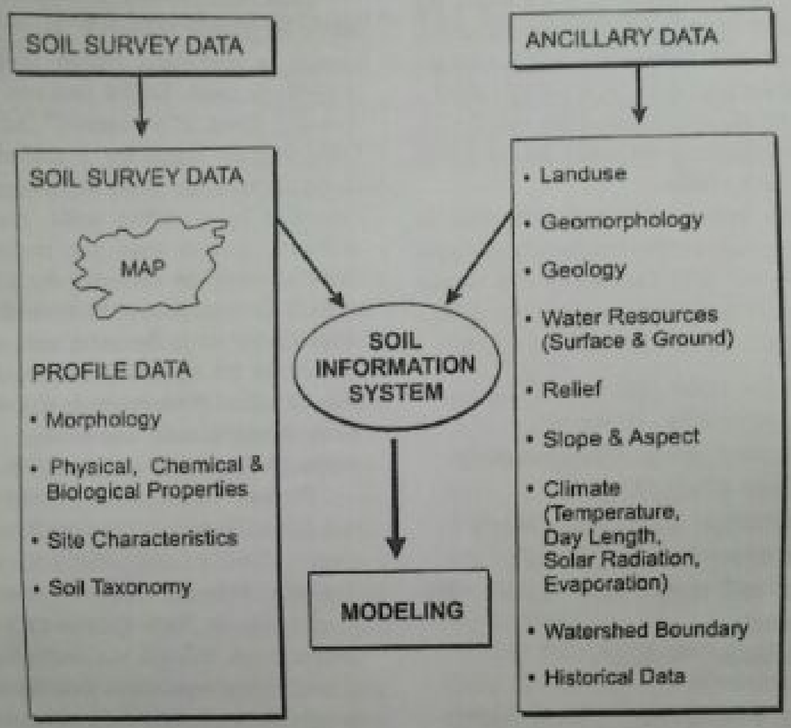


Figure 3. Soil information system: A concept

GPS technology can be applied for mapping soils. Soil surveyors, traditionally, use large-scale base maps or air photos or high resolution imagery to ascertain location on the ground. However, in flat terrain or heavily vegetated areas, these aids are of limited use. GPS has come to the aid of soil mappers to precisely record the geographic co-ordinates of each soil observation point on the ground. The co-ordinates (latitudes and longitudes) given by the GPS receiver for each observation, i.e. auger or profile location are transferred to the base map which in turn help in delineating the location of soil bodies as the surveyor traverses the landscape.

13.3. Soil Information System (SIS)

It is a computerized database system where soil and related data can be organized, stored, retrieved, analysed and processed to make it accessible to the end-users in the form of maps and tables. The SIS is based on a database obtained through remote sensing and ground survey in combination with GIS and Decision Support System (DSS). The concept of Soil Information System is depicted in Figure 3.

The International Society of Soil Science (ISSS), now called as the International Union of Soil Sciences (IUSS), officially endorsed the proposal on the establishment of uniform World Soil and Terrain Digital Database (SOTER) at a scale of 1:1,000,000 during its 13th Congress at Hamburg in 1986. The main function of this database is to provide the necessary data for improved mapping, modelling and monitoring of changes in world soil and terrain resources and presenting a wide range of accurate and timely interpretative analysis for decision- and policy-makers.

Besides soil survey data (soil map and profile data), ancillary data in respect of landuse, geology, geomorphology, water resources, topography (relief, slope, and aspect) and watershed (Figure 3) is collected and used for developing soil information system. Land use/land cover information is essential for the development and management of natural resources. Land use refers to human activities associated with a specific piece of land (i.e.

land used for growing crops, orchards and fodder. On the other hand, land cover is related to the types of features present on the earth, for example houses, factories, lakes, and vegetation).

Information about rocks and minerals is essential to understand the kinds of parent materials from which the soils developed. Both consolidated and unconsolidated materials beneath the solum that influence the genesis and behaviour of soil are described in standard terms (For more details, *see* Chapters 2 and 13).

The information about the land forms is essential as different types of soils are formed on various landforms, e.g. hills, valleys, terraces, flood plains, sand dunes, etc. Before the start of soil survey, aerial photographs or satellite imagery are interpreted to delineate various land forms or physiographic or geomorphic units. The soils developed on various geomorphic units are studied and soil-physiographic relationship is developed.

Information about water resources (surface-water and groundwater) in an area is essential for their rational use and development. Under intensive cultivation, the limited surfacewater resources and precipitation cannot meet the irrigation water requirement, therefore, the groundwater is tapped to augment the surfacewater potential.

Relief, slope and aspect (topography) information is recorded during the course of soil survey as they influence soil information (*see* Chapter 2 for more details).

Development programmes concerning optimum utilization of natural resources are being implemented on watershed basis. Watershed is a natural hydrological entity that covers a specific area within which the entire rainfall runoff, ultimately passes through a specific channel at a particular point. The information about the natural resources of a watershed is very important for watershed management planning. The various parameters of the watershed, viz. stream network (drainage), physiography, land use, vegetation/forest cover and snow cover can be mapped and monitored using remote sensing data.

The SIS developed using soil survey and ancillary data has immense potential in planning judicious management, conservation and sustainable use of soil, land and crop resources. The manifold advantages of the SIS such as ease of handling of voluminous data, reproduction of maps derived suitability and other interpretative maps, easy linkage with other georeferenced coverage to generate new composite overlays, cost-effective and time-saving periodic updation of maps/information and capabilities of quick monitoring and impact assessment of development measures make it a useful tool for generating action plans and its implementation for land resource management of a region/watershed.

14. Agro-ecological Approach for Sustainable Agricultural Development

14.1. Concepts

A proper understanding of potential and limitations of natural resources is necessary for sustainable agricultural development at local, regional and country level. The farm output depends largely on the components like climate, soil and landforms. Therefore, for efficient crop planning in an area and transfer of technology, information concerning requirements of the crops and kinds of soils, their extent, geographic distribution and the local agro-climatic conditions is essential. The delineation of homogeneous agricultural environments, called agro-ecological regions (AER) on the basis of soil characteristics, landforms, climate and biodiversity helps in better crop selection, agro-technology transfer for sustainable land use and land resource planning (Sehgal *et al.*, 1992).

An agro-climatic region is a land unit in terms of major climate and growing period which is climatically suitable for certain range of crops and cultivars (FAO, 1983), whereas an agro-ecological region is characterized by distinct ecological responses to macroclimates as expressed in vegetation and reflected in soil, fauna and aquatic systems. Therefore, an agro-ecological region is the land unit on earth's surface carved out of agro-climatic region when superimposed on different landforms and soil conditions that act as modifiers to the length of growing periods (LGP) and crop environmental needs.

14.2. Concept of Bio-climate and Length of Growing Period (LGP)

14.2.1. Bio-climate

The climate of a particular region is determined by meteorological parameters like rainfall, temperature and potential evapotranspiration (PET). For all practical purposes, the quantification of climatic parameters is done within one metre below the soil surface and few metres above the soil surface where biological life sustains. The bio-climatic profile consists of elements which describe the temperature and moisture condition in the study region. Bio-climatic analysis is done using various climatic and moisture balance indices such as Moisture Index (Im), Aridity Index (Ia), Humidity Index (Ih), Summer Concentration (Sc), and Thermal efficiency type (Thornthwaite and Mather, 1955).

The Water Balance Technique (Thornthwaite and Mather, 1955; Carter and Mather, 1966) and PET values (Penman, 1948) are being used to prepare a bio-climate map. It accounts for the monthly as well as annual water surplus (WS) and water deficites (WD), determining the moisture index (Im) in a particular ecosystem receiving specific amount of rainfall and having specific potential evaportranspiration. The concept of water balance has been illustrated in a flow diagram (Figure 4).

Water deficit and water surplus are calculated from the potential and actual evaportranspiration values. Water deficit is the amount by which the available moisture fails to meet the demand for water and is computed by substracting the potential evapotranspiration from the actual evapotranspiration for the period in question. Water surplus is the excess remaining after the evaporation needs of the soil have been met (i.e. when actual evaportranspiration equals the potential evapotranspiration) and soil storage has been returned to the water-holding capacity level.

The annual water surplus and water deficit help to compute humidity index (Ih), aridity index (Ia) and moisture index (Im), which stand

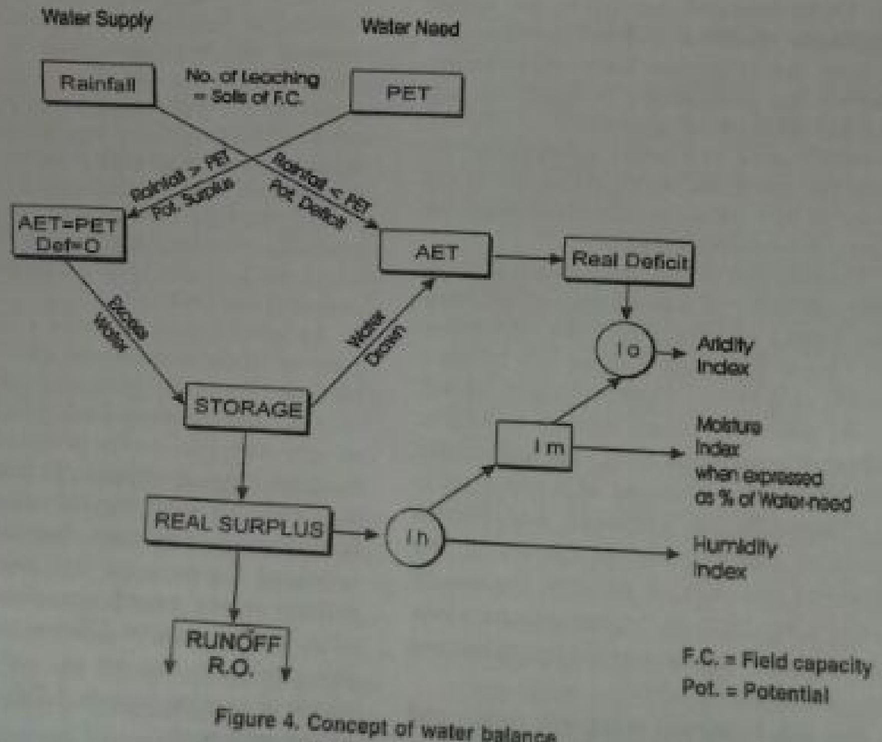


Figure 4. Concept of water balance
After Sehgal (1987)

for a balance between moisture availability and extent of dryness. The moisture index with positive and negative values would indicate moist or dry climate and seasonal variations in effective moisture. The assessment of Im, thermal efficiency (PET) and summer concentration leads to classify the bio-climatic types of an area. Thermal efficiency is indicative of radiation energy received and summer concentration represents the percentage of solar radiation in summer. The criteria evaluating Moisture Index (Im) are given in Table 5.

14.2.2. *Length of Growing Period (LGP)*

The growing period or the moisture availability period for crop growth is the period (in days) when the precipitation (P) exceeds 50% of the PET plus the time required to evapotranspire an assumed 100 mm of stored moisture from the deep soil profile. It may be defined as the duration (in days) when both water (moisture in soils) and temperature permit crop growth (Higgins and Kassam, 1981). During a normal growing season, crop experiences a humid period (when P > PET), a moist period (when P > 0.5 PET but < PET) and moderately dry to dry period (when P < 0.5 PET). The concept of growing period has been illustrated in Figure 5.

Table 5. Moisture Index limits used for bio-climatic classification

Symbol	Type description	Moisture index (Im)
E	Arid	<- 80
D_1	Semi-arid (dry)	- 60
D_2	Semi-arid (semi-dry)	- 40
D_3	Semi-arid (moist)	- 20
C_1	Sub-humid (dry)	0
C_2	Sub-humid (moist)	+ 20
B_1	Humid 1 (1st degree)	+ 40
B_2	Humid 2 (2nd degree)	+ 60
B_3	Humid 3 (3rd degree)	+ 80
B_4	Humid 4 (4th degree)	+ 100
A(1)	Perhumid 1 (1st degree)	> + 100

Source: Sehgal *et al.* (1992)

Based on the values of P and PET of humid, moist and moderately dry periods, the LGP in a year is determined. Moist period is the duration at the beginning of rainy season when most of the crop establishment operations are usually done. The humid period not only meets

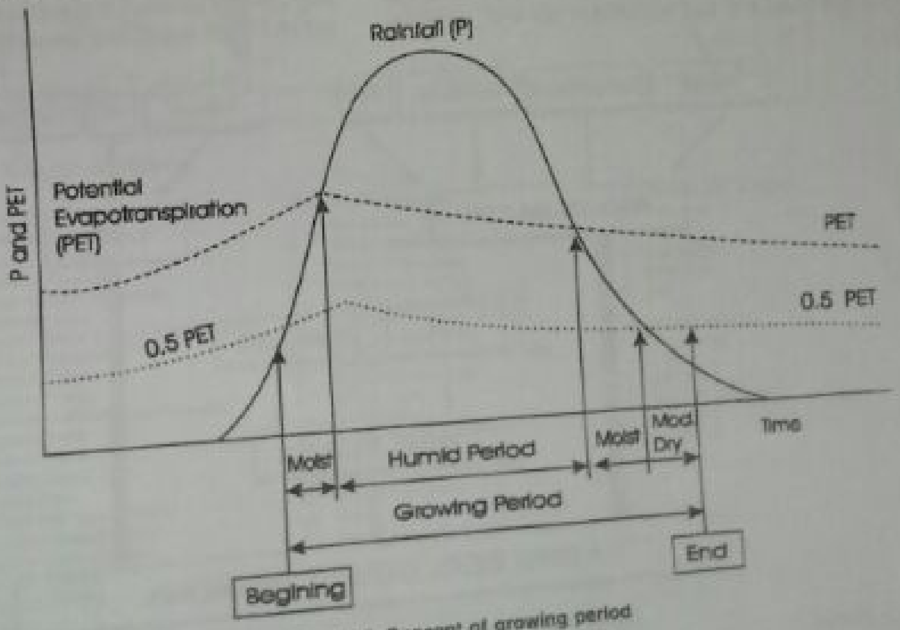


Figure 5. Concept of growing period
Source: Higgins and Kassam (1981)

the full evapotranspiration demand of crop, but also replenishes moisture deficit in the storage (soil profile). During the moderately dry to dry periods, the crop progressively starts drawing moisture from the storage in soil profile.

National Bureau of Soil Survey and Land Use Planning (NBSS&LUP) has delineated 20 agro-ecological regions (AER) in the country using the FAO concept (FAO, 1978) by integrating maps depicting length of growing period, bioclimate and soil scape (soil-physiography) as per the scheme outlined in Figure 6.

The NBSS&LUP has adopted LGP criteria for the delineation of AERs in the country as LGP is the direct indicative of moisture availability of a given landform rather than the total rainfall (Figure 7). For example, both Ratnagiri in western Maharashtra and Nagpur in eastern Maharashtra have LGP of 180-210+ days in a year, but the total annual rainfall of Ratnagiri is more than 2000 mm, whereas that of Nagpur is only 1100 mm. Therefore, the crop planning has to be based on LGP rather than on the total rainfall. The extent, distribution, soils and major constraints of the identified AERs (Sehgal *et al.*, 1992) have been furnished in Table 6.

For sub-regionalization of each AER, the NBSS&LUP has broadened the soil, physiography and LGP concept with the inclusion of soil quality attributes such as soil depth, texture, mineralogy in combination with available water capacity (AWC), cropping pattern and vegetation type (Velayutham *et al.*, 1999). The bioclimatic concept has been extended from Thornthwaite and Mather moisture index to synchronized bioclimate-LGP concept. Arid (E) has been bifurcated into Hyper arid (Eh) with LGP less than 60 days and Typic arid (Et) with LGP 60-90, and Semi-arid (D) into Dry semi-arid (Dd) with LGP 90-120 days and Moist semi-arid (Dm) with LGP 120-150 days, whereas Sub-humid (C) has been bifurcated into Dry sub-humid (Cd) having LGP 120-150 days and Moist sub-humid (Cm) with LGP of 210-240 days. In this way, 20 agro-ecological regions (AER) map prepared in 1992 has been refined by using the length of growing period (LGP) of 30-day interval, starting from LGP of 60 days and, then 90, 120, 150, 180, 210, 240, 270, 300 and > 300 days to generate 60 agro-ecological sub-regions (AESR) map. In the map, the AESRs have been shown with numbers (1.1, 5.2, 8.3, etc.) and mapping unit within each of delineation by alpha numeric codes e.g. in K6Dm4, K stands for physiography (Deccan plateau), 6 stands for soils (mixed red and black soils), Dm stands for bio-climate

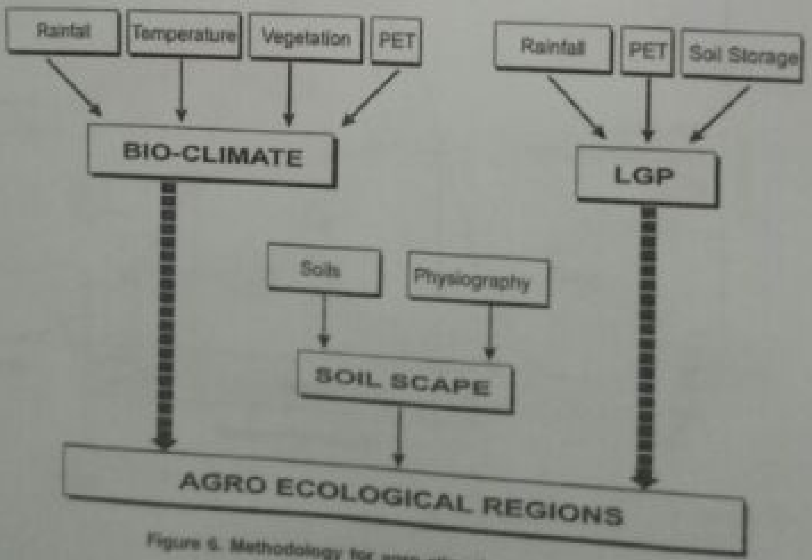
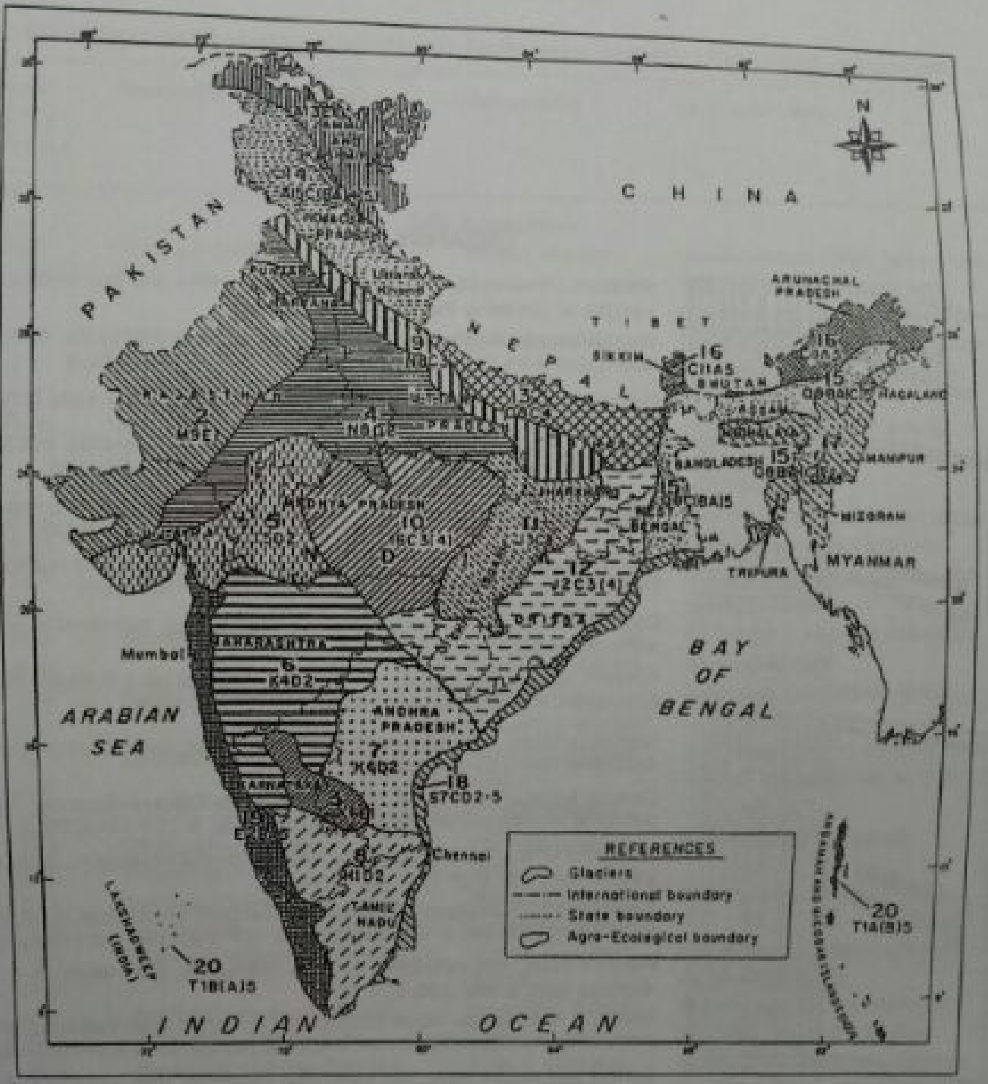


Figure 6. Methodology for agro-climatic regionalization
After Sehgal (1992)



Key for decoding

PHYSIOGRAPHY (1st Letter)

A	Western Himalayas
B	Central Himalayas
C	Eastern Himalayas
D	North-Eastern Himalayas
E	Western Ghats
F	Eastern Ghats & Tamil Nadu Uplands
G	Central Highlands
H	Eastern Plateau
I	Deccan Plateau
J	Kachchh and Kathiawar Peninsula
K	Western Plain
L	Northern Plain
M	Eastern Plain
N	Gujarat Plain
O	Bengal and Assam Plains
P	Western Coastal Plains
Q	Eastern Coastal Plains
R	Eastern Islands
S	Western Islands

SOILS (1st Numeral)

1.	Red Loamy Soils
2.	Red & Lateritic Soils
3.	Red & Yellow Soils
4.	Shallow and Medium (inclusion Deep) Black Soils
5.	Medium & Deep Black Soils
6.	Red and Black Soils
7.	Coastal Alluvium-derived Soils
8.	Aluvial-derived Soils
9.	Desert Soils
10.	Tarai Soils
11.	Brown and Red Hill Soils
12.	Saline and Alkali Soils
13.	Shallow and Skeletal Soils
14.	Grey Brown Soils
15.	Brown Forest and Podzolic Soils
16.	Sandy and Littoral Soils

CLIMATE (2nd Letter)

A	Perhumid
B	Humid
C	Sub-humid
D	Semi-Arid
E	Arid

GROWING PERIOD (2nd Numeral)

1	< 90	(Days)
2	90-150	
3	150-180	
4	180-210	
5	> 210	

Example:	K	6	E	1
	Physiography	Soils	Climate	LGP

Contd...

LEGANDS

Arid Ecosystem	1	Western Himalayas, cold arid ecoregion , with shallow skeletal soils and Length of Growing Period (LGP) <90 days
	2	Western Plain, Kachchh and part of Kathiawar Peninsula, hot arid ecoregion with desert and saline soils and LGP<90 days
	3	Deccan Plateau, hot arid ecoregion, with red and black soils & LGP<90 days
Semiarid Ecosystem	4	Northern Plain and Central Highlands including Aravallis, hot semi-arid ecoregion , with alluvium-derived soils & LGP 90-150 days.
	5	Central (Malwa) Highlands, Gujarat Plains & Kathiawar Peninsula, hot semi-arid ecoregion, with medium & deep black soils & LGP 90-150 days
	6	Deccan Plateau, hot semi-arid ecoregion with shallow and medium (with inclusion of deep) black soils, & LGP 90-150 days
	7	Deccan (Telangana) Plateau and Eastern Ghats, hot semi-arid ecoregion, with red & black soils & LGP 90-150 days.
	8	Eastern Ghats,TN uplands and Deccan (Karnataka) Plateau, hot semi-arid ecoregion with red loamy soils & LGP 90-150 days
Subhumid Ecosystem	9	Northern Plain, hot subhumid (dry) ecoregion, with alluvium-derived soils and LGP 150-180 days
	10	Central Highlands (Malwa, Bundelkhand, & Eastern Satpura), hot subhmid ecoregion, with black and red soils, & LGP 150-180(to 210) days
	11	Eastern Plateau (Chhattisgarh), hot subhumid ecoregion, with red & yellow soils and LGP 150-180 days
	12	Eastern (Chhotanagpur) Plateau and Eastern Ghats, hot subhumid ecoregion, with red & lateritic soils, & LGP 150-180 (to 210) days
	13	Eastern Plain, hot subhumid (moist) ecoregion, with alluvium-derived soils & LGP 180-210 days
	14	Western Himalayas, warm subhumid (to humid with inclusion of perhumid), ecoregion with brown forest and podzolic soils, & LGP 180-210+days
Humid-Perhumid Ecosystem	15	Bengal and Assam Plain, hot subhumid ecoregion (moist) to humid (inclusion of perhumid) ecoregion, with alluvium – derived soils & LGP 210 + days
	16	Eastern Himalayas, warm perhumid ecoregion, with brown and red hills soils & LGP 210 + days
	17	North-eastern Hills (Purvachal), warm perhumid ecoregion, with red and lateritic soils & LGP 210 + days
Coastal Ecosystem	18	Eastern Coastal Plain, hot subhumid to semi-arid ecoregion, with coastal alluvium - derived soils & LGP 90-210 + days
	19	Western Ghats & Coastal Plain, hot humid- perhumid ecoregion, with red, lateritic and alluvium-derived soils, & LGP 210 + days
Island Ecosystem	20	Islands of Andaman-Nicobar and Lakshadweep hot humid to perhumid Island ecoregion, with red loamy and sandy soils, & LGP 210 + days

Figure 7. Agro-climatic regions of India

Table 6. Agro-ecological regions: Extent, distribution, soil orders and major constraints

AER No.	Area (M ha) (% of TGA)	Distribution (Major states)	Soil order(s)	Major constraints
Arid Ecosystem				
1.	15.2 (4.7)	Jammu & Kashmir	Entisol, Aridisol	Severe climate (Cryic Temperature regime). Shallow, sandy and gravelly soils. Nutrient imbalance
2.	31.9 (9.7)	Rajasthan, Gujarat, etc.	Entisol, Aridisol	Deforestation, Soil salinity. Nutrient deficiency viz. N,P, Zn and Fe
3.	4.9 (1.5)	Andhra Pradesh, Karnataka, etc.	Inceptisol, Vertisol, Alfisol	Soil erosion. Subsoil sodicity
Semi-arid Ecosystem				
4.	32.2 (9.8)	Haryana, Uttar Pradesh, Rajasthan, Part of Punjab, etc.	Inceptisol, Alfisol, Entisol	Low plant available water capacity (PAWC). Imperfect drainage.
5	17.6 (5.4)	Gujarat, Madhya Pradesh, etc.	Inceptisol, Vertisol, Aridisol	Imperfect drainage. Salinity and alkalinity hazards
6.	31.0 (9.5)	Karnataka,Maharashtra, etc.	Entisol, Inceptisol, Vertisol	Soil erosion. Nutrient deficiency viz. N, P and Zn
7.	16.5 (5.2)	Andhra Pradesh	Entisol, Alfisol, Inceptisol, Vertisol	Soil erosion, Nutrients deficiency viz. N, P and Zn
8.	19.1 (5.8)	Karnataka, Tamil Nadu, etc.	Entisol, Inceptisol, Alfisol	Soil erosion, low to medium PAWC, Nutrient deficiency viz. N, P and Zn
Sub-humid Ecosystem				
9.	12.1 (3.7)	Bihar, Uttar Pradesh, Punjab, etc.	Inceptisol, Alfisol, Entisol	Waterlogging, soil salinity
10.	22.3 (6.8)	Madhya Pradesh and Maharashtra	Vertisol, Entisol, Inceptisol, Alfisol	Soil erosion
11.	14.1 (4.3)	Madhya Pradesh, Jharkhand, etc.	Entisol, Inceptisol, Alfisol	Soil erosion, nutrient deficiency viz. N, P, Zn and B
12.	26.8 (8.2)	Andhra Pradesh, Jharkhand, Orissa, West Bengal, etc.	Inceptisol, Alfisol, Ultisol	Soil erosion, nutrient deficiency (viz. N, P, Zn) and P fixation
13.	11.1 (3.4)	Jharkhand and Uttar Pradesh	Entisol, Inceptisol, Alfisol	Flooding and imperfect drainage. Salinity and/or sodicity. Deficiency of N, P and Zn
14.	18.2 (5.6)	Jammu & Kashmir, Uttarakhand, Himachal Pradesh, etc.	Inceptisol,Alfisol, Mollisol, Ultisol	Severe climate (Cryic temperature regime). Soil erosion. Soil acidity
Humid-Perhumid Ecosystem				
15.	12.1 (3.7)	Assam, Tripura and West Bengal	Inceptisol, Alfisol	Flooding, low base status, soil acidity
16.	9.6 (2.9)	Arunachal Pradesh, Assam, Sikkim and West Bengal, etc.	Inceptisol, Alfisol, Ultisol, Mollisol	Soil erosion, deforestation
17.	10.6 (3.3)	Assam, Manipur, Mizoram, etc.	Inceptisol, Ultisol, Alfisol	Soil erosion, depletion of soil nutrients
Coastal Ecosystem				
18	8.5 (2.6)	Andhra Pradesh, WestBengal, Tamil Nadu, etc.	Entisol, Inceptisol, Vertisol	Poor drainage, soil salinity and/or sodicity
19.	11.1 (3.6)	Karnataka, Kerala, Gujarat, etc.	Inceptisol, Ultisol,	Depletion of plant nutrients, waterlogging
Island Ecosystem				
20.	0.8 (0.3)	Union Territory (Andaman & Nocobar Islands and Lakshadweep groups of Islands)	Alfisol, Inceptisol, Entisol	Soil erosion

*TGA= Total Geographical Area.

(Semi-arid, moist) and 4 stands for length of growing period (120-150 days), respectively. This map was published in colour on a 1:4.4 million scale, showing distinct boundaries of the country.

The AESR map prepared in 1999 by NBSS & LUP, Nagpur, can only cater to the needs of regional planning. However, for the implementation of development projects at the district level, delineation of agro-ecological units by incorporating details of soils and soil degradation status, crop zones, LGP and biodiversity is necessary. The degree of success in the growth of crop is largely dependent on how well its optimum length of growth cycle fits with the LGP. Curtailment of the growth cycle is naturally reflected by decreased yields and the same is true for enforced extended growth cycle.

15. References

AISLUS (1971) *Soil Survey Manual*. All India Soil and Land Use Survey, Department of Agriculture and Cooperation, Ministry of Agriculture, Government of India. IARI Campus, New Delhi.

Carter, D.B. and Mather, R.J. (1966) Climatic classification for environmental biology. In: *Climatology* **19**, (4).

FAO (Food and Agriculture Organization) (1978) *Report on the Agro-Ecological Zones Project*, Vol. 1, Methodology and Results for Africa, FAO, Rome, 15 p.

FAO (Food and Agriculture Organization) (1978) (1983) *Guidelines: Land Evaluation for Rainfed Agriculture*, Soil Bulletin 52, FAO, Rome, 237 p.

Higgins, G.M. and Kassam, A.H. (1981) The FAO-Agro-ecological zone approach to determination of land potential, *Pedologie* **31**(2), 147-168.

ISRO (2000) *National (Natural) Resources Information System (NRIS): Node Design and Standards*. Indian Space Research Organization Publication : ISRO – NNRMS-SP:72.

Penman, H.L. (1948) Natural evaporation from open water, bare soil and grass. *Proceedings of Royal Society, Series A* **193**, 120-145.

Sehgal, J.L. (1987) *Agroclimatic Environments and Moisture Regimes in North-West India – Their Application in Soils and Crop Growth*. Bulletin No. 17, NBSS&LUP (ICAR), Nagpur.

Sehgal, J.L. (1992) *Soil Series Criteria and Norms*. Technical Bulletin. NBSSLUP Publication No. 36, NBSS&LUP (ICAR), Nagpur.

Sehgal, J.L., Saxena, R.K. and Vadivelu, S. (1987) *Field Manual - Soil Resource Mapping of Different States in India*. Technical Bulletin No. 13, NBSSLUP Publication No. 13.

Sehgal, J., Mandal, D.K., Mandal, C. and Vadivelu, S. (1992) *Agro-Ecological Regions of India*, Second Edition, Technical Bulletin, NBSS&LUP Publication No. 24, 130 p.

Soil Survey Staff (1951) *Soil Survey Manual*. US Department of Agriculture. Handbook No. 18. Oxford & IBH Publishing Co., New Delhi.

Soil Survey Staff (1975) *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. United States Department of Agriculture – Soil Conservation Service, Agriculture Handbook, 436, US Government Printing Office, Washington DC, USA.

Soil Survey Staff (1990) *Keys to Soil Taxonomy*. Agency for International Development, US Department of Agriculture, Soil Management Support Services (SMSS). Technical Monograph No. 18. Virginia Polytechnic Institute and State University, USA.

Soil Survey Staff (1993) *Soil Survey Manual*. US Department of Agriculture. Handbook No. 18. US Government Printing Office, Washington DC, 20402, USA.

Soil Survey Staff (1999) *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. Second Edition. United States Department of Agriculture – National Resources Conservation Service, Agriculture Handbook, 436, US Government Printing Office, Washington DC, USA.

Soil Survey Staff (2003). *Keys to Soil Taxonomy*. United States Department Agriculture and Natural Resources Conservation Services, Washington DC, USA.

Chapter 5

Physical Properties of Soils

D.K. DAS and R.P. AGRAWAL

1. Introduction

The physical properties of a soil play an important role in determining its suitability for crop production. The characteristics like supporting power and bearing capacity, tillage practices, moisture storage capacity and its availability to plants, drainage, ease of penetration by roots, aeration, retention of plant nutrients and its availability to plants are all intimately connected with the physical properties of a soil. Soil as a medium of plant growth should also be physically fertile. The soil which supports plants is a variable mixture of solids (mineral and organic matter), liquid (water) and gases (air) and is called a three phase system (Figure 1). In a representative silt loam soil, about 50% of the total volume is occupied by solids. At optimum moisture for plant growth, 25% of the volume is occupied by water and 25% by air. The inorganic solid phase is composed of discrete mineral particles of various shapes and sizes as well as of amorphous compounds such as hydrated iron and aluminium oxides. The proportion of amorphous material is generally small. The large soil particles are generally visible to the naked eye, whereas the smaller ones are colloidal and can be seen only with the aid of an electron microscope. The liquid phase, consisting of soil water, also contains dissolved salts and thus it is called 'soil solution'. The gaseous phase consists of soil air of varying composition of oxygen and carbondioxide, different from that of atmospheric air.

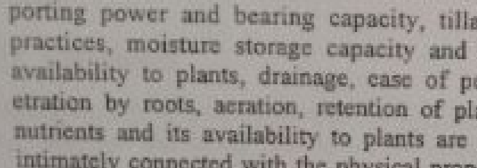


Figure 1. Different components of soil — A schematic diagram

2. Relationships between Different Soil Constituents

The quantitative interrelationships of the solid, liquid and air components of soil are defined in terms of many useful physical parameters of soil (Hillel 1998). These are briefly stated below:

2.1. Particle Density

The particle density (ρ_s), also called density of soil solids, is the ratio of total mass of soil solids (M_s) to total volume of soil solids (V_s) and is expressed in g/cm^3 or Mg/m^3 (Mg = mega gram i.e. 10^6 g)[Equation (1)] :

$$\rho_s = M_s / V_s \qquad ...(1)$$

The particle density depends on the chemical and mineralogical compositions of soil.

In most mineral soils, the particle density varies between 2.60 and 2.70 Mg/m³. The mean value of 2.65 Mg/m³ is usually used for all practical purposes. The presence of organic matter lowers the value of ρ_s.

2.2. *Dry Bulk Density*

The dry bulk density (ρ_b) or simply the bulk density is the ratio of the mass of oven dried soil solid particles (M_s) to the total volume of the soil (V_t). This volume includes the volume of soil solids (V_s), soil water (V_w) and soil air (V_a). The latter two, i.e. volume of water and volume of air constitute the total volume of pores (V_f). So bulk density is expressed by Equation (2):

$$\rho_b = \frac{M_s}{V_t} = \frac{M_s}{V_s + V_w + V_a} \qquad ...(2)$$

The bulk density is expressed as g/cm³ or Mg/m³. Bulk density of soil is influenced by soil texture, structure, organic matter content and land management practices. The bulk density of the coarse-textured soils varies from 1.40 to 1.75 Mg/m³ and of fine-textured soils normally ranges from 1.10 to 1.40 Mg/m³. Increase in organic matter content lowers the bulk density of soil. High bulk density indicates compactness of the soil. The bulk density is generally higher in lower profile layers. In swelling soils, bulk density decreases with increase in moisture content and vice versa. In soils, the value of ρ_b is obviously lower than the value of ρ_s.

2.3. *(Total) Wet Bulk Density* (ρ_t)

It is the ratio of the total mass of a moist soil (M_t) to the total volume of the soil (V_t) [Equation (3)]:

$$\rho_t = \frac{M_t}{V_t} = \frac{M_s + M_w}{V_s + V_w + V_a} \qquad ...(3)$$

where, M_w and V_w are the mass and volume of soil water, respectively and V_a is the volume of soil air. The wet bulk density depends more on the wetness or moisture content of a soil.

2.4. *Specific Volume* (v_b)

It is the ratio of total volume of soil (V_t) to the total mass of dry soil (M_s), i.e. it is the volume of unit mass of dry soil and is expressed as cm³/g or m³/Mg [Equation (4)].

$$v_b = V_t / M_s = 1/\rho_b \qquad ...(4)$$

The v_b is an index of the degree of compaction or looseness of a soil. A higher value of v_b indicates lower levels of compaction. The values of v_b for agricultural soils may vary from 0.55 to 0.70 m³/Mg for coarse-textured soils and from 0.70 to 0.90 m³/Mg for fine-textured soils.

2.5. *Porosity* (f)

It is the ratio of total volume of pore spaces (V_f) to the total volume of soil (V_t) and is expressed as a fraction or as percentage [Equation (5)]:

$$f = \frac{V_f}{V_t} = \frac{V_w + V_a}{V_s + V_w + V_a} \qquad ...(5)$$

Thus, porosity refers to the volume fraction of pores and is an index of the relative pore volume in a soil. Its value usually varies from 0.30 to 0.60, i.e. 30% to 60%. Although the size of individual pores may be bigger in coarse-textured soil than in fine-textured soils, the former is less porous than the latter .

Porosity is related to bulk density of the soil and is given by Equation (6):

$$f = \frac{\rho_s - \rho_b}{\rho_s} = \left(1 - \frac{\rho_b}{\rho_s}\right) \qquad ...(6)$$

Two types of pore spaces — macro and micro — occur in soils without any clear-cut demarcation. Usually, pores larger than about 0.06 mm in diameter are considered as macropores and those smaller than this, as micropores.

2.6. *Capillary and Non-capillary Porosity*

The pore spaces in a soil contain both capillary and non-capillary pores which remain filled with water when the soil is saturated. As the soil dries, the non-capillary pores are emptied first. Thus, capillary porosity is determined by finding the volume of water retained in soil at 50-cm tension. The non-capillary porosity is determined by subtracting the value of capillary porosity from the total porosity value. Capillary pore spaces are important since water retained in these pore spaces are used by the plants for their growth.

2.7. *Void Ratio* (e)

As a result of swelling, shrinkage or compaction, a soil undergoes a change in volume. The V_t in Equation (2) is not constant. In such case, it is advantageous to use the void ratio, e, instead of porosity.

The void ratio is the ratio of total volume of pores (V_f) to the total volume of soil solids (V_s) and is expressed as a fraction, as shown in Equation (7):

$$e = \frac{V_f}{V_s} = \frac{V_w + V_a}{V_s} = \frac{V_f}{V_t - V_f} \qquad ...(7)$$

Void ratio is also an index of the relative volume of soil pores. While porosity is largely used in agriculture, void ratio is preferred in engineering. The value of 'e' usually varies between 0.3 and 2.0.

2.8. *Gravimetric Water Content or Mass Wetness* (w)

It is the mass (or weight) of water (M_w) relative to the mass of oven dried soil particles (M_s) and is expressed as a fraction or as percentage, as per Equation (8):

$$w = \frac{M_w}{M_s} \qquad ...(8)$$

2.9. *Volumetric Water Content or Volume Wetness* (θ)

It is the ratio of total volume of water occupied in the pore spaces at a specific time to the total volume of soil and is expressed as a fraction or percentage (Equation 9):

$$\theta = \frac{V_w}{V_t} = \frac{V_w}{V_s + V_f} \qquad ...(9)$$

Volumetric water content (also called volume wetness) can be computed from gravimetric water content by multiplying the value with bulk density, as shown in Equation (10):

$$\theta = w\rho_b \qquad ...(10)$$

The volumetric water content of soil has wider applications in computation of quantity of water added to soil through irrigation or rain or the amount of water used by crop or lost through drainage, etc. The volumetric water content, expressed as a fraction, when multiplied by the depth of soil gives the amount of water in terms of depth.

In swelling soil where volume is not constant, water content is also expressed as the ratio of the volume of water (V_w) to the volume of soil solids (V_s). This is called water volume ratio or liquid ratio (v_w).
Thus,

$$v_w = \frac{V_w}{V_s} \text{ or } v_w = w.\rho_s \qquad ...(11)$$

2.10. *Degree of Saturation* (θ_s)

It is the ratio of the volume of water present in the soil at a particular time to the volume of pores [Equation (12)]:

$$\theta_s = \frac{V_w}{V_f} \qquad ...(12)$$

2.11. *Air-filled Porosity* (f_a)

It is the ratio of volume of air space to the total volume of soil, as given in Equation (13):

$$f_a = \frac{V_a}{V_t} = \frac{V_a}{V_s + V_w + V_a} = (f - \theta) \qquad ...(13)$$

This index is a measure of the relative air content of the soil. It is a useful criterion of soil aeration.

3. Soil Texture

3.1. *Definition*

The soil solid phase as a whole can be characterized in terms of the relative proportions of its particle size groups (soil separates).

Table 1. Classification of soil particles according to size

System	Soil particles and sizes (diameter, upper limit, in mm)							
International Union of Soil Sciences (IUSS)	Clay	Silt	Sand					Gravel
			Fine sand		Coarse sand			
	0.002	0.02	0.2		2.0			> 2.0
United States Department of Agriculture (USDA)	Clay	Silt	Sand					Gravel
			Very fine	Fine	Medium	Coarse	Very coarse	
	0.002	0.05	0.10	0.25	0.50	1.0	2.0	> 2.0

The relative size-range of soil particles is expressed by the term 'texture' which refers to the fineness or coarseness of the soil. Soil texture is defined as the relative proportions of the different particle size fractions, specifically referred to as sand, silt and clay. This has both qualitative and quantitative connotations. Qualitatively, it refers to the "feel" of the soil material, whether coarse and gritty, or fine and smooth. An experienced soil classifier can tell the soil texture by feel, i.e. by kneading or rubbing soil with his fingers. Quantitatively, soil texture refers to the relative proportions of various sizes of particles in a given soil. The soil texture, typically permanent, is an intrinsic attribute of the soil and the one most often used to characterize its physical make-up, having a bearing on soil behaviour.

3.2. Textural Fractions

The traditional method of characterizing sizes of particles in soils is to divide the array of possible particle diameters into three conveniently separable size ranges known as textural fractions or separates, namely sand, silt and clay. The actual procedure of separating out these fractions and of measuring their proportions is called particle size analysis or mechanical analysis. The results of this analysis yield the mechanical composition of a soil, a term often used interchangeably with soil texture.

3.3. Particle Size Classification

The different particle size fractions are named and classified arbitrarily. Many particle size classification schemes exist, each of which having different class limits for each size fraction. The classifications of the International Society of Soil Science (ISSS), renamed as the International Union of Soil Sciences (IUSS) and the United States Department of Agriculture (USDA) are widely followed (Table 1). The IUSS system has a geometric proportion with a common ratio of 10, starting from the clay fraction 0.002 mm, while the USDA system has a geometric progression with a common ratio of 2 in the sand fraction, except an irregularity in the fine sand fraction (0.10-0.25 mm). For particles of larger than 2.0 mm diameter, US Soil Conservation Service (1981) provides further subdivisions into gravel, cobble, stone, etc. based on their shape and size (Table 2). The 75 mm upper limit is commonly used for engineering purposes. Various rock fragments up to 250 mm can be obtained by the appropriate sieves with the required apertures.

3.4. Characteristics of Different Size Fractions / Soil Separates

The largest group of particles generally recognized as soil material is sand which is defined as particles ranging in diameter from 2 mm to 0.02 mm (IUSS classification) or to 0.05 mm (USDA classification). The sand fraction in IUSS classification is further subdivided into coarse (2-0.2 mm) and fine sand (0.2-0.02 mm) or as coarse, medium and fine sand. Sand particles are primary minerals generally rounded or angular. Sand feels gritty when pressed between the fingers and the particles are generally visible to the naked eye. The particles of sand have relatively lower surface area per unit mass or volume. Sand particles can hold little water. Soils dominated by sand are prone to drought.

Table 2. Size and shape classes for rock fragments

Shape and size*	Name	Quality
Rounded, subrounded, angular or regular or irregular		
0.2-0.5 cm diameter	Fine gravel	Fine gravelly
0.5-2.0 cm diameter	Medium gravel	Medium gravelly
2.0-7.6 cm diameter	Coarse gravel	Coarse gravelly
7.6-25 cm diameter	Cobble	Cobbly
25-60 cm diameter	Stone	Stony
> 60 cm diameter	Boulder	Bouldery
Flat		
0.2-15 cm long	Channer	Channery
15-38 cm long	Flagstone	Flaggy
38-60 cm long	Stone	Stony
> 60 cm long	Boulder	Bouldery

*If significant for classification or interpretation, the shape of fragments is indicated such as "angular gravel" or "irregular boulders"

Source: USDA Soil Conservation Service (1981)

The next fraction is silt (0.02-0.002 mm as per IUSS classification) which is intermediate in size between sand and clay. Mineralogically and physically, silt particles generally resemble sand particles but have a greater surface area per unit mass.

The clay fraction, with particles ranging from 0.002 mm or 2 microns downward, is the colloidal fraction. Clay particles are plate-like or needle-like and are secondary minerals generally belonging to a group of minerals known as the aluminosilicates. Clay is the fraction that has negative charge and greater surface area per unit mass and influences the soil behaviour most. Clay particles absorb water and hydrate, thereby causing the soil to swell upon wetting and shrink upon drying.

The relatively inert sand and silt fractions can be called the soil "skeleton", while the clay by analogy, can be regarded as the "flesh" of the soil. Together, all the three fractions of the solid phase constitute the *matrix* of the soil.

3.5. Influence of Soil Separates on Properties and Behaviour of Soils

The soil separates have profound influence on different properties and behaviour of soils. Different soils contain different proportions of sand, silt and clay and exhibit properties as determined by the dominant fraction.

Coarse-textured or sandy soils are loose, low water retentive, well-drained, well-aerated, easy to cultivate and are called *light* soils. On the other hand, fine-textured or clayey soils tend to absorb and retain much more water, become plastic and sticky when wet, tight and cohesive when dry, difficult to cultivate and are called *heavy soils*. These expressions can be misleading as coarse-textured soils are generally more dense, that is, have a lower porosity than the fine-textured soils, and thus are heavier in weight per unit volume (at least in the dry state). Energy required for tillage is much less in light soils as compared to that in heavy soils.

In general, sandy soils have low water and nutrient holding capacity, low organic matter content, little or no swelling and shrinkage, poor sealing properties for ponds and dams, high leaching of nutrients and pollutants. The fine sands are easily blown by wind, while coarse sands resist erosion by water.

The silty soils have medium to high water and nutrient holding capacity, moderate aeration, slow to medium drainage, medium to high organic matter content, usually good supply of plant nutrients, moderate leaching of pollutants and nutrients. These soils are easily blown by wind and are susceptible to water erosion, easily compacted, have little swelling and shrinkage and are moderately difficult to till after rain.

A loam soil contains a balanced mix of coarse and fine particles with properties intermediate among those of a sand, a silt, and a clay. A loam soil is often considered to be the optimal soil for plant growth and agriculture. Its capacity to retain water and nutrients is superior to that of sand, while its drainage, aeration and tillage properties are often favourable than those of clay.

The clayey soils have high water and nutrient holding capacity, are poorly aerated, have very slow drainage unless cracked, high to medium organic matter content, medium to excellent supply of plant nutrients and high swelling and shrinkage. These soils resist wind erosion; aggregated clays also resist water erosion. They have excellent sealing properties. They are easily compacted and retard leaching of nutrients and pollutants.

3.6. *Mechanical Analysis*

Mechanical analysis is the procedure for determining the particle size distribution of a soil sample. Particles can be separated into different size groups by sieving through graded sieves, up to a particle of approximately 0.05 mm diameter. For separating the finer particles, the method of sedimentation is generally used. The first step is to disperse the soil sample in an aqueous suspension. After complete dispersion, the settling velocity of the particles or the density of the suspension from which the particles are settling is measured based on the principle of Stokes' law.

3.6.1. Stokes' Law and its Derivation

According to Stokes' law, the terminal velocity of a spherical particle settling under the influence of gravity in a fluid of a given density and viscosity is proportional to the square of its radius, and is represented by Equation (14):

$$v = \frac{2}{9}\frac{r^2 g}{\eta}(\rho_s - \rho_f) \qquad \text{...(14)}$$

where, v is the settling velocity (cm/sec or $m \times 10^{-2}$/sec); r is the equivalent spherical radius of the particles (cm or m); η is the viscosity of the suspending fluid (g/cm sec or kg/m sec); g is acceleration due to gravity (cm/sec^2 or m/sec^2); ρ_s and ρ_f are the densities of the solid particles and of the fluid (g/cm^3 or Mg/m^3), respectively.

Derivation — A particle falling in the fluid encounters a frictional resistance (F_r) proportional to the product of its radius (r), velocity (v) and viscosity of the fluid (η). The resisting force due to friction F_r was shown by Stokes (1851) to be

$$F_r = 6\pi\eta r v \qquad \text{...(14a)}$$

Initially, as the particle begins to fall, its velocity increases. Eventually, a point is reached at which the increasing resistance force equals the constant downward force and the particle then continues to fall without acceleration at a constant velocity, known as terminal velocity (V_t).

The downward force due to gravity (F_g) is given by Equation (14b):

$$F_g = 4/3\pi r^3(\rho_s - \rho_f)g \qquad \text{...(14b)}$$

where, $4/3\pi r^3$ is the volume of the spherical particle.

Setting the two forces (i.e. F_r and F_g) equal, we can obtain Stokes' law as,

$$6\pi\eta r v = 4/3\pi r^3(\rho_s - \rho_f)g \qquad \text{...(14c)}$$

$$\text{or} \quad v = \frac{2}{9}\frac{r^2 g}{\eta}(\rho_s - \rho_f) \qquad \text{...(14)}$$

In the case of systems where ρ_s, ρ_f, η and g are constant at specified temperature, the Equation (14) simplifies to Equation (15):

$$v = K.r^2 \qquad \text{...(15)}$$

where, K is a constant. Thus, the sedimentation velocity being directly proportional to the square of the radius of the particles is a sensitive function of r. In aqueous medium at 20 °C, K assumes a value of 34,700.

If d is the diameter of the particle and assuming that the terminal velocity is attained almost instantly, we can obtain the time t needed for the particle to fall through a height h using Equation (16a) or (16b):

$$v = \frac{h}{t} \text{ or } t = \frac{h}{v} = \frac{9h\eta}{2r^2 g(\rho_s - \rho_f)} \qquad \text{(16a)}$$

$$t = \frac{18h\eta}{d^2 g(\rho_s - \rho_f)} \qquad \text{(16b)}$$

3.6.2. Assumptions and Limitations of Stokes' Law

The use of Stokes' law for measurement of the size of soil particle depends on certain simplifying assumptions. These are:

(i) Particles must be spherical, smooth and rigid. But, soil colloidal particles are plate-shaped and fall slower than spherical particles of the same mass.

(ii) Size of particles must be large compared to the size of molecules of the liquid, so that the medium can be considered homogeneous, i.e. no Brownian movement occurs. The larger limit of particles exhibiting Brownian movement is approximately 0.0002 mm.

(iii) Fall must be unhindered. Particles falling very near the wall of the vessel (0.1 mm distance) are slowed down in their descent. Many fast falling particles may drag finer particles down along with them (concentration up to 3% solids).

(iv) Particles must be of uniform density.

(v) The suspension must be still without any turbulence. Particles of > 0.08 mm diameter settle quickly and cause turbulence and therefore, are analysed by other procedure.

3.6.3. Methods of Mechanical Analysis

The international pipette method and the hydrometer method are usually used for mechanical analysis of soils.

The international pipette method is based on the principle of sedimentation, i.e. different sized particles having different weights fall at different velocities. If a sample of soil suspension is taken at a given depth at a particular time, this will contain all the particles which are still in suspension at that depth.

The pipette method is regarded as a standard method for particle size analysis because of its accuracy. It is, however, very time consuming and difficult to be employed in the mechanical analysis of a large number of samples.

The hydrometer method is based on the principle that there is a continuous decrease in the density of soil suspension over time at the rate the particles fall below the level of the hydrometer. So by knowing the density of the soil suspension at required times with a calibrated hydrometer, the proportion of different sized particles can be determined.

The hydrometer method is rapid but less accurate than the pipette method. This method gives erroneous results in soils having high $CaCO_3$ and organic matter contents and high salinity.

3.7. Textural Classes

Within each of the three broad groups of soil textural classes, i.e. *Sandy soils*, *Loamy soils* and *Clayey soils*, specific textural classes have been devised based on the relative proportion of sand, silt and clay particles.

In sandy soils, the sand separate comprises more than 70% of the material by weight (less than 15% of the material is clay). Two specific textural classes are recognised— *sand* and *loamy* sand.

The loamy soil group contains many subdivisions. An ideal loam is defined as a mixture of sand, silt and clay particles that exhibit the properties of these separates in about equal proportions. The various subdivisions are: *sandy loam*, *loam*, *silt loam*, *silt*, *sandy clay loam*, *silty clay loam*, *clay loam*, etc. To qualify for the modifier sandy or silt, a soil must have at least 40 to 45% of these separates. Thus, a loam in which sand is dominant is classed as sandy loam.

In clayey soils, the characteristics of clay separate are distinctly dominant. Three distinct classes exist, viz. *sandy clay* (at least 35% clay), *silty clay* (at least 40% clay and 40% silt) and *clay* (at least 40% clay and less than 40% silt).

The particle size distribution and textural classes of some bench mark soil series of India are given in Table 3.

The textural class is determined on the basis of different proportions of sand, silt and clay by using the textural triangle. There are many textural classification schemes; however, the International Society of Soil Science (ISSS 1929), renamed as International Union of Soil Sciences (IUSS) and USDA Soil Survey Staff (1975) schemes are the most common ones.

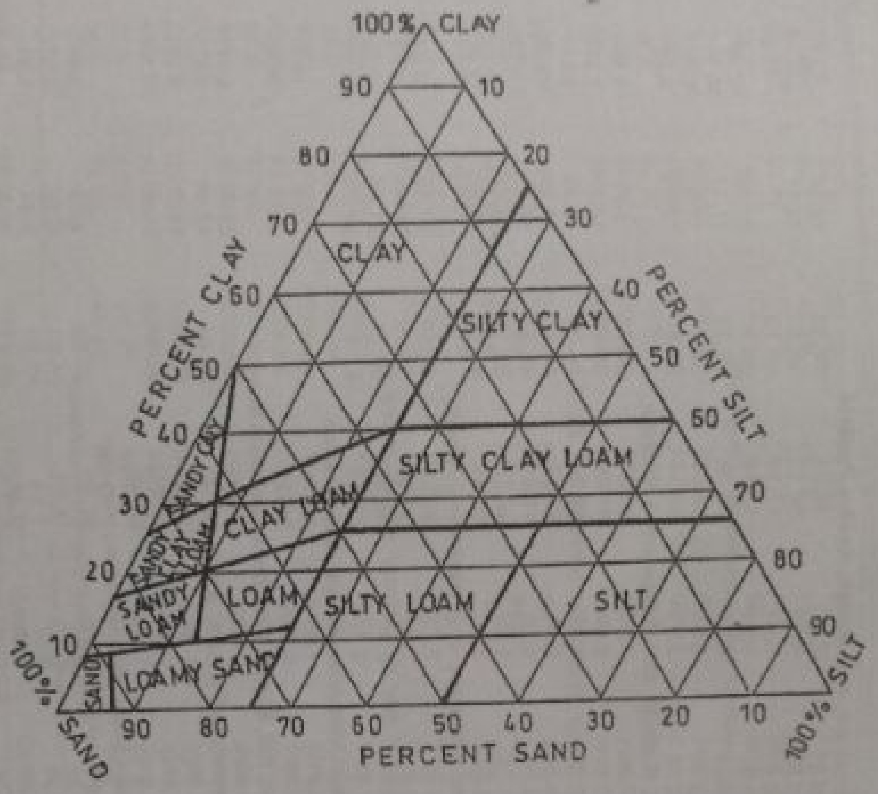


Figure 2. Triangular textural diagram (sand, silt and clay particle sizes of 2-0.02 mm, 0.02-0.002 mm and < 0.002 mm, respectively) (ISSS scheme, 1929)

To illustrate, assume that a soil contains 50% sand, 20% silt and 30% clay. The lower left apex of the triangle (ISSS Scheme, Figure 2) represents 100% sand and the right side of the triangle represents 0% sand. Find the point of 50% sand at the bottom edge of the triangle and follow the diagonally leftward line rising from that point and paralled to the zero line for sand. Then identify the 20% line for silt which is parallel to the zeroline for silt, namely, the left edge of the triangle. The point at which the two lines intersect and meet the 30% line for clay is the point we are seeking. In this particular example, it happens to fall within the realm of "clay loam".

Other textural terms such as peat (fabric), muck (sapric) and mucky peat (hemic) are used to describe organic materials. Peat is a raw undecomposed organic material, while muck is well-decomposed organic material. Mucky peat is an intermediate material between muck and peat. Deposits or soils containing 20% to 50% organic matter are called muck; and those containing more than 50% are classified as peats. In Malaysia, the term organic clay is used in addition to the above terms and the differentiation between the various organic textural classes is essentially based on per cent weight loss after ignition.

The naturally aggregated primary soil particles are separated by removing cementing agents, such as organic matter, calcium carbonate, or iron oxides. Organic matter is removed by oxidation with hydrogen peroxide and $CaCO_3$ is removed by treating with hydrochloric acid. The particles are made discrete by deflocculating the clay with a chemical dispersing agent such as sodium hexametaphosphate and by mechanical agitation (shaking, stirring or ultrasonic vibration). Failure to disperse the soil completely results in flocs of clay or aggregates settling as primary particles and vitiating the results of mechanical analysis.

Table 3. Particle size distribution, textural class, aggregate stability and soil water retention of some benchmark soil series of India'

Horizon	Depth (cm)	Particle size distribution (%)			Textural class	Aggregate stability			Soil water retention (w/w, %)		
		Sand	Silt	Clay		*MWD	**MWD	***% WSA	0 bar	0.33 bar	15 bar
Patancheru series (Udic Rhodustalfs), Patancheru, dist. Medak, Andhra Pradesh											
Ap	0-10	72.1	5.1	22.8	SCIL	0.67	0.63	28.21	36.59	13.36	5.94
A	10-20	75.9	2.6	21.5	SCIL	0.80	0.46	30.75	35.52	13.46	5.60
BA	20-30	65.7	1.3	33.0	SCI	0.61	0.49	57.19	46.73	-	-
Bt1	30-49	50.3	3.8	45.9	CI	3.61	0.99	63.50	59.37	21.59	12.66
Bt2	49-102	36.2	2.6	61.3	CI	5.54	4.34	65.38	65.26	27.41	18.28
BC	102-145	79.8	2.6	17.7	SL	3.49	2.86	84.90	37.27	12.83	6.28
C	145-160	81.1	1.3	17.7	SL	2.13	2.21	90.37	39.87	12.94	6.35
Vijayapura series (Typic Kandiustalfs), vill. Kodihilli, dist. Bangalore, Karnataka											
Ap1	0-7	59.7	7.6	32.7		0.50	0.61	28.35	30.68	9.21	4.17
Ap2	7-21	59.7	5.1	35.2	SCI	0.51	0.60	61.09	38.40	14.13	7.61
Bt1	21-46	54.6	5.1	40.3	CI	0.72	0.87	63.06	44.22	17.33	9.61
Bt2	46-87	54.6	5.1	40.3	CI	3.31	1.44	62.97	44.56	17.53	10.23
Bt3	87-104	54.6	12.7	32.7	CIL	2.70	1.62	71.60	43.26	19.41	11.65
2C1	104-135	67.3	5.1	27.8	SCIL	1.15	1.15	44.78	50.11	24.14	11.77
2C2	135-152	67.3	3.1	30.1	SCI	3.87	2.93	59.89	47.11	23.88	12.93
Trivandrum series (Ustoxic Dystropepts), vill. Srikriyam, dist. Trivandrum, Kerala											
Ap	0-9	60.8	5.2	34.1	SCI	4.87	2.23	60.36	48.53	17.85	12.68
A	9-25	42.7	5.2	52.1	CI	5.32	2.25	50.06	53.19	26.43	19.36
B	25-52	32.4	5.2	62.4	CI	5.57	2.35	40.87	60.58	31.00	22.86
C	52-84	24.7	5.1	70.1	CI	4.29	2.72	70.71	61.67	34.78	24.92
Torkewadi series (Lithic Ustorthents), Rahuri, dist Ahmednagar, Maharashtra											
A	0-15	56.01	15.9	28.09	CIL	4.81	1.95	82.13	59.4	26.8	14.3
C	15-23	56.01	5.3	38.69	CI	5.35	4.75	98.21	62.6	38.1	26.0
Madhpur series (Typic Endoaqualfs), vill. Madhpur, P.O. Bhatar, dist. Burdwan, West Bengal											
Ap	0-14	52.6	26.1	21.5	SiL	0.69	0.76	50.63	40.65	21.71	5.07
BA	14-38	52.5	18.2	29.3	CiL	0.60	0.33	42.92	43.64	17.99	6.97
Bt1	39-98	44.7	18.2	37.1	CI	1.30	1.48	74.57	48.49	19.26	9.21
Bt2	98-150	29.0	23.4	47.5	SiCI	1.39	1.48	77.84	53.96	24.67	12.42
P4/5		29.0	23.4	47.5	SiCI	1.49	1.61	55.66			
Lukhi series (Typic Ustochrepts), vill. Uncha Majra, dist. Gurgaon, Haryana											
Ap	0-8	72.5	10.1	17.4	SL	0.48	0.69	21.30	44.53	8.75	3.97
A	8-27	77.5	5.1	17.4	SL	0.25	0.33	13.63	44.81	6.62	3.62
Bw1	27-55	80.1	2.5	17.4	SL	0.28	0.32	7.73	42.01	6.81	3.81
Bw2	55-86	80.1	2.5	17.4	SL	0.23	0.28	7.48	43.41	6.95	3.72
Bw3	86-125	80.1	2.5	17.4	SL	0.14	0.16	4.11	43.98	7.37	3.90
Bw4	125-178	83.9	3.8	12.3	SL	0.17	0.20	9.18	45.28	7.99	4.06

Contd

Table 3. Particle size distribution, textural class, aggregate stability and soil water retention of some benchmark soil series of India* — *Contd*

Horizon	Depth (cm)	Particle size distribution (%)			Textural class	Aggregate stability			Soil water retention (w/w, %)		
		Sand	Silt	Clay		*MWD	**MWD	***% WSA	0 bar	0.33 bar	15 bar
Mataur series (Dystric Eutrochrepts), vill. Kanra, dist. Kangra, Himachal Pradesh											
Ap	0-21	50.6	35.9	13.5	SiL	0.66	0.72	35.66	54.46	25.73	5.71
BA	21-38	53.1	25.7	21.2	SiL	0.66	0.45	49.78	45.31	19.41	6.72
B1	38-68	55.7	18.0	26.3	ClL	1.15	0.99	53.18	46.28	19.17	8.67
B2	68-90	58.3	17.9	23.8	ClL	0.55	0.53	37.40	44.34	18.98	7.77
Zarifa Viran series (Typic Natraaqualf), Karnal, Haryana											
A1	0-5	48.6	34.3	17.1	SiL	0.86	1.22	39.94	45.85	24.13	5.49
A2	5-24	46.1	34.3	19.7	SiL	0.31	0.26	26.76	45.70	24.18	6.70
Bt1	24-56	43.5	33.0	23.5	SiL	0.31	0.38	14.57	46.25	25.28	8.53
Bt2	56-85	37.2	38.1	24.7	SiL	0.40	0.31	9.25	47.46	28.58	9.58
BCk	85-118	43.5	34.3	22.1	SiL	0.48	0.25	13.38	47.20	28.97	9.11
Ck	118-140					1.93	0.26	8.42	45.44	29.10	10.68
Khatki series (Typic Haplustalfs), Modipuram, dist. Modipuram, Uttar Pradesh											
Ap	0-12	54.9	12.7	32.4	ClL	0.24	0.25	7.53	48.60	21.41	5.44
A3	12-28	52.4	22.9	24.7	ClL	0.19	0.21	5.40	44.97	18.59	5.72
B11	28-45	42.2	21.6	36.2	ClL	0.22	0.26	15.27	48.51	20.93	9.40
B21t	45-78	38.4	21.8	40.0	Cl	0.25	0.31	24.14	49.99	22.54	10.58
B22t	78-99	35.9	22.9	41.2	Cl	0.39	0.44	31.80	49.54	23.52	13.68
B3t	99-140	35.9	22.9	41.2	Cl	0.40	0.39	44.90	49.17	25.34	12.45
Jagdishpur Bagha series (Typic Ustifluvent), vill. Balua, dist. Muzaffarpur, Bihar											
Ap	0-15	57.8	24.0	18.3	L	0.35	0.43	0.43	49.99	27.45	3.80
A3	15-42	59.0	18.9	22.0	L	0.13	0.23	0.23	48.96	29.15	4.14
B_1	42-76	54.0	15.1	30.9	ClL	0.10	0.14	0.14	48.01	29.14	2.50
C_{11}	76-115	38.9	24.0	37.2	ClL	0.12	0.22	0.22	40.88	15.67	1.37
C_{12}	>115	40.1	21.4	38.4	ClL	0.08	0.19	0.19	52.04	31.04	6.11
Coimbatore series (Vertic Ustropepts), TNAU Campus, dist. Coimbatore, Tamil Nadu											
Ap	0-15	49.5	5.3	45.3	Cl	0.38	0.29	17.72	49.63	24.10	13.62
Bw	15-42	49.5	5.3	45.3	Cl	0.55	0.84	65.53	46.84	24.13	14.48
BC	42-75	49.5	5.3	45.3	Cl	0.46	0.43	55.47	48.41	23.92	14.19
Ck	75-124	43.6	20.3	36.1	ClL	5.76	3.34	81.41	52.75	36.29	17.52

*K.M. Hati and A.K. Misra (personal communication);
S = Sand, Si = Silt, Cl = Clay, L = Loam; SClL = Sandy clay loam; *MWD, Mean weight diameter of water stable aggregates including the sand and concretions; **MWD, Mean weight diameter of water stable aggregates excluding sand and concretions; ***%WSA, per cent water stable aggregates of more than 0.25 mm diameter

4. Specific Surface of Soils

The specific surface is an important property of soil grains that results from their sizes. It is defined as the amount of surface area per unit weight or volume of a soil and is expressed in cm^2/g or per cm^3. This is important for chemical and physical reactions.

Specific surface of a sphere is given by Equation (17):

$$\frac{4\pi r^2}{4/3\pi r^3} = \frac{(\text{Surface area})}{(\text{Volume})} \text{ or } \frac{3}{r} \qquad \ldots(17)$$

The specific surface of different soil particle size fractions, soil types and clay minerals are given in Table 4. Besides size, the shape and type of clay minerals also affect specific surface of a soil. The specific surface of soil greatly influences the physical and chemical properties such as retention of water at high suction, swelling, plasticity, soil strength, cation exchange capacity and availability of nutrients.

5. Soil Structure

5.1. *Definition*

Physically, a soil is a mixture of inorganic particles, decaying organic materials, water and air. The inorganic primary particles of various sizes (sand, silt or clay fraction) generally cluster together to form complex and irregular patterns of secondary particles which are called aggregates or peds. **The term 'soil structure' refers to the arrangement of these primary and secondary particles into a certain structural pattern.** Soil structure greatly influences many soil physical processes such as water retention and movement, porosity and aeration, transport of heat, etc. The various soil management practices such as tillage, cultivation, application of fertilizer and manures, amendments (liming, gypsum) and irrigation, bring about changes in soil structure that influences other soil properties, thereby affecting root growth, water and nutrient uptake, crop growth and yield.

5.2. *Classification of Soil Structure*

Classification of soil structure for field description is based on (i) the type, as determined by the shape and arrangement of the peds; (ii) the class, as differentiated by the size of the peds; and (iii) the grade, as determined by the distinctiveness and durability of the peds.

5.2.1. Type of Soil Structure

Based on the shape and arrangement of peds or aggregates, soil structure is classified into four principal types— platelike, prismlike, blocklike and spheroidal with a few subtypes in each category (Figure 3).

(i) *Platelike* — In this structural type the aggregates are arranged in relatively thin horizontal plates, or leaflets. *Platy* structure is often formed from parent materials and can also result due to compaction by heavy machinery on clayey soils.

(ii) *Prismlike* — Two subtypes of this structure, **prismatic** and **columnar**, are vertically oriented aggregates, occurring commonly in

Table 4. Specific surface of soil particles and clay minerals

Size fraction	Diameter (mm)	Specific surface (cm^2/g)	Clay minerals**	Specific surface (m^2/g)
Coarse sand	2-0.2	45	Kaolinite	37-45
Fine sand	0.2-0.02	446	Illite	120-170
Silt	0.02-0.002	4,458	Montmorillonite	580-750
Ultra clay	10^{-4}-10^{-4}	9,890,000 (9.8millions)	Chlorite	130-180
Sandy loam soil		10-40*	Vermiculite	780-900
Loam soil	-	50-100*		
Clay soil		150-250*		

* (in m^2/g).

**Source: Raman and Ghosh (1974)

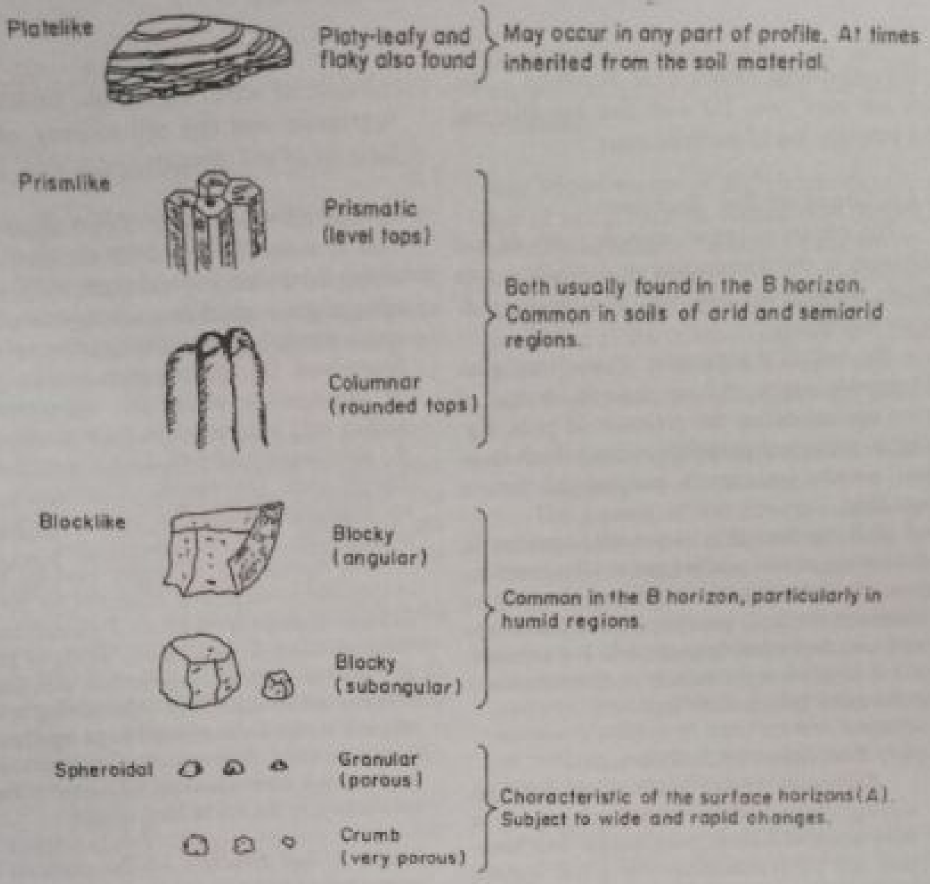


Figure 3. Various structural types found in soils

Source: Brady and Weil (1996)

subsurface horizons of semi-arid and arid regions. These are distinct features in well developed profiles of arid soils of Rajasthan. They are also associated with swelling type of clay and occur in poorly drained soils of humid region too. The prisms having rounded tops are called *columnar* structure, and mostly occur in subsoils of salt-affected/ sodic soils (natric horizon). The *prismatic* structures have the tops of the prisms angular and are relatively flat horizontally.

(iii) *Blocklike* — These aggregates look like blocks, irregularly six-faced and having their three dimensions more or less equal. The peds have sizes varying from 1 cm to 10 cm. Two subtypes of structure exist— *angular blocky* and *sub-angular blocky*. In the former, the cubes have sharp edges and distinct rectangular faces, whereas in the latter, some rounding of the cubes occurs. Formation of these types of structures is influenced by root penetration, soil drainage and aeration. They mostly occur in the subsoil.

(iv) *Spheroidal* — Two subtypes structures— *granular* and *crumb*— are put in this category. When the peds or aggregates are relatively non-porous, they are called *granules* and the porous granules are termed as *crumbs*. These two types of structures occur in surface soils high in organic matter, in grasslands and are greatly influenced by farming practices.

5.2.2. Class of Soil Structure

The various sizes of peds are designated by class of soil structure as very fine, fine, medium, coarse and very coarse. For example, crumb structures having peds less than 1 mm size are very fine, 1-2 mm size are fine and 2-5 mm size are of medium class.

5.2.3. *Grade of Soil Structure*

The degree of aggregation or grade of soil structure is the distinctness or durability with which the peds are observed, viz. weak, moderate and strong.

The soil is structureless when there is no observable aggregation or there is no line of cleavage indicating the presence of peds, e.g. *single grain* soil particles in sand dune areas and *massive* structure in puddled soil from a rice field.

Soil structure is a very useful criterion in field study of soil profiles and in soil classification. Soil structure is described in terms of shape of the peds present, their relative size and also degree of development. For example, a soil horizon might have a weak, medium and subangular blocky structure.

5.3. *Evaluation of Soil Structure*

Evaluation of soil structure assumes special significance in view of its important role in influencing soil-plant relationships. Soil structure can be evaluated by direct and indirect methods (Biswas and Mukherjee, 1987).

5.3.1. *Direct Methods*

The direct methods including the microscopic evaluation in the laboratory and the macroscopic or field observation, characterize the shape, size and arrangement of soil aggregates. In the microscopic method, thin soil sections are examined under a petrographic microscope with respect to the nature, shape and size of the secondary particles and voids. In the macroscopic method, when a chunk of dry soil excavated from soil in the field is allowed to fall gently on a smooth surface, it breaks into small pieces of peds of different sizes and shapes. The shapes and sizes can also be assessed in soil profiles from the cleavages and the *type*, *class* and *grade* of soil structure are determined.

5.3.2. *Indirect Methods*

The indirect methods of soil structure evaluation involve measurement of (i) size distribution of stable aggregates, (ii) stability of aggregates, and (iii) soil property which is a function of soil structure.

(i) Size Distribution of Stable Aggregates— This is measured by dry or wet sieving techniques. In dry sieving technique, the proportion of aggregates which is stable against vibrating action simulating the scouring effect of wind is determined. Dry sieving gives a better picture of aggregation where the aggregates are weakly held together in the moist condition and the mechanical action of sieving is sufficient to destroy them. Dry sieving of aggregates gives an important index for characterizing the susceptibility of soils to wind erosion.

The wet sieving technique used by Yoder (1936) and subsequently improved by the Soil Science Society of America's Committee on Physical Analysis (van Bavel, 1953) is generally used for determining the size distribution of water stable aggregates. Wet sieving is well adapted to the separation of large aggregates. It can be used to screen out aggregates as small as 0.1 mm, although 0.25 mm is more satisfactory as the lower limit of size.

Indices of Soil Structure — The quantum of aggregates of different size-ranges is to be represented by a single value index. Size and weight being the variables, weighted mean diameter is the widely used index. The commonly used forms of mean diameter are: *Mean Weight Diameter*, *Change in Mean Weight Diameter*, *Geometric Mean Diameter*, and *Log Standard Deviation* and *Coefficient of Aggregation*. Also, the *percentage of aggregates greater than 0.25 mm* diameter is a widely used index. In wetland rice soil, the change in mean weight diameter of water stable aggregates before and after puddling as well as clay dispersion are useful for characterizing the change in soil structure due to puddling.

The Mean Weight Diameter (MWD) — It gives an estimate of weighted percentage of average size of all the aggregates. The proportion by weight (w_i) of a given size fraction of aggregates to the total sample weight (W) is multiplied by the mean diameter x_i of that fraction. The sum of these products for all size fractions is called the MWD.

$$\text{MWD} = \sum_{i=1}^{n} x_i w_i \qquad \text{...(18)}$$

where, n is the number of fractions; $w_i = \frac{W_i}{W}$, where, W_i is the weight of each size fraction and W is total sample weight.

MWD may also be estimated by plotting the accumulated percentages of aggregates against average diameter and determining the area atop the curve.

Change in Mean Weight Diameter (CMWD): It takes into account the stability of aggregates both under dry and wet sieving conditions. The difference between the two MWDs is the CMWD. Within certain limits, the lower the CMWD, the better is the soil structure.

In *Geometric Mean Diameter* (GMD) the weight of the aggregates in a given size fraction (W_i) is multiplied by the logarithm of the mean diameter (x_i) of that fraction. The sum of these products for all the size fractions is divided by the total weight of the sample (W). GMD is calculated using the Equation (19):

$$\text{GMD} = \exp\left[\sum_{i=1}^{n} W_i \log x_i / W\right] \qquad \text{...(19)}$$

The mean weight diameter of water stable aggregates and the percentage of water stable aggregates greater than 0.25 mm diameter of some bench mark soil series of India are given in Table 3.

(ii) Measurement of Stability of Aggregates: This is evaluated by the degree to which soil aggregates resist dispersion. Several indices have been developed such as *Stability index, Structural coefficient, Dispersion coefficient* and *Water drop method*. Stability index is the difference between per cent clay and silt as determined by the mechanical analysis and that obtained by suspension of soil sample in water. The greater the difference, the better is the soil structure. Structural coefficient is given by (D–S)/S, where, D is the percentage of particles of < 0.25 mm diameter, as determined by the mechanical analysis and S is the percentage of aggregates smaller than 0.25 mm diameter, as determined by wet seiving technique. The higher is this value, the better is the structure.

(iii) Measurement of Soil Properties: A number of soil properties such as bulk density, rate of water infiltration, hydraulic conductivity, rate of aeration and aeration porosity, available soil water and degree of compaction are used as indicators of the suitability of soil structure for plant growth. A compacted soil with higher bulk density indicates poor structural condition.

5.4. Genesis of Soil Structure and Stable Aggregate Formation

The genesis of soil structure refers to the causes and methods of formation of the structural units or aggregates. An aggregate consists of grouping of a number of primary particles into a secondary unit. The mechanism of formation of these aggregates involves several factors such as vegetation, soil fauna, microorganisms, impact of cations, clay particle interactions in relation to moisture and temperature as well as organic matter and clay-organic matter interactions (Baver *et al.*, 1972).

5.4.1 Flocculation and Aggregate Formation

In elucidating aggregate formation it is important to distinguish between flocculation and aggregate formation, which are not synonymous. In a suspension, the primary particles with high electrokinetic (zeta) potential repel each other. With the addition of flocculating agent, the zeta potential is lowered, the particles collide, mutually attract each other and settle in the form of a *floccule*. The floccules are stable as long as the flocculating agent is present. However, aggregate formation requires the cementation or binding together of flocculated particles so that the primary particles do not disperse in water. Hence, stable aggregate formation or granulation is flocculation plus its cementation which are brought out by interplay of several factors as stated below.

5.4.2. Effects of Vegetation, Soil Fauna and Microorganism

During growth plant roots pushing through the soil, tend to compress soil particles into small aggregates and break the larger aggregates present in the soil. Root exudates help to bind the soil particles together into aggregates. The dehydration of soil in the vicinity of the root system as water is absorbed by the plants, causes local shrinkage and formation of surfaces of fracture.

The beneficial effects on soil aggregation and its stability originating from the integrated activity of soil fauna, microorganisms and vegetation are: (i) The fungi and actinomycetes cause mechanical binding of the aggregates by the mycelia they produce. (ii) The metabolic processes of the mircroorganisms synthesize complex organic molecules which have cementation effects. Thus, microbial products like bacterial gum and microbial polysaccharides, play a major role in soil aggregate stabilization. (iii) Products of microbial decomposition of organic materials added to soil, such as humic acids, colloidal proteins and cellulose materials, also produce stabilizing effect.

5.4.3. Cationic Effects

Aggregate formation is highly influenced by the nature of cations adsorbed by the soil colloids. For example, when Na^+ is a dominant adsorbed ion, as in soils of arid and semi-arid areas or alkali soils, the colloids continue to repel each other causing a dispersed condition. In such a state, soil particles do not come together to form structural aggregates. The soil becomes impervious to water and air. The divalent cations like Ca^{2+}, Mg^{2+}, on the other hand, form electropositive links between the electronegative soil particles which cause the individual colloidal particles to come together and form small aggregates called floccules. Calcium affects the production and decomposition of organic matter in soil. It also helps in the binding action between organic colloids and clay particles.

5.4.4. Effects of Organic Matter

Organic matter plays an important role in stimulating the formation and stabilization of granular and crumb type aggregates. The role of organic root exudates, microbial and decomposition products of organic materials in formation of stable aggregate has already been stated. Organic compounds, such as polysaccharides, chemically interact with the clays and form bridges between the individual soil particles and bind them together in water-stable aggregates.

5.4.5. Clay Particle and Clay-Organic Interaction

Three distinct groups of soil colloidal materials have cementation and aggregating effects— clay particles themselves, inorganic colloids of iron and aluminium oxides, and organic colloids.

Clay-to-clay particle interaction, particularly under dry condition, takes place due to electrostatic and van der Waal's forces. It leads to the formation of secondary particles. Aggregate formation is dependent upon the linkages established by polyvalent exchangeable cations on the clay particles. Polyvalent cations, in fact, serve as bridges to form clay-organic complexes, leading to the formation of stable aggregates.

Colloidal hydrated iron oxides become almost completely irreversible on dehydration, which is an important factor in the production of stable aggregates. This is observed in lateritic soils which have high degree of aggregation. Sesquioxides form complexes with humus, resulting in stable aggregate formation.

The bonding of organic polymers to clay surfaces by cationic bridges, hydrogen bonding, van der Waal's forces and sesquioxide-humus complexes are important mechanisms by which organic colloids stabilize soil structure. As a result of the interaction, the properties of the colloidal clay surfaces change. The organic colloids existing in association with the clay particles may tend to reduce wetting by water as well as swelling.

5.5. Management of Soil Structure

Soil structure management aims at the improvement and maintenance of soil structure, which are the major challenges in cultivated lands. The general principles relevant to structural management are :

(i) Tillage activities need to be restricted to periods of optimum soil moisture condition to ensure least destruction in the soil structure;

(ii) Adoption of suitable tillage or minimum tillage to reduce the loss of aggregates stabilizing organic matter;

(iii) Soil surface should be kept covered with crop residues or plant litter to protect aggregates from the beating action of rain and to add organic matter to soil;

(iv) Incorporation of crop residues and animal manures (FYM, poultry manure, etc.) into the soil would stabilize soil aggregates through supply of decomposition products;

(v) Suitable cropping systems, application of phosphate fertilizers, inclusion of grasses and sod crops in rotation, would enhance organic matter and favour stable aggregation;

(vi) Green manuring and cover crops are good sources of organic matter.

Each soil has distinct problems and requires specific soil and crop management practices for soil structure management.

6. Dynamic Properties of Soils

The behaviour of soil to an applied stress is expressed in terms of dynamic properties of soils which are mostly governed by two physical properties—adhesion and cohesion. *Adhesion* refers to the attraction of two dissimilar phases, e.g. attraction of a liquid phase such as water on the surface of a solid phase such as soil solid particles. *Cohesion* is the attraction between two similar phases. In soils, cohesion is bonding of the particles due to attractive forces between them due to van der Waal's forces, electrostatic attraction of negatively charged clay surfaces and positively charged clay edges, cationic bridges, cementation effects of organic matter and surface tension forces, etc. Cohesion also takes place between the water molecules which remain as films between the adjacent soil particles.

6.1. Soil Consistency

The manifestations of the physical forces of cohesion and adhesion acting within the soil at various moisture contents are designated by the term *soil consistency*. It includes such properties of the soil as resistance to compression, friability, plasticity, stickiness, etc. Thus, most soils exhibit four forms of consistency from progressively wet to dry soil moisture conditions.

(i) Hard or Harsh Consistency — It has the pronounced characteristics of hardness. At low moisture contents, the soil remains very hard and coherent due to cementation effect between the dried particles.

(ii) Friable Consistency — As the moisture content of the soil increases, water molecules are adsorbed on the surface of the soil particles and decrease the coherence. The soil mass becomes friable. Friability characterizes the ease of crumbling of soils. The range of soil moisture contents in friable condition is optimum for tillage operation.

(iii) Plastic Consistency — It is manifested by the properties of toughness and the capacity of the soil to be moulded into any desired shape.

(iv) Sticky Consistency — It is evident by the property of stickiness to various objects. The sticky point is defined as the moisture content of the soil when it ceases to stick to any foreign object.

6.2. Soil Plasticity

Soils containing more than 15% clay exhibit plasticity. Plasticity is defined as the property which enables a clay/ soil to take up water, to form a mass that can be deformed into any desirable shape and to maintain the shape after the deformation pressure is removed. Plasticity results from the plate like nature of the clay particles and the combined binding/ lubricating effect of the adsorbed water. With the adsorption of water, thin films are formed around the soil particles. With applied pressure or force, the particles slide over each other and are held in that condition by the tension of the moisture films even after the pressure is removed.

6.2.1. Indices of Plasticity and their Significance

Plasticity is exhibited over a range of moisture contents referred to as plasticity limits. There are three indices called the *lower plastic limit* or simply the *plastic limit*, the *upper plastic limit* or *liquid limit* and the *plasticity index*. The *plastic limit* is the lowest moisture content at which a soil can be deformed without cracking. It is the upper limit of moisture content for tillage operation for most crops, except rice. Tilling at moisture content above plastic limit results in smearing and puddling of soil. The *upper plastic limit* or *liquid limit* is the moisture content at which a soil ceases to be plastic. It becomes semifluid and tends to flow like a liquid under an applied force. This limit is used for classfication of soils for engineering purpose. The difference in the moisture content between the upper and lower plastic limits is the range over which a soil remains plastic and is called *plasticity index*. Soils with high plasticity index are difficult to plough. Soils with expanding lattice clays such as smectites have high liquid limits and plasticity indices. Clay minerals like kaolinite have low liquid limit values. Soil plasticity is determined by an apparatus called *liquid limit device*.

7. Soil Compression and Compaction

7.1. Soil Compression

Soil compression is defined as the change in volume of a soil under an applied stress, whereas soil compaction refers to the increase in density of a soil as a result of applied pressure or load.

Volume of soil is consisted of solid phase and the voids between it, which are occupied by water and air. So compression denotes the decrease in the void ratio per increment of applied pressure or load. The increase in compression with each increment of pressure can be expressed by the Equation (20):

$$y = ae^{-bx} \qquad \text{...(20)}$$

where, 'y' is the amount of compression, 'x' is the pressure, and 'a' and 'b' are constants.

Compression takes place with increment in pressure due primarily to particle orientation and changes in the size of the colloidal micelles. Soils may be compressed under low or high pressures. In the latter case, the soil expands when the pressure is removed because of swelling of the micelles. Compression is affected by the nature of clay mineral, the type of exchangeable cations and the degree of polarity of organic liquids. With increase in soil moisture content, compression increases to a maximum and then decreases with further increase in moisture content. The maximum compression occurs within the plastic range of soil moisture content.

7.2. Soil Compaction

Soil compaction encompasses compression plus increase in density of a soil. It is the dynamic behaviour of soil. The degree of compaction depends upon the nature of soil, amount of energy applied, water content and extent of manipulation of the soil. Compaction is also associated with the rearrangement of the soil solid particles so that soil water and soil air are compressed within the pore spaces.

Because of the imcompressible nature of soil particles and high internal friction, dry soils cannot be compacted to high densities. An increase in water content decreases cohesion between the particles and internal friction, thereby facilitating compaction. During compaction, the density of soil under a load increases with the increase in soil water content up to a certain limit, beyond which further addition of water does not increase compaction because of imcompressible nature of soil solids and water. This critical limit is called *procter moisture content*.

In many soils, a compacted layer is commonly found at the bottom of the zone of ploughing. This layer is termed as *'plough-sole'*. Age-old practices of tillage using wooden plough and other farm practices including use of tractor and other heavy farm equipments, create the problem of surface and sub-surface compaction in soils. These hard compact layers often restrict root penetration and growth and also reduce water and nutrient uptake by crops. Soil compaction changes soil moisture retention, soil thermal regimes and mechanical resistance in soils. For enhancing crop growth, the management practices should be able to break the hard compacted layer and should avoid or reduce its formation. However, in highly coarse-textured sandy, loamy sand soils, compaction may be a boon to enhance the water retention and reduce percolation loss of water. This is elaborated in this chapter in section No. 11.1.

8. Soil Crusting

Soil crust is the thin compact layer of higher bulk density formed at the soil surface following dispersion of natural soil aggregates as a result of wetting and impact of raindrops and its subsequent rapid drying due to radiant energy of the sun.

8.1. Crust Formation

When the raindrops strike the exposed dry soil surface, there is disintegration and dispersion of the aggregates. The finer clay particles move down along with infiltrating water and clog the pores immediately beneath the surface, thereby sealing the soil surface. Also, the dispersed soil may remain in suspension, coarse particles start to settle out, but fine clay particles remain in suspension. As the water evaporates, clay settles on the top of coarse particles, forming a crust on drying. The soil particles tend to pull together due to surface tension forces and form a dense strong soil layer with decreased porosity. Soil crusting is a major structural feature of soils of arid and semi-arid regions. The problem of soil crusting is severe in silty clay loams of older alluvial terraces and levees due to their unstable soil structure.

8.2. Properties and Effect of Soil Crust Formation

The strength of soil crust increases linearly during the first few days of formation. The thickness of the soil crust formed depends on the size of raindrops. Larger raindrops with more energy destroy the original soil structure to a greater depth. The crusts formed are thicker, varying from 4 to 5 mm. In a cloddy soil where fine particles removed from the surface of the clods by rain-drop impact are deposited in inter-clod areas, the crusts become thinner and weaker. The thickness of the crust also depends upon the amount and type of clay and silt present in the soil. The small amount of kaolinite type of clay tends to form a thin crust which curls up and breaks on drying, but montmorillonite type of clay tends to form a thick and hard crust. The thickness of the crust increases with increase in clay content.

In addition to size and intensity of rain drops, soil factors that result in low aggregate stability, encourage soil crust formation. Soils having low organic matter, high silt content and high exchangeable sodium percentage are more susceptible to soil crusting.

Soil crusts have high bulk density, low non-capillary porosity and high soil strength. Stratification and orientation of soil solids and subsequent cementation of particles when dry, make the crusts harder than the rest of soil mass. Infiltration through the crusted layer and the hydraulic conductivity of the crusted layer become very low (Figure 4). As a result, run-off and soil loss increase. Redistribution of water in the soil profile is also greatly affected. The most important effect of soil crusts is the emergence and growth of seedlings which are greatly affected. These, in turn, affect crop stand and ultimately the yield.

8.3. Evaluation of Soil Crust Strength

Soil crust strength can be evaluated by using a penetrometer, balloon pressure technique or modulus of rupture test. A pocket penetrometer or a cone penetrometer is pushed steadily into the soil and the penetration resistance is measured by directly taking the reading on the body or dial of the penetrometer, which is expressed in kg/cm². In balloon pressure technique, the balloons are burried in the soil and the pressure required to inflate the balloon at a

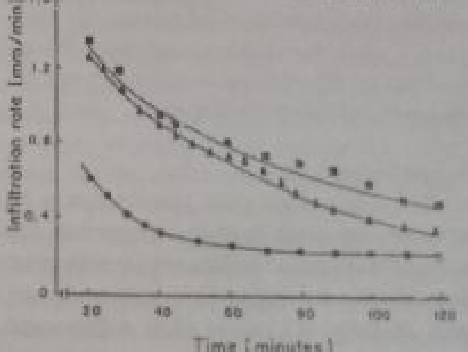


Figure 4. Infiltration rate versus time relationships for crusted, cultivated and FYM-treated soils

point where it ruptures the soil surface is measured. In the modulus of rupture test, a briquette of soil sample is prepared and the force required to rupture it is measured. For a rectangular briquette, the modulus of rupture S (dynes/cm^2) is calculated from the Equation (21):

$$S = \frac{3FL}{2bd^2} \qquad \ldots(21)$$

where, F (in dynes) is the breaking force, L is the distance (in cm) between the bars supporting the briquette, b is the width, and d is the thickness of the briquette (in cm).

8.4. *Reducing Soil Crust Formation and Increasing Emergence of Seedlings*

Soil surface-covers such as mulches, plant residues, etc. reduce the impact of raindrops, prevent dispersion and breakdown of structure of the surface soil, reduce crust formation and maintain high infiltration rate. Close growing crops, grasses and legumes that cover the soil to a large extent also prevent crust formation. Crusting tendency of soils can be reduced by using farmyard manure and chemical amendments like cationic, anionic and non-ionic surfactants and use of gypsum in salt-affected soils, which in turn enhance the seedling emergence (Figure 5). Crop rotation and residue management increase the soil organic matter content and help in formation of low strength crusts. Keeping the soil moist by irrigation also helps in reducing the crust formation. In crust-affected soils, seedling emergence can be enhanced by mechanical breaking of the crusts, planting seeds in furrows and dibbling 2-3 seeds per hill where the seedlings exert more thrusts to overcome the physical restraints imposed by the overlying soil crusts (Figure 6).

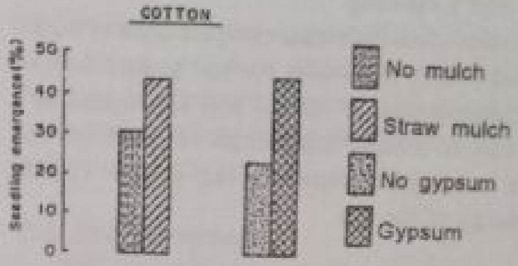


Figure 5. Seedling emergence of cotton as influenced by mulch and gypsum application in crust-affected soils

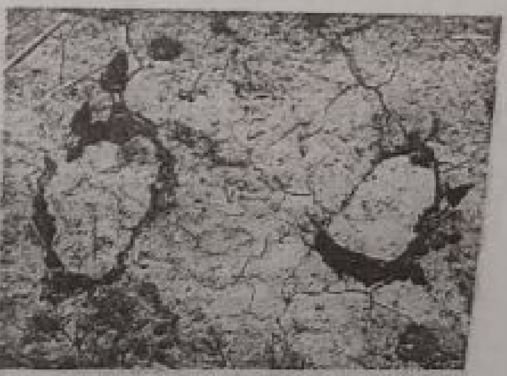


Figure 6. Seedlings exert thrust to emerge through soil crusts

9. Soil Colour

Soil colour often gives a ready clue to its condition and some important properties. For example, dark-colour surface soils absorb more solar radiation than light-colour soils and so get warm up faster. Colour is often a diagnostic mode of major soil classification and interpretation. Soil colour is included in the description of a soil profile. Surface colour that differs from that of the parent material usually serves as an indicator of the processes involved in the soil formation. It is also indicative of other factors such as presence of excessive salts (white), erosion, etc.

9.1. *Determination of Soil Colour*

For determination of soil colour, standard system using Munsell colour chart is used. A small piece of soil is compared to the standard colour chips in the soil colour book. Each colour chip is described by three components of colour, namely, hue, value and chroma. Hue is the dominant spectral colour. Value refers to the relative brightness (lightness or darkness) of the colour and is a function of the total amount of light reflected. Chroma is the relative purity or strength (intensity) of the dominat spectral colour.

9.2. *Factors Influencing Soil Colour*

Soil colour varies from place to place in the landscape. Even adjacent soils have different surface horizon colours such as in black and red soils existing side by side in Andhra Pradesh. Within a soil profile, the colour may change with depth through the different horizons. Even within a single horizon, colour may vary from spot to spot.

Most important factors influencing the soil colour are mineralogy and chemical constituents, organic matter content, soil moisture, soil structure and particle size. Red, yellow or brown colours are mostly related to different degrees of oxidation, hydration and diffusion of iron oxides in the soils, which form coatings on the soil particles. Thus, yellow, red and brown colours are mostly due to the presence of goethite, hematite and maghemite, respectively. The dark colour of soils is associated with a combination of several factors, including the content and decomposition of organic matter, impeded drainage condition and the presence of titaniferous magnetite. Minerals like manganese oxide and glauconite impart black and green colour, respectively to soils. Carbonate, such as calcite accumulating in soils of semi-arid regions imparts a white colour to it. Under prolonged wet anaerobic conditions, the coatings of reduced iron oxide change the soil colour to gray or bluish, a condition of *gley* formation.

10. Soil Physical Constraints that Affect Crop Production

In India, about 85.65 Mha of land produce very low yield of crops due to unfavourable soil physical conditions. The major soil physical constraints identified are: low water retention and high permeability, slow permeability, surface and sub-surface mechanical impedence, hardening and shallow depth of soils, which either restrict crop growth or reduce efficiency of basic inputs, such as water, fertilizer, etc. (Figure 7). The physical properties of representative physically-constrained soils are given in Table 5 (Gupta *et al.*, 1984).

10.1. Low Water Retentive and Highly Permeable Soils

These soils covering about 10.77 Mha, occur mostly in Rajasthan and some parts of Haryana and Punjab. The high permeability of these soils are associated with their sand and loamy sand texture. The soil has low bulk density but high hydraulic conductivity and infiltration rates (as high as 36.5 cm/h for some profiles), which indicate high permeability and low water retention capacity of the soil. The fertilizer and water use efficiency of these soils is very low and the nutrient losses are very high. These problems do not encourage the farmers to use high levels of inputs. The major crops grown on these soils are: pearl millet (*bajra*), maize, wheat and barley in the western parts of India and sorghum, maize, *ragi* (minor millets) and sugarcane in the southern parts of India.

10.2. Slowly Permeable Soils

The slowly permeable soils of about 9.43 Mha occur in Madhya Pradesh, Maharashtra and also in parts of Rajasthan, Uttar Pradesh, Bihar, and Tamil Nadu. The infiltration rate is as low as 0.2 cm/h and hydraulic conductivity is less than 0.15 cm/h for some of the profiles, which indicate slow permeability of the soil and possibility of submergence during rainy season. The very low permeability, which is associated with black clay soils, creates oxygen stress in the root zone due to stagnation of water. The prevailing anaerobic conditions cause the accumulation of carbon dioxide and other by-products in this zone, which restricts the root growth. These black clay soils are sticky when wet and very hard when dry often forming big cracks (Figure 8), thus, these soils could be cultivated or tilled only within a limited soil moisture range.

In the valley lands where the topography is flat, weed menace associated with humid conditions combined with limited range of workability of these soils, prevents the farmers from growing crops during the rainy period. On the undulating and rolling topography, the poor vegetative cover during most of the rainy season exposes the surface soil to the impact of high intensity storms causing loss of large quantities of soil along with run-off water. Under these conditions, a shallow or deep *havali* system is followed by the farmers on low lands to conserve maximum amount of moisture in the soil for the cultivation of crops

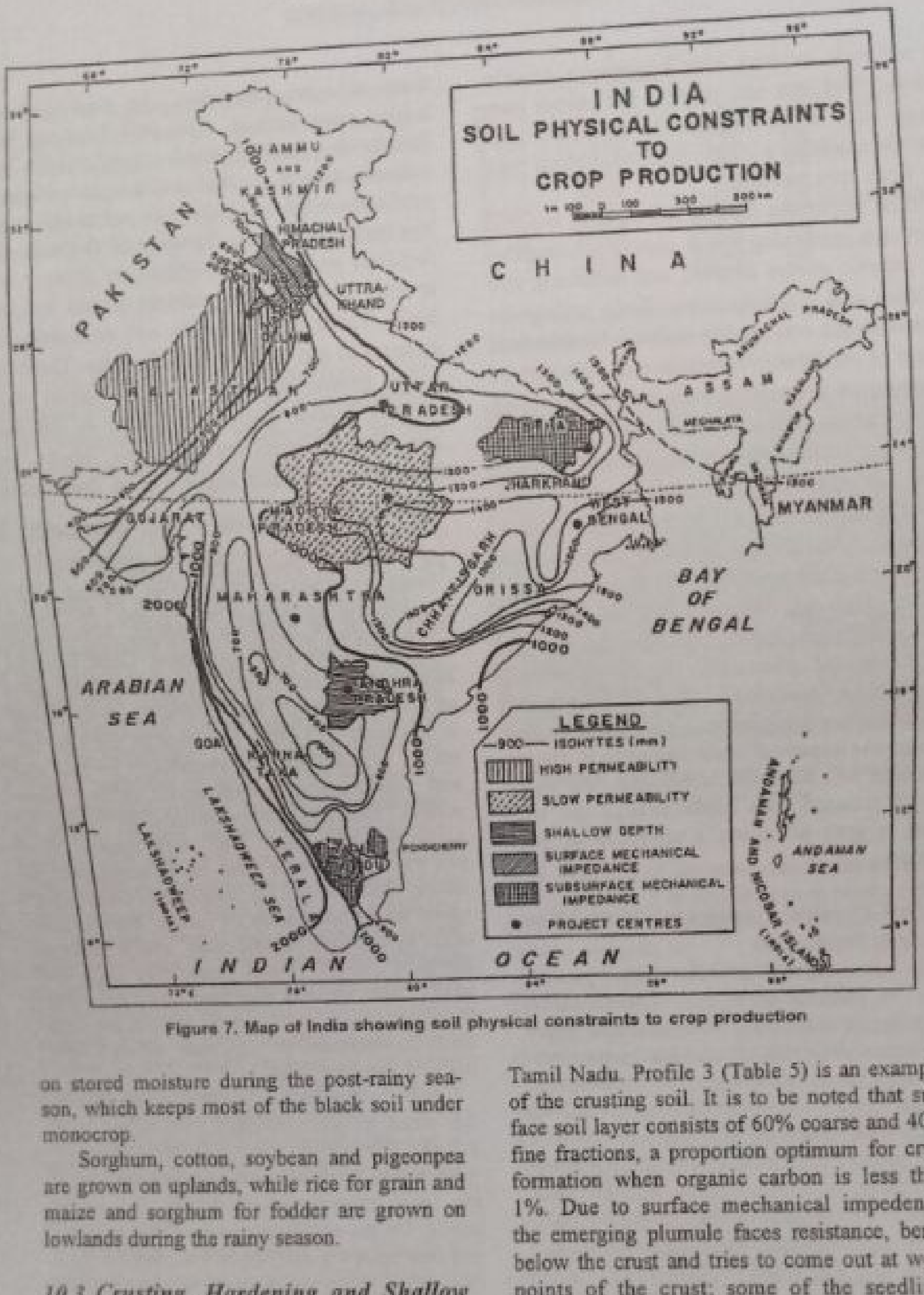


Figure 7. Map of India showing soil physical constraints to crop production

on stored moisture during the post-rainy season, which keeps most of the black soil under monocrop.

Sorghum, cotton, soybean and pigeonpea are grown on uplands, while rice for grain and maize and sorghum for fodder are grown on lowlands during the rainy season.

10.3. Crusting, Hardening and Shallow Soils

The shallow soils cover about 25.02 Mha, hardening soils 20.35 Mha and soils having other physical constraints 9.45 Mha. These soils occur widely in Haryana, Andhra Pradesh and Tamil Nadu. Profile 3 (Table 5) is an example of the crusting soil. It is to be noted that surface soil layer consists of 60% coarse and 40% fine fractions, a proportion optimum for crust formation when organic carbon is less than 1%. Due to surface mechanical impedence, the emerging plumule faces resistance, bends below the crust and tries to come out at weak points of the crust; some of the seedlings injure their tips and fail to emerge. The emergence of pearl millet, cotton and jute seedlings is adversely affected if the crust is formed on the soil surface by the occurrence of rainfall within 48 hours of sowing.

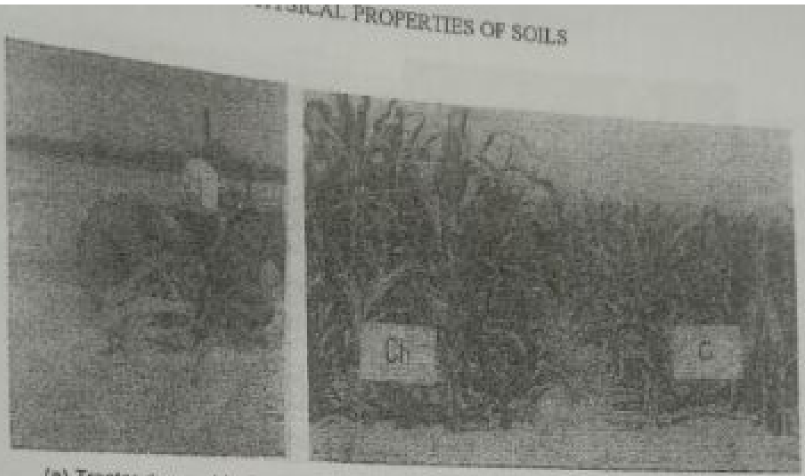
Table 5. Physical characteristics of representative physically constrained soils

Depth (cm)	Sand (%)	Silt (%)	Clay (%)	Organic carbon (%)	Texture (USDA) classification)	Bulk density (Mg/ m³)	Moisture retention (g/100 g) at			Soil water retention between 0.1 and 15 bars	Infiltration rate (cm/h)
							1/10 bar (%)	1/3 bar (%)	15 bar (%)		
					(a) Highly permeable soil						
Profile 1. Jobner sand, desert soil, Entisol/ Ustipsamment, Jobner (Jaipur), Rajasthan										12.9	36.5
0-40	89.9	4.5	4.5	0.11	S	1.50	9.8	3.8	1.1		
40-95	86.9	6.3	6.8	0.15	LS	1.56	10.0	4.2	1.5		
95-158	87.9	5.9	5.3	0.06	S	1.55	9.3	4.0	1.3		
158-215	88.9	5.8	4.2	0.06	S	1.54	9.9	3.9	1.1		
					(b) Slowly permeable soil						
Profile 2. Black clay soil, Vertisol, Kuthuliaya (Rewa), Madhya Pradesh											
0-15	25	30	45	0.31	C	1.36	44.5	36.8	21.7	24.9	0.2
15-30	27	27	45	0.24	C	1.48					
30-45	25	30	45	0.16	C	1.49	46.0	38.1	21.7		
45-60	35	40	25	0.17	L	1.52					
60-75	32	40	27	0.24	L	1.55	48.5	39.8	21.9		
75-90	35	42	22	0.11	L	1.52					
					(c) Crusting soil						
Profile 3. Barwala sandy loam, Inceptisol/Ustochrept, University Farm, Hisar, Haryana											
0-15	59.7	24.1	16.2	0.54	SL	1.52	35.1	29.2	8.1	21.5	0.4
15-45	69.8	13.8	16.4	0.60	SL	1.59	35.3	29.8	8.5		
45-80	68.3	13.9	17.8	0.77	SL	1.55	35.8	30.1	8.7		
80-120	69.4	12.8	17.8	0.81	SL	1.54	36.1	31.2	8.8		
120-185	70.0	14.0	16.6	0.77	SL	1.67	37.1	31.5	8.9		
					(d) Hardening soil						
Profile 4. Hyderabad red sandy clay loam, Alfisol/ Paleustalf, Hyderabad, Andhra Pradesh											
0-10	62.4	12.5	23.5		SCL	1.5		18.7	8.9	5.8 (for 32 cm)	2.8
10-32	47.2	14.8	35.7		SCL	1.7		22.2	10.8		
32-65	Loose calcareous gravelly and murrumy layer admixed with soil										
					(e) Highly subsurface mechanical impedence soil						
Profile 5. Pichanur red clay loam, Entisol/ Ustorthent, Pichanur (Coimbatore), Tamil Nadu											
0-12	66	11	23	0.60	SCL	1.55	25.6	16.2	6.8	15.9	4.0
12-32	37	13	48	0.35	C	1.62	40.6	31.8	16.8		
32-108	55	22	22		GSCL	1.72	25.6	18.4	10.5		

S= Sand, LS = Loamy sand, L = Loam, SL = Sandy loam, C = Clay, CL = Clay loam, SCL = Sandy clay loam,
GSCL = Sandy clay loam mixed with gravel

The red sandy clay loam 'Chalka soil' of Andhra Pradesh dries out very quickly due to low water-retention capacity and becomes very hard. The maximum root growth of most of the crops is confined to the surface layer, and the crop growth suffers due to hardening. The crops grown on this soil include pigeonpea, maize, castor, groundnut and sorghum. The yield of groundnut is very low due to reduced size of pods caused by hardening of the soil.

Profile 4 (Table 5) is an example of hardening of soils. The surface soil layer contains about 62% coarse and 36% fine fractions, a proportion optimum for hardening and crust



(a) Tractor-drawn chiseller (b) Maize growth with (Ch) and without (C) Chiseling

Figure 11. Chiseling technology for management of soils having mechanical impedance at shallow depths

organic manures. To improve the structure and reclaim highly dispersed sodic (alkali) soils, application of gypsum/pyrites in combination with organic manures, green manuring or incorporation of crop residues is successful.

11.5. Chalka Soils

The 'crop residue recycling technology' involving incorporation of high C:N ratio organic materials, viz. powdered groundnut shells or paddy husk has been recommended for the management of red sandy loam '*Chalka*' soils, which become very hard on drying. The 'seed line mulch technology' involving application of FYM or wheat *bhusa* @ 25 q/ha on seedlines immediately after sowing of pearlmillet and cotton seeds has been useful for the management of soils suceptible to form a crust.

11.6. Rice-Wheat Cropping System

Management of soil structure and enhancing and sustaining crop yields under rice-wheat cropping system in alluvial soils of Indo-Gangetic Plains are formidable tasks. Deep ploughing along with incorporation of crop residues, FYM, leguminous green manure crops during puddling for growing of paddy have been found successful (Das *et al*., 1997). For subsequent wheat crops, tillage must be timed carefully with respect to soil moisture content to reduce and avoid large clod formation. Adoption of resource conservation technology such as zero tillage or minimum tillage for growing wheat subsequent to rice has been found to be very useful.

12. References

Baver, L.D., Gardner, W.H. and Gardner, W.R. (1972) *Soil Physics*. Wiley Eastern Limited, New Delhi, 498 p.

Biswas, T.D. and Mukherjee, S.K. (1987) *Textbook of Soil Science*. Tata McGraw-Hill Publishing Company Limited, New Delhi, 314 p.

Brady, N.C. and Weil, R.R. (1996) *Elements of the Nature and Properties of Soils*. Prentice Hall, New Jersey, USA.

Das, D.K., Kumar, S., Aggarwal, P. and Garg, R.N. (1997) Tillage to improve soil physical properties and crop production. *AICRP on Tillage Requirements of Major Indian Soils for Different Cropping Systems*, IARI (ICAR), New Delhi, 129 p.

Gupta, R.P., Kumar, S. and Singh, T. (1984) Soil management to increase crop production. Consolidated Report 1967-1982; *AICRP on Imporvement of Soil Physical Conditions to Increase Agricultural Production of Problematic Areas*, ICAR, New Delhi. 104 p.

Hillel, D. (1998) *Environmental Soil Physics*. Academic Press, San Diego, USA; 771 p.

ISSS (International Society of Soil Science) (1929). Minutes of the First Commission Meetings. International Congress of Soil Science, Washington, 1927. *Proceedings of International Society of Soil Science* 4, 215-220.

Sample Problems

(In calculation of numericals, use either C.G.S. units or S.I. units. The two units should not be mixed up)

Question 1. A soil sample collected with a cylindrical core sample having height of 15 cm and diameter of 6 cm, has a fresh weight of 675.0 g and oven dry weight of 550.0 g. Calculate bulk density, specific volume, porosity, void ratio, mass wetness, volume wetness, degree of saturation, water volume ratio, airfilled porosity and depth of soil water in 0-15 cm soil layer (assume particle density $\rho_s = 2.65$ g cm^{-3}).

Solution:

Volume of core sample i.e. soil core $V_t = \pi r^2 h = 3.143 \times (3)^2 \times 15 = 424.3$ cm^3

Weight of moisture = Fresh weight of soil (M_t) – Oven dry weight of soil (M_s)

$= 675.0 - 550.0 = 125.0$ g

(i) using Eq. (2), Bulk density $\rho_b = \frac{M_s}{V_t} = \frac{550.0}{424.3} = 1.30\,\text{g cm}^{-3}$;

(ii) using Eq. (4), Specific volume $v_s = \frac{V_t}{M_s} = \frac{424.3}{550.0} = 0.77\,\text{cm}^3\text{g}^{-1}$;

(iii) using Eq. (6), Porosity $f = 1 - \frac{\rho_b}{\rho_s} = 1 - \frac{1.30}{2.65} = 0.509$ or 50.9%;

Alternately, $f = \frac{\text{Volume of pores}}{\text{Total volume}} = \frac{V_f}{V_t} = \frac{V_t - V_s}{V_t}$

where, V_s = volume of soil solid $= \frac{M_s}{\rho_s} = \frac{550.0}{2.65} = 207.6\text{cm}^3$

So $f = \frac{424.3 - 207.6}{424.3} = \frac{216.7}{424.3} = 0.511 \text{or} 51\%$

(iv) using Eq. (7), Void ratio $e = \frac{V_f}{V_s} = \frac{424.3 - 207.6}{207.6} = 1.044$

(v) using Eq. (8), Mass wetness or gravimetric water content

$$w = \frac{M_w}{M_s} = \frac{125}{550} = 0.227 \text{ or } 22.7\%$$

(vi) using Eq. (10), Volume wetness or volumetric water content

$$\theta = w.\rho_b = 0.227 \times 1.30 = 0.295 \text{ or } 29.5\%$$

Alternately, $\theta = \frac{V_w}{V_t} = \frac{M_w/\rho_w}{V_t} = \frac{125/1.00}{423.3} = \frac{125}{423.3} = 0.295$

where, ρ_w = Density of water

(vii) Water volume ratio or liquid ratio (using Eq. 11) $v_w = \frac{V_w}{V_s} = \frac{125}{207.6} = 0.602 \text{ or } 60.2\%$

(viii) Degree of saturation (using Eq. 12) $\theta_s = \frac{V_w}{V_f} = \frac{0.295}{0.509} = 0.58 \text{ or } 58\%$

(ix) Air filled porosity (using Eq. 13) $f_a = \frac{V_a}{V_t} = \frac{V_t - V_s - V_w}{V_t}$

$$= \frac{424.3 - 207.6 - 125.0}{424.3} = \frac{91.7}{424.3} = 0.216$$

Alternately, (using Eq. 13) $f_a = (f - \theta) = 0.510 - 0.295 = 0.215$

(x) Depth of soil water in 15 cm soil layer

Depth of soil water (W_d cm) = Volume wetness θ × profile depth (d)

i.e. $W_d = \theta . d = 0.295 \times 15 = 4.43$ cm

Question 2. If bulk density of soil is 1.5 Mg m^{-3}, find the weight of 0.10 metre depth of 1 ha of soil.

Solution:

Volume of 1 ha of soil up to 0.10 cm depth

$V_t = 10000 \times 0.1\ m^2 = 1000\ m^3$

So weight of soil ($M_s = V_t \cdot \rho_b$) = $1000 \times 1.5 = 1500$ Mg = 15×10^5 kg

Question 3. Derive the following useful interrelationships:

(a) Mass wetness (w) and volume wetness θ i.e. $\theta = \frac{w \rho_b}{\rho_w}$

(b) Porosity (f) and void ratio (e), $e = f/(1-f)$, $f = e/(1+e)$

(c) Degree of saturation (θ_s) and volume wetness (θ), $\theta_s = \theta/f$

(d) Airfilled porosity (f_a) and volume wetness (θ), $f_a = (f_s - \theta) = f(1 - \theta_s)$

Question 4. A soil 80 cm deep has a volume water content θ = 0.12. Find out the quantity of water that must be added to bring the volume water content to 0.30

(Ans: 14.4 cm)

Question 5. A soil core sample has a fresh weight of 1450 g and ovendry weight of 1200 g. The bulk density of soil is 1.2 Mg m^{-3}. Find out the volumetric water content of the soil sample

(Ans: 25.0%)

Question 6. A soil has a bulk density and particle density of 1.30 and 2.60 Mg m^{-3}, respectively. The volumetric water content is 30.0%. What is the air-filled porosity of the soil?

(Ans: 20.0%)

Question 7. The gravimetric water content in 0-15 cm soil layer in a wheat field is 15.5% before irrigation and 35.5% after irrigation. If the bulk density of the soil is 1.5 Mg m^{-3}, find out the amount of irrigation water stored in the 0-15 cm soil depth.

(Ans: 4.5 cm)

Question 8. Calculate the volume of water present in the root zone i.e. 1 m depth of 1 ha soil having mass water content 20% and bulk density 1.4 Mg m^{-3}.

(Ans: 2800 m^3)

Question 9(a). Using Stokes' law calculate the settling velocity of particles corresponding to the upper size limit of clay fraction (2 μm diameter) at 20°C.

Solution:

We assume particle density $\rho_s = 2.65 \times 10^3$ kg m^{-3}, fluid density ρ_f at 20°C = 1.00×10^3 kg m^{-3};

Viscosity η of fluid at 20°C = 1.00×10^{-3} pascal sec (Pa s) and g = 9.81 m sec^{-2}

N.B. (Note in this example the values of all the parameters are in SI units)

We use Eq. 5.14 and substitute the values of η, g and r

$$v = 2(\rho_s - \rho_f)\, gr^2/9\eta$$

$$= 2(2.65 \times 10^3 - 1.00 \times 10^3) \times 9.81 \times (1.00 \times 10^{-6})^2/9(1.00 \times 10^{-3})$$

$$= 3.30 \times 1.09 \times 10^{-6} = 3.6 \times 10^{-6}\ m\ sec^{-1}$$

Question 9(b). Also calculate the time required for collection of sample through pipette at 10 cm depth of the suspension for determination of clay content.

$$\text{Using Stokes' law, } v = \frac{h}{t} \text{ or } t = \frac{h}{v} = \frac{10\times10^{-2}}{3.6\times10^{-4}} = 7\,\text{hr}\ \ 43\,\text{min}$$

Question 10(a). Using Stokes' law, calculate the time required for all sand particles (diameter > 0.05 mm) to settle out of a depth of 15 cm in an aqueous suspension at 20°C (η = 0.01005 poise).

Here we use C.G.S. units;
h = 15 cm, η at 20°C = 0.01005 poise = 0.01005 g cm^{-1} sec^{-1}, ρ_s = 2.65 g cm^{-3},
ρ_f at 20 °C = 1. g cm^{-3}, g = 981 cm sec^{-1}, r = 0.0025 cm;
Use equation (16a) to calculate t

(Ans: 67 sec)

Question 10(b). Also calculate the time required for all silt (diameter > 0.02 mm) of the above sample to settle out of 15 cm.

(Ans: 1 hr 9 min 51 sec)

Question 11. The following data were obtained by wet sieving of 25 g of clod soil samples (initially retained on 5 mm sieve) by Yoder's wet sieving method for aggregate analysis. Calculate the mean weight diameter and geometric mean diameter of the water stable aggregates.

Sieve size class (mm)	5.0	5.0-2.0	2.0-1.0	1.0-0.5	0.5-0.25	0.25-0.10
Weight of aggregates retained in each size (g)	14.08	5.31	2.25	2.15	1.06	0.15

Solution:

Sieve size class (mm) (X_i)	Weight of aggregates retained (g) (W_i)	Mean diameter of each size fraction (mm) (x_i)	Fraction of total weight (w_i)	$x_i w_i$	$\log x_i$	$W_i \log x_i$
5.0	14.08	5.0	0.563	2.816	0.699	9.841
5.0-2.0	5.31	3.5	0.212	0.743	0.544	2.889
2.0-1.0	2.25	1.5	0.090	0.135	0.176	0.396
1.0-0.5	2.15	0.75	0.086	0.065	-0.125	-0.269
0.5-0.25	1.06	0.375	0.040	0.016	-0.426	-0.452
0.25-0.1	0.15	0.175	0.006	0.001	-0.757	-0.114
Summation	25.0			3.776		12.291

$$\text{Using Eq. (18), MWD} = \sum_{i=1}^{n} x_i w_i = 3.78\,\text{mm};$$

$$\text{Using Eq. (19), GMD} = \exp\left[\frac{12.291}{25}\right] = \exp(0.492) = 1.64$$

Question 12. Calculate the mean weight diameter and geometric mean diameter of water stable aggregates of a black clay soil. The sieve sizes and the weight of aggregates retained in each sieve after wet sieving of 25 g of clod soil sample are given below:

Sieve size class (mm)	5.0	5.0-2.0	2.0-1.0	1.0-0.5	0.5-0.25	0.25-0.10
Weight of retained aggregates (g)	19.10	1.39	0.25	0.125	0.062	0.025

(Ans: MWD = 4.03 mm; GMD = 1.76 mm)

Soil Water

S. KAR, M.C. OSWAL and USHA KIRAN CHOPRA

1. Introduction

Water is an essential constituent of the terrestrial ecosystem. In soil, water is held on the surfaces of its particles as adsorbed water and in the intervening voids as pore water. Soil water plays a significant role in controlling the energy balance of the soil and of its oyerlying micro-environment by modifying its radiation exchanges and thermal properties. Its content in variable amounts changes the volume of the air-filled pore space and regulates the gaseous exchange in the upper layer of the soil. When present in available form, soil water helps plants to carry out their basic physiological and metabolic processes, including nutrient absorption from soil. It dissolves salts and nutrient compounds and controls their transport processes in soil. Water is also important for various activities of the soil microorganisms leading to the decomposition of organic matter and fixation as well as release of soil nutrients. The presence of water in different amounts in soil governs its mechanical, physical and chemical properties. The soil properties like consistency, compactibility, strength, swelling and shrinkage are strongly influenced by its water content. In order to understand the availability of soil water as well as its role in controlling various soil properties and processes, it is essential to study the energy status, retention and movement of water in soil.

2. Hydrological Cycle

The cyclic process in which water moves from the earth's surface to the atmosphere and back to the earth is termed as the hydrological cycle. In this process, water is evaporated from the water bodies such as ponds, rivers and oceans, as well as from the wet surfaces of soils and plants. The water vapour migrates to the atmosphere where it gets condensed. The condensed water vapour falls as precipitation or snow on the soil surface. A portion of this precipitation runs off laterally over and below the soil surface and reaches the water bodies from where it is evaporated again. The hydrological cycle thus formed is repeated continuously. The remaining portion of precipitation enters the soil and is subjected to the processes of infiltration, redistribution, percolation, drainage and seepage. The net result of these flow processes is water storage or loss from the soil profile. The entire soil profile is categorized in three zones. The upper zone known as 'Soil Water Zone/ Root Zone' extends from the soil surface to the lower boundary of the root zone. Below the soil water zone, there is an unsaturated zone called 'Vadose Zone' which extends up to the upper boundary of the next zone which is called 'Capillary Fringe Zone' and is a near saturated zone occurring over the water table. The recharging of groundwater depends on the water flow through this unsaturated Vadose Zone.

3. Structure and Properties of Water

A molecule of water is formed from two atoms of hydrogen and one atom of oxygen by sharing of electrons in a tetrahedral charged structure (Figure 1). An excess of negative

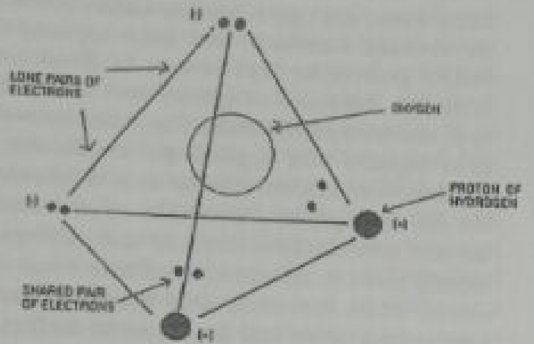


Figure 1. Tetrahedral charge structure of water molecule

charge prevails on one side of the molecule due to the unshared lone pair of electrons, while an excess positive charge prevails on the other side of the molecule due to the unscreened positive charge of the protons of hydrogen atoms, as shown in this figure. Such an arrangement makes a water molecule a dipole which helps it to link with the neighbouring molecules through H-bonding, which is a weak bond developed between hydrogen and an electronegative element like oxygen.

In the liquid state, a molecule of water is surrounded by five or more water molecules, while in the solid state, each molecule of water is linked to four other molecules in a hexagonal type arrangement. This makes the water lighter in solid form than in liquid state. Since water molecules are dipoles, liquid water has a high dielectric constant, whose value is about 80 for the pure water. Due to H-bonding, liquid water has a higher values of melting point, boiling point, dielectric constant, specific heat and viscosity as compared to other compounds of similar molecular weight. However, soil water, due to the presence of solutes, has a lower freezing point and higher boiling point than pure water. The depression in freezing point is sometimes used to characterize the wilting point of soil.

Soil water is a viscous fluid. It experiences resistance from the surface of soil particles while it moves through soil pores. The frictional force (F) is proportional to surface area (A) and velocity gradient ($\delta v/\delta x$), as given by Equation (1):

$$F = \eta A \frac{\partial v}{\partial x} \qquad \ldots(1)$$

If force (F) is expressed in dynes, surface area (A) in cm^2, velocity of water (v) in cm/sec, and thickness of water layer (x) in cm, then coefficient of viscosity (η) has the dimension of dyne/sec/cm^2 or poise. At 25 °C, the viscosity of water is 0.0089 poise.

4. Soil Water Content

The amount of water present in soil at a particular point of time is called soil water content. When it is expressed on mass basis, it is called water content by weight (θ_w) and is equal to the mass of water divided by the dry mass of the soil expressed as per cent. It is also expressed on volume basis as volume of water divided by the total volume of soil (θ_v) multiplied by 100. One form can be converted to the other form using the relation (2):

$$\theta_v = \theta_w \text{ (Bulk density of soil/ Density of water)} \qquad \ldots(2)$$

4.1. Soil Moisture Constants

There are many terms that are associated with the water that is retained by the soil. The common ones are saturation water content, field capacity, permanent wilting point and hygroscopic coefficient.

(1) ***Saturation Water Content*** (θ_s): Soil is a porous medium and when all the pores of the soil are filled with water, it is referred to as saturated soil. The saturation water content refers to the water content at which all the pore spaces, including the expanding space of clay minerals, are filled with water and the matric suction is zero. It can be determined with the help of tension plate assembly where the soil core/soil sample is brought in contact with water column at a height equal to the mid-point of the sample and the moisture held by soil in equilibrium with the water column is determined. Since in saturated condition all the pores are filled with water, the total porosity would be same as the volumetric water content. Therefore, it can also be calculated using the formula (3) for porosity:

$$\theta_s = P = [1 - BD/PD] \times 100 \qquad \ldots(3)$$

where, P is the porosity in per cent, BD is bulk density (g/cm^3) and PD is particle density (g/cm^3) of soil. PD is generally assumed to be 2.65 g/cm^3.

(2) ***Field Capacity (FC)***: When a saturated field soil is allowed to drain, soil moisture decreases exponentially with time. After 2 to 3 days, the moisture in the surface layer attains a steady state. At this stage, the water retained in the soil is known as 'Field Capacity'. Thus, field capacity is defined as the amount of water retained in the soil after the downward water movement from a pre-saturated soil has 'materially' ceased. The matric potential corresponding to this moisture content varies with soil texture from -1/10 bar for coarse-textured soils to -1/3 bar for medium- to fine-textured soils. It is, therefore, best estimated in the field by saturating the root zone and determining the soil moisture after free internal drainage ceases. It takes about 24 to 36 hours for coarse and 2-3 days for medium-textured soils in field.

(3) ***Permanent Wilting Point (PWP)***: It refers to the soil moisture content at which plant roots cannot extract water at a rate sufficient to meet the transpirational needs. At this moisture content, plants lose cell turgidity and show symptoms of wilting. The permanent wilting point (PWP) corresponds to the water content of soil when an indicator plant such as sunflower growing on it get wilted so severely that they do not regain their turgidity even when placed in water vapour-saturated atmosphere. The matric potential corresponding to this moisture content is -15 bar. Water retained by soil at -15 bar matric potential is generally estimated in the laboratory using either a pressure plate or a pressure membrane apparatus.

(4) ***Hygroscopic Coefficient***: It is the water held by the soil at -31 bar (3.1 MPa) soil moisture matric potential and is determined using a pressure membrane assembly meant for higher pressures.

4.2. Forms of Soil Water

In soil, the water is held in three forms, depending on the tenacity with which it is adsorbed: (i) Gravitaty or Free water, (ii) Capillary water, and (iii) Available water. The water between saturation and field capacity is called *gravity* or free water. It flows out under the influence of gravity and is considered unavailable to plants. The water between field capacity and hygroscopic coefficient is referred to as *capillary water* as it is bound to the soil by capillary forces. The water held by the soil between field capacity and permanent wilting point is termed as *available water* for plants, while the water held between permanent wilting point and hygroscopic coefficient is considered unavailable to plants.

Some plants are able to survive even at moisture contents below PWP for short periods. The availability of capillary water to plant roots depends on pore diameter which controls the pressure of water as measured by the height of water in capillaries. The narrower the capillary pore, the lower is the water pressure and lesser is its availability. With further increase in water content, when the macropores also get filled with water, it is more and more loosely held. This loosely held water has a suction ranging from 0.01 to 0.05 bar. Under the action of gravitational force, this loosely held water is liable to move downwards and hence it is known as *Gravity water* or free water. The drainage or deep percolation loss of water following irrigation results from the downward movement of the gravity water.

5. Energy Status of Soil Water

Knowledge of different forms of soil water content is not sufficient to characterize the state of water in soil. For example, the water content at which water is 100% available to plants in sandy loam soil, may impose permanent wilting to plants in clay soil. This is due to the higher energy with which water is retained in clay than in sandy loam soil. The energy status of water governs not only its availability but also its mobility in various transport processes occurring in soil. Information on energy status of water is thus essential to quantitatively describe the net force of retention and flow of water in soil.

5.1. Concept of Water Potential

The total energy of water in soil is the sum of potential energy as well as kinetic energy.

Since kinetic energy is very small in comparison to its potential energy, the potential energy is considered as the total energy of soil water, which is expressed by soil water potential. Potential energy represents the capacity to do work. According to the International Soil Science Society, total soil water potential is defined as the amount of work done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of water at a specified elevation at atmospheric pressure to the soil water at the given point. Pure free water has the maximum capacity to do work. Water in soil being held by adsorptive, osmotic and pressure gradient forces, has relatively lower capacity to do work. Rather, work needs to be done on water to move it from one position to another in soil against the force fields to which it is subjected. Therefore, the water potential is negative in soils. Extraction of water by plant roots is an example of work being done on soil water. If a unit quantity of water is moved against a force field (F) from the reference point (R) to a point in question (P), the total work done (W) is given by Equation (4):

$$w = \int_R^P F.ds.\cos\theta \qquad ...(4)$$

where, F is the force acting on water, ds is the infinitesimal distance along the flow path, and θ is the angle subtended by the flow path with the direction along which the force is acting.

5.2. *Components of Total Soil Water Potential*

The force fields acting on soil water are short-range adsorptive forces (matric forces), osmotic forces if salts are present, pressure gradient force, and the universal gravitational force. For each of the forces there is an associated potential. This means that the force of gravity will lead to gravitational potential (Ψ_g), hydrostatic and gas pressure gradient force to pressure potential (Ψ_p), osmotic force to osmotic potential (Ψ_{os}), and adsorptive force including air-water interfacial tension to adsorptive or matric or capillary potential (Ψ_m). The total potential (Ψ_t) of soil water at any point of equilibrium would be equal to the algebraic sum of all the component potentials (Scott, 2000):

$$\Psi_t = \Psi_g + \Psi_p + \Psi_m + \Psi_{os} \qquad ...(5)$$

Each of the component potentials may be defined, in principle, in terms of work done against the respective force field, as described by Equation (5). At macroscopic level, the total potential is often considered as the sum of gravitational and pressure potentials only, the latter incorporating the matric and pressure potential components.

5.3. *Units of Water Potential*

Water potential may be expressed on volume, mass or weight basis. If E is the potential energy of a volume V, mass m and density ρ of soil water, then different forms of water potential are as follows:

Potential energy per unit volume =

$$\frac{E}{V} = \frac{mgh}{\left(\frac{m}{\rho}\right)} = \rho gh$$

$$= \Psi \text{ (pressure unit)}$$

Potential energy per unit mass =

$$\frac{E}{m} = \frac{mgh}{(m)} = gh$$

Potential energy per unit weight =

$$\frac{E}{mg} = \frac{mgh}{mg} = h \text{ (head unit)}$$

5.4. *Hydraulic Head Concept*

In the study of soil water movement, the soil water potential is generally expressed on weight basis i.e., in head units. The hydraulic head (H) term used in the Darcy's equation consists of soil water pressure/suction head (h) and gravity head (z) at the point in question, as shown in Equation (6):

$$H = h + z \qquad ...(6)$$

The water pressure head at a particular soil depth can be measured by a piezometer in saturated soil and by a tensiometer in unsaturated soil, as shown in Figure 2. It is positive in

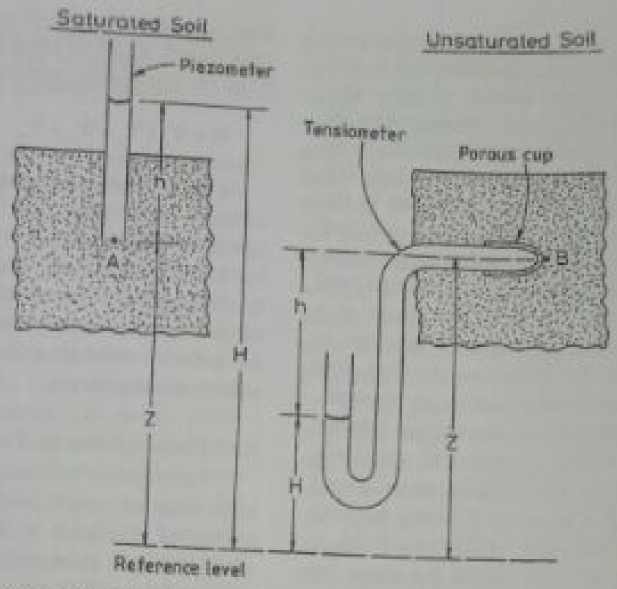


Figure 2. Concept of hydraulic head for saturated and unsaturated soils

saturated soil and negative (tension head) in unsaturated soil. The gravity head is estimated from the difference in elevation between the reference level, chosen arbitrarily at the soil surface, and the soil depth under consideration. It is positive above the reference level and negative below the reference level. The distribution of gravity, water pressure and total pressure heads in a 20-cm soil column which is immersed up to 5 cm in water, is illustrated in Figure 3.

6. Soil Water Availability to Plants

Veihmeyer and Hendrickson (1955) gave the concept of plant available water. The soil water storage capacity is generally expressed in terms of Available Water Capacity (AWC). The total available water in soil is regarded as the difference between soil moisture content of "field capacity" and "wilting point", i.e.

$$AWC = FC - PWP \quad ...(7)$$

It is customary to express available water capacity in terms of head dimensions so that it is analogous with irrigation, evaporation rate and rainfall units, which are generally expressed in terms of head units. The available water capacity of a specific layer can be expressed in 'cm' if bulk density values and depth of soil layer of the root zone are known as illustrated below:

$$D_w = [(FC - PWP) \times BD \times D_s]/100 \quad ...(8)$$

Table 1. Units, dimensions and common symbols of soil water potential

Form	Symbol	Common name	Dimensions	SI units	CGS units
Energy/volume	ψ	Soil water potential, suction or tension	$ML^{-1}T^{-2}$	$N\ m^{-2}$ (Pa*)	erg cm^{-3}
Energy/mass	μ	Chemical potential	L^2T^{-2}	$J\ kg^{-1}$	erg g^{-1}
Energy/ weight	h	Pressure head	L	m	cm

*Pa = Pascal = 10^{-5} bar

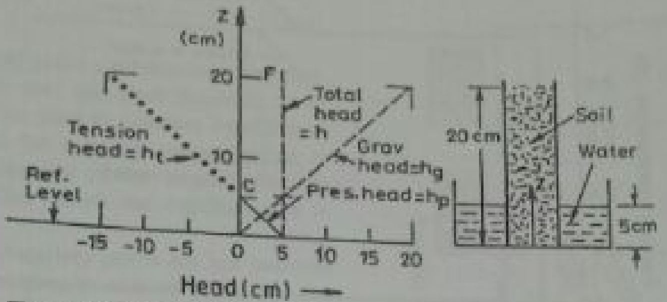


Figure 3. Gravity, tension and total pressure head distribution in soil

where,

D_w = Depth of water contained in a specific soil layer (cm)

D_s = Thickness of that soil layer (cm)

BD = Bulk density of the particular soil layer (g cm^{-3})

FC = Soil water content at Field Capacity (% by wt), and

PWP = Soil water content at Permanent Wilting Point (% by wt).

The profile water available capacity (D_p) is computed by first determining D_w for each soil layer and then adding it for all the soil layers.

$$D_p = \Sigma D_w \qquad ...(9)$$

However, the values of BD, FC and PWP vary from soil to soil and location to location, even under the same major soil group for the simple reason that there will be a variation in soil depth and available water capacity limits also. Therefore, as already emphasized earlier, the actual determination has to be made for the specific soil under consideration.

A school of thought led by Veihmeyer and Hendrickson (1955) maintain that soil moisture is available for plant growth equally over the range between field capacity and permanent wilting point. According to this view, transpiration/plant growth is unaffected by the magnitude of soil moisture content unless PWP is reached when the water suddenly becomes non-available. Another view point states that the water held by the soil becomes progressively less available with a linear decreasing trend as the water content decreases from FC to PWP (Thornthwaite and Mather, 1955). Yet another school of thought maintains that transpiration/plant growth is unaffected by the magnitude of soil moisture content from field capacity up to certain threshold value below which almost linear decrease in rate of transpiration occurs (Ritchie *et al.*, 1972). The magnitude of the threshold value would, however, depend upon a variety of factors related to soil, plant and atmospheric conditions. In general the threshold potential is -1.0 bar for a coarser soil and -5 bar for a finer soil. This approach is the one generally subscribed to by most workers for its practicality. All the three approaches have been depicted in Figure 4.

7. Water Retention by Soils

In an unsaturated soil, water is retained by the forces of capillarity and adsorption. The adsorption of water occurs through the hydration of dry soil matrix surface and exchangeable cations. With further intake of water, it gets into the soil pores where surface tension and radius of curvature of the air-water interface determine its content. The capillary concept is, therefore, often utilized to describe water retention in soil pores. In coarser soils, capillary action is more important than adsorption, while the reverse is true in the finer soils.

7.1. Capillarity and Water Retention

The water rises in a capillary tube when it is partially immersed in water. If the tube is made of glass or quartz, water makes contact with the tube at an angle close to zero, and forms a concave air-water meniscus (Figure 5a). Since the pressure at the water

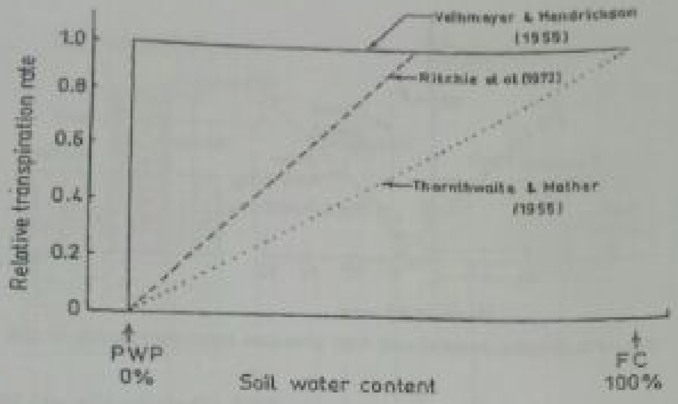


Figure 4. Concepts of soil water availability to plant

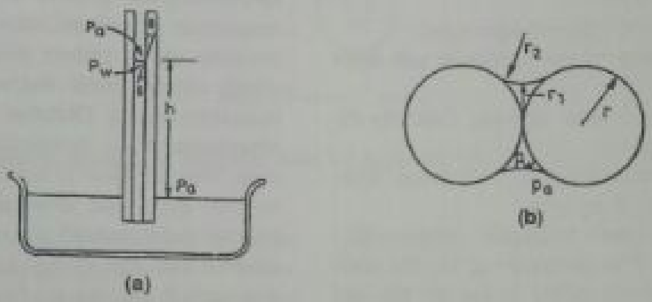


Figure 5. (a) Capillary rise of water, and (b) Principal radii of curvature at air-water interface in soil

surface outside tube is the atmospheric pressure, water rises in the tube till the hydrostatic pressure of water column equals the pressure difference across the air-water meniscus. The force due to the hydrostatic pressure of the rise in capillary water is $\pi r^2 h \rho_w g$, where, r is the radius of the capillary tube, h is the height of water in capillary, ρ_w is the density of water and g is the acceleration due to gravity. This downward force is balanced by the upward force due to surface tension which is estimated as $2\pi r \gamma \cos\theta$, where, $2\pi r$ is the circumference of the capillary tube, γ is the surface tension and θ is the contact angle between water and glass tube.

At equilibrium, the downward force will be balanced by the upward force as represented in Equation (10):

$$\pi r^2 h \rho_w g = 2\pi r \gamma \cos\theta \qquad \text{...(10)}$$

From this, water height (h) can be determined by Equation (11):

$$h = \frac{2\gamma \cos\theta}{r \rho_w g} \qquad \text{...(11)}$$

For water and glass interface, θ being close to zero, cosθ may be taken as equal to unity and Equation (11) simplifies to Equation (12):

$$h = \frac{2\gamma}{r \rho_w g} \qquad \text{...(12)}$$

or

$$h = \frac{4\gamma}{d \rho_w g} \qquad \text{...(13)}$$

where, d is the diameter of capillary tube.

Considering $\theta = 0$, $\gamma = 72.75 \times 10^{-3}$ N/m, $\rho_w = 0.998$ Mg/m³ and $g = 9.81$ m/s² at 20 °C, we get the simplified Equation (14):

$$h = \frac{0.279}{d} = \frac{0.3}{d} \qquad ...(14)$$

Equation (14) gives a relation between capillary rise (h in cm) and pore diameter (d in cm) which can be used to estimate capillary or matric potential for a known pore size, or to calculate the diameter of the largest effective pore of a soil for which the capillary rise is known.

In soil, the capillary pores are not uniform in size. The variation in size of pores and the degree of saturation lead to air-water interfaces of varying radii of curvature, which also affect the matric potential by altering the pressure difference across the air-water interface. The relation of the curvature of an air-water interface to the pressure difference across the interface is given by the Laplace equation [Equation (15)]:

$$P_w - P_a = 2\gamma\left(\frac{1}{r_1}+\frac{1}{r_2}\right) = \frac{2\gamma}{r} \qquad ...(15)$$

where, P_w and P_a are the pressure of water below and above air-water meniscus, respectively and r_1 and r_2 are the principal radii of curvature (Figure 5b) formed above and below the air-water meniscus. The mean radius of curvature r is positive if it lies within the water phase and negative if it lies in the gas phase.

7.2. Soil Moisture Characteristics

The retention and release of water in soil are the functions of its energy status. When a field soil is saturated with irrigation or precipitation, its water potential becomes zero. Subsequently, on cessation of water application, soil water depletes due to drainage and evaporation, at a faster rate initially from the non-capillary pores where the potential of water is relatively high and then at a slow rate from capillary pores where water potential is relatively low. In other words, water retention in a soil is a function of its matric potential or suction. A soil moisture characteristic (SMC) curve describes the functional relationship between soil water content and its energy status in terms of its matric potential under equilibrium conditions. The SMC is an important property of a given soil controlled by soil pore size distribution which is strongly affected by texture, structure and organic matter content.

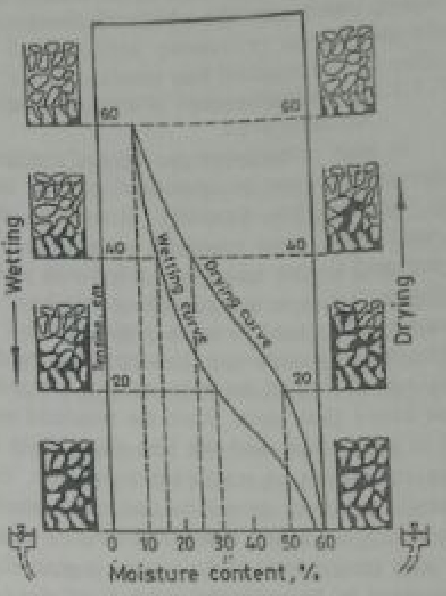


Figure 6. Drying and wetting curves of soil illustrating the effect of hysteresis (redrawn from Ghildyal and Tripathi, 1987)

7.2.1. *Soil Water Hysteresis*

It is important to note that the SMC relationship is not unique but depends on the sorption or desorption pathways of soil water. When SMC is obtained by taking an initially saturated soil which loses its moisture to become dry, desorption pathway is followed and a unique SMC curve is obtained. However, when a dry soil is gradually wetted and its suction-water content relationship is obtained, sorption pathway is followed which may be a different curve from the desorption curve with the coinciding end points (Figure 6). At a particular suction, the soil water content during drying is generally higher than during wetting. This phenomenon is referred to as 'Hysteresis'. The hystersis in SMC occurs due to various factors: (1) the ink bottle effect resulting from non-uniformity in shape and size of interconnected pores, (2) different contact angles for advancing and receding water menisci, (3) entrapped air during adsorption, and (4) swelling and shrinking of soils under wetting and drying, which alter the pore size distribution and porosity.

7.2.2 Determination of Water Retention Characteristics

In high potential range, the soil moisture characteristics can be determined with the help of 'Tension Plate Assembly'. It consists of a fritted glass plate which is connected to a graduated U-type manometer filled with water (Figure 7). Let us now consider a thin layer of air-dry soil placed over the plate. The soil is saturated by maintaining the water level on the right hand limb of the manometer at par with the fritted glass plate. Since the saturated soil is in equilibrium with the free water level in the manometer at atmospheric pressure, the matric suction or tension on soil water would be zero. Considering this initial level of water as zero suction level, if the suction is gradually increased by lowering the water level slowly, there would be no outflow from soil water system till a critical suction is reached at which point the largest soil pore would begin to drain. This critical suction is known as '*air entry suction*'. As suction is further increased, more water will drain out of the soil pores having smaller radii. The radius of pore drained at a given suction can be estimated by Equation (12) or (14). If an increase in the suction from h_1 to h_2 results in a release of V volume of water, then V must be the pore volume having the largest and smallest effective pore radii of r_1 and r_2, respectively. Thus, the moisture retention characteristics of a soil can be utilized to calculate the pore volumes of different size ranges.

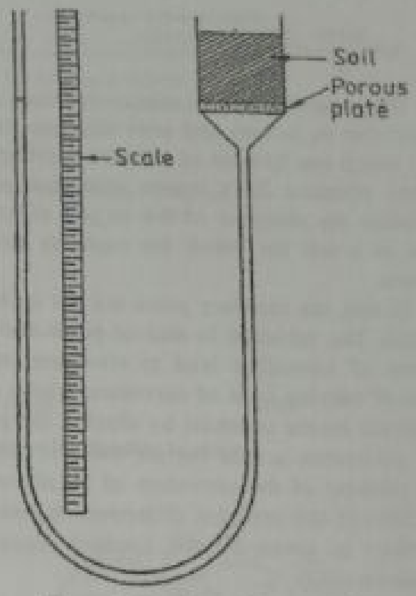


Figure 7. Tension plate assembly

In the broad range of suction from 0 to 15 bar, the soil moisture characteristics can be determined with the help of 'Pressure plate apparatus'. It consists of a ceramic plate cell on which the soil cores are saturated, and a pressure chamber in which positive pressure is applied by an air compressor to desorb water from initially saturated soil. The significance of moisture characteristics, as depicted by curves in Figure 6, is best visualized by the variation in slope of the curve with changes in suction or moisture content. The inverse of the slope of the water retention characteristic curve is referred to as 'Differential Water Capacity', C_θ ($c_\theta = -\frac{\partial\theta}{\partial\psi}$) which is a function of water content. A vertical trend indicating a steep slope of the curve, represents a soil containing large number of pores of the same size range, while a horizontal trend with little slope reveals that the soil has almost no drainable pore in the suction range under consideration. The trend with intermediate slope represents soil with uniform distribution of pores which results in uniform release of water per unit change of suction.

7.2.3. Factors Affecting Soil Water Retention Characteristics

In the high potential range (low suction), water retention is largely controlled by the capillary effect as a function of pore size distribution of soil. Since the size distribution of pores is a consequence of arrangement of particles, water retention at relatively high potentials is primarily controlled by soil structure and related properties. On the other hand, water retention in low potential range (< –1 bar) depends primarily on adsorption of water as a function of surface properties of soil matrix.

Hence, soil texture is more important than soil structure in controlling water retention at low potentials.

Water retention is higher in finer (clayey soils) than coarser (sandy soils) soils, irrespective of water potential (Figure 8). The release of water is, however, gradual in a finer soil and relatively rapid in a coarser soil. Among the mechanical separates, clay and silt fractions are strongly related with water retention. With increase in clay content, water retention increases at all potentials. Water retention is favoured by organic matter content, particularly at high potentials, but at low potentials, organic matter has little influence on water retention. The water retentivity of the soil is also affected by the nature of clay minerals. Black clay soils having smectitic clay minerals retain more water than the kaolinitic lateritic soils (Biswas and Mukherjee, 1990). The presence of soluble salts in soil solution also influences retention and release of water in soil. The diurnal and seasonal variations of soil temperature also affect the water retention of a soil.

8. Measurement of Soil Water

Soil moisture is measured directly by the gravimetric method and indirectly by several improved techniques based on the determination of electrical resistance, slowing down of fast neutrons, gamma-ray attenuation, electrical capacitance and backscattering coefficient as a function of water content.

8.1. Gravimetric Method

In the thermogravimetric method, soil is sampled from the field location, weighed immediately to determine fresh weight and dried in an oven to a constant weight at 105 °C for 24 hours to estimate the oven-dry weight. The per cent soil water on weight basis is estimated by dividing the water content by oven-dry weight of soil and multiplied by 100.

8.2. Electrical Resistance Method

One of the non-destructive, indirect and *in-situ* measurement methods of soil water content is the widely used electrical resistance method, using gypsum or nylon or fibre glass blocks. In this method, two platinum electrodes are embedded in a gypsum block which is placed in soil and allowed to equilibrate. The electrical resistance of the porous block is then measured using a Wheatstone bridge. The electrical resistance of the soil decreases with increase in water content. A calibration curve

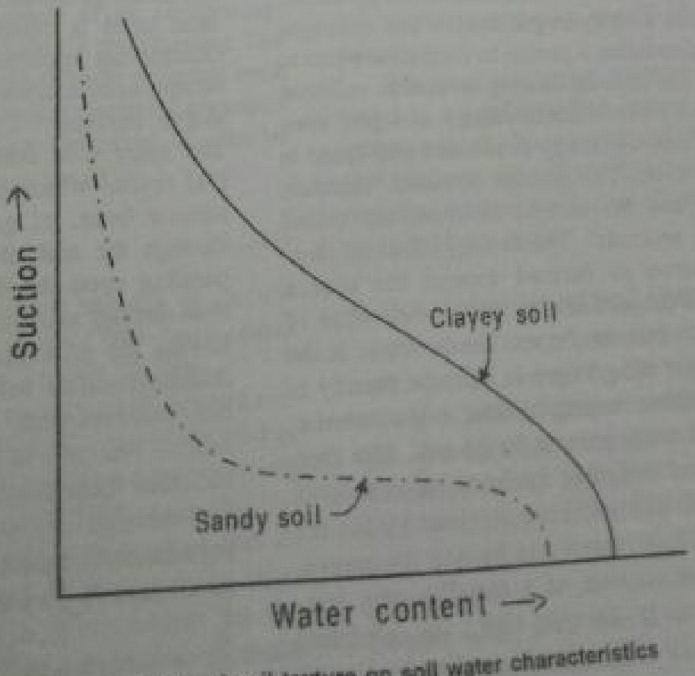


Figure 8. Effect of soil texture on soil water characteristics

between the resistance and water content of the block is determined. The soil water content is obtained with the help of this calibration curve, from the measured electrical resistance of the block.

8.3. Nuclear Techniques

The neutron scattering and gamma-ray attenuation are two nuclear techniques that can be used for non-destructive *in-situ* measurement of soil water content.

8.3.1. Neutron Moisture Meter

In the neutron scattering technique, the instrument used is called Neutron Moisture Meter. It consists of Amerecium and Beryllium or Radium and Beryllium as the neutron source. In case of Amerecium and Beryllium, fast neutrons are produced from the following reactions:

$$\underset{\text{Americium}}{{}^{241}_{95}\text{Am}} \longrightarrow \underset{\text{Neptunium}}{{}^{237}_{93}\text{Np}} + \underset{\text{Alpha particle}}{{}^{4}_{2}\text{He}} \qquad \text{...(16)}$$

$$\underset{\text{Beryllium}}{{}^{9}_{4}\text{Be}} + \underset{\text{Alpha particle}}{{}^{4}_{2}\text{He}} \longrightarrow \underset{\text{Carbon}}{{}^{12}_{6}\text{C}} + \underset{\text{Fast neutron}}{{}^{1}_{0}n} \qquad \text{...(17)}$$

The instrument works on the principle that when fast moving neutrons are emitted from the source into the soil, they undergo elastic collisions with hydrogen nuclei. The hydrogen nucleus contains a proton having mass equal to that of the neutron causing an elastic collision in which neutron loses energy and gets scattered. Thus, its energy is reduced and speed is slowed down. The process is called 'thermalization' and the slowed neutrons are called 'thermal neutrons'. The density cloud of thermal neutron so formed around the source (probe) is proportional to the concentration of hydrogen nuclei in the soil. Since water is the main source of hydrogen in soil, the density of slowed neutron is proportional to the volume-fraction of water present in the soil. The thermalized neutrons (fast neutrons slowed down by repeated collisions) are captured by the detector, which is placed just behind the source. The detector consists of a metallic tube containing boron tri-fluoride (BF_3) gas enriched with Boron-10 (^{10}B). The thermalized neutrons react with the boron producing alpha particles according to the following reaction.

$$\underset{\text{Boron}}{{}^{10}_{5}\text{B}} + \underset{\substack{\text{Thermal}\\\text{neutron}}}{{}^{1}_{0}n} \longrightarrow \underset{\text{Lithium}}{{}^{7}_{3}\text{Li}} + \underset{\text{Alpha particle}}{{}^{4}_{2}\text{He}} \qquad \text{...(18)}$$

The detector is able to register thermal neutrons only. The alpha particles thus produced in the detector cause ionization which generates electrical pulses. These pulses are amplified by a pre-amplifier housed in the probe itself and then transmitted to the scalar. The scalar counts the number of pulses (or neutrons) in a pre-set time interval and these counts are converted to moisture content with the help of a calibration curve. The neutron moisture meter has to be calibrated for each soil. The advantages of this method are : it is rapid, less laborious, non-destructive and repeated measurements can be made at the same depth. The probe has to be lowered into the soil through an aluminum access tube. The main disadvantages are that it cannot be used for surface soil moisture content measurement since neutrons escape into the atmosphere and secondly, it is costly.

8.3.2.Gamma Ray Attenuation Method

In the gamma-ray attenuation technique of soil moisture content measurement, the instrument used is called Gamma-ray scanner or Gamma-ray apparatus. It consists of two aluminum access tubes and in one of these, radioactive isotope ^{137}Cs (Caesium-137) is used. The other tube consists of thallium-activated NaI crystal which acts as a detector. When a narrow beam of gamma-radiation is passed through the soil, the rays are attenuated depending upon the thickness of soil mass, the bulk density and water content of the soil, following the principle of Beer's law. So if the thickness and the bulk density of soil are known, the water content of the soil can be determined. The ratio of the monochromatic gamma radiation flux passing through wet soil (N_w) and through dry soil (N_d) of thickness x is exponentially related to the volumetric soil water content (θ_w) by the following relation (19):

$$\frac{N_w}{N_d} = \exp(-\theta_w \mu_w x) \qquad \text{...(19)}$$

where μ_w is the mass attenuation coefficient for water. This relation provides measurement of soil water content θ_m from Equation (20):

$$\theta_m = -\frac{l_n(N_m/N_d)}{\mu_w x} \quad \ldots(20)$$

The gamma-ray apparatus is most useful in laboratory studies where moisture content of very small depth increments of soil can be determined. However, practical difficulties limit its use in the field.

A number of isotopes are used for soil water movement studies. The isotope tritium (^3H) is largely used for water movement, groundwater recharge and pollution studies. In hydrological studies and seepage, stable isotopes deuterium (^2H) and oxygen (^{18}O) are used.

8.4. Time Domain Reflectometry

A recent development in soil moisture measurement is the Time Domain Reflectometry (TDR). It is based on the estimation of dielectric constant (ε) of soil as varied by its water content. Water has a dielectric constant of 80. The apparent dielectric constant (K_a) of a soil varies from 3 to 6 in the dry state and from 20 to 30 in the wet state. The K_a measured in conjunction with a conversion table, can help estimate the volumetric moisture content. K_a is estimated by measuring the time (t) taken for the propagation of an electromagnetic pulse launched along a transmission line or wave guide of length (L) and reflected back to the origin. Based on the electrical transmission line theory of high frequency, K_a can be estimated from the Equation (21):

$$K_a = \left[\frac{ct^2}{2L}\right] \quad \ldots(21)$$

where, c is the propagation velocity of electromagnetic wave in vacuum or free space (equal to the velocity of light, i.e. 3 × 10^8 m/s).

The wave guide, consisting of two parallel electrode rods spaced at 50 mm, is inserted in the soil. The recently developed microprocessor based and calibrated TDR units directly give the volumetric moisture content and dielectric constant of soil. TDR is gaining more and more acceptance as it overcomes the disadvantages of other time-consuming methods. The technique is also useful in measuring soil salinity in terms of electrical conductivity. TDR gives an average soil moisture content of the soil depth equal to the length of the wave guide.

8.5. Microwave Remote Sensing of Soil Moisture

All the above methods give point measurement of soil moisture content. The recently developed microwave remote sensing technique enables us to estimate soil moisture content over a region or large area. The basic principle is that the microwave backscattering coefficient is greatly influenced by the dielectric property of the soil which is mostly a function of soil water content. The dielectric constant is less than 5 for dry soil, whereas it is 80 for water. As the soil water content increases, the dielectric constant of soil also increases which decreases the emissivity (e) of the wet soil. The decrease in emissivity is approximately linear with increase in soil water content. The back scatter coefficient (σ) is a function of emissivity and is given by Equation (22):

$$\sigma = f\,(e, \lambda, \phi, \rho, \alpha, R, V) \quad \ldots(22)$$

where, e is the emissivity, λ is the wavelength of incident radiation, ϕ is the incident angle, ρ is the polarization, α is the aspect angle, R is the surface roughness and V is the volume scatering.

The back-scattering coefficient (σ) measured in a satellite-based sensor synthetic aperture radar (SAR) is correlated with soil moisture content. However, this technique gives moisture content estimation of only surface soil up to 5-10 cm depth.

9. Determination of Soil Water Potential

Total soil water potential at a point can be estimated by measuring its component potentials: gravitational, pressure, matric and osmotic. The gravitational potential is estimated from the difference in elevation between the reference level and the level of the point in question. The pressure potential is generally considered as hydrostatic pressure in saturated or flooded soil. It is measured by the height of water in the Piezometer, sunk up to the soil depth in question. The Piezometer, a metallic tube with perforated bottom, is subtended to the soil depth. A water level recorder is used to determine the height of water in the piezometer. In an unsaturated soil, the matric potential forms the major component of the total soil water potential. It is evaluated with the help of a tensiometer up to –0.8 bar and below this, the matric potential can be estimated along with osmotic potential by soil psychrometer.

9.1. *Tensiometer*

The tensiometer consists of a porous cup made of ceramic material and a mercury manometer attached to the water-filled cup through a water reservoir tube. The porous cup has high conductance, low response time and air entry pressure of about 1 bar. Following saturation when the porous cup is placed in soil and equilibrated, water tends to move out of the cup under the suction exerted by soil. As a result, a vacuum pressure develops in the cup dialyzate and to make up this, mercury rises in the manometric tube attached to the water reservoir tube. The vacuum in the porous cup is actually the matric potential (Ψ_m) of soil water. The soil matric potential (Ψ_m) is determined from the relation (23):

$$\Psi_m = -12.6\, Z_{hg} + Z_c \qquad \ldots(23)$$

where Z_{hg} is the height of mercury in manometer and Z_c is the distance between the mercury surface in cup and centre of the ceramic cup.

The use of tensiometer is confined up to the minimum matric potential of –0.8 bar, but much of the available water is retained by soil between –0.8 bar and –15 bar, particularly in fine-textured soils. Beyond the tensiometric range, the matric potential is usually interpolated from the water retention characteristics curve using the measured soil moisture content.

9.2. *Soil Psychrometer*

The total soil water potential is determined with the help of a soil psychrometer. As liquid water in soil is always in equilibrium with its vapour in the ambient air, the energy status of water vapour (vapour potential) must be equivalent to the energy status of soil water. At constant temperature and gravitational potential, the vapour potential is largely contributed by matric and osmotic potentials. The vapour potential of soil can be precisely estimated with the help of a Thermocouple Psychrometer.

The soil water potential (ϕ) is related to the vapour pressure of soil water (p) and vapour pressure (p_o) of pure and free water at the same temperature and air pressure by the relation (24):

$$\phi = RT \ln (p/p_o) \qquad \ldots(24)$$

where, R is the universal gas constant and T is absolute temperature. This relation is used to determine soil water potential by a thermocouple psychrometer.

10. Movement of Soil Water

10.1. *Principles of Soil Water Movement*

Water is a highly dynamic component in soil system. It moves in all the three phases: solid, liquid and vapour. In a flooded or saturated soil, water moves in liquid phase, while in a partially dry or unsaturated soil, it moves in both liquid and vapour phases. Movement in solid phase as commonly occurring in the frozen soil is also believed to take place close to the clay surface. In liquid and vapour phases, water movement is governed by the principles of fluid flow.

10.1.1. *Bernoulli's Equation*

In a flooded soil when water moves under hydrostatic pressure difference through a uniform horizontal pore, the energy state of water at any point is characterized by its velocity (v), mass density (ρ_w), pressure (P) and height above the reference level (Z). At any point on the steady state flow path, the energy state of water, either potential or kinetic, remains constant. In other words, the sum of pressure head ($h_p = P/\rho_w g$), potential head ($mgZ/mg = Z$) and velocity head [$(1/2\ mv^2)/mg = v^2/2g$] remains constant for a non-viscous fluid under steady state flow. This relation is known as

Bernoulli's equation and is represented by Equation (25):

$$Z+\frac{v^2}{2g}+\frac{P}{\rho_w g}=\text{Constant} \quad \text{...(25)}$$

10.1.2. *Poiseuille's Law*

Poiseuille's law is used to describe the flow of a fluid showing viscous properties. It states that flow of water through a narrow tube is directly proportional to the fourth power of its radius and pressure difference, and inversely proportional to the viscosity and length of tube, it is given by the Equation (26):

$$q=\frac{\pi r^4}{8\eta}\left(\frac{\Delta p}{L}\right) \quad \text{...(26)}$$

where, Δp is the pressure difference in dynes cm^{-2}, r is the radius of tube in cm, L is the length of tube in cm, and h is the coefficient of viscosity of the liquid in dyne-sec cm^{-2} (poise). This relation is true only when the flow is steady or streamline and pressure is constant along the cross-section of the tube.

10.1.3. *Reynolds Number*

Laminar flow, sometimes known as streamline flow, occurs when a fluid flows in parallel layers, with no disruption between the layers. The Reynolds number is an important parameter that describes whether flow conditions lead to laminar or turbulent flow. The ratio of inertial forces of flow to frictional forces is called Reynolds Number (R) which is a dimensionless number given by Equation (27):

$$R=\frac{\rho_w v d}{\eta} \quad \text{...(27)}$$

where, v is the flow velocity, ρ_w is the density of water, η is the coefficient of viscosity of water and d is the pore diameter. In the case of flow through a straight pipe with a circular cross-section, Reynolds Numbers of less than 2300 are generally considered to be of laminar type; however, the Reynolds Number upon which laminar flows become turbulent is dependent on the flow geometry.

It has been experimentally observed that when water flows through a straight tube of diameter d, the flow remains streamline when R ranges between 1000 and 2000 and the critical velocity v_c, which is the maximum velocity up to which the flow is streamline, corresponds to a highest value of R called, the critical Reynolds number, R_c

$$R_c=2000=\frac{\rho_w v_c d}{\eta} \quad \text{...(28)}$$

or

$$v_c=\frac{2000\eta}{\rho_w d} \quad \text{...(29)}$$

When flow velocity ρ_w is less than v_c, the flow remains streamline.

When the tube is either curved or its diameter varies with its length, the value of R_c reduces. In a porous medium such as soil, due to tortuous structure of pore space, deviations from Equation (28) occur at much smaller values of R_c which is less than one ($R_c<1$). Thus, for a water flow through a soil, the critical velocity for streamline flow is given by the relation (30):

$$R_c=\frac{\rho_w v_c d}{\eta}=1 \quad \text{...(30)}$$

or,

$$v_c=\frac{\eta}{\rho_w}\cdot\frac{1}{d} \quad \text{...(31)}$$

For water, we can approximate

$$\frac{\eta}{\rho_w}=10^{-6}\frac{m^2}{s} \rightarrow v_c=10^{-6}\frac{1}{d}\frac{m^2}{s} \quad \text{...(32)}$$

This gives the effect of pore diameter on the critical velocity of streamline flow. For example:

(a) When pore diameter d = 1 mm = 10^{-3} m

$$v_c=10^{-6}\times\frac{1}{10^{-3}}\frac{m^2}{s}\frac{1}{m}=10^{-3}\frac{m}{s}=1\frac{mm}{s}$$

That is flow will be laminar as long as the flow velocity is less than 1 mm/s

(b) For smaller pores of diameter = 10 µm = 10^{-5} m

$$v_c=\frac{10^{-6}}{10^{-5}}\frac{m}{s}=10^{-1}\frac{m}{s}=10\frac{cm}{s}$$

This indicates that the flow will be laminar for velocities less than 10 cm/s. Thus, as the pore diameter decreases, the flow remains streamline for higher flow velocity.

10.2. *Types of Flow*

Soil water may be in static equilibrium or dynamic equilibrium state.

10.2.1. Static Equilibrium State of Soil Water

It occurs when mass and energy fluxes in the soil are zero, which implies that the total soil water potential remains constant and does not change with different positions or timings. Therefore, no flow of water occurs under static equilibrium. However, natural soils are never in the static equilibrium state as the water fluxes occur due to rainfall, evaporation and water uptake by roots.

10.2.2. Dynamic Equilibrium State of Soil Water Flow

The state of dynamic equilibrium can occur either as a steady state or stationary flow or transient or non-stationary flow of soil. The steady state flow of soil water system is that where water flux at a given point is constant with time. This means all temporal changes are negligible. If ϕ (t,x,y) is a time (t) and space (x,y) variable representing the state of the soil water (called state variable), then for a steady state or stationary system, we have

$$\frac{d\phi}{dt}=0$$

The water flow in a saturated soil exhibits a steady state flow.

The transient or non-stationary flow of soil water is that where water flux changes with space (position) and time. This implies

$$\frac{d\phi}{dx}\neq 0 \text{ and } \frac{d\phi}{dt}\neq 0$$

The unsaturated flow, in general, is transient flow. The behaviour of soil water flow is different under saturated and unsaturated soils.

10.3.1. Saturated Flow

From the above description it is clear that Poiseuille's law that describes steady state water flow through a tube of uniform radius cannot estimate non-steady state water flow through a large volume of soil containing pores of varying radius and length. In such a heterogeneous system, the flow of water can be reasonably estimated only when the flow velocity is considered as a function of macroscopic water pressure gradient of the soil volume.

Let us consider some laboratory cases of water flow through homogeneous soil. Figure 9(a) shows a saturated horizontal soil column of length x and cross-sectional area A being subjected to h units of pressure head at the inflow end. The outflow end is open to the atmosphere. Considering the reference level z units below the axis of the column, the hydraulic heads at the inflow and outflow ends are h+z and z, respectively. The pressure difference between the ends is h. Thus, gravity head

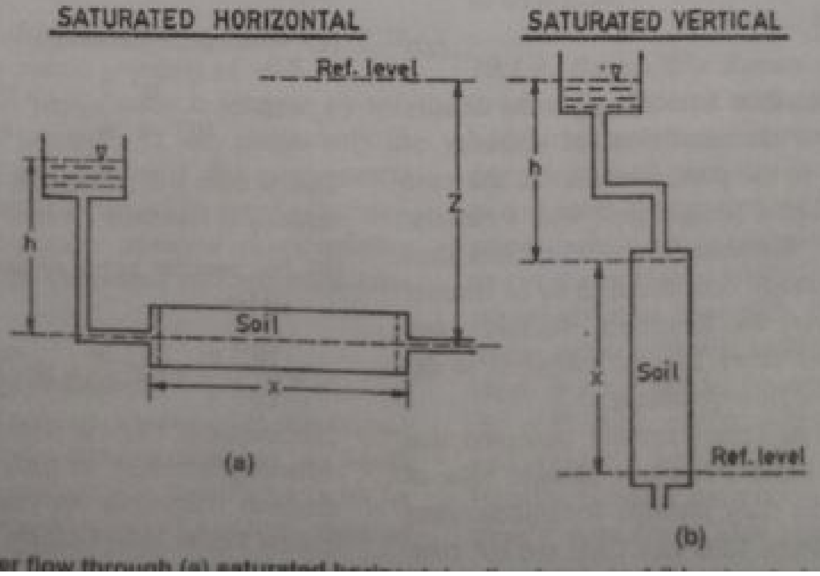


Figure 9. Water flow through (a) saturated

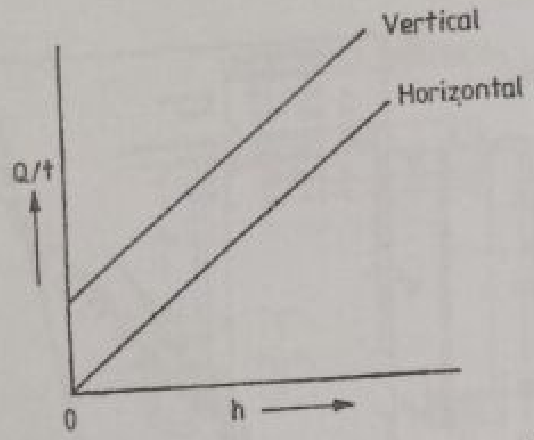


Figure 10. Relationship between water flux and pressure head for saturated orizontal and vertical soil columns

gets nullified in the case of horizontal flow. Water flows under the hydraulic gradient of h/x and the volume of water flowing per unit time (Q/t, water flux) can be measured by collecting the outflow in a measuring cylinder. Now, if we increase the inflow pressure head to 2h and 3h, we will see that outflow increases to 2Q and 3Q, respectively and when it is reduced to 0, the outflow decreases to 0. Similarly, when the cross-sectional area is increased from A to 2A and 3A, the outflow increases to 2Q and 3Q, respectively. In contrast, when the length of the column is increased to 2x and 3x, the outflow decreases to (1/2)Q and (1/3)Q, respectively. Based on these experimental results, we may conclude that Q/t is directly proportional to h and A and inversely proportional to x, i.e.

$$\frac{Q}{t} \propto \frac{Ah}{x} \qquad \text{...(33)}$$

Now, if we keep A and x constant and vary h, we will find that Q/t has a 1:1 linear relationship with h (Figure 10). The ratio of hydraulic head to the length of soil column (h/x) is known as 'hydraulic gradient'. The slope of the straight line plot between (Q/At) and h/x is the proportionality constant which is termed as 'saturated hydraulic conductivity' (K) and defined as the velocity (Q/At) of water per unit hydraulic gradient. K is a function of pore size distribution of soil and the properties of flowing water such as density and viscosity. So, the mathematical relationship between Q/t and Ah/x for saturated horizontal flow may be expressed by Equation (34) as:

$$\frac{\theta}{t} = -\frac{KAh}{x} \qquad \text{...(34)}$$

The negative sign on the right hand side of Equation (34) indicates that the flow always takes place in the decreasing direction of hydraulic head, implying that the change in head is decreasing with increasing length x.

For saturated vertical soil column (Figure 9b) with the reference level taken at the outflow end, we find that the outflow is not zero even at zero water pressure head. When the water level in the reservoir is brought at par with the inflow end, the hydraulic head becomes equal to x at that end. Since the reference level is considered at the outflow end which is open to the atmosphere, both gravity and water pressure heads become equal to zero at the outflow end. From this, it is clear that the vertical flow in saturated soil is strongly influenced by the gravity. For h water pressure head, as shown in Figure 9b, the flow equation may be given by Equation (35):

$$\frac{Q}{t} = -KA\frac{h+x}{x} \qquad \text{...(35)}$$

where, Q/t is the water flux, A is the cross-sectional area of the column; (h+x) is the hydraulic head difference between inflow and outflow ends, x is the length of the soil column, and K is the saturated hydraulic conductivity. The ratio of hydraulic head difference to length of soil column is known as 'hydraulic gradient'.

The equations thus developed are known as Darcy's Law which states that water flux or flow velocity is proportional to hydraulic gradient and the flow occurs in the direction of decreasing water pressure.

If Q/t is measured for varying h and plotted against (h + x) / x, a straight line relationship will be obtained, but unlike horizontal flow, it will not be 1:1. The Q/t will not be zero even at h=0. The above equations for saturated flow may be generalized in terms of velocity of water (v) and hydraulic gradient (*i*) by the Equations (36) and (37) for horizontal and vertical flows, respectively:

Horizontal flow:

$$\frac{Q}{At} = v = -K\left(\frac{h}{x}\right) = -Ki \quad ...(36)$$

Vertical flow:

$$\frac{Q}{At} = v = -K\left(\frac{h+x}{x}\right) = -Ki \quad ...(37)$$

Thus, the general form of Darcy's law for horizontal as well as vertical flow can be written as per Equation (38):

$$q = -K\frac{dH}{dx} \quad ...(38)$$

where, $q = Q/At$ = Soil water flux, and dH/dx is the hydraulic gradient and H is total hydraulic head where $H = h+x$

In a saturated flow, the hydraulic conductivity K is constant as long as pore structure does not change and it is generally represented by K_s, called saturated hydraulic conductivity.

The assumptions made in Darcy's Law are:

(i) Kinetic energy of water is negligible,
(ii) Mass density of water is constant;
(iii) Viscosity of water is constant, and
(iv) Flow is streamline.

The first assumption implies that when the flow velocity is small, the kinetic energy of soil water is negligible in comparison to its potential energy; hence the total energy of soil water is due to its potential energy only. The second and third assumptions are valid so long as temperature does not vary so that the density and viscosity of water remain constant. The fourth assumption is valid when flow velocity is smaller than the critical velocity for streamline flow. As flow velocity increases, the flow becomes turbulent. However in soils, the flow velocity does not reach the critical limit of streamline flow because even when velocity tends to increase, the forces exerted by the turbulent flow on the soil matrix are so strong that matrix begins to disintegrate. Its fragments block the pore space, increasing the resistance to flow and thus reducing the velocity.

10.3.2. Unsaturated Flow

The soils are usually not saturated because some of the pores are always filled with air. In an unsaturated soil, the water pressure is subatmospheric and it decreases with increase in the degree of unsaturation. The hydraulic conductivity is at its maximum in the saturated soil where all the pores are filled with water and take part in water transmission. With increasing degree of unsaturation, more and more soil pores become devoid of water and do not participate in water transmission, and as a result, the saturated hydraulic conductivity, K, decreases with decrease in either water content or water pressure. The functional relationship of K with either water content (Figure 11) or water pressure is essentially needed to estimate K for a specific water content. As observed with saturated soil, the flow velocity of water in unsaturated soil is also a function of hydraulic gradient.

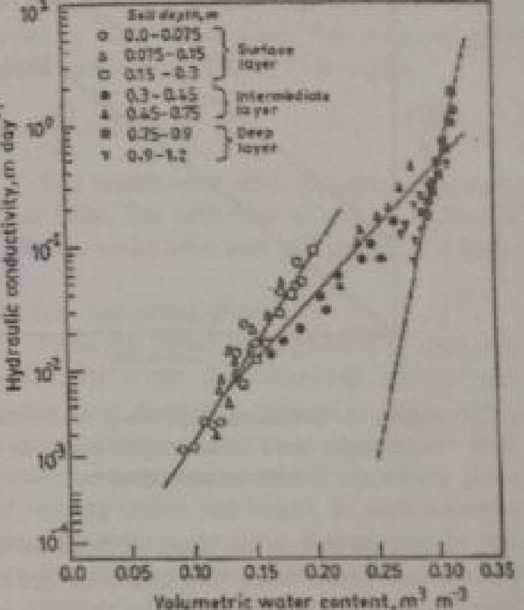


Figure 11. Relationship of hydraulic conductivity with water content for different layers of the Kharagpur lateritic sandy loam soil

Let us take some laboratory cases to derive flow equation describing water flow through homogeneous unsaturated soil. Figure 12a shows an initially unsaturated horizontal soil column subjected to negative water pressures of $-h_1$ and $-h_2$ units, respectively at inflow and outflow ends through porous plates. As a result, the left half of the column is equilibriated to $-h_1$ water pressure and the right

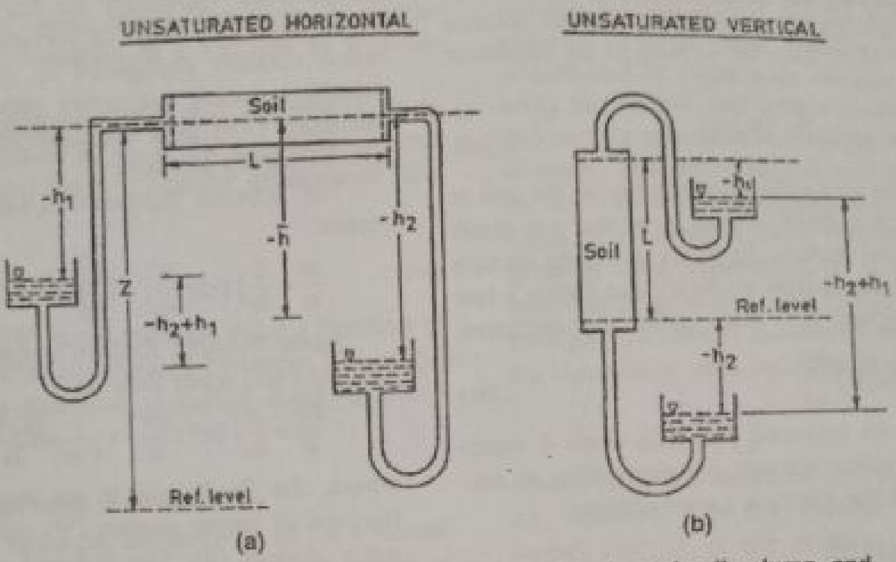


Figure 12. Water flow through (a) unsaturated horizontal soil column, and (b) unsaturated vertical soil column

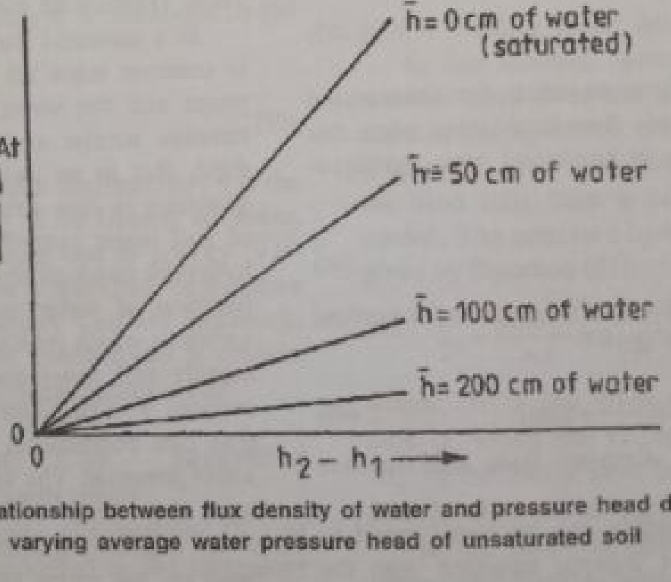


Figure 13. Relationship between flux density of water and pressure head difference for varying average water pressure head of unsaturated soil

half is equilibrated to $-h_2$ water pressure. Water at the central portion of the column would have a mean water pressure of $-\bar{h} = -(h_1+h_2)/2$. As there is no difference in gravity head between inflow and outflow ends, the total pressure head is actually the water pressure head which is higher at the inflow than at the outflow end. Water flows due to pressure head difference, $-h_1+h_2$. The larger the difference in pressure head, the higher is the flow rate, provided the mean pressure head, $-\bar{h}$, is kept constant. If the flow velocity (Q/At) is plotted against h_2-h_1, the graph is linear and the slope of the graph decreases with the decrease in $-\bar{h}$ (Figure 13). The slope of the graph is proportional to the unsaturated hydraulic conductivity which is a function of $-\bar{h}$. It decreases with the decrease in $-\bar{h}$. Based on these relations, the flow equation for horizontal flow of water through unsaturated soil may be expressed as per Equation (39):

$$\frac{Q}{At} = -K(\bar{h})\frac{h_2-h_1}{x} \quad ...(39)$$

Applying the same principles, the flow of water through unsaturated vertical soil column (Figure 12b) may be expressed by Equation (40):

$$\frac{Q}{At} = -K(\bar{h})\frac{(h_2 - h_1) + z}{z} \quad \text{...(40)}$$

$K(\bar{h})$ is the hydraulic conductivity of the soil at mean suction head of soil layer and z is depth of soil from surface. In terms of matric suction (ψ) it can be written as $K(\psi)$ and Darcy's law for unsaturated soil water movement becomes:

$$q = -K(\psi)\frac{dH}{dz} \quad \text{...(41)}$$

$K(\psi)$ can be replaced by $K(\theta)$ when θ represents the water content corresponding to suction ψ and the Darcy's Law becomes:

$$q = -K(\theta)\frac{dH}{dz} \quad \text{...(42)}$$

For three dimensional flow, the Darcy's equation can be written as:

$$q = -K(\theta)\,\Delta H \quad \text{...(43)}$$

A general flow equation for unsaturated transient and steady flow is obtained when this Darcy's equation is combined with the equation of continuity

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q}{\partial x} \quad \text{...(44)}$$

which results in Richard's equation for vertical flow of soil water as:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(K(\theta)\frac{\partial\Psi}{\partial z} + K(\theta)\right) \quad \text{...(45)}$$

where, z is the distance from soil surface downwards.

Richard's equation can be transformed to either Ψ-based or θ-based equation with the use of another paramter defined as hydraulic diffusivity, D(θ), as:

$$D(\theta) = \frac{K(\theta)}{C(\theta)} \quad \text{...(46)}$$

with $C(\theta) = -\frac{\partial\theta}{\partial\Psi}$ which is differential water capacity and is the slope of θ–Ψ function.

There are few advantages in using D(θ) in this flow equation. It reduces the determination of two parameters K(θ) and C(θ) to one parameter D(θ). It also transforms the equation into a form for which analytical solutions are possible, especially for horizontal water movement. The range of variation in D(θ) is much smaller compared to that of K(θ), thereby reducing the errors introduced while assuming D(θ) as a constant value of effective diffusivity of a layer.

The θ-based Richard's equation then becomes:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\theta)\frac{\partial\theta}{\partial z}\right) - \frac{\partial K(\theta)}{\partial z} \quad \text{...(47)}$$

Again Ψ-based Richard's equation for unsaturated soil water flow is:

$$\frac{\partial\Psi}{\partial t} = \frac{\partial}{\partial z}\left(D(\Psi)\frac{\partial\Psi}{\partial z}\right) + \frac{1}{C(\Psi)}\frac{\partial K(\Psi)}{\partial z} \quad \text{...(48)}$$

where, the D, K and C are expressed as a function of matric suction (Ψ) rather than soil water content (θ). The θ based equation has the advantage that D(θ) does not vary much with θ in comparison to variation in K with Ψ.

10.4. Hydraulic Conductivity

In a saturated flow, hydraulic conductivity is constant since all the pores are filled with water and the water geometry in soil pores remains similar as long as pore structure is rigid. But in an unsaturated flow, the water geometry in pore structure becomes a function of soil water content. Therefore, unsaturated hydraulic conductivity has to be measured as a function of water geometry. There are two variables which represent this geometry in soil, namely soil water content, θ, and soil water suction, Ψ. The different values of water content can correspond to same value of soil water potential. The soil water potential represents the energy status of water due to air-water interface geometry and it does not depend on the entire geometry of water phase, whereas volumetric water content (θ) depends on the entire volume geometry of water phase. Hence volumetric water content (θ) is a better index for geometry of water phase than soil water suction (θ). Moreover, when hydraulic conductivity K is expressed as function of Ψ, hystersis is observed, i.e. K(Ψ) depends on the direction of processes (wetting or drying). This is so again because Ψ is an index of surface geometry which differs while drying and wetting. However, when K is expressed as a function of θ, the hysteresis of K(θ) is negligible because θ is an index of entire geometry of volume of water phase. Therefore, K is also generally expressed as a function of θ rather than Ψ.

Buckingham (1907) was the first to study dependence of K on θ. He proposed the one dimensional Darcy's Law for unsaturated flow as:

$$q = -K(\theta)\frac{dH}{dx} \qquad ...(49)$$

where, $K(\theta)$ is not constant but is a function of θ, termed as unsaturated hydraulic conductivity. This equation is called Buckingham – Darcy's Law.

10.5. Intrinsic Permeability

The hydraulic conductivity (K) is a parameter which is dependent on the properties of the medium (soil) as well as the fluid (water). Intrinsic permeability (κ) is dependent on the properties of the medium (soil) only.

Intrinsic permeability κ, is calculated by the Equation (50):

$$\kappa = \frac{K\eta}{\rho_w g} \qquad ...(50)$$

where, K is the hydraulic conductivity; η is the viscosity of water; ρ_w is the density of water, and g is the acceleration due to gravity. Values of 'η' and 'ρ_w' at a specified temperature can be obtained from any standard handbook. The unit of hydraulic conductivity is cm/sec, while that of intrinsic permeability is cm^2.

10.6. Measurement of Saturated Hydraulic Conductivity

Saturated hydraulic conductivity (K_s) of soil is generally determined in the laboratory on undisturbed soil core with the help of either a Constant Head permeameter or Falling Head permeameter. The constant head method is used for soils having high values of hydraulic conductivity, e.g. coarse-textured soils like sandy loam. Since the flow rates of fine-textured soils like clay are very low, the falling head method is generally used.

10.6.1. Constant Head Method

In this method, a constant head is maintained at the soil surface of a previously saturated sample and water is allowed to flow through the sample till the measured outflow attains a constant value. Undisturbed soil core is collected by depth core sampler. After covering the bottom end with a muslin cloth, the soil core is saturated by placing it in a trough of water. The saturated soil core is then placed in the constant head permeameter where water is allowed to flow through it under a constant head maintained by a constant head reservoir (Figure 14). The steady state outflow through soil core is measured and the following equation is used for calculating K_s:

$$K_s = \frac{Vx}{At(\Delta H + x)} \qquad ...(51)$$

where, K_s is the saturated hydraulic conductivity, cm/sec; V is the volume of water collected, cm^3; A is the cross-sectional area of the sample, cm^2; t is the time interval of collection, sec; and $(\Delta H + x)$ is the pressure head difference between inflow and outflow ends.

10.6.2. Falling Head Method

In this method, instead of maintaining a constant head over the saturated soil sample surface, the hydraulic head is allowed to fall (Figure 15). The time interval, t, during which the head falls from a height H_1 to H_2 is recorded. The saturated hydraulic conductivity is given by Equation (52).

$$K_s = \frac{2.3.a.x}{A.t}\log\left(\frac{H_1}{H_2}\right) \qquad ...(52)$$

where, 'a' is the cross-sectional area of the tube in which the falling head is measured; 'H_1' is the initial head in cm; and 'H_2' is the final head in cm, after time t (sec).

In the methods described above, undisturbed core samples brought to laboratory from the field are used for the measurement. Under field conditions, the Auger Hole method is generally adopted if the water table is close to the surface (within 1.5 to 2 metres). It is a very quick method and the calculations are kept to the minimum through the use of nomograms.

10.7. Estimation of Unsaturated Hydraulic Conductivity

The experimental measurement of unsaturated hydraulic conductivity $K(\theta)$ is difficult

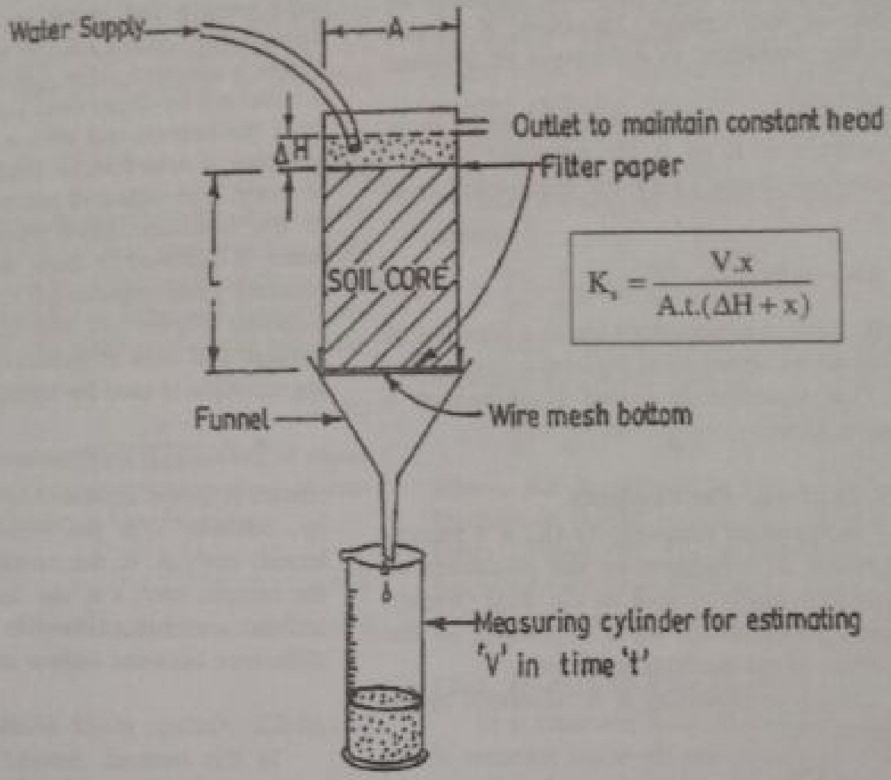


Figure 14. Constant head method for estimating saturated hydraulic conductivity

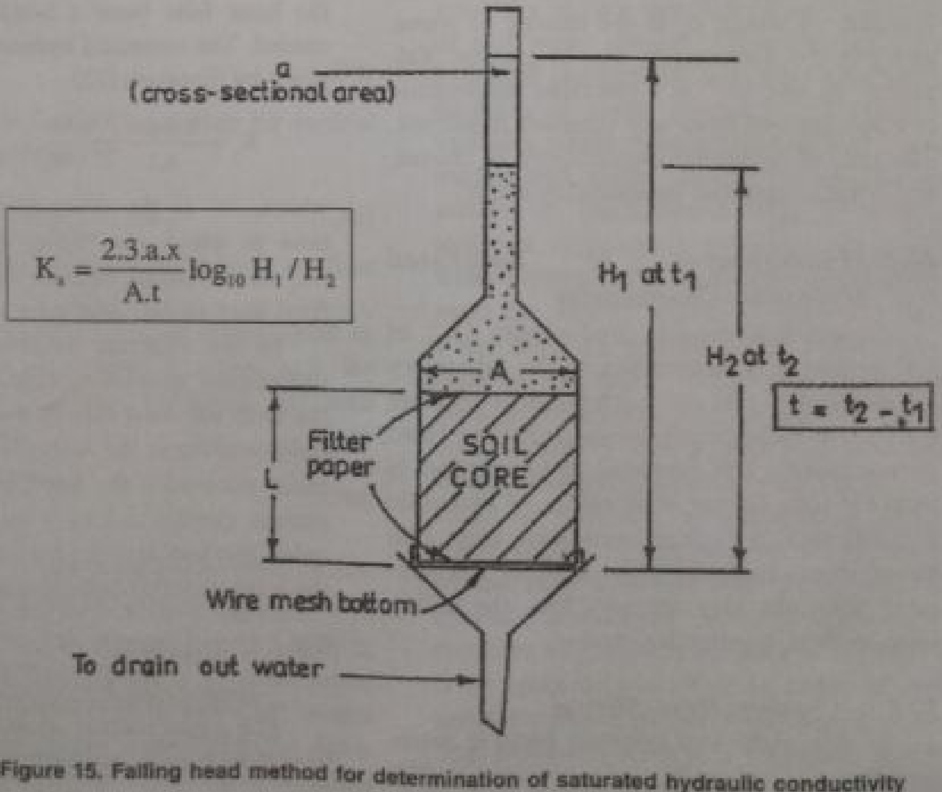


Figure 15. Falling head method for determination of saturated hydraulic conductivity

and complicated. Moreover, at very low θ values, the measurement errors are very high. Therefore, several methods have been developed to estimate $K(\theta)$. These estimations are based on three methods:

(i) Empirical and purely statistical regression methods are commonly known as Pedotransfer Functions (PTF), where $K(\theta)$ is expressed as a function of several measurable soil properties:

$$K(\theta) = f \text{ (Bulk density, \% of silt, \% of clay, organic carbon,...)}$$

This method has local applications and the accuracy is sufficient for the soil mapping purpose.

(ii) Methods based on dynamics of water flow in soil, where $K(\theta)$ is determined by monitoring θ and Ψ as a function of space and time, e.g. instantaneous profile method (Hillel, 1982).

(iii) Method based on parameterization by assuming simplified geometric structure for pore space, the parameters of the geometry are deduced from the measured Ψ–θ characteristics and $K(\theta)$ is then estimated from these parameters.

10.7.1. Empirical and Pedotransfer Function Methods

Several empirical equations have been reported in literature for dependence of K on θ or Ψ which can be used to compute K. The most widely used relations are:

$$K(\Psi) = a\,\Psi^{-n} \qquad \text{...(53)}$$

$$K(\Psi) = \frac{a}{(b + \Psi^{n})} \qquad \text{...(54)}$$

$$K(\theta) = a\,\theta^{n} \qquad \text{...(55)}$$

and

$$K(\theta) = K_s\, s^{n} \qquad \text{...(56)}$$

where, $K(\Psi)$ and $K(\theta)$ are matric suction (Ψ) and soil water content (θ) dependent unsaturated hydraulic conductivities, K_s is saturated hydraulic conductivity of the same soil, s is degree of saturation and a, b and n are empirical constants, which differ from soil to soil. To estimate $K(\theta)$ or $K(\Psi)$, first the empirical constants are determined for a given soil from a set of experimentally determined data and then these empirical constants are substituted in the equation to compute unsaturated hydraulic conductivity at any required θ or Ψ.

Apart from these empirical relations, several pedotransfter functions (PTFs) have been developed where K is expressed by a regression equation and easily-measurable soil properties influencing K are taken as independent variables. For example,

$$K(\Psi) = a_1 \text{ (Bulk density)} + a_2 \text{ (\% clay)} + a_3 \text{ (\% silt)} + a_4 \text{ (\% organic carbon)} + \text{.......} \qquad \text{...(57)}$$

Here, the statistical methods are used to first determine the regression coefficients a_1, a_2, a_3, a_4, ... from a measured data set on K and other soil properties and then with these estimated coefficients, this equation is used to calculate the unsaturated hydraulic conductivity for a given situation with the help of measured soil properties. Such equations are called pedotransfer functions (PTFs).

10.7.2. Instantaneous Profile Method

This method is based on monitoring the transient state internal drainage of a soil profile. The assumptions made in this method are: (i) no evaporation is occurring from the soil surface, and (ii) no horizontal movement of soil water is taking place. In this method, a bare field is selected in which access tube for neutron moisture meter and several tensionmeters in different soil layers are installed. The field is irrigated and is then covered with a plastic sheet. Measurements of soil water content and matric suction of various layers in the profile are made at regular intervals. The rate of change of moisture content per unit time multiplied by the layer thickness determines the drainage flux of that layer. The drainage flux at a depth Z_s can be calculated using Equation (58):

$$q_s = -\frac{dW}{dt} = -Z_s \frac{d\theta}{dt} \qquad \text{...(58)}$$

The cumulative drainage flux from all the layers gives the total drainage flux from the bottom of the profile. Under the irrigated conditions of nearly saturated profile, the downward flux is equal to the hydraulic conductivity $K(\theta)$. So that we get

$$K(\theta) = q_s \qquad \text{at depth } Z_s \qquad \ldots(59)$$

The instantaneous matric suction measurements allow the determination of hydraulic gradient across the profile. Knowing the flux and hydraulic gradient, the hydraulic conductivity can be calculated.

10.7.3. Parameterization Method

- Parameterization of a Function

If a function or a relation between two variables is represented by an equation having a number of parameters, then how these parameters are to be chosen so as to correctly represent a measured set of data variables is called parameterization.

- Parameterization of Unsaturated Hydraulic Conductivity Models

The objective of parameterization of unsaturated hydraulic conductivity is to interpolate and sometimes extrapolate its value. The various unsaturated hydraulic conductivity models are based on different Ψ–θ relationships. A model given by Mualem expresses the conductivity as:

$$K(\theta_G) = K_s\theta_G^a\left[\frac{\int_0^{\theta_G} h^{-(1+b)}d(\theta_G)}{\int_0^1 h^{-(1+b)}d(\theta_G)}\right]^2 \qquad \ldots(60)$$

where, $h = -\frac{\Psi}{\rho g}$ which represents the suction head in terms of matric suction Ψ, density of soil water (ρ) and acceleration due to gravity (g) and

$$\theta_G = \frac{\theta - \theta_r}{\theta_s - \theta_r}$$

where, θ is the actual volumetric water content, θ_s is the saturated water content and θ_r is the residual water content which can be taken as wilting point water content, as this water is not available to plant roots for uptake.

If h is expressed in terms of θ_G using van Genuchten model (1980), the unsaturated hydraulic conductivity can be expressed as per Equation (61):

$$K(\theta_G) = K_s\theta_G^a\left[1 - \{1 - \theta_G^{\frac{n}{n-1}}\}^{1-\frac{1}{n}}\right]^2 \qquad \ldots(61)$$

In terms of suction head h, it can be written as

$$K(h) = K_s\left[1 + (\alpha h)^n\right]^{-\left(1-\frac{1}{n}\right)a}\left[1 - (\alpha h)^{n-1}\{1 + (\alpha h)^n\}^{-1+\frac{1}{n}}\right]^2 \qquad \ldots(62)$$

These two equations are called Mualem-van Genuchten model for unsaturated hydraulic conductivity with five parameters to be determined, namely θ_s, θ_r, a, n and α from a measured data set. Once these parameters are known, this model can predict $K(\theta_G)$ or $K(h)$, as the case may be.

Another model for unsaturated hydraulic conductivity is given by Burdine (1953) where he used the Brooks-Corey model (1966) for h–θ_G relation to solve Mualem equation (60). It is called Burdine-Brooks-Corey model for unsaturated hydraulic conductivity and is expressed as Equations (63) and (64):

$$K(\theta_G) = K_s\theta_G^{a+2+\frac{2}{\lambda}} \qquad \text{for } \theta_G < 1 \qquad \ldots(63)$$

or

$$K(h) = K_s\left(\frac{h}{h_e}\right)^{-2-\lambda(a+2)} \qquad \text{for } h > h_e$$

$$= K_s \text{ for } h \le h_e \qquad \ldots(64)$$

where, h_e being air entry suction head and λ is a parameter related to the pore geometry of soil. This model needs four parameters θ_s, θ_r, a and λ to be determined for predicting unsaturated hydralic conductivity.

Table 2 shows that the values of saturated water content (θ_s) and air entry suction (h_e) are lowest for sand and are due to large continuous structures in the pore space and narrow distribution of pore sizes. In the case of silt and loam, there is a wider range of pore radii so the maximum radius of water-filled pores decreases more rapidly in fine soil than in coarse-textured soil at moderately high values of θ. This results in comparatively rapid decrease of $K(\theta)$ with decreasing θ.

11. Field Water Balance

The field water balance is the balance sheet of inputs and outputs of water in the

Table 2. Mualem–van Genuchten and Brooks-Corey parameters for typical sand, silt and loam soils

Parametrs	Sand	Silt	Loam
Mualem–van Genuchten parameters			
α (m^{-1})	2.3	0.7	1.6
n	4.17	1.30	1.25
θ_r	0.03	0.01	0.00
θ_s	0.32	0.41	0.43
	-1.10	0.00	0.00
K_s ($10^{-6} ms^{-1}$)	2.2	1.0	0.3
Brooks-Corey parameters			
h_b (m)	0.43	1.43	0.63
λ	3.17	0.30	0.25

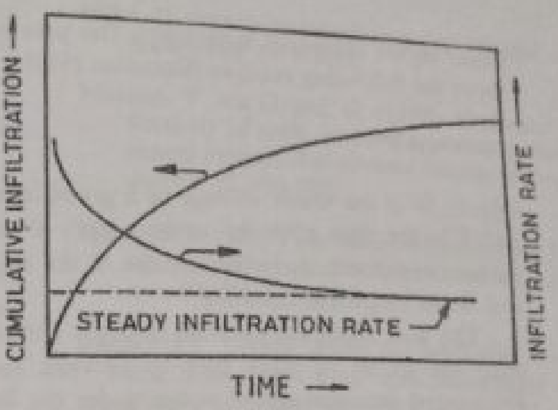


Figure 17. Temporal changes of cumulative infiltration and infiltration rate

soil-water zone. It begins with the entry of rainfall and/or irrigation water by the process of infiltration, continues with the retention and movement of water in the rooting zone and ends up with its withdrawal by drainage, evaporation and plant uptake. The water balance equation for the root zone depth (z) in the time period t_2–t_1 may be expressed as Equation (65):

$$\int_{t_1}^{t_2}(i - v - q)dt - \int_0^z \int_{t_1}^{t_2} r_z dzdt = \int_0^z \int_{t_1}^{t_2} \frac{\partial \theta}{\partial t} dzdt \quad \text{...(65)}$$

where, 'i' is the rate of infiltration, q is the soil evaporation rate, v is the deep percolation rate at depth z and r_z is the rate of water uptake by plant roots.

11.1. Infiltration

Infiltration is the entry of water at the soil-air interface due to sorption and vertical flow of water through the soil profile. The process is of great practical importance as it determines how much of rainstorm or irrigation water enters the soil, and how much overflows the land surface as runoff. The infiltration rate is helpful in deciding the supply rate of irrigation water. It also helps assess the textural and structural conditions of surface soil and transmission characteristics of the sub-surface layers.

When water is ponded in an Infiltrometer (iron cylinder) sunk in soil up to 15-cm depth, the measured rate of entry of water is faster initially due to higher sorption of water by the dry soil. Subsequently, as the top layer becomes saturated, the sorptivity function decreases to the minimum and the rate of entry slows down considerably. It is then controlled by the rate of advancement of the wetting front through sub-surface layer as a function of unsaturated hydraulic conductivity and matric potential gradient in the wetting zone. Irrespective of soil texture, the infiltration rate reduces exponentially with time and attains a steady rate after a long lapse of time (Figure 17). The final steady rate is also known as basic infiltration rate. The infiltrability of soils is the maximum flux of water that enters the soil with free water ponded at the surface. The total depth of water that infiltrates the soil in a time period gives a measure of cumulative infiltration.

The time changes of 'cumulative infiltration' (I) and 'infiltration rate' (i) are better described by the following equations [Eqs (66) and (67)] given by Phillip (1957):

$$I = s\,t^{1/2} + Kt \quad \text{...(66)}$$

$$i = \tfrac{1}{2}\,s\,t^{-1/2} + K \quad \text{...(67)}$$

where, s is the sorptivity function and K is the hydraulic conductivity of the soil corresponding to the water content prevailing in the transmission zone that occurs below the top saturated zone.

Water infiltrated into soil distributes and redistributes itself in the profile. During the initial stages, the water content of the upper layers of the profile decreases and that of the lower layers increases with time. The process obeys the following relation [Equation (68)]:

$$W = AT^{B} \qquad \ldots(68)$$

where, W is the water storage in a given layer and T is the time after the water supply at the surface is cut off. A is the storage at the start and B, an empirical constant.

The soil moisture content changes further with the lapse of time due to upward and downward movements of water under the influence of evaporation and gravitational pulls. These changes in soil moisture profile are referred to as the 'redistribution of soil water'.

11.2. Soil Water Evaporation

Following infiltration, water present in the soil evaporates as long as there is gradient of vapour pressure between soil surface and atmosphere and supply of heat to meet the latent heat of vaporization (580 cal/g of water at 20 °C). The process continues only when there is a continuous supply of water from sub-surface layers to the soil surface, which is influenced by soil hydraulic conductivity. Initially when the soil is wet, evaporation takes place at a constant but rapid rate, depending on the atmospheric evaporativity. The period of this constant rate of drying varies from a few hours in a coarser soil to a few days in a finer soil. When the soil surface dries out, the evaporation rate reduces exponentially with time and is controlled by the rate at which water is supplied by the sub-surface layer to the evaporation site. This stage of drying, which is known as 'falling rate stage' persists for a few days in a coarser soil and for few weeks in a finer soil. After this stage when the surface layer becomes so dry that the conduction of liquid practically ceases, evaporation continues at a very slow and steady rate, depending on the vapour diffusivity of the surface layer and adsorptive forces holding water in the sub-surface layer.

The evaporation loss during the drying cycle can be approximated from the soil water depletion in a time period after deducting the deep percolation loss. In areas where the water table is very shallow, the rate of soil water evaporation is strongly influenced by the depth of water table. For the steady upward flow from shallow water table, the evaporation rate is directly proportional to saturated K and inversely proportional to the depth of water table.

11.3. Deep Percolation

It is the downward flux of water below the maximum root zone depth. In field water balance this forms an important component. It is an unavoidable loss following irrigation, which is more in the coarser than in fine-textured soils. Under steady state conditions, as in a flooded soil, saturated hydraulic conductivity of the least permeable layer in the root zone helps approximate the deep percolation loss. However, under transient unsaturated condition, as is commonly prevalent in irrigated fields, it can be estimated by the 'Instantaneous profile method' (Hillel, 1982).

11.4. Root Water Uptake

A growing plant extracts soil water for its metabolic and physiological processes. Water enters the root through root hairs, moves through the xylem tissue to the leaf and leaves the plant through stomata to the atmosphere. This flow in soil–plant–atmosphere continuum (SPAC) takes place along the direction of decreasing water potential, which is at its minimum in the atmospheric boundary layer surrounding leaf. The rate of flow is moderated by the resistances in SPAC. Under the steady state conditions, the rate of water extraction (r_s) by plant roots from an unit volume of soil can be described by Equation (69):

$$R_s = (\psi_{soil} - \psi_{plant}) / (R_{soil} + R_{root}) \qquad \ldots(69)$$

where, ψ_{soil} and ψ_{plant} are the total water potential in soil and plant, and R_{soil} and R_{root} are the resistances in soil and plant, respectively.

Soil and leaf water potentials are precisely determined by Soil psychrometer and Leaf psychrometer, respectively. The hydraulic resistance in soil is inversely proportional to K(θ) and rooting density 'L' in the soil layer, while in plant root, it is considered to be the sum of resistances to water absorption and conduction. Knowing transpiration rate (considered equivalent to water uptake), ψ_{soil}, ψ_{plant}, and R_{soil} (=1/BK(θ)L), R_{root} can be approximated.

Resistances in plant including those in root xylem and leaf-air boundary, are higher than soil resistance under wet soil water regime while the reverse is true under dry soil water regime.

12. References

Biswas, T.D. and Mukherjee, S.K. (1990) *Textbook of Soil Science*. Tata McGraw-Hill Publishing Company Limited, New Delhi.

Brooks, R.H. and Corey, A.T. (1966) Properties of porous media affecting fluid flow. *Journal of Irrigation and Drainage Division*, Proceedings, American Society of Civil Engineering (IR$_2$). **92**: 61-88.

Buckingham, E. (1907) Studies on the movement of soil moisture. *Bulletin 38, USDA, Bureau of Soils,* Washington, DC.

Burdine, N.T. (1953) Relative permeability calculations from pore size distribution data. *Petr. Trans. Am. Inst. Mining Metall Eng*. **198**, 71-77.

Darcy, H. (1856) *Les Fontaines de la Ville de Dijon*, Dalmont, Paris.

Ghildyal, B. P. and Tripathi, R.P. (1987) *Soil Physics*. Wiley Eastern Limited, New Delhi.

Hillel, D. (1982) *Introduction to Soil Physics*. Academic Press, New York.

Mualem, Y. and Dagan, G. (1978). Hydraulic conductivity of soils: Unified approach to the statistical models. *Soil Science Society of America Journal* **42**, 392-395.

Phillip, J. R. (1957) The theory of infiltration.1: The infiltration equation and its solution. *Soil Science* **83**, 345-357.

Ritchie, J.T. (1972) Model for predicting evaporation from a row crop with incomplete cover. *Water Resources Research* **8**, 1204-1213

Scott, H. Don, (2000) *Soil Physics : Agricultural and Environmental Applications*. Iowa State University Press, Ames.

Thornthwaite, C.W. and Mather, J.R. (1955) The water budget and its use in irrigation. In: *Water* (A. Stefferud, Ed), U.S. Department Agro. Yearbook. U.S. Government Printing Office, Washington. pp 346-358.

van Genuchten, M.T (1980). A closed form of equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society of America Journal* **44**, 892-898.

Veihmeyer, F.J. and Hendrickson, A.H. (1955) Does transpiration decrease as the soil moisture decreases. *Transactions of American Geophysical Union* **36**, 425-428.

Sample Problems

Question 1. The gravimetric (g g^{-1}) soil water content (%) at field capacity (FP) and Plant Wilting Point (PWP) for four layers of a sandy loam soil along with each layer bulk density (BD) are given as follows:

Soil depth (cm)	Water content (% by weight)		
	BD	FC	PWP
0-15	1.59	15.4	4.7
15-30	1.54	17.9	6.0
30-60	1.51	19.4	6.6
60-90	1.49	18.5	6.2

To determine the available water capacity (AWC) of the soil up to 90 cm depth, first the AWC of each layer in terms of depth of water (D_w) is calculated using Equation (8) and then is added for all the four layers to get the AWC for whole soil profile as given in Equation (9)

$$Dw = \{(FC\text{-}PWP) \times BD \times D_s\}/100 \qquad \ldots(8)$$

Soil layer	Thickness of soil layer, D_s (cm)	FC-PWP (%)	D_w (cm)
0-15	15	10.7	2.55
15-30	15	11.9	2.75
30-60	30	12.8	5.80
60-90	30	12.3	5.50

AWC for 0-90 cm soil layer = ΣD_w=16.60 cm. Therefore, total AWC of soil up to 60 cm depth is 16.6 cm.

Question 2. A soil core of length 7 cm and diameter 10 cm is filled with undisturbed soil sample. Water is ponded on the top of the saturated soil sample to a constant height of 3 cm. Under steady state condition, 42 cm^3 water was collected from the bottom of the sample into measuring cylinder in 30 minutes. The saturated hydraulic conductivity is calculated as follows:

Length of core L = 7 cm
Diameter of core = 10 cm
Radius of core (r) = 5 cm
Cross sectional area of core (A) = πr^2 =
$\pi \times (5)^2 = 78.57$ cm^2
Water height above soil (ΔH) = 3 cm
Water collected in 30 minutes = 42 cm^3
Time (t) = 30 min = 0.5 hr

Using Equation (51), we get

$$K_s = \frac{VL}{At(\Delta H + L)} = \frac{42 \times 7}{78.57 \times 0.5(3+7)}$$
$$= 0.748 \text{ cm hr}$$

The measured saturated hydraulic conductivity is 0.748 cm/hr.

Question 3. The average volumetric FC of soil A and soil B are 29.4% and 22.5%, respectively, in upper 60 cm of soil layer. The average volumetric PWP for soil A is 10.4%, whereas for soil B is 6.9%. Calculate the AWC for both the soils, in cm of water in upper 60 cm layer and find which soil is better for plant growth.

(Ans: AWC for soil A = 11.4 cm,
AWC for soil B = 9.36 cm
Soil A is better than soil B)

Question 4. A soil sample in a soil core of height 5 cm and cross-sectional area of 10 cm^2, saturated with water has a head of 5 cm of water maintained on its surface. If the volume of water collected at the outflow end is 8 cm^3 in 10 minutes, calculate the saturated hydraulic conductivity of the soil.

(Ans: K_s = 2.4 cm/hr)

Chapter 7

Soil Air and Soil Temperature

R.P. TRIPATHI and V.S. TOMAR

1. Introduction

The physicochemical properties that affect plant growth and biological processes in most of the soils depend on soil air and soil temperature. The aerobic respiration in roots of plants and soil microorganisms involves continuous consumption of oxygen and evolution of carbon dioxide. An inadequate gas exchange may decrease the growth and yield of crops because metabolic processes of roots are impaired immediately if this exchange is interrupted. The rate of organic matter decomposition and mineralization of the organic form of nitrogen increases with temperature. The biological decomposition as well as uptake and movement of water and nutrients practically cease under very low temperature conditions. Similarly, seed germination, emergence, root and shoot growth are slow under low temperature conditions. It is possible to optimize soil air and soil temperature through cultural practices.

2. Soil Air

2.1. Soil Air and its Composition

A typical mineral soil, on volume basis, contains half solid and half pore space. The solid part is consisted of mineral material and organic matter, and space part is filled with water and air interchangeably (Figure 1). In a saturated soil (after heavy rains or irrigation), most of the pore spaces get filled by water, therefore, under such conditions there is a deficiency of aeration in the soil and plant growth may suffer (except rice). As soil dries, air enters into the pore spaces and in a dry soil most of the pore spaces get filled with air. The proportion of air-filled pore spaces, thus, is subjected to rapid and large fluctuations, depending upon soil moisture content.

Soil aeration refers to the exchange of oxygen and carbon dioxide between the soil pore spaces and the atmospheric air (Figure 2). The process controls the deficiency of oxygen consumed during respiration of plant roots and soil microorganisms and prevents toxicity of carbon dioxide evolved during respiration in the soil air. The term *'air capacity'* is used to describe aeration status of soil. According to

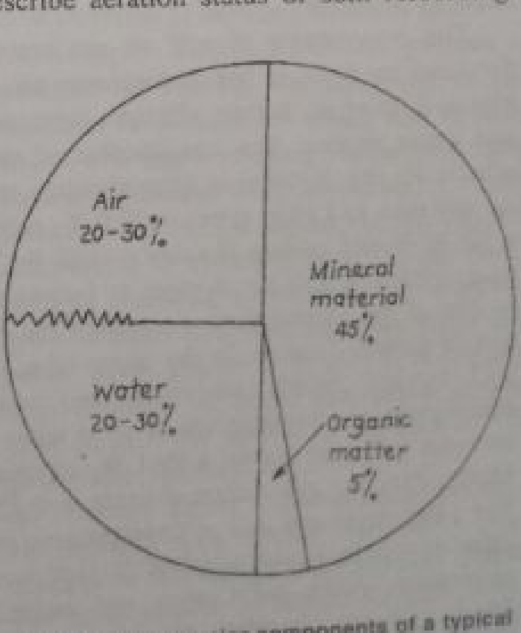


Figure 1. Four major components of a typical mineral soil on volume basis

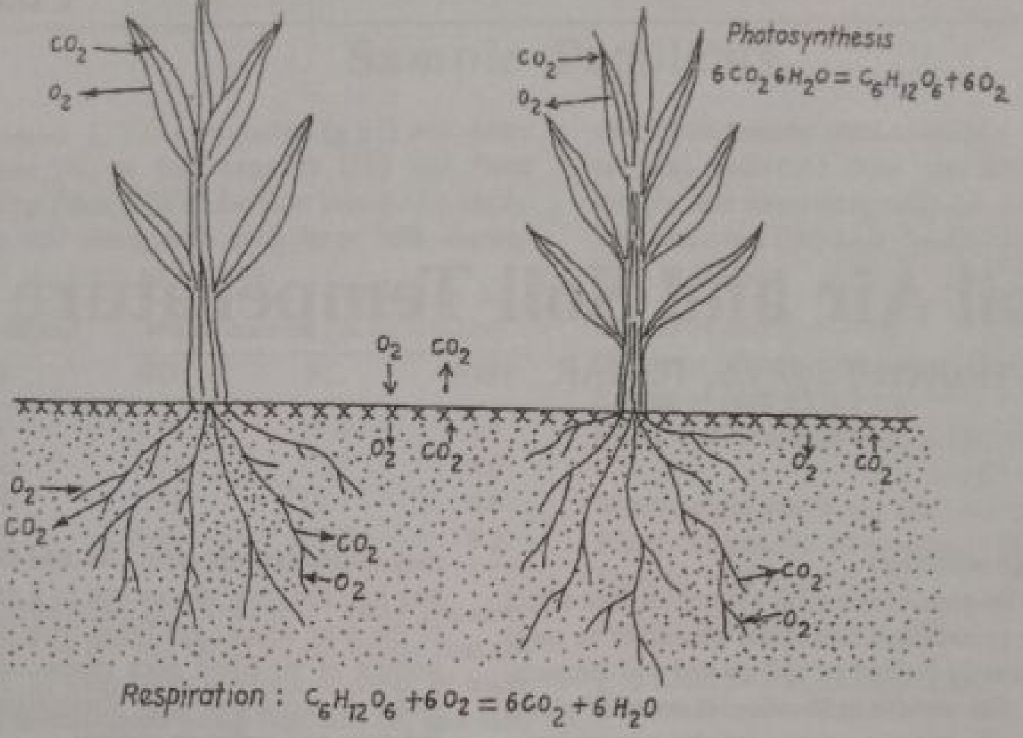


Figure 2. Consumption of oxygen and production of carbon dioxide during respiration by roots of higher plants and microbes. Arrows indicate the exchange of gases between atmospheric air and soil air

Kohnke (1968), the air capacity of soil refers to the volume of pore space filled with air when the soil is under a tension of 50 milli bar. This value is also called as 'non-capillary porosity'. It corresponds to a pore size of 0.06 mm or larger in diameter.

The components of soil air are basically the same as those of the atmospheric air, viz. nitrogen, oxygen, carbon dioxide, inert gases and water vapour. The composition of atmospheric air (by volume) is approximately taken as: nitrogen and inert gases = 79.0 %, oxygen = 20.79 % and carbon dioxide = 0.03 %. The composition of soil air is more or less similar to that of the atmospheric air, except the content of carbon dioxide. In the upper layer of a soil having stable structure and ample macropores, the composition of air is taken as: nitrogen and inert gases = 79.1 %, oxygen = nearly 20.6 %, and carbon dioxide = 0.25%. The plant roots and soil microorganisms consume oxygen and evolve carbon dioxide during respiration. Also, plant roots and microbes decay with the liberation of carbon dioxide. The content of carbon dioxide in soil air is, therefore, several times higher and that of oxygen is slightly less than that in the atmosphere. The composition of soil air is dynamic and varies largely with soil moisture content and from place to place within a given soil.

2.1.1. Factors Affecting Composition of Soil Air

The composition of soil air undergoes rapid changes due to continuous consumption of oxygen by plant roots and soil microorganisms and the production of carbon dioxide and also with renewal of soil air by atmospheric air. Several factors, such as soil characteristics, crop, tillage, organic matter, biological activity, season, etc. affect the composition of soil air.

(i) Soil Respiration – The respiration by roots of higher plants and soil microorganisms is the major cause of consumption of oxygen and production of carbon dioxide in soil. The rate of respiration is influenced by temperature, water supply and the type and amount of respiring tissues.

(ii) Soil Organic Matter – The addition of organic matter to the soil through manures, crop

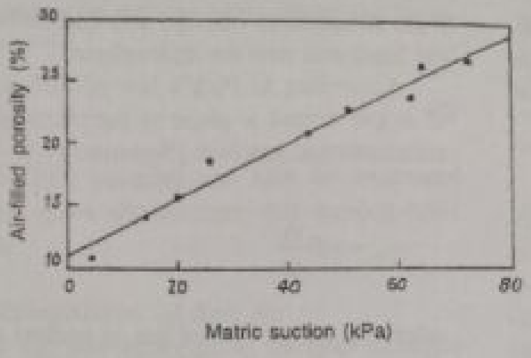


Figure 3. Relationship between air-filled porosity and matric suction in clay loam soil
Source: Bhushan and Sharma (1999)

residues, sewage sludge or other organic materials may affect soil air composition to a large extent. During the process of microbial decomposition of these organic materials, the concentration of oxygen decreases and that of carbon dioxide increases.

(iii) Soil Moisture – The pore spaces in the soil at a given time may be filled by either water or air or by both. These pores get filled by water immediately after heavy rains or irrigation and the oxygen content may fall nearly to zero. When the water from the soil drains or evaporates from the soil surface, air enters into the soil pores (primarily macropores) and oxygen content increases (Figure 3).

(iv) Macropores – The macropore content also affects the total air space and gaseous exchange. Soil texture, bulk density, aggregate stability and organic matter content, all influence the macropore content and, in turn, soil aeration.

(v) Soil Depth – The soil air composition varies with depth. Subsoils are usually more deficient in oxygen than top soils (Figure 4). The total pore space as well as the macropore space are generally much lower at lower depths. The upper layer of a soil having stable structure and large number of macropores, may have oxygen content of nearly 20%. But this oxygen content may become very low, or to near zero, in lower horizons of a poorly-drained soil with few macropores or in waterlogged areas.

(vi) Soil Heterogeneity – It is generally possible that poorly-aerated areas may be found in some well-drained and well-aerated soils. The poorly-aerated portion may be a heavy-textured or compacted soil layer, or it may be located inside a soil structural unit where tiny pores may limit rapid air exchange.

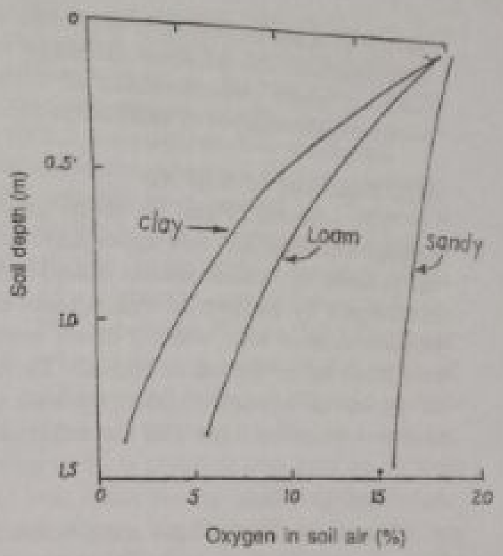


Figure 4. Changes in oxygen content with depth in soils of different texture

(vii) Biological Activity – The carbon dioxide concentration increases with increase of biological activities in soil.

(viii) Effect of Cropping – Cropped fields generally have higher concentration of carbon dioxide than fallow fields. The biological activity associated with crops tends to increase the concentration of carbon dioxide in soil.

(ix) Tillage – Tillage affects the gaseous composition by altering the exchange of gases, which is faster in tilled fields than in untilled fields. The depth of tillage influences the gaseous composition. A shallow tillage encourages formation of carbon dioxide in the top soil in comparison to a deep tillage. Tillage practice, such as puddling or compaction, required to grow lowland rice, decreases the transmission pores (pores greater than 50 µm in diameter). This decrease in larger size pores results in poor aeration for upland crops, such as wheat grown after lowland rice.

(x) Seasonal Effect – The composition of soil air changes with season, primarily due to changes in soil temperature and soil moisture. The concentration of carbon dioxide increases after rains, probably because of increased nitrification and decomposition of organic mat-

ter in addition to slow gas exchange. In summer months, the gaseous exchange is faster causing higher concentrations of oxygen and lower concentrations of carbon dioxide.

2.2. *Renewal of Soil Air*

The renewal of soil air or the gaseous exchange between soil and atmosphere is a natural process by which carbon dioxide in soil is exchanged by oxygen of atmospheric air. Carbon dioxide is also washed down beyond the root zone by irrigation or rainfall. The renewal of gases is brought about by two mechanisms— the mass flow and the diffusion.

2.2.1. *Mass Flow*

The mass flow of air occurs due to total pressure gradient of gas. The pressure gradients cause movement of entire mass of air from a zone of high pressure to that of low pressure. Such a flow of air may occur from atmospheric air to soil air or vice-versa. It may also occur from one pocket of soil to another pocket. When water enters the soil, the soil air is forced out. The rain water is often saturated with oxygen. One cm of rain over an area of one hectare contains 4,339 g (3,000 litres) of oxygen at 20 °C at atmospheric pressure. On the other hand, when water is lost from the soil by evaporation, plant uptake and deep drainage, the air is drawn into the soil. Changes in other meteorological factors, such as temperature, wind action and barometric pressure also affect the gaseous exchange. However, the renewal of gases by mass flow is less important than the diffusion in determining the total exchange that occurs.

2.2.2. *Diffusion*

Diffusion is a process by which each gas molecule tends to move in the space occupied by another gas molecule as a result of difference in partial pressures of the gas. The atmospheric air contains a higher amount of oxygen than the soil air which contains more carbon dioxide than the former. Thus, the partial pressure of oxygen is higher in the atmospheric air than in the soil air, and the partial pressure of carbon dioxide is higher in the soil air than in the atmospheric air, even though the total pressure in the atmosphere and soil may be the same. This results in the movement of oxygen from atmosphere into the soil and carbon dioxide from soil into the atmosphere.

According to Fick's law of diffusion, flux of a gas across a plane is proportional to the concentration gradient [Equation (1)]:

$$q_{gi} = -D\frac{dc}{dx} \qquad \text{...(1)}$$

where, q_{gi} is the flux of gas gi normal to the x-direction, D is the diffusion coefficient, and dc/ dx is the concentration gradient of that gas. The flux refers to the amount of gas diffusing in a unit time across a plane of unit area. D depends on the structure and moisture content of the soil. It is approximately proportional to a fraction of the total soil volume occupied by the gas-filled pores. The rate of diffusion increases with temperature. The diffusion of gases through dry soils of fine texture is higher than that through dry soils of coarse texture. If the soils are moist, the behaviour is reversed because then the air-filled pores will be more in a coarse-textured soil than in fine-textured soil. The value of D in soils is usually between 0.6 and 0.8.

2.3. *Factors Affecting Soil Aeration*

The process of soil aeration controls the levels of oxygen and carbon dioxide in soil. Factors, such as drainage of excess water and macropore content determine the soil areation status. The soil macropore content is influenced by texture, structure, bulk density and organic matter content of a soil. Respiration by plant roots and soil microorganisms, and incorporation of large quantities of manure, crop residues or other organic materials alter the composition of soil air. The soil aeration also changes with season, soil heterogeniety and depth of soil.

2.4. *Characterization of Soil Aeration Status*

Several methods have been used to characterize the soil aeration status; some of these are described below:

2.4.1. *Air-filled Porosity*

The air-filled porosity (f_a) of a soil can be determined by finding total pore space (S) and water content on volume basis (θ) by the Equation (2):

$$f_a = S - \theta \qquad \ldots(2)$$

The air-filled porosity can also be evaluated using air-space pycnometer and tension-table methods.

2.4.2. *Composition of Soil Air*

The oxygen and carbon dioxide contents of the gas samples, taken from soil air, can be determined using chemcial or gas analyzer or gas chromatographic methods.

2.4.3. *Amount of Carbon Dioxide Given off at Soil Surface*

The amount of carbon dioxide is determined for a definite area of the soil surface. It gives an idea about the microbial and root activity in the soil.

2.4.4. *Air Permeability*

Air permeability of soil is a parameter which indicates its readiness to transmit gases when a difference in pressure builds up. The percentage, size and continuity of pores within the soil and occupied by the air determine the permeability of a soil. Special instruments like 'Air Permeameter' and 'Gasometer' are used to determine air permeability under field and laboratory conditions. This method gives a better idea of renewal of soil gases than the actual availability of oxygen to plant roots.

2.4.5. *Apparent Diffusion Coefficient*

Diffusion is the principal mechanism in the exchange of gases between the soil and the aerial atmosphere. Oxygen normally diffuses into the soil and carbon dioxide diffuses out into the atmosphere. The apparent diffusion coefficient is used to characterize a soil for its aeration status.

2.4.6. *Redox Potential*

The oxidized or reduced state of chemical elements in the soil gives an idea about its aeration status. In a well-aerated soil, the oxidized states of elements such as those of iron as ferric (Fe^{3+}), manganese as menganic (Mn^{4+}), nitrate (NO_3^-), and sulphate (SO_4^{2-}) dominate. In poorly-aerated soils, the reduced forms of these elements such as ferrous iron (Fe^{2+}), manganous manganese (Mn^{2+}), oxides of nitrogen (NO_x) and sulphide (S^{2-}) are found.

The oxidation-reduction (redox) potential (E_h) of a soil can be measured by inserting a standard electrode and a platinum electrode in the soil solution. It provides a measure of the tendency of a system to reduce or oxidize substances. The values of E_h can be interpreted up to a certain extent in terms of the ratios of certain oxidizing and reducing organic and inorganic materials present in the soil system. The value of E_h is positive and may be as high as + 0.5 volt under strong oxidizing conditions. It is low and even negative when elements are present in reduced forms. E_h values have the advantage of providing a measure of intensity of reduction under conditions having no molecular oxygen. It is, therefore, useful in characterizing the soil aeration in the range below which the diffusion of oxygen is zero, for example, in rice fields where soils remain saturated or submerged under water. The oxidized and reduced forms of certain elements in soil and the E_h values at which change in the form occurs are given in Table 1.

2.4.7. *Oxygen Diffusion Rate (ODR)*

Analogous to measuring of oxygen movement to a plant root, the rate of oxygen diffusion is measured by inserting a small platinum wire electrode into the soil. The apparatus consists of a platinum microelectrode which is pushed into the soil to a desired depth and an

Table 1. The oxidized and reduced states of certain elements in soil and the redox potential (E_h) at which change in the form occurs

State of element		E_h at which change in form occurs (V)
Oxidized	Reduced	
O_2	H_2O	0.38 to 0.32
NO_3^-	N_2	0.28 to 0.22
Mn^{4+}	Mn^{2+}	0.28 to 0.22
Fe^{3+}	Fe^{2+}	0.18 to 0.15
SO_4^{2-}	S^{2-}	-0.12 to -0.18
CO_2	CH_4	-0.20 to -0.28

Source: Patrick and Reddy (1978)

electrical cell is formed by inserting into the soil nearby a saturated calomel electrode. The calomel electrode through a galvanometer is connected to the positive terminal of an external source of direct current. Its negative terminal is connected to the platinum micro-electrode. The result of measurement is an electric current which is related to the rate of oxygen flux at the electrode surface and is given by Equation (3):

$$i \times 10^{-6} = n\,FAf \qquad \ldots(3)$$

where, i is electric current in microamperes, n is number of electrons required for reduction of one molecule of oxygen which is 4 , F is the Faraday's constant, A is the surface area (in cm^2) of platinum electrode, and f is the flux or ODR to the electrode surface in number of molecules of oxygen per second per cm^2. The ODR is calculated by Equation (4):

$$\begin{aligned} \text{ODR} &= (i \times 60 \times 32 \times 10^{-6}) / (4 \times 96500 \times A) \\ &= (0.497 \times 10^{-8} \times i) / A \ (g\ cm^{-2}\ min^{-1}) \qquad \ldots(4) \end{aligned}$$

The requirement of ODR of a number of crops ranges from 22.9×10^{-8} to 39.5×10^{-8} g cm^{-2} min^{-1} (Table 2). In general, the critical ODR value of soils in which roots of many plants will not grow is of the order of 20×10^{-8} g cm^{-2} min^{-1}.

2.5. *Effects of Soil Aeration*

The soil aeration may have direct and indirect effects on activities of higher plants and soil microorganisms. The level of soil aeration determines the forms of several inorganic elements, soil reactions, microbial decay of organic matter, symbiotic nitrogen fixation and, in turn, the soil properties and plant growth. The harmful effects of poor aeration are described below:

Table 2. Critical oxygen diffusion rate (ODR) for different crops

Crop	ODR (10^{-8} g cm^{-2} min^{-1})
Maize	30.0
Soybean	28.4
Wheat (unirrigated)	26.9
Wheat (irrigated)	26.3
Peas (Unirrigated)	32.9
Peas (irrigated)	39.5
Toria	22.9

Source: Sinha and Ghildyal (1982)

(i) The growth of plants, particularly the root growth, is adversely affected by poor aeration. Waterlogging for more than 48 hours at knee-high and tasselling growth stages may retard the growth of maize crop. In some cases, an abnormal development of roots may occur, e.g. the de-shaping of sugarbeet and carrots.

(ii) The absorption of nutrients and water gets reduced in poorly-aerated soils.

(iii) Under anaerobic conditions, the formation of inorganic compounds, which are toxic to plants, is favoured.

(iv) The changes in soil oxygen content may affect the susceptibility of plant roots to diseases, and/or the virulence of soil-borne disease organisms, or both, resulting in increased disease incidence.

(v) The absence of sufficient oxygen in soil affects the nitrogen fixation by *Rhizobium*.

(vi) The soil organic matter cannot be decomposed properly by soil microorganisms unless sufficient oxygen is present in the soil. If anaerobic bacteria decomposes soil organic matter, toxic substances like sulphides, methane, organic acids, etc. are liberated which are harmful for plant growth.

(vii) In well-aerated soils, oxidized forms of most of the elements are present, e.g. nitrate-N. These are desirable for most common crops. The reduced forms, specifically of iron and manganese, may be present at toxic levels in the moist acidic soils.

2.6. *Management of Soil Aeration*

The proportion of air-filled pore spaces and the ease with which exchange of gases can take place between atmospheric air and soil air influence the aeration status. The practices encouraging the air-filled pore spaces and the gaseous exchange may improve the aeration status. The variation in tendency of plant species to tolerate water stagnation/ poor aeration can also be utilized to harness better crop yields from such soils. The soil aeration under field conditions can thus be optimized by the following methods:

2.6.1. Improving Soil Structure

An increase in the volume of air-filled pores can be attained by improving soil structure. The practices that help in the maintenance of stable soil aggregates such as use of animal manure, green manure, plant residue manure and growing of legume crops, will, in turn, encourage better soil aeration.

2.6.2. Drainage

An aerobic soil environment can be maintained by providing surface and /or sub-surface drainage. The soil pores get filled with water due to continuous seepage from canals, presence of perched or high water tables, and after heavy rains or irrigation. The drainage of such fields is essential for the supply of sufficient oxygen.

2.6.3. Cultivation

A light cultivation of soil or interculture operation does not only control weeds but helps in the exchange of gases, specially in heavy-textured poorly-drained soils. After rains, an impermeable layer/ crust is formed at the soil surface that hinders the gaseous exchange. A light cultivation will break it and help in improving soil aeration.

2.6.4. Plant Adaptations

Plant roots, in general, are adapted to aerobic respiration. However, some of the plant species develope mechanisms such as increase in the air space of roots (root porosity) or internal aeration through leaves and cortex cells and thereby can grow even in oxygen-deficient soils. The selection of crop species, therefore, is important for growing crops in waterlogged or poorly-drained soils. For example, rice thrives well in submerged soil conditions. Also, soybean crop can tolerate temporarily waterlogged soil conditions better than maize, pigeonpea and other deep-rooted crops.

3. Soil Temperature

3.1. Soil Temperature and Heat

Heat is a form of energy and temperature is a measure of the heat energy. The heat energy refers to kinetic energy of random motion (vibration) of molecules of a substance. The speed of vibration is directly proportional to temperature. The unit of heat is calories or joules and that of temperature is °C. Solar radiation is the source of soil heat. The flux of heat into- and out of the soil determines the soil thermal regime, which is characterized in terms of soil temperature.

3.2. Variation of Soil Temperature

Soil temperature at any time of the day depends on the ratio of solar energy absorbed to energy lost and therefore, depends on the angle of incidence of sun rays on the soil surface. Since position of the sun from the earth and hence the accompanying energy varies both diurnally and annually, the soil profile experiences diurnal and annual variations of soil temperature.

3.2.1. Dirunal Variation

When sun rays strike the ground, the energy it carries is partly absorbed by the soil, thereby the surface soil gets heated and a difference of temperature between the surface soil (hotter) and the subsurface soil (cooler) is created. This difference of temperature causes the heat to flow downward. The difference between maximum and minimum temperature is highest at the soil surface and decreases with depth and becomes small beyond 0.25 m depth (Figure 5).

While the soil surface starts getting heated just after sunrise, heating at lower depth starts much later. The surface soil is heated rapidly between 10.00 and 13.00 hours of a day. But the soil temperature at 0.05 m and 0.1 m depths is maximum around 15.00 and 17.00 hours, respectively. The temperature maxima at still lower depths occur during night time when the upper part of the profile starts cooling rapidly. The progressive delay in the time of occurrence of maximum temperature with depth indicates slow rate of penetration of heat wave into the soil. This time lag in the heating of soil profile 'at a particular depth is a function of thermal properties, moisture content and structure of the soil. Figure 5 shows that the cooling period of daily cycle in the surface layers is more than twice as long as the warming period. The average temperature of daily cycle in the surface layer of the soil is generally higher than the average air temperature. In

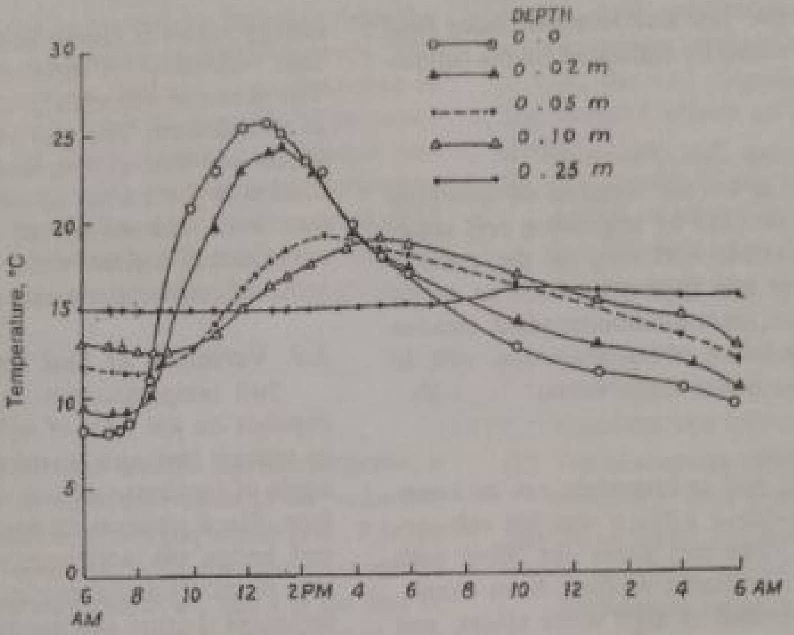


Figure 5. Diurnal variation of soil temperature in a bare soil on a clear day
Source: Tripathi and Ghildyal (1979)

general, the dirurnal changes in soil temperature are lower in moist than in dry soils, in compacted than in loose soils, and in deeper than in shallower soil layers.

3.2.2. Annual Changes

The annual variation of thermal regimes of a Mollisol is depicted in Figure 6. The annual temperatures at 0.05m depth show a progressive increase from 5th week of the year to 24th week (onset of south-west monsoon), followed by a rapid decrease until the 28th week and thereafter a partial stability until the 35th week (end of monsoon). The soil temperature tends to increase slightly during 35th and 42nd weeks of the year and then declines, completing the annual cycle. The sudden drop in temperature between 24th and 41st weeks is associated with excessive wetness due to rainfall, rise in groundwater table and cloudy conditions. The annual changes in soil temperature may occur even beyond 1.0 m depth.

3.3. Factors Affecting Soil Temperature

Soil temperature is controlled largely by the factors which affect energy balance on and within the soil. Incoming and outgoing solar radiations, albedo, soil colour, slope, soil composition, mulches and vegetation, irrigation and drainage and evapòration affect the soil temperature.

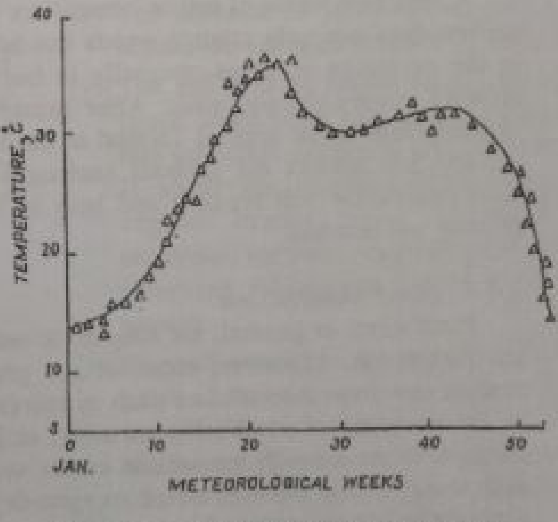


Figure 6. Annual variation in average soil temperature at 7 AM and 2 PM in a bare soil at 0.05 m depth

3.3.1. Solar Radiation

The amount of solar energy received by soil depends on the constituents of the atmosphere and the aspect. Clouds, water vapour, dust particles and aerosols reduce the solar energy reaching the soil surface. Only 35-40% of the solar radiation actually reaches the earth in cloudy humid regions and about 75% in the cloudless arid regions. Out of this, only 10 % is actually absorbed to warm the soil and the remaining is lost as heat of vaporization from soil or leaf surfaces and is re-radiated as long wave radiation or reflected back to sky. The aspect differences result in warmer south-facing and west-facing slopes than their respective opposite faces.

3.3.2. Albedo

Albedo, which generally, refers to short wave reflectivity, is related to colour and angle that sun rays make with the striking surface. A higher albedo at the dawn and dusk than at other times of the day (Figure 7) and in winter than in summer, causes a continuous change in absorption of heat by soil which, in turn, leads to diurnal and annual variations in soil temperature. A lower albedo means a higher absorption of solar energy. Moist soils which are relatively darker have lower albedo than drier ones.

3.3.3. Soil Colour

Light-colour soils have twice the albedo of the dark-colour soils. Consequently, a dark-colour soil is warmer on clear days because it absorbs more solar energy than a light-colour soil. However, since a good absorber is also a good emitter, dark-colour soils emit more energy during the night and, thus, allow a wider daily variation in temperature. Also, dark-colour soils are generally associated with high humus and high retention of water and, therefore, have high heat capacity.

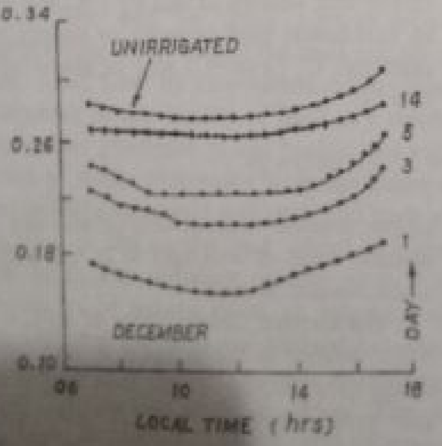


Figure 7. Diurnal variations of bare soil albedo on 1, 3, 5 and 14 days after irrigation and in unirrigated plot

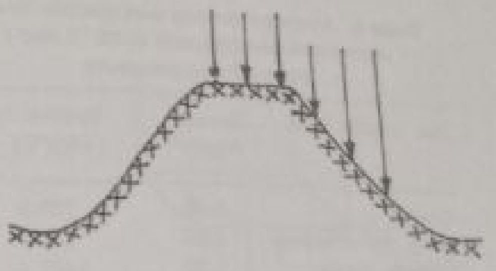


Figure 8. Soil area warmed by sun rays striking on level surface and on slopes

3.3.4. Slope

Solar radiations reaching soil perpendicular to the soil surface level heat it more as these concentrate on a smaller area than when the same amount of radiations reach a slope where these get spread on a larger area (Figure 8). This is the reason why temperature on crest of the ridges is higher than on sides and furrows. In northern hemisphere, a south-facing slope is warmer than the north-facing slope and horizontal surface. Similarly, a west-facing slope is warmer than the east-facing slope.

3.3.5. Soil Composition

The specific heat of soil components (Table 3) indicates that the soils rich in humus will be cooler than those poor in organic materials in an environment for the same incoming radiations. Because of high thermal conductivity, a compacted soil is cooler than a loose soil.

3.3.6. Mulches and Vegetation

Mulches affect the amount of heat received by the soil and its dissipation. For example, straw mulch highly immobilizes the air within the mulch and since the still air has very low thermal conductivity, heat flow from surface of the mulch to the soil and vice-versa is reduced to a large extent. This, in turn, lowers the maximum temperature and raises the minimum temperature of a soil. The vegetation has effects

Table 3. Average density and specific heat of some soil constituents at 25 °C and 1 atmosphere pressure

Soil constituents	Density (Mg/m³)	Specific heat (J/kg/°C)
Quartz	2.65	733.2
Many soil minerals	2.65	733.2
Organic matter	1.30	1927.4
Water	1.0	4190.0
Ice at 0 °C	0.917	2107.6
Air	0.0012	1005.6

similar to those of the straw mulch. Both, the canopy height and thickness of vegetation affect the heat flux into and out of the soil. A thick vegetation highly resists the movement of air within its canopy and therefore, insulates the soil surface against appreciable heat fluxes.

3.3.7. Irrigation and Drainage

For the same incoming solar radiation, irrigation stabilizes the soil temperature by increasing its heat capacity, whereas the drainage, which leads to the removal of water from soil, decreases its heat capacity and raises the soil temperature. A dry soil is heated more easily than a wet soil because the specific heat of water is nearly five-times that of dry soil. The high specific heat of water prevents the sudden changes in temperature of irrigated soils during winter. The effect of irrigation on temperature also depends on the relative temperature of irrigation water and soil, atmospheric condition, and nature and extent of crop cover.

3.3.8. Evaporation

The evaporation causes significant cooling of soil because the heat of vaporization of water is high (580 kcal/kg of water at 25 °C). The low temperature of a wet soil is due partially to its high heat capacity and evaporative cooling.

3.4. Thermal Properties of Soil

The thermal properties of soil are characterized in terms of heat capacity, specific heat, thermal conductivity and thermal diffusivity.

The heat capacity of a soil is defined as the ratio of heat supplied to a body to the corresponding rise in its temperature, and is given by Equation (5):

$$C = \frac{\Delta Q_h}{\Delta T} \qquad \ldots(5)$$

where, C is the heat capacity, ΔQ_h is the quantity of heat supplied to the body and ΔT is the increase in its temperature (°C).

The heat capacity per unit mass of a body is called the specific heat (c) and is defined as the quantity of heat required to raise the temperature of a unit mass of the substance through 1 °C [Equation (6)]:

$$C = \frac{\text{Heat capacity}}{\text{Mass}} = \frac{\Delta Q_h}{m\Delta T} \qquad \ldots(6)$$

where, m is the mass of the substance.

The values for specific heat of some soil constituents and certain soils are given in Tables 3 and 4, respectively. If m_1, m_2, m_3 and m_4 are masses of mineral matter, organic matter, water and air in soil and c_1, c_2, c_3 and c_4 are their specific heats, respectively, then the heat capacity C (J/kg/°C) is given by Equation (7):

$$C = m_1c_1 + m_2c_2 + m_3c_3 + m_4c_4 \qquad \ldots(7)$$

The heat capacity of a soil is expressed as quantity of heat required to raise the temperature of unit volume of soil by 1 °C and is known as volumetric heat capacity or simply the heat capacity, C_v (J/m³/°C) and is given by Equation (8):

$$C_v = \rho c(1 + \theta) \qquad \ldots(8)$$

where, ρ is the bulk density of soil and c is the specific heat of soil and θ is the volumetric water content in soil

The thermal conductivity of soil, λ (J/m/s/°C) is defined as the quantity of heat passing in a unit time through a unit area of soil under a unit temperature difference between the faces. The thermal conductivity of a soil depends on its water content, texture, structure, mineralogical composition, organic matter content and compaction, but is independent of the tempera-

Table 4. Specific heat of some soils

Soils	Specific heat (J/kg/°C)
Kharagpur sandy clay loam	963.7
Phoolbagh clay loam	967.9
Beni silty clay loam	936.6
Haldi loam	867.3

tures encountered by it. The thermal conductivity of soil increases with increase in the water content, till about 50% of the soil saturation. In general, the thermal conductivity of soil varies in the order: sand > loam > clay > peat. The thermal conductivity of quartz, water and air are in the ratio 333 : 23 : 1.

The thermal diffusivity, α (m²/s), is defined as the ratio of thermal conductivity, λ, to heat capacity, C_v [Equation (9)]:

$$\alpha = \frac{\lambda}{C_v} \qquad \ldots(9)$$

The thermal diffusivity, α, is a measure of the rate at which changes in temperature occur in the body.

3.5. Heat Flow in Soil

The heat flow in a soil occurs mainly through conduction. The flow of heat by convection and radiation is of little importance. In conduction, heat flows by the transfer of momentum of individual faster-moving molecules at the hot end to their slower-moving neighbouring molecules at the cool end by collision. Heat flow by convection occurs due to differences in the density of fluid as a result of temperature gradients in soil profile and vaporization of water molecules from the hotter end of a partially saturated soil pore and their condensation at the cooler end, where it releases latent heat. The radiative transfer of heat in a soil occurs in empty pores as thermal radiation. The heat flow in a soil through conduction is described by Fourier's law, according to which the heat flux in a slab of soil (Figure 9) is directly proportional to the ratio of temperature difference $(T_1 - T_2)$ between the hot and cool ends of the slab and its thickness $(x_1 - x_2)$. The proportionality constant is given by thermal conductivity λ (J/m/s/°C). The Fourier's law in the differential form is written by Equation (10):

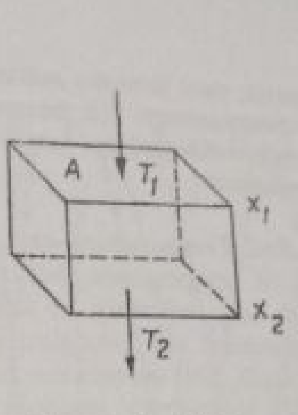


Figure 9. A soil slab

$$q_h = -\gamma \frac{dT}{dx} \qquad \ldots(10)$$

where, q_h is the heat flux (J/m²/s) or quantity of heat, Q_h (J), passing through a unit area, A (m²), of soil in unit time, t (s); and dT/dx is the temperature gradient, °C/m. The negative sign indicates that flux and gradients are in opposite direction. A principal factor influencing the heat conduction in a soil is its water content. A dry and loose soil is a poor conductor of heat than a wet and compact soil.

3.6. Measurement of Soil Temperature

The measurement of soil temperature is based on the change in thermometric properties of a subsidiary system which is in thermal equilibrium with the soil. The thermometric methods are both contact and non-contact types. The contact type methods include measurement using thermometers in which thermal expansion of a solid, liquid or gas is measured, the corresponding examples are: bimetallic strip thermometers, mercury or alcohol (in glass) thermometers, and constant pressure or constant volume thermometers. Other common contact type thermometers are based on the changes in electrical properties of the material with change in temperature; common examples being thermistors and thermocouple thermometers. In most of the commonly used contact methods, changes in temperature are recorded using mercury (in glass) thermometers, thermistors and thermocouple thermometers. The non-contact type methods include optical pyrometers, total intensity radiometers and infrared thermometers. All thermometers must be carefully calibrated and installed at the point of measurement.

3.7. *Soil Temperature and Plant Growth*

Soil temperature influences the plant growth only indirectly, by affecting the physical, chemical and biological processes in soil and plants. Soil temperature influences the cell activity, biochemical reactions and phsiological processes necessary for the overall growth and development of a plant. Seeds of most of the crop species germinate within a reasonable time period between 10 °C and 35 °C. Unfavourable temperatures prevent the emergence of many tender seedlings.

Some plant species are adapted to low temperatures only while some are adapted to high temperatures. At higher soil temperatures, the roots of low-temperature crops start decaying and are damaged by soil organisms and plant diseases, but the reverse is not true for a high temperature crop. The temperature optima for root growth of most crop species is between 20 °C and 25 °C. The optimum temperature for the root growth is often lower than for shoot growth. The optimum temperature for the activity of most of the microorganisms is between 25 °C and 35 °C. A low temperature produces toxic substances which are injurious to plants, whereas a high decomposition rate at optimum temperature usually results into the products which are beneficial to plants.

For many crops, particularly cereals, the shoot meristem, which is the site of temperature perception, lies below the soil surface for an appreciable period. The plant growth is initiated only when the minimum (or base) temperature is reached and the rate of growth increases up to the optimum temperature, followed by a decline at the maximum tolerable temperature. The minimum and maximum temperatures differ in different crop species and at different stages of crop growth.

3.8. *Management of Soil Temperature*

The primary source of heat energy to soil being the sun, practices encouraging the heat absorption and the flow in soil and those discouraging heat loss to the atmosphere would help in heat storage and temperature rise. The soil temperature under field conditions can thus be altered by mulching and vegetation, tillage, compaction, irrigation, and drainage.

3.8.1. Mulching and Vegetation

Mulches, such as straw, tend to buffer the extremes in soil temperatures (Figure 10). Mulching with the polyethylene sheets raises the soil temperature during the day. Mulches also conserve soil moisture. A polyethylene film laid on a soil surface is transparent only to the incoming radiations. The outgoing radiation is inhibited due to the condensation of water vapour and the presence of dust particles on soil side of the polythene surface.

The vegetation, like straw mulch also intercepts both the incoming as well as the outgoing radiation from soil and therefore, reduces the temperature fluctuations. The overall effect depends on the proportion of shaded soil. Air temperature above a crop is lower than at the surface of soil on a clear night. Soil under vegetation warms up more slowly in winter than a bare soil. A soil under a dense vegetative canopy may remain at a uniform temperature in surface layers.

3.8.2. Tillage

Tillage of any kind affects the pore-size distribution and wetness and consequently, the soil temperature. The tillage forming ridges often create a range of soil temperature regime (Figure 11). Rapid changes in soil temperature are observed on ridges, whereas furrows are generally cooler. In poorly-drained soils, ridging improves the soil aeration. A tilled-surface soil is generally warmer during the day time because of its higher porosity and low thermal conductivity than a compacted-untilled soil.

3.8.3. Soil Compaction

Compaction in soil brings its particles closer and regulates temperature in the root zone. The abrupt changes in soil temperature, as observed in the surface layers of a loose soil, do not occur in a compact soil due to its high thermal conductivity. The soil compaction under field conditions is often achieved using a tractor or bullock-drawn roller.

3.8.4. Irrigation and Drainage

The soil water controls absorption of solar radiation, loss of heat energy to the atmosphere,

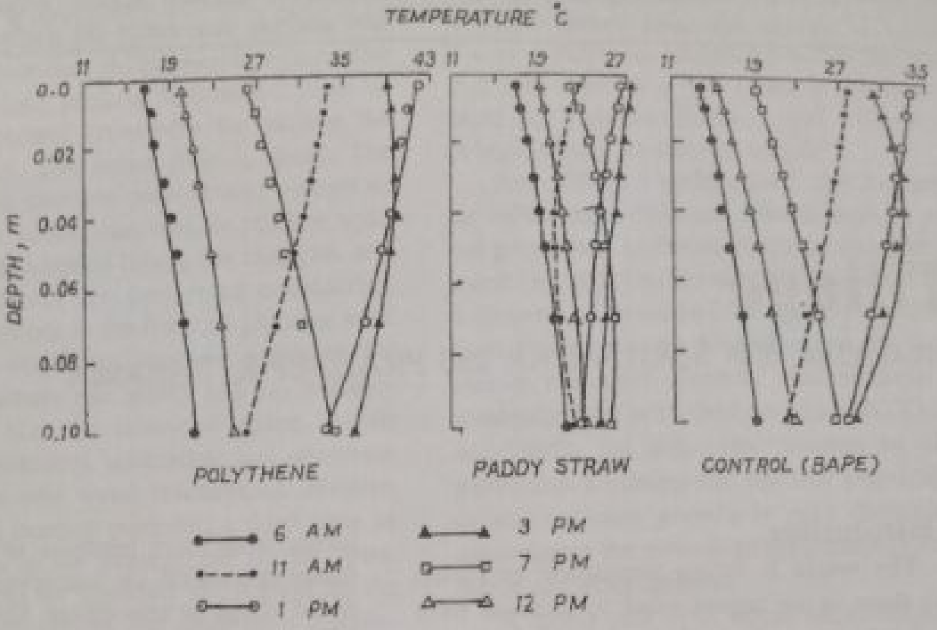


Figure 10. Effect of mulches on soil temperature profiles, 19 days after irrigation
Source: Tripathi and Katiyar (1984)

Figure 11. Soil temperature at 0.02 m depth in a north-south running ridge-furrow system at Pantnagar (Uttrakhand)

and movement of heat in soil. Water resists changes in soil temperature because of its high specific heat and high heat of vaporization. A wet soil conducts heat faster than a dry soil and therefore, temperature fluctuations in the surface layers are rapid and approach extreme values easily in drier soils than in wetter soils. The dry soils prevent heat flow and encourage temperature fluctuations. The raising of soil temperature by draining of waterlogged fields and stabilizing soil temperature by irrigating dry fields are the common practices used by farmers. Irrigation reduces both rise and fall of soil temperature due to high heat capacity of water and evaporative cooling. The irrigation in summer causes large evaporative cooling of the surface soil.

4. References

Bhushan, L. and Sharma, P.K. (1999) Effect of depth, bulk density and aeration status of root zone on productivity of wheat. *Journal of the Indian Society of Soil Science* **47**, 29-34.

Kohnke, H. (1968) *Soil Physics*, McGraw-Hill, Inc. New York, USA.

Patrick, W.H. (Jr.) and Reddy, C.N. (1978) Chemical changes in rice soils. In: *Soils and Rice*, The International Rice Research Institute, Los Baños, Laguna, Philippines, pp 361-379.

Sinha, A.K. and Ghildyal, B.P. (1982) Soil aeration. In: *Review of Soil Research in India*, Part I. Indian Society of Soil Science, New Delhi.

Tripathi, R.P. and Ghildyal, B.P. (1979) Thermal regimes of Mollisols under high water table conditions. *Agricultural Meteorology* **20**, 493-505.

Tripathi, R.P. and Katiyar, T.P.S. (1984) Effect of mulches on the thermal regimes of soil. *Soil and Tillage Research* **4**, 381-390.

Tillage

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1. Introduction

The world is facing population explosion and there is an urgent need to increase farm productivity and total food production on sustainable basis, without compromising on natural resources and environment. Several innovations have gone into improving the efficiency of farming systems, with variable results. Tillage is the first step in agriculture. It is an integral part of crop production. Tillage includes operations of planting; weed control; incorporation and mixing of crop residues, manures, fertilizers and pesticides in soils; compaction of loose soils; managing the problems of soil compaction; terracing; mulching and other soil management practices.

The basic objective of tillage is to modify soil conditions in such a way that they become most favourable for the growth and yield of the crop. Tillage practices, in fact, started when man first planted seed in soil. In the process of evolution, tillage has gone through modifications, starting with manipulation of soil with simple hand-operated stone and wooden tools, through animal-drawn metal tools to modern heavy tractors and machines. Each tillage system has its own merits and demerits. Tillage requirements are highly crop- and site-specific. Tillage is the single most expensive component of crop production. The cost of cultivation may be reduced if tillage operations are minimized. This, along with the introduction of chemicals (herbicides) in agriculture in the 1960s, led to the change in the concept of tillage from more intensive to less intensive tillage operations, i.e. conservation tillage, reduced tillage or zero tillage. However, several health risks are associated with chemical agriculture. Further, climate change and global warming have become issues of immediate concern to mankind. In modern agriculture, increased emphasis is being laid on carbon sequestration in soils to arrest global warming and climate change. These factors have once again forced the agricultural scientists to rethink about the utility of tillage in safe crop production.

2. Definition

Tillage is usually defined as the mechanical manipulation of soil aimed at improving soil conditions affecting crop production. In the true sense of words 'mechanical manipulation' includes all soil-handling operations in agriculture, earth moving, mining and construction. However, from crop production point of view, tillage is a process in which human, animal and/or machine energy is applied for physical manipulation of soil to provide soil environment favourable for the growth and production of plants (food, pulses, oilseeds, fibre, timber, medicinal and aromatic plants, etc.). During tillage operations, the soil is subjected to shearing, compressive and tensile stresses. A pure shear stress changes the shape without changing the volume, whereas pure compression leads to the change in volume without changing shape. Practically, shear and compression usually occur together in soil during tillage.

Traditionally, tillage process involves a series of primary (to break soil surface into clods) and secondary (to pulverize, repack and/or smoothen soil surface) operations; some of them are operated repeatedly for getting the desired result. The entire field is tilled. The entire process involves considerable amount of time, energy, labour and cost. In modern agriculture, the concept of tillage has changed, and tillage operations are performed considering two distinct zones in the field: (i) planting zone (where soil conditions must be optimum for seed to germinate and grow), and (ii) management zone (i.e. the inter-row space, which must have maximal infiltration and minimum soil erosion and weed infestation). Modern mechanized farming considers a third zone as well, which is the passage for tractor tyres. Special tracks are specified in the field for the movement of tractors and other farm machinery to avoid their random movement. It is more relevant in the case of row crops, nurseries and orchards.

3. Soil Tilth

Discussion on tillage shall remain incomplete if no description is made about the soil tilth, as tillage is done to obtain a desirable soil tilth. Tillage is the cause and tilth is the effect.

Soil tilth is generally used to describe the fitness of soil to plant growth and yield. The Soil Science Society of America defined soil tilth as "the physical condition of soil as related to its ease of tillage, fitness as a seedbed, and its impedance to seedling emergence and root penetration".

The significance of soil tilth in crop production was known to the farmers long ago. The first recorded evidence about soil tilth was given by Fitzherbert in his book on '*Boke of Husbandry*' (1523), as reported by Keen (1931). Fream (1890) narrated a childhood story to explain the concept of soil tilth. An old man while on death bed called his sons and told them that if they would diligently dig the garden, they would find hidden treasures. The sons dug the garden deeper than the garden had ever been dug before, but alas! they found no treasures. But, the subsequent harvest from the well-dug garden was so large that the sons could understand the meaning of their father's saying. They gradually realized that the words of their father were the words of wisdom. Fream (1890) also illustrated the importance of soil tilth through a 17th Century quote, attributed to Thomas Tusser, that "*Good tilth brings seeds, ill tiltures weeds*".

Soil tilth is a multifaceted soil characteristic reflecting a dynamic combination of physical properties, including texture, structure, permeability and consistence that can be sustained, improved or destroyed by many chemical and biological processes depending on the way the soil is managed. Several studies have been conducted and published on soil tilth (Karlen *et al.*, 1990), yet soil tilth remains an elusive parameter. Defining the optimal physical state of soil for crop growth is very difficult, and determining the most feasible means to achieve this is still more difficult.

A good soil tilth must meet biological, agronomical and engineering objectives. A soil with good tilth, in general, should have the following elements (Slipher, 1932; Yoder, 1937):

- Offers minimum resistance to root penetration
- Permits free intake and moderates retention of water (rainfall and irrigation)
- Provides an optimal soil-air supply with moderate gaseous exchange between soil and atmosphere
- Holds to a minimum the competition between air and water for occupancy of the pore space volume
- Provides maximum resistance to erosion
- Facilitates the placement and coverage of green manures and organic residues
- Promotes microbial activity, and
- Provides stable traction for farm implements.

The following methods help in improving/sustaining soil tilth:

- Adoption of reduced tillage for crop production
- Management of crop residues
- Use of environmentally acceptable drainage programmes
- Implementation of nutrient management practices that maintain favourable fertility and soil pH

- Stimulation of soil microbial and faunal populations
- Use of animal manures, green manures, cover crops or rotations involving grasses, forages or other crops with fibrous root system, and
- Organic farming.

4. Objectives of Tillage

The objectives of tillage and tillage operations vary with soil, crop and climatic conditions, and are also influenced by the socio-economic status of the farming community. The basic objective of tillage is to obtain good soil tilth for crop growth and yield. Tillage benefits crop production primarily by improving soil environment (mainly the air-water relations in the seed bed), controlling weeds and reducing mechanical impendence to root growth.

Accordingly, there are three primary aims of tillage:

(i) preparation of seed bed with optimum tilth,
(ii) weed control, and
(iii) reduction of mechanical impedance to root growth.

Other objectives of tillage may include, soil incorporation of organic/inorganic matter (crop residues, manures, fertilizers, amendments), soil moisture conservation, modification of infiltration characteristics, land shaping for irrigation, terracing on sloping lands, compaction of loose soils, breaking hard pans in root zone, breaking surface soil crust, control of insect, pests and disease incidence, etc. The overall objective of tillage, however, is to minimize erosion losses by wind and water, maximize soil carbon sequestration, and maximize crop productivity on sustainable basis.

The different objectives of tillage are explained below:

4.1. Seedbed Preparation

As stated above, the basic objective of tillage is to prepare seedbed with soil tilth optimum for plant growth and yield. Tillage improves soil environment by imparting desirable soil-air-water relations in the seedbed. Seedbed with proper tilth has a bearing on planting, germination and water and air exchange. Tillage is often performed to loosen, granulate or otherwise condition soils. Soil may be cut, loosened, or broken-up to facilitate placement of seed in soil at the desired depth, and improve water and air movement through soil. Soil may be loosened to promote drying or reduce mechanical strength of soil mass.

Proper timing with respect to soil water content is very essential to obtain good soil tilth. Tillage operations performed in dry moisture range consume higher energy and lead to clod formation, while those in the wet range, destroy soil aggregation and cause soil compaction. As a thumb rule, for proper soil aggregation, tillage should be performed when soil is in the friable consistency.

It may be noted that improvement in soil tilth is the short-range effect of tillage. In the long-run, especially without proper organic residue management, tillage causes deterioration of soil tilth by increasing rate of oxidation of organic matter. Use of heavy tractors and machinery for tillage in mechanized farms generally leads to soil/sub-soil compaction, thus, destroying soil tilth.

4.2. Weed Control

Tillage is frequently performed to control weeds during crop production. Tillage helps in controlling weeds in many ways. It destroys the growing weeds by uprooting, smothering, desiccating or decomposing them. Tillage improves the effectiveness of herbicides. It brings out the regenerative organs of a plant (seeds, stolens, etc.) to the surface where they may dry out and lose their viability. The exposed seeds of weeds may also be picked up by birds. Tillage also reduces or eliminates the weed competition by way of enhancing the crop growth. Gajri *et al.* (1999) reported that disc plough (soil inversion) and disc harrow were more effective for weed control than strip-till system.

The advent of chemical herbicides about five decades ago appeared to reduce the significance of tillage as a method of weed control. However, due to health hazards associated with the use of chemical herbicides (application hazards, residual effects, environmental hazards) and consumers becoming more and more health conscious, and increasing cost of herbicides, the interest now again is shifting towards weed-controlling aspects of tillage.

4.3. Incorporation of Materials in Soil

Several organic and inorganic materials, such as crop residues, manures, fertilizers, soil amendments, etc. are incorporated and/or mixed with soil in tillage operations. Special implements have been designed to invert soil for covering materials lying on soil surface.

4.4. Soil Mixing

Sometimes, tillage operations are performed to mix soil layers to enhance soil drying, or to mix moisture throughout the depth of soil, or to distribute amendments such as soil conditioners, fertilizers, and soil-stabilizing materials. In special cases, several layers of soil may be mixed to form more desirable soil texture.

4.5. Land Shaping

Tillage is also performed for changing configuration of surface soil for planting (ridge-furrow system or raised/sunken bed system), irrigation, drainage and landscaping, etc. Ridge-furrow planting is recommended in rainfed areas for *in-situ* moisture conservation. Under the excessive water conditions (marshy lands), raised-bed planting helps in improving the air-water relations of soil; thus, benefitting crop growth and yield.

4.6. Segregation

Tillage may be used to move soil or other materials from one layer of soil to another such as when deep tillage is used to create clods and to move these clods to the surface to minimize wind erosion. Segregation may also involve the removal of materials from the soil. Rock picking and root harvesting are some of the examples.

4.7. Soil Compaction

Soil compaction is required to decrease infiltration and improve water retention in loose (sandy) soils. It is also used for decreasing percolation losses in water channels or water reservoirs. The soil compaction may be achieved by using cement/metal rollers or even road rollers (Sharma *et al.*, 1995). Soil compaction benefits crop production in sandy soils.

4.8. Soil Conservation

On sloping landscapes in high rainfall areas, cultivated soils are exposed to excessive water erosion. It is recommended to go for terracing if the soil slope exceeds 7%. Tillage (cut-fill) operations are performed for terracing such lands to reduce soil erosion hazards. Use of organic mulches is another way of soil conservation in hills. Special tillage practices are required that help in planting seeds and at the same time maintain mulches at the soil surface.

5. History of Tillage

Mother Nature is a no-till farmer since it never leaves the earth bare of vegetation provided basic ingredients of soil, sunlight and moisture, are not limiting. Every plant species has evolved a method of reproducing itself in an environment suited to its needs with no aid from man in the form of tillage. For thousands of years, man used only his hands and possibly a sharpened stick to dig a hole or cut a slot in the soil, bury a seed and cover it with earth.

Slowly, man started to realize that tilling the land produced more crop yield than without tillage. It probably happened as fertilizers were not known and tillage increased mineralization of organic nitrogen. For centuries, farmers kept on tilling their lands with primitive ploughs. The eighteenth century saw the introduction of more specialized metal tools for performing tillage operations, like disks, harrows and cultivators. Thomas Jefferson (1784) gave a mathematical formula for a moldboard plough for complete turning over of the soil prior to planting. In 1796, Charles Newbold patented a cast iron plough. John Deere (1837) introduced manufactured steel plough. The early horse- or oxen-drawn plough was a boon to farmers of the time. The use of moldboard plough spread westward in the United States to break the virgin prairie sod in the 19th century. The horse-drawn straddle row cultivator was another notable tillage development of the 19th century. Development of steam engine and later, gasoline and diesel engines, led the farmers to turn from horses to horse power in the early 20th century. The tillage tools became larger and much sophisticated and more productive.

In early-1930s, the United States experienced economic depression. Simultaneously, millions of hectares of land, which formerly was protected by natural grasses, was now exposed to the fury of weather — thanks to the cultivation by moldboard ploughs. The 1930s experienced the country's first environmental crises, when tonnes and tonnes of soil was blown into the atmosphere by wind erosion. It was observed that although there were several other factors responsible for the dust storm, excessive tillage by moldboard plough and the lack of soil conservation techniques had compounded the problem.

Edward H. Faulkner has criticized moldboard ploughing in his 1934 book, *Plowman's Folly*. Faulkner received very little scientific support at that time. It was in the late-1940s and early-1950s that scientists began to experiment with a revolutionary alternative to tillage. Introduction of herbicides gave farmers for the first time a practical tool other than plough, disc, cultivator and hand hoeing to control weeds and to prepare environment-friendly seedbed. The introduction of Paraquat herbicide became the chemical foundation for no-tillage farming, which led the scientists to think that crop could directly be drilled into the soil with minimum of tillage or no-tillage at all, when weeds are controlled chemically. However, increased health hazards associated with the use of chemicals in agriculture, including herbicides, has once again shifted focus on tillage as a weed control practice.

6. Different Tillage Systems

Depending upon the kind, amount and sequence of soil disturbance during seedbed preparation, tillage systems may be grouped into two classes:

(a) Conventional tillage, and
(b) Conservation tillage.

Tillage terminology can be better explained with a triangle given in Figure 1. At the broad base is the conventional tillage, indicating a full tillage package. As the triangle narrows down, tillage operations are reduced, and it leads to reduced or minimum tillage. At the upper point of triangle, tillage is eliminated completely, indicating no-tillage.

6.1. Conventional Tillage

It refers to a tillage system that has been developed traditionally and followed by the farming community depending on the type of soil, climate, crop and socio-economic status. It includes a combination of primary and secondary tillage operations normally performed to grow a particular crop in a given geographical area. It is also known as 'clean tillage'. SSSA (1987) defines conventional tillage as "a process of ploughing and cultivation which incorporates all residues and prevents growth of all vegetations, except the particular crop desired during the growing season".

Conventional tillage is a system that involves maximum tillage operations in two steps: primary tillage (cutting and loosening of soil, and incorporating crop residues/stubbles into the soil), and secondary tillage operations (pulverization and compaction of pulverized soil, sowing, weed controlling, intercultural opera-

Figure 1. Flow chart showing three basic types of tillage systems

tions, etc.). Although definition of conventional tillage (clean tillage) by SSSA (1987) emphasizes the soil incorporation of all crop residues, conventional tillage also includes systems in which crop resides are either removed as cattle feed, fuel or fencing material, etc. or heaped in the field and burnt before tilling the land.

The resource-poor farmers in developing countries traditionally use animal-drawn ploughs consisting of an iron blade attached to a wooden log (country-side ploughs) for opening the land, followed by wooden planks, with or without iron nails, etc., moved repeatedly on the soil surface to break clods, pulverize soil, level field and slightly compact the soil surface. Seed is sown by broadcast method or in rows behind the plough manually or by using animal-drawn seed drills. Light planking follows the sowing of seed. Because of limited draught power, animal-drawn ploughs cause very little soil inversion. In the mechanized farming, as in developed countries, tractor-drawn heavy ploughs, such as moldboard or disc ploughs, are used for cutting, loosening and inverting the soil. The large clods are broken and soil is pulverized by using disc harrows or rotary tillers. In the conventional tillage system, the soil surface at the time of seeding is practically bare.

The advantages of conventional tillage are:

- No or minimum interference of crop residues with sowing
- Incorporation of fertilizers and other amendments is facilitated
- Water infiltration is improved
- Increased roughness at soil surface enhances temporary surfacewater storage, thereby decreasing surface runoff and increasing infiltration
- Decreases mechanical impedance in the seedbed and root zone
- Incorporation of organic residues favours microbial activity in soil
- Controls proliferation of insect, pests and diseases by burying residues in soil

Following are the disadvantages of conventional tillage:

- ◆ It makes soils loose and bare, and thus increases their vulnerability to erosion by water and wind
- ◆ Repeated conventional tillage results in decline in soil organic matter content
- ◆ Destroys soil aggregates and deteriorates soil structure
- ◆ It is cost-intensive

6.2. Conservation Tillage

Conservation tillage in a broad sense is any tillage system that is less intensive than conventional tillage. In this tillage system, either land preparation and sowing operations are combined in one operation or tillage operations for land preparation are eliminated altogether. The most important component of conservation tillage is the retention of crop residues on soil surface.

The Conservation Technology Information Centre (CTIC, 1993) in Indiana, USA, has defined conservation tillage as "tillage and planting system in which at least 30% of the soil surface is covered by plant residue after planting to reduce erosion by water. Where soil erosion by wind is the primary concern, at least 1000 lbs of flat small grain residue per acre (1120 kg/ha) should be put on the surface during the critical erosion period". Conservation tillage is an umbrella-term and covers all tillage systems that are less intensive than the conventional tillage. The CTIC has sub-divided the conservation tillage into four systems:

(a) no-tillage (also called no-till, zero tillage, slot planting, sod planting, ecofallow, chemical fallow, direct drilling),

(b) reduced tillage,

(c) stubble mulch tillage, and

(d) ridge tillage (Parr *et al.*, 1990).

Some researchers are of the view that reduced/minimum tillage or zero/no-till system should be considered under conservation tillage only if it does not destroy or burry all plant residues, and leaves some residues as soil cover.

The success or failure of conservation tillage depends on the use of herbicides, crop residue management and efficiency of planting equipments to place seed in soil below the residues.

Conservation tillage has the following advantages:

- Protects land against erosion
- Increases soil organic matter content
- Increases activity of soil flora and fauna
- May control some plant diseases by encouraging activity of organisms in soil that are antagonistic to pathogens, or may modify soil environment to favour multiplication of selected organisms
- Improves soil structure and associated soil physical properties
- Lowers cost of cultivation

Conservation tillage may have the following limitations:

- The residues left on soil surface may lead to phytotoxicity or allelopathy on the subsequent crops.
- It may increase the incidence of plant diseases by providing crop residues left on soil surface as a source of overwintering disease inoculums, thereby increasing overwintering survival rate of pathogen propagules.
- Sowing of seed under crop residues requires special planting equipments.

6.2.1. No-tillage

It is a method of planting crops that involves no seedbed preparation, other than the opening of soil for placing seed at the intended depth (SCSA, 1982). In this tillage system, crop residues are largely retained on soil surface by avoiding their removal, burning or animal grazing, etc. The seed is placed in soil by cutting small slits or by punching holes in the soil. In extreme cases, seed is broadcast on untilled soil, and the germination is assured by covering seed with residues (protected against bird damage) and by regulating optimum soil moisture regime through irrigation management. This practice is being followed in the rice growing areas of Nepal, where wheat is broadcast-seeded in wet/moist soil in the stubbles of recently harvested rice crop (called surface seeding of wheat). In the north-western parts of India, planting of linseed immediately after rice harvest in untilled-wet soil is a common practice. The weeds are controlled by use of herbicides or maybe manually.

One may argue that if crops can be grown without tillage why farmers traditionally have been performing different tillage operations to grow crops. Different tillage operations are performed by farmers basically for two reasons: weed control and optimum environment for seed to germinate and grow. In no-till system, herbicides replace tillage for weed control, and favourable environment to seeds is provided by the decaying residues on soil surface, which act as mulch to conserve soil and moisture and improve soil productivity (chemical, physical and biological). Thus, basic needs of plants are met without tillage.

No-tillage system has the following advantages:

- Saving in time, energy and labour and cost effective
- Favours timely sowing of crops; particularly useful in rice-wheat system, where wheat yields are low due to its late planting because of large turn-around time between rice and wheat caused by slow drying of wet rice soils
- Increase in soil organic matter content
- Moisture conservation, by decreasing evaporation losses and increasing water retention properties of soil
- Soil conservation, by reducing water and wind erosion
- Improvement in soil structure and associated soil physical properties (aeration, water retention, infiltration, drainage)
- Favours prolific growth of soil flora and fauna beneath the litter. Increased earthworm and arthropod activities have been observed under no-till system. They improve soil quality in the long-run by improving soil structure and other soil properties.
- In hot climates, residues with no-tillage system keep the soil temperatures cooler than under conventional/clean-tillage system by increasing albedo. Crop residues also reduce heat fluxes into soils as the thermal conductivity is lower of residues than of a mineral soil. It benefits crops by avoiding high temperatures that may be detrimental to root growth and soil faunal activities.

No-till may have the following limitations:

- May increase incidence of crop diseases and pests by providing excellent substrate to inoculums and overwintering source in the form of residues.
- In cold climates, surface residues under no-till system may delay soil warming, resulting in delayed planting, germination, emergence and crop establishment.
- Reduction in the surface flow of water and the reduced evaporation rates under crop residues may increase excess water conditions and problems related to poorly-drained soils. Enhanced nitrate leaching and denitrification have been observed in wet soils under no-till system (Rice and Smith, 1982; Blevins *et al.*, 1985).
- No-till without adequate surface cover (or no crop residues) decreases infiltration and increases surface flow of water, especially in hard-setting soils or sandy soils in arid/semi-arid regions due to formation of surface crusts after rainstorms.
- In many developing countries, particularly in tropics and semi-arid tropics, major constraint to adoption of no-till system is non-availability of adequate surface residues, as there is increased demand of crop residues as cattle feed, fuel or fencing material. Under such circumstances, it is emphasized that the farmers may remove crop residues selectively, grow fodder for animals, put waste lands under forage cultivation, adopt alley cropping or may control the number of livestock.
- Requires special equipments for sowing of seeds under crop residues.

6.2.2. Reduced Tillage

As the name indicates, it is a tillage system less intensive than the conventional tillage. Attempts are made to minimize tillage operations by elimination of one or more tillage operations from a conventional tillage programme. Only those tillage operations are performed that are absolutely necessary for crop production under a given set of soil, crop and climatic conditions. The primary and secondary tillage operations are generally combined together. Land preparation and seeding is completed in one operation. Ploughing is normally eliminated, but the total field surface is still worked by tillage equipment. Crop residues may or may not be retained on the soil surface. The crop residues are retained on soil surface for as long as possible if the objective is to conserve soil and soil moisture during the rainy season. Under irrigated conditions, reduced tillage system may be practised after removing residues from the soil surface. The practice has largely been adopted in alluvial soils of the Indo-Gangetic Plains, where wheat is planted with minimum tillage operations in lean fields. Several variations of minimum tillage system are in use globally, varying in degree from almost no tillage to nearly full conventional tillage (Unger, 1984).

The major advantages of minimum tillage over conventional tillage may include:

- Better soil and water conservation
- Time, energy, labour and cost effective

Some of the disadvantages associated with minimum tillage system are:

- Lower soil temperature which may delay germination, emergence and crop establishment
- Poor seed placement in the presence of residues on surface
- Problems of diseases and pests

6.2.3. Stubble Mulch Tillage

Any tillage operation that retains crop residues on the soil surface may be classified under mulch tillage. It may be defined as "tilling the soil so that plant residues or other materials are retained to cover soil surface" (SCSA, 1982). It is also known as mulch farming, mulch tillage or ploughless farming.

In the stubble mulch tillage system, the soil surface is undercut to control weeds and prepare the seedbed in such a way that most of the crop residues are retained on the soil surface. Generally, subsurface tillage implements such as chisel ploughs are used for the purpose. The presence of residues may also enhance infiltration and reduce evaporation. Stubble mulch tillage was basically designed to control wind erosion, but it controls water erosion as well.

Advantages of stubble mulch tillage are:

- Wind and water erosion control
- Cost effectiveness

The main disadvantages of this system are:

- Difficulty in performing tillage operations in the presence of large amounts of residues
- Difficulties in tilling due to higher soil water content as a result of reduced evaporation and poor weed control when precipitation occurs soon after tillage

6.2.4. *Ridge Tillage*

Ridge tillage is "a method of land preparation whereby the top soil is scraped and concentrated in a defined region to deliberately raise the seedbed above the natural terrain" (Lal, 1990). In this system, crops may be grown on ridges or in furrows depending on the moisture status of soil. In poorly-drained, wet soils, crops are grown on ridges with one or more rows per ridge. In dry and sub-humid areas, crops are grown in furrows, which favour *in-situ* moisture conservation by retaining rain water and decreasing evaporation losses.

Ridge tillage is adapted to a wide range of soils, crops, rainfall regimes, ecological environment, socio-economic and cultural conditions (Lal, 1990). On sloping landscapes (with usually less than 7% slope), ridge planting is particularly effective in soil and water conservation. The excess surface water can easily and safely be disposed off on sloping lands via furrows. However, ridge-furrows need to be properly designed and constructed, otherwise surface water may concentrate at spots above the ridges and cause their breaching, resulting in severe erosion damage.

The concept of raised-bed planting, another variation of ridge tillage, was recently experimented by Rice-Wheat Consortium (CIMMYT, IRRI, NARS and IARCs) in the Indo-Gangetic Plains cultivating rice-wheat system. The objective was to economize on irrigation water in rice cultivation and increase water productivity in wheat cultivation. In this system, rice is planted on permanently maintained broad-beds (30-40 cm), 2-3 rows per bed. Irrigation is applied in furrows at the defined intervals. The furrows cover about 40% area in the plot. Since rice is sown on about 60% area of the plot, there is a saving of seed. The yield loss due to decrease in plant population is expected to be compensated by realizing the border effect on raised beds. The studies have revealed that growing rice on permanently raised beds, under aerobic conditions, may save 35-40% of irrigation water, without showing a significant yield decline. The saving in water could be more if the technology is combined with (LASER) field levelling and better farm layouts. Coupled with water saving, bed planting shows promise for better N management. The technique, however, needs thorough investigation to answer several questions on irrigation scheduling, weed control, suitability of rice cultivars for bed planting, salt dynamics in normal and salt-affected soils, etc.

Significant advantages of ridge-tillage system are:

- Effective soil and water conservation technique on sloping lands
- Effective in conserving water in root zone in semi-arid to sub-humid regions; in dryland farming helps to conserve soil moisture
- On shallow soils, cultivating crops on ridges increase the effective rooting volume
- On poorly-drained soils, ridge tillage improves soil aeration and makes possible the successful cultivation of upland crops

The limitations of ridge tillage include:

- Requirement of special equipments for establishing ridges
- If ridges are maintained permanently for few cropping seasons, especially under low rainfall conditions, there is danger of increase in salt content in ridges due to upward flux of salts along with water
- Involves additional cost if ridges are to be constructed every cropping season

6.3. *Special Tillage Systems*

6.3.1. *Wet Tillage*

Wet tillage is performed during cultivation of aquatic/semi-aquatic plants. Rice is the best example. For transplanting rice, the most common rice culture, land is repeatedly tilled under wet conditions to convert soil into a soft mud, a process called puddling. Buehrer and Rose (1943) defined puddling as "the destruction of aggregated condition of the soil by mechanical manipulation within a narrow range of moisture content above and below the field capacity, so that soil aggregates lose their identity and the soil is converted into a structurally more or less homogeneous mass of ultimate particles". According to Bodman and Rubin (1948), "puddling is the mechanical reduction in the apparent specific volume of soil".

The objective of puddling is to destroy macro-pores so as to minimize percolation losses from the rice fields. Puddled soil is a two phase system: the solid phase and liquid phase; and the gaseous phase is practically eliminated (some air may be present occluded in blocked soil pores). During puddling, at least three processes occur that reduce percolation, viz. (i) destruction of soil aggregates and elimination of macro-pores, (ii) clogging of soil pores by clay particles, and (iii) sub-soil (below puddled soil layer) compaction caused by implements used during puddling. Maximum destruction of soil aggregates occurs when tillage is performed at moisture contents between field capacity and saturation. At this moisture content (a) cohesion within soil aggregates is minimal, which helps in increasing the shear planes in soil aggregates, and (b) cohesion between the aggregates is maximum, which restricts the movement of aggregates along each other and along the implement, thereby, transferring maximum energy for the destruction of aggregates. The clay particles suspended in the floodwater during puddling settle down with time, blocking the soil pores further to reduce percolation. Thus, best puddling occurs when tillage is performed at moisture contents at or slightly below saturation, followed by tillage under submerged conditions.

The ease and degree of puddling depend on soil wetness, soil type, nature of implement used and cultural practices. Saturation/near-saturation moisture content, less aggregate stability, more silt and clay, more swelling-type clay minerals, and rotary implements give a better soil puddle.

Puddling causes a series of physical, chemical and biological changes in soil (Sharma and De Datta, 1986). The soil effects of puddling may be classified into two categories: short-term and long-term effects.

A. Short-term Effects of Puddling

- Destruction of soil structure due to destruction of soil aggregates
- Elimination of macro-porosity
- Decreases in bulk density; bulk density increases with time, and the puddled soil when dry has higher bulk density
- Soil strength follows the trend in bulk density; puddled soil has very low soil strength/mechanical impedance, which increases as the puddled soil dries
- Reduction in saturated hydraulic conductivity and percolation rates
- Increase in soil water retention
- Decrease in evaporation losses; puddled soils take more time to dry than the aggregated soils
- Puddling and water submergence increase the volumetric heat capacity, decrease the thermal conductivity, thermal diffusivity and net heat flux into soil, and increase the time lag for peak temperatures between air and soil. In one study, the time lag between temperature peaks at soil surface (flood-water) and at 30-cm depth of puddled soil was about 11 hours as compared to less than 7 hours in an upland soil (Singh and Sandhu, 1979; Sharma and De Datta, 1991).
- Reduction in gaseous exchange between soil and outer atmosphere
- Decrease in soil redox potential, resulting in increased solubility of several plant essential elements in soil (P, Fe, Mn, etc.), and increase in the production and accumulation of (phyto-toxic) organic acids
- The pH of puddled and submerged soils tends towards neutrality

B. Long-term Effects of Puddling

- Development of sub-surface hard pans due to physical compaction caused by tillage implements and human beings, and illuviation/precipitation of Fe and Mn in the subsurface layers

Puddling is favourable for the rice crop as it eases transplanting, reduces percolation of water and nutrients, increases water stagnation, controls weeds and increases availability of essential plant nutrients in soil. However, puddling is not favourable to upland crops that follow wetland rice, as puddling destroys soil structure, increases draft power requirement to pulverize puddled dry soils, impairs drainage, creates aeration problems and increases mechanical impedance to root growth. Hence, if soils are not properly managed, yields of upland crops in a wetland (lowland)-upland culture are lower than in upland-upland culture.

6.3.2. Soil Compaction

Soil compaction refers to the increase in density of a soil by a dynamic load. Under natural conditions in certain soil types, hard pans or compact horizons are recognized in a soil profile. These hard layers of soil may be genetically induced or developed due to improper tillage practices. In recent years with the increase in mechanization, uses of heavy machines have created a serious problem of artificial soil compaction. The compact soil layer at the bottom of the zone of ploughing is termed as the 'plough sole'. Since all agricultural implements may create some kind of sole under favourable soil conditions, these have been named on the implement used, i.e. the plough soles, sub-soiler soles, disc harrow soles, traffic soles, etc. The last one is caused by the tractor tyres and vehicular traffic. The artificial compaction is achieved by rolling, tamping, vibrating or compressing the soil by any means. The degree and depth of compaction depend on soil texture, antecedent soil moisture, load applied at soil surface., i.e. compactive energy and the amount of soil manipulation. Sandy soils are difficult to compact than clayey soils. Under the same load, the depth of compaction is more in sandy than clayey soils. Dynamic load is more effective in compacting soils than static load. Best moisture content for achieving maximum compaction under a given load is at or slightly below the field capacity (called 'proctor moisture content'). The state of compaction affects the soil-air-water-temperature relationships profoundly and determines the physical, chemical and biological properties of soil.

In the compaction process, soil particles are packed together in a closer state of contact indicated by a change in bulk density, porosity, etc. Under a dynamic load sufficient to increase the density of soil, the following changes may occur:

(a) Compression of soil solids
(b) Compression of liquid and gas within the soil pore spaces
(c) Change in liquid and gas contents in the soil pore spaces
(d) Re-arrangement soil solids.

Soil compaction generally interferes with root development of crops and is, therefore, not a desirable soil property. However, under certain conditions, soil compaction is desirable. Soils in the arid and semi-arid regions are generally coarse-textured, having high permeability, low water-retention and low fertility. Same is true with sandy soils of other eco-systems. The crops suffer in these soils mainly from the moisture stress. Compaction of such soils has shown positive effect on the crop yields. Soil compaction compresses larger pores into smaller ones, favouring water retention over drainage.

The benefits of compaction of sandy soils may be summarized as follows:

- Soil compaction reduces
  - o Nutrient and water losses through leaching
  - o Time taken for each irrigation
  - o Depth of water required in each irrigation
  - o Evaporation losses, and
  - o Soil erosion
- Soil compaction increases
  - ◆ Soil moisture retention
  - ◆ Effectiveness of fertilizer
  - ◆ Nutrient uptake by crops
  - ◆ Uniformity in seed germination, and
  - ◆ Soil productivity
- Single time compaction lasts for two to three crops (Majumdar, 1994).

Soils may be compacted by moving manual, animal-drawn or tractor-drawn metal or cement rollers on soil surface. Majumdar (1994) used a 500-kg or 1000-kg iron roller for soil compaction. A loamy sand soil with 3-6 passings of iron roller at proctor moisture content could be compacted to more than 45 cm depth. Heavy machines like road roller can also be used to compact soils. In one study, nine passes of 12-tonne road roller (Dynapac CA25) at about 12% soil moisture content compacted a loamy sand soil (75% sand content) up to a depth of about 60 cm, thereby decreasing percolation rate from 11.8 mm/d in control to 1.4 mm/d and increasing rice yield from 0.9-2.4 Mg/ha to 2.1-4.0 Mg/ha (Sharma *et al.*, 1995).

Majumdar (1994) observed increase in bulk density and soil water retention, and decrease in saturated hydraulic conductivity of a sandy loam soil with 6 passings of 1000-kg iron roller at about proctor moisture content (Table 1). The 8 passings of the roller decreased infiltration rate of soil from 33.2-37.2 cm/h to 20.0-23.1 cm/h, and increased yield of various crops (pearl-millet, guar, cow-pea, wheat, barley, raya, taramira, fenugreek, tomato and chilli) by about 2-64% (Majumdar, 1994; Majumdar *et al.*, 2000). The increase in yield was attributed to the favourable root growth and nutrient availability in compacted soil. In another experiment, soil compaction increased water expense efficiency of mung-bean and pearl millet by about 15-31% (Yadav and Jakhar, 2001; Yadav and Yadav, 2004). Yadav and Vyas (2006) have reported about 14% increase in N-use efficiency and 21% increase in apparent nitrogen recovery by wheat in a loamy sand soil due to soil compaction.

7. Tillage and Edaphic Environment

Tillage influences soil organic matter content and soil structure, and all other associated soil properties, including bulk density, soil strength, porosity, pore-size distribution, infiltration, hydraulic conductivity, water retention, drainage, evaporation, gaseous exchange, soil thermal properties, soil microbial activities, etc. Tillage may loosen or compact the soil, and may enhance or retard the rate of decomposition of soil organic matter. Tillage systems also influence hydrological soil properties, such as surface runoff and soil erosion, through their influence on soil structure. The modifications in hydraulic properties of soil significantly influence nutrient availability, leaching losses, groundwater contamination, plant growth and crop productivity.

7.1. Soil Organic Matter Content

Continuous ploughing and other tillage operations stimulate organic matter losses by increasing aeration, changing temperature and

Table 1. Effect of compaction on the average bulk density, saturated hydraulic conductivity and moisture content of a loamy sand soil immediately after compaction

Soil properties	Depth (cm)	Treatmant	
		Control	Compaction*
Bulk density (Mg/m³)	0-15	1.48 ± 0.02	1.57 ± 0.02
	15-30	1.51 ± 0.02	1.63 ± 0.03
	30 -45	1.53 ± 0.01	1.59 ± 0.03
Saturated hydraulic conductivity (cm/h)	0-15	10.85 ± 0.02	8.16 ± 0.02
	15-30	9.20 ± 0.02	6.60 ± 0.02
	30-45	8.75 ± 0.01	6.80 ± 0.01
Soil moisture content (%)	0-15	8.11 ± 0.02	13.09 ± .02
	15-30	8.61 ± 0.02	13.79 ± .01
	30-45	8.96 ± 0.04	13.66 ± 0.01

*Compaction of loamy sand soil by 8 passes of 500-kg iron roller

Source: Majumdar (1994)

Table 2. A comparison of soil organic matter distribution after 10 years of no-till and conventional tillage in corn production

Soil depth (cm)	Soil organic matter content (%)		
	No-tillage	Conventional tillage	Untreated bluegrass sod
0-5	4.82	2.40	6.55
5-15	2.34	2.31	2.98
15-30	1.15	1.22	1.55

Source: Blevins (1986)

moisture conditions, and thus favouring microbial decomposition. In addition, soil aggregate disruption by tillage exposes once protected organic matter to decomposition. Hence, intensive conventional tillage system in the long-run may cause a decline in soil organic matter (SOM) content. Conservation tillage systems, with proper residue management, are expected to maintain or increase SOM by maintaining relatively slow decomposition rate of surface residues. Blevins (1986) observed significantly higher SOM under no-till system compared to conventional tillage system in operation for 10 years (Table 2).

7.2. *Bulk Density, Porosity and Mechanical Impedance*

The magnitude of change in bulk density of soil depends on its antecedent properties, time of measurement, depth and intensity of tillage operations. Soil loosening decreases and compaction increases the bulk density. The studies on the impact of tillage systems on soil bulk density have shown variable results. For example, Lal (1976) reported that no-tillage with mulch improved the soil porosity, soil structure and water transmission in an alfisol. Sharma and Acharya (2000) have observed lower bulk density at maize harvest under conservation tillage (no-till + mulch) than conventional tillage, and the effect was associated with higher soil organic carbon (SOC) under the former than the latter (Table 3). Blevins *et al.* (1985), on the other hand, did not observe any change in bulk density of a medium-textured soil under no-till system, Gantzer and Blake (1978) reported increase in bulk density of a fine-textured soil with no-tillage compared to conventional tillage. The soil strength generally follows the bulk density trends.

The total porosity of soil would change with tillage depending on the change in bulk density. Tillage operations increase macro-porosity, while the compaction increases the micro-porosity of the soil. Kar (1994) has reported significant increase in the proportion of non-capillary pores with conventional tillage compared to zero-tillage in a lateritic sandy loam soil (Table 4). In contrast to conventional tillage, conservation tillage system results in a more continuous pore system because of increase in earthworm activity, old root channels and vertical cracks between peds.

Table 3. Soil organic carbon (SOC) and bulk density at maize harvest under different tillage treatments

Treatment	SOC (g/kg soil)	Bulk density (Mg/m³)
Conventional tillage*	9.0	1.37
Zero tillage + Residue mulch**	12.3	1.31
LSD (0.05)	1.1	0.04

Notes: *Fine seed-bed prepared by digging top 12-15 cm soil layer manually
**Furrows opened in mulched plots with hand plow for sowing
Lantana was used as mulch at about 15 Mg/ha fresh biomass
These data were recorded after four maize-wheat cropping cycles
Source: Adopted from Sharma and Acharya (2000)

Table 4. Effect of tillage treatments on proportion of non-capillary pores in tilled layer of a lateritic sandy loam soil

Treatment	Non-capillary porosity (%)
No tillage	15.8
Moldboard plough + Disk harrow	42.9
Disc harrow	37.1
LSD (0.05)	0.9

Source: Kar (1994)

7.3. Soil Hydraulic Properties

A change in the soil pore system (total porosity, pore geometry, continuity of pores), and maintenance of surface residues and associated soil faunal activities (e.g., earthworms) due to tillage system have significant effect on all soil hydraulic properties, viz. infiltration, hydraulic conductivity, water retention, drainage and evaporation. Tillage affects the soil-water status as well as capacity of plants to utilize it. Tillage also influences the soil wetness through weed control.

Conservation tillage is an excellent technique of *in-situ* soil moisture conservation in rainfed areas (Sharma and Acharya, 2000). Creation of dust mulch through repeated tillage is a common practice of moisture conservation in rainfed farms (Sharma *et al.*, 1990). However, zero tillage with mulch has been found superior to dust mulch in soil moisture conservation in a sandy loam soil; best results have been observed with conventional tillage with mulch (Table 5).

7.4. Soil Aeration

The impact of tillage system on soil aeration depends on the tillage intensity, residue management, soil aggregate stability and soil moisture regime. Tillage operations generally increase total soil porosity. The conventional tillage supports a relatively higher proportion of macro-pores in the tilled layer, which is able to restore adequate water-free porosity more rapidly after heavy rains or irrigation by rapidly draining the excess water. A decrease in the volume-water fraction increases the volume of air-filled pores, thus, enhancing soil aeration. In fact, the effect of tillage on aeration of soil is directly linked to its effect on enhancing the infiltration rates, which considerably reduces the time for which the surface remains sealed to air exchange. Tillage is likely to affect the oxygen diffusion rate (ODR) in soil.

In soils with stable structure, conventional tillage generally supports higher aeration status than conservation-tillage system; more so under the excess water conditions (say close proximity to water table). In one study at IRRI, Philippines, aeration porosity was significantly higher with conventional than with no-till system in two different soils (Table 6).

Table 5. Effect of different tillage systems on soil moisture conservation and grain yield of wheat

Treatment	Soil moisture at sowing of wheat (mm)		Soil moisture conservation over control (mm)	Wheat grain yield (Mg/ha)
	0-7.5 cm soil	0-45 cm soil	0-45 cm soil	
Control	4.8	33.8	-	1.22
Dust mulch*	8.3	64.6	30.8	1.53
No-tillage + mulch**	9.3	75.2	41.4	1.46
Conventional tillage + mulch	14.4	101.5	67.7	1.88
LSD (0.05)	-	-	-	0.23

Notes: *Soil was repeatedly ploughed with animal-drawn plough, followed by planking at about weekly interval, starting from maize harvest in the last week of September until sowing of wheat in mid-November (local farmers' practice)

**Dry leaves of sal (*Shorea robusta*) were used as mulch at about 10 Mg/ha

Tillage and mulch treatments were imposed after maize harvest in the last week of September

Source: Adapted from Sharma *et al.* (1990)

Table 6. Effect of tillage systems on air-filled porosity of soil at 4 weeks after planting (4 WAP) and at harvest of summer mungbean

Tillage treatments	Air-filled porosity (%)			
	Clay loam		Sandy loam	
	4 WAP	At harvest	4 WAP	At harvest
No-till	8	12	27	27
Minimum tillage	17	21	43	42
Conventional tillage	18	25	44	39
LSD (0.05)	4	7	5	3

Note: Clay loam soil had water table with 20 cm depth
Source: Sharma *et al.* (1988)

7.5. Soil Temperature

The changes in soil surface roughness and the presence of plant residue cover under a given tillage system influence the thermal regime of soils. A change in bulk density alters the specific heat capacity of a soil, primarily by changing the relative amount of mineral matter and water per unit volume of soil. Compaction increases thermal conductivity of the soil because of decreased porosity and increased contact between particles. Tillage systems, that maintain better soil cover with residues, lower the soil temperature by increasing albedo. In a field study, no-tillage maintained 1-2 °C higher maximum soil temperature at 10-cm soil depth than tillage with moldboard plough; the application of surface mulch maintained 2-8 °C higher maximum temperature in each case (Sharma, 1991). The presence of residue mulch at soil surface keeps the soil cooler, irrespective of tillage.

7.6. Chemical and Biological Properties

Tillage influences several chemical and biological properties of a soil by modifying its hydro-thermal regime and aeration status. The conventional tillage (plough-based) incorporates crop residues and distributes organic matter almost evenly throughout the ploughed layer. Conservation tillage does not completely incorporate the crop residues and results in accumulation of organic matter at the soil surface.

The effect of tillage on biological and chemical properties of a soil depends upon climate, soil type, quantity and quality (mainly C/N ratio) of crop residues and management practices. Ploughing of a soil often stimulates nitrogen mineralization. Conservation tillage system tends to immobilize the applied nitrogen. The phosphorus distribution under different tillage systems follows a similar pattern as depicted by the organic matter.

Tillage operations disturb soil and affect several changes in soil microbial population and their activities. The effect of tillage on soil microbes occurs largely through the changes it brings about in organic matter, moisture and thermal regime of soil. Since different tillage systems affect these factors differently, they have different effects on soil biology. The microbial population generally follows the trend of soil organic matter contents. Conservation tillage increases the heterotrophic bacteria and has higher activity of denitrifiers than the conventional tillage. There is a higher earthworm activity under conservation tillage than conventional tillage.

8. Tillage Effects on Crop Growth and Productivity

Tillage affects seed germination, emerging of seedlings, root growth, and crop yields as described below.

8.1. Seed Germination and Seedling Establishment

The basic requirements for seed germination and seedling emergence are that the seed should be placed at a uniform depth, in good contact with soil so that it may absorb water easily, yet the soil must be well aerated. The soil above the seed must remain sufficiently loose for the seedling to grow up through the soil. Further, the soil must maintain optimum temperature in the seed zone. As described earlier, tillage influences all these factors significantly.

The tillage system that leaves residues on the surface of soil may cause low soil temperature and may delay seed germination and seedling emergence in temperate regions. However, in tropics and sub-tropics, reduction in super-optimal temperature with residue may facilitate seed germination.

Tillage practices also play an important role in controlling the surface crust and in affecting the seedling emergence. Any tillage system that results in a stratified seedbed with finer aggregates in the seedbed covered by coarser aggregates on the surface helps in combating the adverse effects of crusting.

8.2. Root Growth

The access of crop to water and nutrients is directly related to the size of its root system. Tillage affects the depth and density of rooting by modifying the mechanical impedance, continuity, stability and size distribution of pores, air-water dynamics and thermal regime of soil (Prihar *et al.*, 2000). Roots grow into the soil pores or root channels (of decomposed roots of previous crops) which are bigger than the size of the root tips. Where such pores do not exist, the growing roots make their own path by displacing the soil particles. The role of tillage is to provide space for the growing roots in soil profile by reducing the soil strength and increasing the macro-porosity. Loosening of root-restricting hard soil layers by tillage promotes root growth into deeper soil layers. The effect of tillage on root growth is governed by the soil type, climate and other management practices. Deep tillage (sub-soiling to 40-cm depth) has been found to increase the depth and density of maize roots even in sandy soils (Arora *et al.*, 1991).

8.3. Crop Yields

For a tillage system to have a positive effect on growth and yield of crops, there must exist in soil some constraint(s) to crop growth and the tillage operations must be able to alleviate that constraint(s). If the soil has air-water relations and soil strength optimum for plant growth, the type of tillage may not affect much crop growth and yield.

Deep tillage to 40-cm depth without inversion with a chisel has been found beneficial in increasing crop yields of maize, wheat, mustard and sunflower in a coarse-textured soil that exhibited increase in soil strength on drying (Table 7). The magnitude of yield increase was more in summer than winter season, and was more in sand and loamy sand than in sandy loam soils.

Table 7. Effect of deep tillage (chiseling 40-cm deep, 35-40 cm apart) on crop yields in relation to soil type and season

Crop	Season	Soil	Crop yield (Mg/ha)	
			Conventional tillage	Deep tillage
Maize	Winter	Sand	3.2	4.9
		Loamy sand	4.6	5.0
	Summer	Sand	0.9	2.2
		Loamy sand	1.1	2.2
		Sandy loam	4.0	4.6
Wheat		Sand	2.8	3.3
		Loamy sand	4.9	5.3
		Sandy loam	5.2	5.4
Sunflower		Loamy sand	1.7	2.0
		Sandy loam	1.9	2.2

Source: Prihar *et al.* (2000)

In loose sands, compaction of soils increases micro-porosity and consequently, there is an increase in water and nutrient retention, leading to an increase in the yield of crops, as has been observed in pearl millets and other crops (Majumdar, 1994). Depending on the depth and density of sub-surface layer, subsoiling with chisel to 40 cm depth or profile modification increases the infiltration of water and water storage in the profile by breaking the restricting layer. These tillage practices also encourage deeper rooting and hence better mining of water by the crops, resulting in high yields.

Conservation tillage which retains the crop residues at or near the surface is most suited for erosion-prone soils in the arid and semi-arid regions for increasing crop yields. The maximum benefits occur on well-drained soils and in crops grown in rotation. In the rainfed areas where water stress conditions prevail, conservation tillage shall improve crop yields. In poorly-drained soils in humid areas or in soils with shallow water table, especially in temperate regions, conservation tillage generally gives lower yields than conventional tillage by keeping soils relatively cool and wet, which causes aeration stress.

In areas where there is a short window period for planting wheat after rice harvest, direct drilling may benefit wheat crop, provided it advances the seeding time in comparison with conventional tillage. If seeded at the same time, for equivalent yields, the direct drilled wheat crop requires more nitrogen and water input than by the conventional tillage.

It is thus seen that all types of tillage systems are not suited to all soils, crops and climatic conditions. Tillage requirements are highly crop- and site-specific. Therefore, the selection of type of tillage must consider the soil and climate conditions, and identify constraints that limit crop yield. There is a need to assess the soil/crop specific factors that limit the potential yields, which can be modified through tillage. The wrong choice of tillage system would not only lower the crop yields but would unnecessarily cause wastage of money and other resources.

8.4. Tillage Indices

One or more soil properties, alone or in combination, that guide the selection of the type of tillage in a given soil and climate and for a given crop is called a tillage index. Tillage indices help in deciding the type of tillage system that is required under a given set of conditions.

Lal (1985) has developed a rating system to assess tillage requirements for diverse soil conditions in the tropics. Soil and climatic properties considered in developing the rating system include erosivity, erodibility, soil loss tolerance, compaction, soil temperature regime, available water holding capacity, cation exchange capacity, soil organic matter content and crop residue at soil surface at seeding. The minimum and maximum cumulative rating values for all factors range from 14 to 70. No-till is applicable for soils with cumulative ratings of < 30, and conventional tillage system of ploughing and harrowing for soils whose cumulative rating values exceed 45. For soils with intermediate ratings, some form of minimum or reduced tillage is suggested. Separate rating systems are suggested for rice and tropical root crops. These are the tentative guidelines that should be evaluated for local soils and environments.

9. References

Arora, V.K., Gajri, P.R. and Prihar, S.S. (1991) Tillage effects on corn in sandy soils in relation to water retentivety, nutrient and water management and seasonal evaporativity. *Soil Tillage Research* **21**, 1-21.

Blevins, R.L. (1986) Reduced tillage and soil properties. In: *Proceedings of International Symposium on Minimum Tillage*, Bangladesh Agricultural Research Council, Dhaka. pp 34-44

Blevins, R.L., Frye, W.W. and Smith, M.S. (1985) The effects of conservation tillage on soil properties. In: *A System Approach to Conservation Tillage* (Frank M. D'Itri, Ed.). Lewis Publishers Inc., Chelsea, MI, USA. pp 99-110.

Bodman, G.B. and Rubin, J. (1948) Soil puddling. *Soil Science Society of America Proceedings* **13**, 27-36.

Chapter 9

Water Management

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1. Introduction

Water, seemingly, is an abundant natural resource as three-fourths of the earth's surface is covered with it, but only 2.7% of the global water available is fresh water and of this, only 30% is available to meet the water demands of the human and livestock population, both of which are increasing at an alarming rate. The rest of the fresh water is locked up in glaciers and snow covers. On 2.3% of the world's land area, India supports almost 17% of the world's population with only 4% of the world's fresh water resources. Not only the per capita land availability but also the per capita water availability is decreasing day by day. It was more than 5300 m^3 in 1951, but decreased to 1905 m^3 in 1999 and is likely to be less than 1500 m^3 by 2025. The per capita availability of water less than 1700 m^3 is considered as the 'stress' level. Below this level, availability of water is classified as "scarce" and is considered a severe constraint on socio-economic development and environment quality.

Considering the wide spatial and temporal variations in the water availability within the country, several regions are already facing severe water crisis. Rain is the primary source of water — be it irrigated or rainfed agriculture. It is the precipitation in the catchment area that finally fills the dams and reservoirs so vital for irrigation and it is rainfall again that recharges the groundwater.

2. Water Resources

The major source of water to Indian agriculture is the monsoon system. India annually receives about 400 Mha-m of rainwater, which is 4% of the global precipitation. Out of this, 24 Mha-m equivalent rainwater is available for harvesting. Most parts of the country receive rains during south-west monsoon, except parts of Andhra Pradesh and Karnataka which receive rains from the north-east monsoon also.

Extensive efforts were made in the country to develop water resources for agriculture before independence through construction of village tanks and small harvesting structures. Post-independence, the Government of India has developed a huge infrastructure of canal-based irrigation system. Since 1970s, with the availability of electricity supply and water-pumping technologies, there has been a massive increase in groundwater utilization through tubewells, though largely under private investment. Groundwater today provides about 70% of irrigation and 80% of drinking water supplies. At present, the total irrigated area in the country is 60 Mha, out of which the canal system, tanks and tubewells + open wells constitute 30%, 4.1% and 62%, respectively. During the past 60 years, the management of water resources has changed from the hands of some communities (tanks and small water structures) to government (major and medium irrigation projects), and private domain (groundwater). Though, the net irrigated area in the country has reached 60 M ha, there is a wide spatial variability in availability of water to agriculture in view of the rainfall pattern, land forms, and the river systems. The current status of water resources is shown in Table 1.

3. Water Management Issues

Water management is a highly critical area for sustaining agriculture productivity. It is estimated that by 2050 about 22% of geographical area and 17% of the population would be under absolute water scarcity (< 500 M^3 per capita per year), and about 70% of the area and 16% of the population will be on the verge of economic scarcity [illegible] water availability of less than 1000 M^3 per capita per year. Declining water availability to agriculture and use of poor quality waters are the key constraints. The present water management practices are beset with many problems of wastage and poor productivity for unit water used. There has been an alarming increase in the exploitation of groundwater. While in the northern, western and southern states, the groundwater is being fully exploited, the eastern and north-eastern parts, offer some untapped potential.

Table 1. Water resources of India

Estimated utilizable water	
Surface water	690 km³
Groundwater	432 km³
Total	**1,122 km³**
Irrigation potential	
Major and Medium (Surface water)	
Ultimate	58.50 Mha
Created	40.85 Mha
Utilized (up to 2005-06)	33.45 Mha
Minor irrigation	
Surface water	
Ultimate	17.40 Mha
Utilized (up to 2005-06)	8.17 Mha
Groundwater	
Ultimate	64.00 Mha
Utilized (up to 2005-06)	47.93 Mha
Total (surface and ground)	
Ultimate	81.40 Mha
Created (up to 2005-06)	80.01 Mha
Utilized (up to 2005-06)	56.10 Mha
Total (Major and Medium + Minor)	
Ultimate	139.90 Mha
Created (up to 2005-06)	120.86 Mha
Utilized (up to 2005-06)	89.55 Mha

Source: Central Water Commission (2008)

A. WATER MANAGEMENT OF RAINFED SOILS

A major portion of rainfall received in India is restricted to a period from June to September (Table 2), and it is received in four or five major wet spells. This feature of rainfall leads to generation of sizeable runoff with associated soil erosion and nutrient loss. Crops grown on soils with internal poor drainage and on low lands suffer because of surplus water, whereas the post-rainy season crops often encounter water stress. It is mainly because of these reasons that the agricultural productivity of rainfed regions is generally low (Table 3) and unstable. Scientific management of natural resources including rain water is therefore, called for.

An efficient rain-water management is important for enhancement of agricultural productivity on sustainable basis in rainfed soils which are distributed in most of the states of India. They occupy almost 60% of the area cultivated and contribute more than 40% to India's total food production. Out of the total area used for growing different crops in our country, the rainfed soils cover more than 80%

Table 2. Distribution of annual rainfall according to seasons in India

Rainfall	Duration	Approximate percentage of annual rainfall
Pre-monsoon	March-May	10.4
South-west monsoon	June-September	73.7
Post-monsoon	October-December	13.3
Winter or north-east monsoon	January-December	2.6

Source: Reported from India Meteorological Department, Pune

Table 3. Area, productivity and major growing regions of principal rainfed crops

Crops	Area (Mha)	Productivity (t/ha)	Principal growing areas
Sorghum	15.79	0.690	Maharashtra, Karnataka, Madhya Pradesh, Andhra Pradesh
Pearlmillet	10.69	0.522	Rajasthan, Maharashtra, Gujarat, erstwhile Uttar Pradesh
Maize	5.88	1.325	Erstwhile Uttar Pradesh, Rajasthan, Madhya Pradesh, Bihar
Pulses, including chickpea and pigeonpea	23.83	0.540	Madhya Pradesh, Rajasthan, erstwhile Uttar Pradesh, Maharashtra
Oilseeds, including groundnut, rapeseed and mustard	18.87	0.650	Madhya Pradesh, Gujarat, Maharashtra, erstwhile Uttar Pradesh, Rajasthan, Karnataka, Tamil Nadu
Cotton	7.58	0.175	Maharashtra, Gujarat, Karnataka

Source: Abrol (1988)

area of coarse cereals, 87.5% of pulses, 77% of oilseeds, 65% of cotton and 45% of rice. Thus, the nature, characteristics and management of these soils and rain-water resources hold a key to their long-term productivity.

4. Rainfed Soils of India

The rainfed regions of India encounter different soil types besides the climatic features (Figure 1). Consequently, the problems associated with management of soil and water also differ in nature and severity. The characteristics of the major soil groups in the regions with respect to water management problems are briefly described below and have been depicted in Table 4.

4.1. Sierozems (Aridisols)

These soils are found in about 90% of the geographical area in the north-western part of the country, particularly Rajasthan, parts of Gujarat, west Haryana and west Punjab. The rainfall received in these regions is not only low (less than 500 mm) but also varies highly from year to year. These soils are coarse-textured, often containing a layer of calcium carbonate concretions at different depths. The solum thickness varies from 70 cm to 100 cm. These soils can retain only low amounts of water and plant nutrients and are highly prone to wind erosion. The prospects of intensive agriculture and enhanced crop productivity are low owing to their low water and nutrient retention capacity.

4.2. Sub-montane Soils

This group comprises Mollisols, Inceptisols and Entisols. These soils cover about 9% of the geographical area of India and occur on gently sloping lands, lower elements of slope and steep slopes, respectively, in the hill and foot-hill regions of India. The soils are silty loam to loam in texture, generally deep and retain about 300 mm of water per metre of soil profile. A relatively light texture, slopy terrain and moderate to high rainfall (700-2000 mm) make these soils prone to runoff-induced erosion and landslides.

4.3. Alluviums

Alluviums occupying rainfed regions include Inceptisols, Alfisols and Entisols. These soils are predominantly coarse-textured and occupy 21% of the geographical area of India. They are widely spread across the states of Punjab, Haryana, erstwhile Uttar Pradesh, Bihar, Delhi and Rajasthan. The soils are generally deep and possess high production potential, particularly under irrigated conditions.

4.4. Red Soils (Alfisols)

Red soils occupy 20% of the geographical area of our country. These soils are highly prone to crusting after heavy rains. This leads to high runoff and erosion as the annual rainfall in these areas ranges from 700 mm to 2000 mm. The water retention capacity of these soils ranges from 0.15 to 0.20 m^3 per m^3 at field capacity.

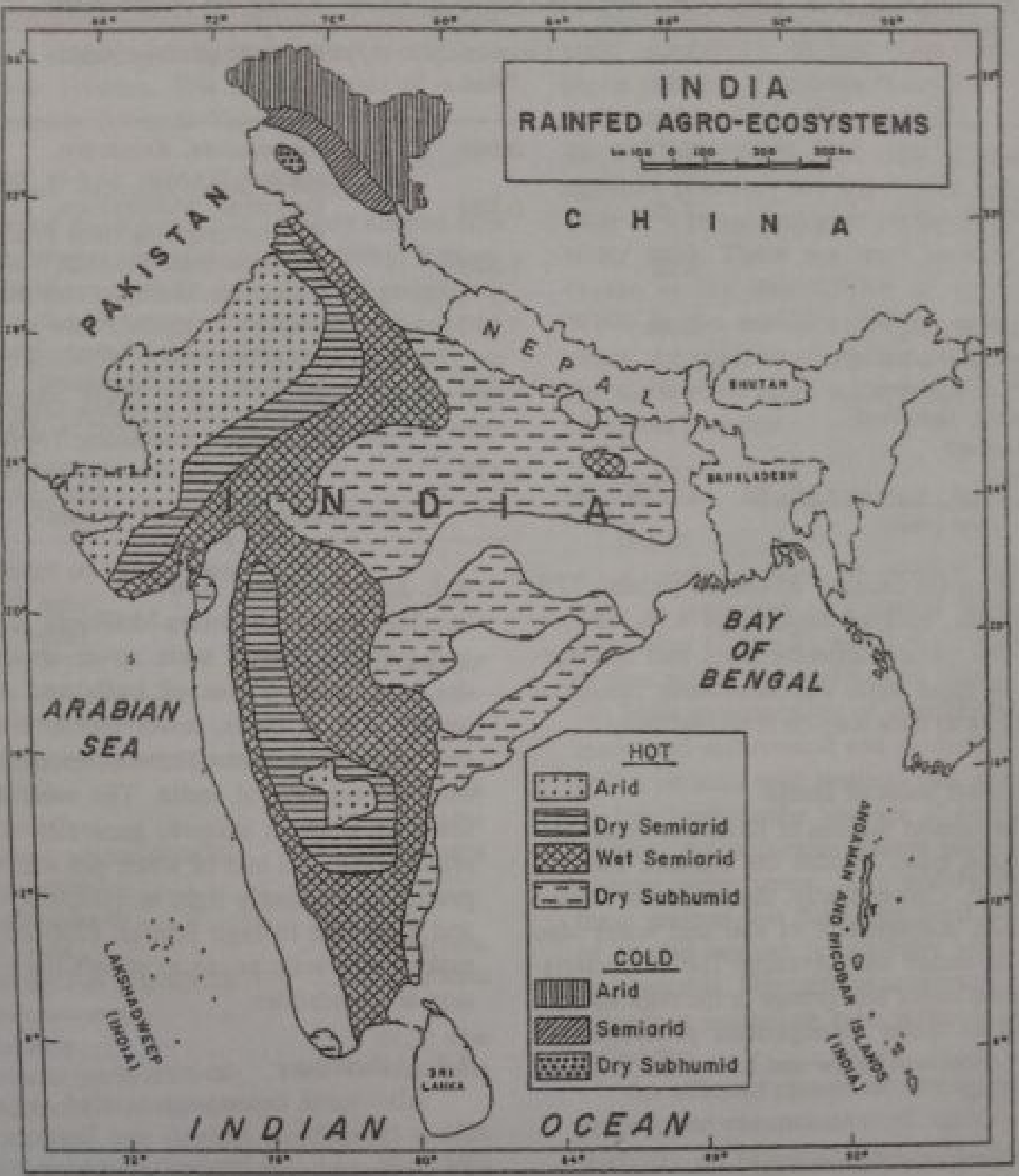

Figure 1. Rainfed agro-ecosystems of India
Source: Sehgal and Sharma (1994)

4.5. Black Soils (Vertisols)

Black soils are mainly Vertisols with associated soils and cover almost 22% of the geographical area of India. They are extensively spread across the states of Maharashtra, Madhya Pradesh, Gujarat, Andhra Pradesh, Karnataka and Tamil Nadu. These soils are characterized by high clay content and consequently, high available water holding capacity. Because of the dominance of the smectite mineral, these soils exhibit swell-shrink properties during a wetting-drying cycle. High water retention and differential water capacity over an extended range of soil water suction render these soils most suitable for rainfed farming. The soils are by and large highly productive. However, they require special soil and rainwater management practices owing to their low permeability, swell-shrink nature and narrow workable moisture range.

5. Water Management Practices

Areas practising rainfed agriculture are diverse with regard to soils and rainfall pattern. The improved farming practices in these areas could enhance the productivity and

Table 4. Distribution, characteristics and water management problems of rainfed soils of India

Agro-ecoregion	Area (Mha)	Soil characteristics	Land utilization type	Water management problems
Western Himalayas	15.2	Shallow, sandy to loamy, skeletal soils with low available water capacity (AWC)	Millets and wheat (in patches)	Severe climatic limitations, viz. cryic temperature regime. Short crop growing period because of low water storage capacity of soils
Western Plains and Kachchha Peninsula	31.9	Deep, sandy and coarse loamy, desert soils with low AWC, deep, loamy, saline and alkali soils with medium AWC	Millets and pulses	Erratic and scanty rainfall, leading to high water deficit. Soil salinity and sodicity, resulting in physiological droughtiness. Acute droughtiness at the time of grain formation
Deccan Plateau	4.9	Deep loamy red soils and clayey, black soils with medium and high AWC	Millets, cotton and oilseeds	High runoff and erosion hazard during stormy cloud bursts. Prolonged dry spells during crop growing period. Poor internal drainage and narrow workable soil moisture range. Subsoil sodicity affecting water availability to plants.
Northern Plain and Central Highlands including parts of Gujarat Plains	32.2	Deep, loamy, alluvium-derived soils with medium AWC and shallow to medium, sandy to loamy, gray brown soils with low to medium AWC	Millets, oilseeds wheat, pulses and maize; sugarcane and cotton under irrigation	Low AWC because of coarse texture. Lowering of groundwater table in some areas because of over-exploitation. Imperfect drainage conditions at some places, leading to surface and sub-surface salinity and/or sodicity
Central (Malwa) Highlands, Gujarat Plains and Kathiawar Peninsula	17.6	Medium and deep, fine loamy and clayey, black soils with medium and high AWC	Millets, wheat, pulses and oilseeds	Droughtiness during the intermittent dry spell period. Imperfect drainage conditions that limit optimum root ramification and oxygen availability in low lying areas.
Deccan Plateau	31.10	Shallow and medium loamy to clayey, black soils with low and medium AWC, inclusive of deep, clayey, black soils with high AWC	Millets, cotton and pulses; Sugarcane under irrigation	Prolonged dry spell periods, adversely affecting the crop growth and leading to crop failure in some areas. High runoff during stormy cloud bursts in the rainy season.
Deccan Plateau (Telangana) and Eastern Ghats	16.5	Medium, red loamy soils with slow and medium AWC and deep, clayey, black soils with medium and high AWC	Millet, and oilseeds, rice; cotton and sugarcane under irrigation	High runoff during rainy season leading to severe soil loss in both red and black soil areas. Under-irrigated agriculture, unjudicious use of irrigation water and imperfect drainage

Contd

Table 4. Distribution, characteristics and water management problems of rainfed soils of India— ***Contd***

Agro-ecoregion	Area (Mha)	Soil characteristics	Land utilization type	Water management problems
				conditions result in high ground-water table, leading to subsoil salinity and sodicity, especially in black soil areas. Frequent droughtiness, resulting in crop failure in some years.
Eastern Ghats (Tamil Nadu uplands) and Deccan Plateau (Karnataka)	19.1	Shallow to deep, loamy, red soils with low to medium AWC; inclusive of deep clayey black soils with high AWC	Millets, pulses and oilseeds; sugarcane and rice under irrigation	High runoff resulting in severe erosion. Coarse soil texture and low to medium AWC, resulting in severe droughtiness during crop growth period.

Source: Sehgal *et al.* (1990)

sustainability. Several water management practices are in vogue in the rainfed regions. They include contour cultivation, bench terracing, strip cropping, and different types of land configurations which are described below. More details about these practices are provided in Chapter 10.

Contour cultivation increases the opportunity time, and hence, the infiltration of rainwater into soil profile. It forms a multitude of mini-barriers across the flow path of runoff which vastly improves the detention and storage *in situ*. Contour cultivation is most effective on deep permeable soils with moderate slopes of 2-7%, whereas it is least effective on flat and steep slopes.

Bench-terracing is practised in areas with 16 to 33% slope (*see* colour plate at the end of this book). It helps in the conservation of soil, rainwater and plant nutrients besides ensuring uniform application of irrigation water. Table top or level terraces are made where rainfall is good. The outward sloping terraces are recommended for areas with relatively low rainfall and shallow soils. The inward sloping bench terraces are suitable for high rainfall tracts with deep permeable soils. Here, even the intra-terrace runoff can be harvested and stored at the end of slope of a terrace in a small dug-out pond. The stored water is conveyed through gravity flow to irrigate the next lower terraces. The percolation losses from these ponds can be reduced by poly-lining (*see* colour plate at the end of this book) as cement-concrete structures develop cracks and fail to store water.

The flat contour cultivation and the ridge and furrow system at 0.4% grade are effective in reducing the runoff and soil loss and in increasing the crop yields (Table 5), particularly on Alfisols having slope less than 1.5%. The raised land configurations (broad bed and furrow, raised and sunken bed system, and narrow ridge and furrow) are useful in low permeable soil regions receiving moderate to high rainfall. They, in some cases, pose the problem of low stability and complications in accommodating certain crop combinations.

For soils occurring on slopes of more than 1.5% in low rainfall regions, a modified contour bund system holds promise (Table 5). This system differs from the original system in that it has outlets in the lower field sections, land is smoothened and the crop is planted on grade instead of on contour. The system allows runoff water to be stored above the bund for a certain period and then released at the desired rate through an outlet. The graded bunds and vegetative barriers spaced at 0.75 m vertical interval have been observed to be useful in conserving soil, rain-water and plant nutrients, besides being cost-effective (Table 6) on moderate slopes.

Table 5. Effect of alternative land surface configurations on crop yield, runoff, and soil loss on Alfisols at ICRISAT Centre, 1981-84

Land treatment	Crop yield (kg/ha) Intercrop system Sorghum	Pigeonpea	Runoff (mm)	Soil loss (tonnes/ha)
(A) Land slope <1.5%				
Broadbed and furrow at 0.4% slope	2740	825	315	3.79
Ridge and furrow at 0.4% slope	2910	870	282	3.02
Flat contour cultivation	2960	875	172	2.05
Flat-on-grade at 0.4% plus ridging up later	2880	840	180	2.78
SE (±)	**151**	**65**	**16**	**0.24**
(B) Land slope > 1.5%				
Contour bund*	2520	710	75	0.97
Modified contour bund with gated outlet*	3020	970	160	0.92
Flat-on-grade with field bunds*	2810	900	215	3.35
Traditional flat with field bunds**	2380	220	256	4.79
SE (±)	**175**	60	**23**	**0.19**

Notes: * Treatment with recommended crop management practices which implies the use of acceptable or recommended variety, cropping system, chemical fertilizer and other practices for weed, pest and insect control.

**Treatment with traditional management practices which implies the use of variety, cropping system, farmyard manure, implement and other practices.

Source: Pathak *et al.* (1989)

6. Structural Conservation Measures

The runoff water is a significant component of water balance of rainfed regions in India (Table 7). The rainfall is usually concentrated over a period of 90 to 100 days and has a few high intensity and stormy events, leading to substantial runoff volumes, depending on topographic features, rainfall characteristics and other associated factors. The volumes of runoff water can be conserved/ stored and recycled for supplemental irrigation during a prolonged dry-spell. The estimated potential of rainwater storage in the various rainfall zones is given in Table 8. The following types of storage structures are commonly used.

6.1. Farm Ponds

These are bodies of water made by constructing an embankment across a water course or by excavating a pit or by combining both. The dug-out ponds are recommended for areas with flat topography, whereas the impounding type farm ponds are developed on water courses in the regions with rolling topography.

The size of farm ponds is governed by the runoff potential. It may be half of the total amount of annual runoff expected from the catchment so that more than one filling can be obtained during the season/ year and the construction cost is minimized. The permissible depth of a pond and side slopes are decided on the basis of soil characteristics of the site such that water retention in the pond and stability of the side walls is ensured. The side wall slopes are kept flatter than 1:1.

Table 6. Effect of inter-terrace land treatments on runoff, sediment and nutrient loss from a Vertisol at Indore

Treatment	Runoff (mm)	Soil loss (kg/ha)	Nutrient loss (kg/ha) N	S
Check	66.4	626.3	13.90	3.94
Graded bund	47.3	387.5	8.25	2.25
Vegetative barrier (Vetiver)	47.3	369.1	7.68	1.54

Source: Annual Progress Report (1993-94)

Among different shapes, cylindrical ponds offer geometrical advantage of maximum storage per unit perimeter. This minimizes seepage losses from side walls and also the cost of

Table 7. Expected water yield in different regions of India

Location	Soil Type	Annual rainfall (mm)	Land slope (%)	Expected water yield (%)
Dehradun (Uttaranchal)	Silty loam	1600	2.4	16.5
Bellary (Karnataka)	Deep black	508	1.2	20.0
Shiwalik region	Sandy loam	1150	10.15	55.0
Anantapur (A.P)	Shallow red	540	3.5	20.0
Hyderabad (A.P.)	Red sandy loam	770	2.5	10.0
Bangalore (Karnataka)	Red soil	830	2.3	20.0
Karnal (Haryana)	Sodic soil	650	0.2	40.0
Chitradurga (Karnataka)	Red loam	612	5.1	10.0
Bunga (Haryana)	Silty clay loam	1116	10.12	50.0

Source: Sastry *et al.* (1985)

Table 8. Estimated potential volume of rainwater storage for small-scale water harvesting structures

Rainfall zone (mm)	Geographical area (Mha)	Rainfall for effective surface storage (%)	Harvestable *runoff* in water harvesting structures (Mha-m)
100-500	52.07	5	0.78
500-750	40.26	6	1.51
750-1000	65.86	7	4.03
1000-2500	137.24	6	14.61
>2500	32.57	4	3.26
Total			23.99

Soruce: Katyal (1997)

lining. All farm ponds must be provided with a spillway for safe disposal of overflows. Spillways provided on ponds with small catchment could be of vegetative nature, whereas for ponds with large catchments, a well-designed mechanical structure has to be provided. Gabion structures (*see* colour plate at the end of this book) (made from boulders arranged in wire mesh boxes) are suitable for black soil region as they are cost-effective and flexible enough to withstand swelling pressures from the underneath soil.

The percolation losses from farm ponds are a major impediment in their popularity in the rainfed regions. Several lining materials have been used with varied success. Bentonite, soil + cement mixture, soil + dung + straw slurry and plastic lining could be used. "Silpaulin", a plastic of 120 g/m^2 density, has been observed to be quite promising, cost-effective and reasonably durable (*see* colour plate at the end of this book).

6.2. Low Earthen Dams

These are commonly used in the areas with preponderance of water streams. They are constructed on the basis of engineering principles across water courses for creating a water reservoir.

6.3. Nala Bunds and Percolation Tanks

These structures are constructed across *nalas* for reducing the velocity of runoff, increasing the water percolation and improving the soil moisture regime in the watershed areas.

7. *In-situ* Conservation

The conservation of water where it falls by means of appropriate moisture conservation practices is termed as *in-situ* conservation. Some of the practices include off-season tillage, deep ploughing, mulching, dead furrows, keyline cultivation, compartmental bunding, cover crop strips, and interplot water harvesting.

Off-season tillage facilitates infiltration of rain-water into the soil profile and reduces the weed infestation. Off-season tillage, however, is not suggested for Aridisols as this would accelerate wind erosion. Deep ploughing (> 22 cm), once every two or three years, promotes infiltration of water and improves the productivity of rainfed crops, particularly in soils with hard pan. The deep tillage immediately following the harvest of post-rainy season crops, encourages the conservation of water in the soil and reduces the incidence of insects and pests. Mulching impedes the movement of moisture from the soil into the atmosphere. Vertical mulching (sorghum stalks placed in 40 cm deep, 15 cm wide trenches and protruding 10 cm above the ground level) can be practised to increase infiltration in the low permeable soils. "Dead furrows" laid out at 3 to 6 m interval help in improving the water retention.

The inter-row water harvesting system is a practice suitable for light-textured soils occurring in low rainfall regions. Significant yield increase can be obtained in crops like pearl-millet by adopting this practice in combination with recycling of harvested water for supplemental irrigation.

8. Agronomic Practices for Improving Water Use Efficiency (WUE)

8.1. Tillage Practices

Tillage refers to mechanical manipulation of soils in a way that it promotes root activity and plant growth. The objective of tillage in rainfed/ dryland areas is to provide a suitable environment for seed germination and root growth, control or minimize weed infestation, reduce soil erosion, conserve moisture *in situ* and reduce runoff. The common tillage practices in rainfed/ dryland areas are minimum or optimum or reduced tillage, conservation or mulch tillage and zero tillage.

The goals of minimum/ optimum/ reduced tillage systems include (a) reducing the energy input and labour requirement for crop production, (b) providing optimum seedbed condition rather than homogenising the entire soil surface, (c) minimizing soil compaction, and (d) conserving soil moisture and reducing erosion.

The objectives of conservation/ mulch tillage are: (i) achieving soil and water conservation, and (ii) saving energy through reduced tillage operations. Both systems usually involve continuous soil coverage by crop residues. Conservation tillage modifies energy balance of soil and reduces diurnal changes in soil temperature besides minimizing runoff and evaporative losses.

In zero tillage or no-till system, planting is practised without seedbed preparation. No-tillage system leads to the saving of energy, conservation of soil moisture and early planting of post-rainy season crop. It, however, allows weed infestation of the crop, particularly in the later stages of crop growth.

8.2. Fallowing

Fallowing, whereby the land is cropped in alternate seasons, is traditionally practised in many dry/ rainfed areas. Rainy season fallowing is usually practised with a view to conserve rainwater in the soil profile, restore soil fertility to some extent and reduce incidence of soil-borne pathogens.

The efficiency of fallowing (E) can be computed using Equation (1):

$$E = \frac{M_2}{M_1 + \text{R or Sc (whichever is less)}} \quad \ldots(1)$$

where,

E is the efficiency of moisture conservation,

M_1 is the available moisture at the beginning (mm),

M_2 is the available moisture at the end of the period (mm),

R is the rainfall received (mm), and

Sc is the storage capacity of the profile (mm).

To make fallowing more effective, it should be combined with conservation tillage systems that leave crop residues on the surface.

With increasing pressure on land resources, the practice of fallowing is being abandoned even in the traditional areas. The availability and use of improved dryland technology have enabled farmers to judiciously utilize land and rain-water resources and maintain soil fertility without resorting to fallowing.

8.3. Crops and Cropping Systems versus Water Availability Period

The problems that are common to crop production in rainfed areas are: (i) often inadequate and erratic water availability, leading to moisture stress, (ii) generally poor resource base of the farmers and therefore, low risk-bearing capacity, and (iii) low and unstable crop productivity. The water availability period, which accounts for both rainfall and stored soil moisture and the recommended cropping systems for different dryland areas of India are presented in Table 9.

The water availability period ranges from 75-90 days in the western Rajasthan and parts of Haryana (Aridisols zone) to 260-280 days in the sub-mountain regions of Punjab and Himachal Pradesh in Inceptisols and related soil zone. In regions receiving 350-600 mm rainfall and having 20 weeks of effective growing season, only single cropping (100% cropping intensity) in *kharif* (red and shallow black soils), or in *rabi* (deep black soils) is possible.

Intercropping is possible in regions with 20-30 weeks of effective growing season. In areas receiving more than 750 mm rainfall and having an effective growing season of more than 30 weeks, double cropping is assured.

8.4. Use of Mulches

A mulch is a material used to cover soil surface with the primary objective of reducing water evaporation. Mulches improve the infiltration of rainwater into soil through runoff control (Table 10), increase infiltration, moderate soil temperature, decrease evaporation and reduce weed infestation. Since major difference in moisture content between the mulched and the bare soils occurs near the surface, they are very useful in seedling establishment.

Several types of materials can be used as a mulch in the rainfed areas. However, crop residue mulches are economical, effective and widely used. Stalks of various crops like maize and pearlmillet, groundnut haulm and shells, paddy husk, wheat and soybean straw, twigs of wild bushes, pine needle and coir pith dust have been used successfully as mulches in the rainfed areas. Soil mulch is also of importance in the Indian context. Soil mulch can be practical in both *kharif* as well as *rabi* crops. In *kharif* crops, it helps in reducing the weed population and increasing the infiltration, whereas in *rabi* crops, its main function is to reduce evaporation. Use of soil mulch in black soils is of special significance as it minimizes the evaporation from shrinkage cracks also.

8.5. Planting Geometry

The geometry of crops can be suitably manipulated to offer resistance to the overland flows. An optimum plant population is essential for ensuring efficient use of land and soil water. The excessive plant population may lead to severe moisture stress and consequently, to crop failures under rainfed conditions.

The practices discussed above may require some modification/ refinement depending on the location and site characteristics. Best results in terms of resource conservation, their use efficiency and enhanced biomass production, and sustainability are achieved through adoption of "watershed management" approach.

B. WATER MANAGEMENT OF IRRIGATED LANDS

In areas where natural rainfall is insufficient – in time and space – the crops are supplied with additional water through artificial means to secure potential crop yields. This process is called irrigation. It enhances crop yields by eliminating crop water stress and increasing possibility of using other inputs affecting the crop growth such as fertilizers, tillage, better seeds and plant protection measures. An assurance against crop failures with irrigation gives sufficient confidence to the grower to use these costly inputs. Recognizing that irrigation holds the key to increasing agricultural production, most nations of the world have invested huge sums of money for developing their surface water and groundwater resources for irrigation.

The availability of water for irrigation is by no means unlimited. Moreover, the increasing demand of water for other uses, viz. civic and industrial uses, hydropower generation, navigation, recreation, fisheries, etc. would decrease

Table 9. Water availability period and efficient cropping systems for different dryland areas in India

Soil zone and region	Water availability period (days)	Crop sequences	Intercropping system (with row ratios)
1. Vertisols and related soil zone			
Malwa Plateau (Madhya Pradesh)	210-230	Maize-safflower/chickpea Soybean-wheat	Maize + soybean (2:2) Soybean + pigeonpea (4 or 6:2) Sorghum + pigeonpea (2:1)
	190-210	Sorghum-safflower/ chickpea	
Baghelkhand (Madhya Pradesh)	210-230	Rice-chickpea/lentil	Wheat + chickpea (2:1) Chickpea + linseed (2:1) Sorghum + pigeonpea (2:1)
	190-210	Sorghum-chickpea Black gram/ green gram-wheat Groundnut-chickpea	
Bundelkhand (Uttar Pradesh)	190-220	Sorghum-chickpea Black gram-mustard/ safflower Fodder cowpea-mustard	Pearlmillet + fodder legume (1:1) Sorghum + pigeonpea (2:1)
Vidharbha (Maharashtra)	190-210	Groundnut-safflower Sorghum-safflower	Sorghum + pigeonpea (2:1) Cotton + pigeonpea (2:1 or 2) Cotton + green gram/ cowpea (1:1) Pigeonpea + green gram (1:3) Pearlmillet + pigeonpea (2:1)
	170-190	Greengram-sorghum/ safflower	Sorghum + greengram/ black gram (2:1)
Southern Maharashtra	160-180	Greengram-sorghum /safflower	Pealmillet + pigeonpea (2:1) Sunflower + pigeonpea (2:1 or 2) Groundnut + sunflower (2 or 3:1) Chickpea + safflower (3:1)
	120-130	–	Pearlmilleit + mothbean (2:1)
Southern Rajasthan	160-180	Green gram -safflower	Maize + pigeonpea (1 or 2:1) Sorghum + green gram (2:2) Groundnut + pigeonpea (2:2) Chickpea + mustard (4 or 7:1)
Northern Central Karnataka	130-150	Cowpea-sorghum Greengram-safflower	Pearlmillet + pigeonpea (1:1) Groundnut + pigeonpea (3 or 4:1) Sunflower + pigeonpea (2:1) Chickpea + safflower (2 or 3:1) Sorghum + pigeonpea (2:1) Sorghum + coriander (2:1)
	100-120	–	Safflower + coriander (2:1)
Saurashtra (Gujarat)	130-140	–	Groundnut + castorbean/ pigeonpea (3:1) Pearlmillet + pigeonpea/ castorbean (2 or 4:1)

Contd

Table 9. Water availability period and efficient cropping systems for different dryland areas in India— Contd

Soil zone and region	Water	Crop sequences availability period (days)	Intercropping system (with row ratios)
Southern (Tamil Nadu)	120-130	–	Sorghum + black gram/ cowpea (2:1) Cotton + black gram (2:2)
2. Inceptisols and related soil zone			
Jammu region of Jammu and Kashmir	280-320	Maize-wheat	Maize + green gram/ black gram/cowpea (1:2) Wheat + mustard (4:1) Chickpea + mustard (4:1) Barley + chickpea (2:2)
Sub-mountain region of erstwhile Uttar Pradesh	240-280	Rice-wheat/chickpea Soybean-wheat Maize-chickpea	Maize + soybean (2:8) Rice + soybean (1:2) Wheat + mustard (9:1) Wheat + chickpea (2:2)
Sub-mountain region of Punjab and Himachal Pradesh	260-280	Maize-mustard/chickpea Pearlmillet-chickpea	Maize + blackgram (1:2) Chickpea + mustard (4:1) Pearlmillet + blackgram (2:1)
	240-260	–	Chickpea + mustard (4:1)
Plains of Uttar Pradesh	200-230	Rice-chickpea/lentil Maize-lentil	Maize + blackgram (1:1) Pigeonpea + blackgram (1:1) Pigeonpea + sesame (1:1) Groundnut + pigeonpea (2:1) Barley + mustard (6:1)
	180-200	Pearl millet-chickpea Greengram / black gram mustard/ barley	Pearl millet + greengram /blackgram (2:1) Pearl millet + pigeonpea (2:1) Chickpea + mustard (4-5:1) Chickpea + barley/linseed (2-3:1)
3. Oxisols zone			
Sub-humid region of Orissa	200-220	Rice-horse gram Maize/finger millet-horsegram Groundnut – horse gram	Rice + green gram (2-3:1) Rice + pigeonpea (3-4:1) Finger millet + pigeonpea (1:1) Maize + pigeonpea (2:1)
Sub-humid region of Southern Bihar	160-180	Maize-chickpea/safflower Groundnut-barley Fingers millet/chickpea Rice-chickpea/lentil	Maize + pigeonpea (1:1) Rice + pigeonpea (2-3:1) Groundnut + pigeonpea (2-3:1) Soybean + pigeonpea (3:1) Pigeonpea + black gram (1-2:3)
4. Alfisols and Related soil zone			
Southern Karnataka	190-220	Cowpea-finger millet Soybean-finger millet	Finger millet + pigeonpea (3:1) Groundnut + pigeonpea (4:1) Finger millet + soybean (1:1)
Telangana (Andhra Pradesh)	140-160	–	Sorghum + pigeonpea (3:1)
Rayalaseema (Andhra Pradesh)	110-130	– –	Castor bean+cluster bean(2:2) Groundnut + pearlmillet (6-8:1-2)

Contd

Table 9. Water availability period and efficient cropping systems for different dryland areas in India— Contd

Soil zone and region	Water	Crop sequences availability period (days)	Intercropping system (with row ratios)
			Groundnut + pigeonpea (6-8:1-2) Groundnut + castorbean (2:1)
5. Arid soil zone North-western Gujrat	100-120	–	Pearlmillet + green gram (1:3) Pearlmillet + clusterbean (3:1) Clusterbean + green gram (2:1)
Western Rajasthan and parts of Haryana	75-90	–	Pearlmillet + balck gram/ green gram (2:1) Pearlmillet + clusterbean (1:2) Cluster bean + black gram (1:1) Pigeonpea + black gram (1:2)

N.B. Figures in parentheses are optimum row ratios of inter-cropping system.
Source: Singh (1988)

Table 10. Effect of groundnut shell mulch on crop yield, runoff and soil loss from Alfisols at ICRISAT Center, Patancheru (1981-85)

Rate of groundnut shell mulch (tonne/ha)	Crop yield (tonne/ha) (intercropping)		Runoff (mm)	Soil loss (tonne/ha)
	Sorghum	Pigeonpea		
No mulch	2.07	0.84	282	3.92
2.5	2.27	0.97	147	1.95
5.0	2.42	1.11	88	1.21
10.1	2.63	1.34	26	0.35
SE±	*0.17*	*0.09*	*13*	*0.23*

Source: ICRISAT Annual Report (1986)

the proportion of available water for agricultural usage in future (Table 11). Therefore, the available irrigation water must be used most efficiently. The efficient use of this resource for crop production consists of conserving the resource, allocating it optimally among the competing demands, its application to crop at optimal rates and timing, maximizing the utilization of resource by crop and maximizing returns per unit resource used by the crop. Concerns for sustainability of production systems require that the agricultural practices should not degrade the environmental quality and longtime productivity of the system. Irrigation, unfortunately is only a mixed blessing. Injudicious irrigation is known to cause waterlogging and salinization of soil, which jeopardise the soil productivity. Hence, the objective of water management in irrigated lands is to optimize agricultural production with available irrigation water without adversely affecting the quality of environment and production resources.

9. Basic Aspects of Field Water Management

Water management for irrigation is accomplished in two phases, viz. (i) water development for irrigation, and (ii) utilization of irrigation water for agricultural production. First phase is the realm of civil engineers and is

Table 11. Water requirements for different uses

(Quantity in BCM)

S. No	Uses	Year 1997-98	Year 2010			Year 2025			Year 2050		
			Low	High	%	Low	High	%	Low	High	%
					Surface Water						
1.	Irrigation	318	330	339	48	325	366	43	375	463	39
2.	Domestic	17	23	24	3	30	36	5	48	65	6
3.	Industries	21	26	26	4	47	47	6	57	57	5
4.	Power	7	14	15	2	25	26	3	50	56	5
5.	Inland Navigation		7	7	1	10	10	1	15	15	1
6.	Flood Control		-	-	0	-	-	0	-	-	0
7.	Environment (1) Afforestation		-	-	0	-	-	0	-	-	0
8.	Environment (2) Ecology		5	5	1	10	10	1	20	20	2
9.	Evaporation Losses	36	42	42	6	50	50	6	76	76	6
	Total	**399**	**447**	**458**	**65**	**497**	**545**	**65**	**641**	**752**	**64**
					Groundwater						
1.	Irrigation	206	213	218	31	236	245	29	253	344	29
2.	Domestic & Municipal	13	19	19	2	25	26	3	42	46	4
3.	Industries	9	11	11	1	20	20	2	24	24	2
4.	Power	2	4	4	1	6	7	1	13	14	1
	Total	**230**	**247**	**252**	**35**	**287**	**298**	**35**	**332**	**428**	**36**
	Grand Total	**629**	**694**	**710**	**100**	**784**	**843**	**100**	**973**	**1180**	**100**
					Total Water Use						
1.	Irrigation	524	543	557	78	561	611	72	628	817	68
2.	Domestic	30	42	43	6	55	62	7	90	111	9
3.	Industries	30	37	37	5	67	67	8	81	81	7
4.	Power	9	18	19	3	31	33	4	63	70	6
5.	Inland Navigation	0	7	7	1	10	10	1	15	15	1
6.	Flood Control	0	0	0	0	0	0	0	0	0	0
7.	Environment (1) Afforestation	0	0	0	0	0	0	0	0	0	0
8.	Environment (2) Ecology	0	5	5	1	10	10	1	20	20	2
9.	Evaporation Losses	36	42	42	6	50	50	6	76	76	7
	Total	**629**	**694**	**710**	**100**	**784**	**843**	**100**	**973**	**1180**	**100**

Source: Report of the National Commission for Integrated Water Resources Development (Vol-I), Sept., 1999

beyond the scope of this book. Discussion here pertains only to the second phase. It is assumed that maximum (natural) rainwater is conserved through adoption of soil and water conservation practices and irrigation is applied to supplement the soil-stored water for enhancing crop production. Efficient use of irrigation water calls for an understanding of the soil-water-plant-atmosphere relationships. Water is stored in the root zone and is taken up by plant roots and transported to the foliage to be lost to atmosphere through transpiration (T). Simultaneously, water is also lost through evaporation directly from the exposed soil surface. For optimum photosynthetic activity of the plant, transpiration from its green foliage must meet the atmospheric demand. When actual T falls short of potential T demand of the plant, the rate of photosynthesis is reduced. The potential crop water demand is governed by green-area-index (green surface area per unit land surface) and atmopsheric evaporativity. And the actual T depends on the soil water status and water uptake capacity of the root system. Therefore, in order to develop strategies for efficient water use, it is important to understand the components of the soil-water-plant-atmosphere system. These sub-systems are: (i) soil-water relations, (ii) soil-plant relations, (iii) soil-water-plant relations, and (iv) plant-water relations.

9.1. *Soil-Water Relations*

Soil-water relationships such as water retention and transmission characteristics, plant available water (PAW), infiltration and redistribution of water in the soil, etc. have been described in detail in Chapter 6. However, it may be reiterated that depending upon the texture, structure and mineralogical composition, soils differ widely in moisture characteristics and water transmission properties. The plant available water capacity is low in coarse-textured soils and it increases with proportion of fine particles up to a point and decreases thereafter.

An important parameter of soil-water relationship from field water management standpoint is the infiltrability of soil which determines the suitability of irrigation methods for assuring efficient water-use. Water infiltrated into the soil redistributes itself in the profile. Soils with slow infiltrability and redistribution may cause aeration stress to crops in the event of heavy rain or irrigation. As already stated, when the rate of water application or rainfall intensity exceeds the infiltrability of soil, excess water accumulates at the surface. It cuts off the process of soil aeration till all the water enters the soil and gets redistributed so as to create sufficient water-free pore spaces for renewal of oxygen in the soil air by diffusion.

9.2. *Soil-Plant Relations*

A plant uses soil for germination of seeds and seedling emergence and growth and proliferation of its root system. The first step in seed germination is imbibation of water by seed to start the metabolism. The rate of water movement to the seed is governed by soil wetness and hydraulic conductivity of soil together with the properties of the seed-coat. There exists an optimum range of soil water content in each soil or of soil water potential across soils for germination and seedling emergence. On either side of this range, germination and emergence get delayed or beyond a certain threshold, it does not occur at all. Thus, it is important to ensure adequate soil moisture in the seed-zone for establishment of seedlings.

9.2.1. Root Systems and Water Uptake

The plants take up water and nutrients from the soil through their root system. Like seed germination, the process of root growth also requires an optimum range of soil moisture as would supply enough water to the roots, keep the soil strength low and permit adequate soil aeration. For management of irrigation water, it is important to know that in early stages root systems of crops are confined to shallow surface layers and so is the water uptake from soil. But with time, roots penetrate deeper into the profile and may extend down to 2-3 m even in annual crops. In perennial crops and trees, root systems penetrate several metres and derive water from such depths. Under non-limiting soil water content, the rate of water uptake from various layers of soils is proportional to the fraction of roots located in that layer. However, when the upper layers are wetted, they become the zone of maximum water uptake.

It is generally reported that frequent wetting of soil with small amounts of water tends to promote shallow rooting. Infrequent and heavy applications of water induce deep rooting. Practices such as tillage, mulching, nutrient application which influence depth and density of rooting are likely to influence the water management criteria for best returns.

9.3. *Soil-Water-Plant Relations*

Dynamics of soil-water uptake by plants is an interplay of soil-water status and root proliferation on the one hand, and canopy development and atmospheric evaporation on the other. Water in the soil-water-plant-atmosphere continuum (SPAC) moves in response to difference in the energy status of water in the system. A drop in the energy status of leaf water on exposure to radiation, wind and vapour-unsaturated atmosphere is transmitted through the stem to the root in contact with soil-water. This sets up the energy gradient required for water to flow from soil to root, from root to stem and from stem to leaf and atmosphere.

The flux of water (q) through SPAC can be expressed by Equation (2):

$$q = (\psi_s - \psi_l) / R_t = T \qquad ...(2)$$

where, ψ_s and ψ_l are soil-water potential and leaf-water potential, R_t is the total resistance in the soil and plant, and T is transpiration.

Since during nights, T is negligible, the pre-dawn values of ψ_l are considered equal to ψ_s values. As the day advances, the stomata open up and the evaporative demand increases; the ψ_l decreases till afternoon and then the cycle reverses.

9.4. Plant-Water Relations

To remain fully turgid and photosynthesize at potential rate, the plant foliage must lose water fast enough to meet the atmospheric evaporative demand. When the transpirational loss falls short of the evaporative demand, the plant develops water stress. This stress is reflected in the loss of turgidity, increase in leaf diffusion resistance and decrease in leaf water potential. Before describing the effect of stress on plant growth and yield, it is important to know about the commonly used indices of plant-water stress.

10. Indices of Plant-Water Stress

10.1. Relative Leaf-Water Content

The relative turgidity measured as relative leaf-water content (RLWC) is computed from the fresh weight, turgid weight and oven-dry weight of leaf discs using Equation (3):

$$\text{RLWC} = \frac{\text{Fresh weight} - \text{Oven dry weight}}{\text{Fully turgid weight} - \text{Oven dry weight}} \qquad ...(3)$$

The turgid weight is obtained by keeping the leaf discs in free water till they acquire maximum weight. In the case of rice, the leaf discs must float on water for 4 to 6 hours before water deficiency is completely removed. Plants generally start wilting at RLWC of 75-85%.

10.2. Leaf-Water Potential

Plant-water potential is a measure of energy status of plant water and has three components as depicted by Equation (4):

$$\psi_l = \psi_m + \psi_p + \psi_o \qquad ...(4)$$

where, ψ_l is the leaf water potential, and ψ_m, ψ_p and ψ_o are matric, pressure and osmotic potentials. Since the contribution of ψ_m is not known, it is lumped with ψ_p and the relation is expressed by Equation (5):

$$\psi_l = \psi_p + \psi_o \qquad ...(5)$$

The leaf-water potential, also called the 'xylem water potential', is intimately related to RLWC. For example, in rice leaves, y_l decreases sharply as RLWC falls below 100%; in the intermediate RLWC range, the drop of y_l is gradual, but below RLWC of 35%, the y_l drops sharply (Tomar and Ghildyal, 1973).

10.3. Leaf Diffusion Resistance

In a fully turgid leaf in the presence of day-light, the stomata are fully open. Water vapour and carbon dioxide diffuse across the open stomata without any resistance. Functioning of gaurd cells which regulate the stomatal opening is controlled by the tissue water status. When the plant tissues develop water-deficiency, stomata begin to close partially and may be completely when water stress reaches a critical level. The closure of stomata reduces their apertures which increases the resistance to the diffusion of water vapour and carbon dioxide across stomatal openings and thus, the rate of transpiration from plants decreases. The leaf diffusion resistance (LDR) increases with a decrease in RLWC.

11. Water Use Efficiency

Agronomists define water-use efficiency (WUE) as the ratio of produce per unit area to the water used by the crop [Equation (6)]:

$$\text{WUE} = \frac{\text{Biomass(kg)}/\text{area}(\text{m}^2)}{\text{Water used(m)}} = \text{kg.m}^{-3} \qquad ...(6)$$

The water use in this relationship represents the evapotranspiration (ET) loss from the crop. In other words, it includes direct evaporation from the exposed surfaces in the field.

The plant physiologists define WUE as the ratio of biomass per unit area to transpiration (T) loss from the crop [Equation (7)]:.

$$\mathrm{WUE} = \frac{\mathrm{Biomass(kg)/area(m^2)}}{\mathrm{T(m)}} = \mathrm{kg\,m^{-3}} \qquad ...(7)$$

In this case, water lost as direct evaporation is excluded.

An integrated approach has also been developed to express WUE, where it is defined as biomass (B) or grain (G) yield in terms of evapotranspiration (ET) or transpiration (T) on daily or seasonal basis. For example, WUE (G, ET, s) means WUE is expressed as the ratio of grain yield to seasonal ET. Similarly, WUE (B, T, d) signifies WUE based on biomass per unit T on daily basis.

12. Management of Irrigation Water

In view of the high investment costs of developing irrigation and limited availability of irrigation water, it should be used most efficiently. It has to be transported to the site of use and applied to the fields efficiently at appropriate timing and in adequate amounts.

12.1. Sources of Irrigation

Irrigation water is available as surface storage or as groundwater. The size of surface water reservoirs varies from a few hundred cubic metres in small tanks to several million hectare-metres in big storage dams. Depending upon the site of storage and site of use, the surface water is transported over distances, ranging from a few metres to several hundred kilometres. Groundwater is generally used in the vicinity of the source and conveyed over short distances. The quality of water also differs within and between sources of irrigation.

12.2. Surface Water Conveyance

Water from big reservoirs is carried to the use-site through canal networks consisting of main canal, distributaries and water courses. To permit gravity irrigation, the dug-out canal network is bunded above ground. Therefore, water in the canal system is lost as downward percolation from bottom and seepage through the sides of the channels. In unlined channels, the seepage losses from the main canal, distributory and field water courses depend upon the water transmission properties of the terrain through which the channels pass.

The major loss of water occurs in the field water courses (Table 12). If seepage losses are high, canals are lined with various materials such as bricks, concrete and stones, etc. to reduce the same. Since the seepage near the head-end is more than near the tail-end, it is not necessary to line the entire length of a water course. The lining of 50-60% of the total length may reduce the seepage loss by 80-85%. A plot of the ratio of annual benefits to annual costs versus the length of lining can help decide about the length to be lined for best returns.

Water from pumps and tubewells and small tanks is transported over short distances. The lining of water courses is effective in reducing seepage loss in these cases also. Conveyance of water through underground pipes has also

Table 12. Conveyance losses from unlined, partially-lined and lined irrigation system as per cent of water released from a reservoir

System characteristics	Main canal (%)	Distributory (%)	Field water courses (%)	Total conveyance losses (%)	Conveyance efficiency (%)
Entire system unlined	15	7	22	44	56
Only canal lined	4	7	25	36	64
Canal and distributories lined	4	2	26	32	68
Whole system lined	4	2	6	12	88

Source: Aggarwal and Khanna (1983)

been recommended. It is especially useful for undulating lands and highly permeable soils. Apart from reducing seepage losses, it has the added advantage of saving land. It decreases the maintenance costs and increases the freedom of farm operations.

13. Methods of Water Application

Once irrigation water reaches the field, the next task is to apply it efficiently. Attributes of judicious application are: application efficiency, storage efficiency and distribution efficiency. Water application efficiency (WAE) is the ratio of water stored in the root zone to water delivered to the field during irrigation [Equation (8)]:

$$\text{WAE(per cent)} = \frac{\text{Water stored in root zone}}{\text{Water delivered to field}} \times 100 \qquad \text{...(8)}$$

Water storage efficiency (WSE) is the ratio of actual irrigation water stored in root zone to water storage needed in the root zone before irrigation [Equation (9)]:

$$\text{WSE (per cent)} = \frac{\text{Water stored in root zone}}{\text{Root zone water deficit before irrigation}} \times 100 \qquad \text{...(9)}$$

Similarly, water distribution efficiency (WDE) is given by Equation (10):

$$\text{WDE} = \left(1 - \frac{N}{D}\right) \times 100 \qquad \text{...(10)}$$

where, N is the average numerical deviation in depth of water stored at different locations from the average water storage depth of field following irrigation, and D is the average water storage depth of field.

Irrigation methods fall under two categories, viz.

(i) Surface methods, and
(ii) Pressurized irrigation systems

13.1. Surface Irrigation Methods

In India, systems of surface methods of irrigation have been developed in the past years. Among them, the check basin method of irrigation is the most popular one. It is very important to employ the correct method of water application to minimize adverse effects of irrigation. The soil type, land topography, crops to be grown, quality and quantity of water available for irrigation and other site-specific variations influence the selection of the right method of irrigation. Proper land levelling and channel linings are some of the management options that are generally ignored. Land levelling increases the water application efficiency, which leads to higher yields as well as increase in water-use efficiency (Table 13). It also has a direct impact on the nutrient-use efficiency.

13.1.1. Check Basin Method

It is the most common method of surface irrigation. Several lateral field channels are taken off from the main irrigation channels running along the side of the field. Each lateral channel feeds on both sides, a row of several small plots surrounded by ridges — termed as check-basins. More lateral channels facilitate rapid spread of water over the enclosed area of land. This method is preferred and practised over level lands. The spread time is kept generally low to minimize percolation losses near the head-end of the basin. Therefore, on highly permeable soils, high stream size and small check basins are desirable. However, an adequate water depth to make up the soil water deficit should be ensured.

13.1.2. Border Strip Method

Water is applied from one end (head-end) to borders of suitable dimensions with appropriate slope. The design criteria of borders depend on the infiltrability of soil, border slope, border length and stream size. Narrow strips help achieve uniform advance of water front from head-end to tail-end. Similarly, a gentle and uniform slope improves the water application efficiency. Slope generally varies between 0.05% and 0.5% for most crops. The recommended design dimensions of borders for various soils, stream size and slopes are given in Table 14. During application, water supply is cut off when the advancing water front is still some distance away from the tail end. The cut-back point on the strip length depends on strip width, soil slope, stream size and soil characteristics.

Table 13. Effect of land levelling on yield and water use efficiency

Levelling Index (cm)	Water use efficiency (kg/ha-cm)		Yield (q/ha)	
	Wheat	*Bajra*	Wheat	*Bajra*
1.2	166	120	46.6	37.3
2.0	138	106	42.2	34.2
2.5	128	99	39.3	32.9
3.0	116	92	36.5	31.4
3.7	110	84	34.9	27.9

Table 14. Recommended design dimensions of border strips for various soils, slopes and stream sizes

Place	Soil	Slope (per cent)	Stream size (L/s)	Border dimensions (m)	
				Width	Length
Hissar	Sandy loam	0.1-0.2	14.0	7.5	50-70
Jobner	Sand	0.3-0.5	3.5	1.5	20
Karnal	Clay loam	0.1-0.15	10-12	6.0	70
Kharagpur	Sandy loam	0.3	6-8	3.5	50-70
Kota	Clay loam	0.2	10.0	6.0	50
Pantnagar	Clay loam	0.3-0.4	12-15	7.5	100

Source: Yadav (1982)

13.1.3. Furrow Method

The furrow method of irrigation is generally used to irrigate row crops and vegetables, and is suited to soils in which the infiltration rates are between 0.5 and 2.5 cm/hr. It is ideal for slopes varying from 0.2% to 0.5% and a stream size of 1-2 L/sec. Many of the field crops in which water is applied through check basin or border strip methods, can easily be adapted for furrow irrigation. In areas requiring surface drainage, furrows are more effective. Scientific studies suggest that irrigation in alternate furrows for the cotton and maize crops saves irrigation water to the tune of 30%, without compromising the crop yield

Water from the main or lateral channel is directed into small parallel furrows drawn along the slope in between the small ridges or broad checks. Water in the furrow moves both laterally and vertically to moisten the ridges and sub-soil layers. On soils with slope, furrows are made along the contour. Furrow irrigation is more efficient than either border strip or check-basin methods. Furrow dimensions depend upon the soil type, stream size, land slope, and nature of crop and implements available for cultural operation.

13.1.4. Surge Flow Irrigation

Excessive water intake and deep percolation losses are major limitations for irrigation applications through furrows and border strips. Surge flow irrigation, the intermittent application of water in a series of on and off modes of constant or variable time spans has the potential of reducing intake and percolation losses, increasing the irrigation efficiencies and conserving irrigation water. Field experiments have corroborated that surge flow furrow irrigation results in faster water front advance and reduces the volume of water required to complete the advance phase. The studies have also confirmed that it increases water distribution efficiency, reduces infiltration rate and the cumulative infiltration depth, and consequently, reduces the percolation losses.

13.1.5. Raised Bed Planting Systems

This system of crop establishment is a modification of the ridge-furrow system in which the ridges are replaced by flat beds. Crops are cultivated on the raised beds, while the furrows are used for irrigation. Fertilizers can be placed on the raised beds. The width of raised beds depends on the crops to be grown. This crop establishment technique has many advantages as it improves both water and nutrient use efficiency, requires lower seed rates and the furrows act as the drainage channels in the case heavy rainfall, preventing temporary waterlogging. Several crops, particularly vegetables, are ideally suited for this method of crop establishment.

13.1.6. Water Management in Rice

Rice is the staple food in Asia, the most populous continent, which accounts for almost 90% of the global production and consumption of rice. This demand is likely to increase by 70% in the next three decades for the ever-increasing Asian population. Asia produces more than 530 million tonnes of paddy rice every year. More than 75% of this comes from the irrigated area, which accounts for 55% of the total rice area. Irrigated low land rice constitutes the major economic activity, source of employment and income for the rural population. In India, out of the 43 million hectares under rice cultivation, almost 55% is irrigated, which consumes almost 50% of the irrigation water resources. The growing water shortage has threatened the sustainability of irrigated low land rice production systems, which consume up to 3000-5000 litres of water to produce 1 kg of grain. Fresh water consumption for irrigated low land transplanted rice in most Asian countries is, therefore, very high. Since rice is the staple food of billions, it is imperative to devise technologies for growing rice with less water, reducing dependence on traditional methods of rice growing, without compromise on its production.

Traditionally, it is thought that for maximum production rice needs continuous submergence/flooding throughout its life-cycle. As a result, the water requirement of rice is many times more than other crops. This practice not only leads to wastage of water but also causes environmental degradation and reduces fertilizer-use efficiency. Years of research conducted on varying soils all over the country has conclusively established that irrigating rice one to three days after disappearance of ponded water can save 20 to 30% of irrigation water applied without any significant effect on the yield (Table 15). The saving in irrigation water could be used to bring additional area under irrigation or diverted to dry and/or water scarce areas.

Table 15. Effect of irrigation on the yield of paddy at various locations in the country

Place	Yield (t/ha)			
	Continuous submergence	Irrigation after disappearance of ponded water from the surface		
		One day	Two days	Three days
Pusa (Bihar)	3.59 (81)*	3.47 (60)	3.25 (46)	2.85 (35)
Chiplima (Orissa)	7.21 (99)	6.66 (90)	6.47 (76)	5.85 (68)
Kharagpur (W.B.)	6.11 (197)	5.98 (150)	5.89 (129)	4.99 (108)
Bilaspur (M.P.)	5.86 (98)	5.59 (77)	5.13 (70)	4.62 (56)
Pant Nagar (Uttranchal)	8.09 (121)	7.57 (112)	7.38 (90)	6.92 (60)
Ludhiana (Punjab)	5.52 (190)	5.44 (145)	5.12 (113)	5.20 (96)
Hisar (Haryana)	5.66 (220)	5.15 (196)	4.69 (126)	-
Kota (Rajasthan)	5.41 (145)	5.33 (86)	5.05 (68)	-
Mudrai (T.N.)	6.59 (128)	6.66 (104)	6.18 (82)	-
Bhavanisagar(T.N.)	4.57 (136)	5.09 (110)	4.95 (80)	4.19 (55)
Chalakudi (Kerala)	3.47 (14)	3.44 (63)	3.45 (49)	-

Note: *Figures given in brackets show the quantity of irrigation water applied (cm)

Source: Chaudhary (1997)

Efforts are being made to devise and develop new management options and varieties using modern tools like marker aided selection to increase water-use efficiency of rice crop. Aerobic rice cultivation and location-specific modified versions of system of rice intensification (SRI) are some of the options that are being explored.

13.2. Pressurized Irrigation Systems

Considering the need of irrigation methods that are more efficient, efforts are being made to develop water-efficient methods. Two of these methods are sprinkler and drip irrigation systems, which are more efficient than the conventional surface methods (Table 16). Since these methods need water under pressure, they are classified as 'Pressurized Irrigation Systems', in contrast to the surface irrigation methods which are generally gravity-driven. The pressurized methods of irrigation provide many advantages over the gravity surface irrigation methods in terms of water savings and crop yields. The quality of produce is also considered generally superior than in conventionally irrigated crops. These pressurized irrigation systems are very useful, particularly in undulating and uneven lands or coarse-textured soils. The use of these techniques can be promoted for adoption in irrigation commands also under specific conditions. The total potential area where these methods could be applied is very large (Table 17) and Government of India is making several efforts to popularize these methods among the farmers.

13.2.1. Sprinkler Irrigation

In the sprinkler method of irrigation, water is sprayed through nozzles connected to a network of pipes through which water is supplied

Table 16. Irrigation efficiencies (per cent) under different methods of irrigation

Irrigation efficiencies	Methods of irrigation		
	Surface	Sprinkler	Drip
Conveyance efficiency	40-50 (canal) 60-70 (well)	100	100
Application efficiency	60-70	70-80	90
Surface water moisture evaporation	30-40	30-40	20-25
Overall efficiency	30-35	50-60	80-90

Source: Sivanappan (1998)

Table 17. Theoretical potential area for drip and sprinkler irrigation in India (Mha)

Crop	Area		
	Drip	Sprinkler	Total
Cereals	-	27.6	27.6
Pulses	-	7.6	7.6
Oilseeds	3.8	1.1	4.9
Cotton	7.0	1.8	8.8
Vegetables	3.6	2.4	6.0
Spices & condiments	1.4	1.0	2.4
Flowers & medicinal and aromatic plants	-	1.0	1.0
Sugarcane	4.3	-	4.3
Fruits	3.9	-	3.9
Coconut & plantation crops, Oil palm	3.0	-	3.0
Total	27.0	42.5	69.5

Source: Government of India (2003), Agricultural Statistics at a glance 2003, Directorate of Economics and Statistics, MoA, New Delhi; NHB (2002), Indian Horticulture Database: 2002, National Horticulture Board, MOA, GOI, Gurgaon under pressure. The rate of spray of water can be regulated and natural rainfall may be simulated. This method has higher application and distribution efficiency, effects considerable savings of water, and provides complete control on timing and quantity of irrigation water to be applied. The overall efficiency of sprinkler method can be as high as 60%, compared to around 30% of surface method of irrigation. Water can be applied in quantities required to adequately wet the root zone of the crop, thus avoiding wastage of water through percolation. As the water is conveyed through a closed pipe system, conveyance losses are minimum in sprinkler method of irrigation. Run-off losses are also reduced considerably. As the time of application of water can be controlled effectively, light and frequent irrigations can be applied and a healthy moisture regime can be maintained in the root zone of the crop, thus leading to healthy crops and higher levels of production of good quality produce. Further, this method leads to considerable savings in the irrigation water and increase in area covered and, hence, higher irrigation intensity can be achieved.

In spite of above-mentioned advantages, very limited cropped area in the country is under this method of irrigation. High initial capital cost, and lack of expertise in its operation and maintenance have limited the use of this system to crops. This technology is being gradually adopted in water-scarce areas and for cultivation of high-value export-oriented commercial crops like flowers, vegetables and horticultural crops. The sprinkler method of irrigation could be introduced for a large number of crops, depending upon the soil, slope, water resource, farmers' capacity for investment, etc.

Historically, development of sprinkler irrigation got a boost mainly after the Second World War, with the introduction of lightweight, portable aluminum pipes. At the same time, improved sprinklers and quick couplers, that facilitate convenient uncoupling and recoupling of pipes, has been a remarkable development. The farming community in India was not familiar with sprinkler irrigation till mid-1950s. Even today, the area under sprinkler method of irrigation is too low in India compared to several other countries. It is largely because an average Indian farmer is resource-poor and cannot invest in the system.

The plantation owners in the hills use sprinkler irrigation to irrigate tea, coffee and cardamom crops during dry seasons and dry spells of the monsoon period. In mid-1970s, progressive farmers in the Narmada valley in Madhya Pradesh, southern parts of Haryana and north-east parts of Rajasthan started using sprinkler system to overcome problems of water-shortage, particularly during summers. Adoption of this system gradually spread to larger areas in the states of Haryana, Rajasthan, Madhya Pradesh, Maharashtra and Karnataka. However, in major agricultural states like Punjab, Uttar Pradesh and Bihar, this system of irrigation is extremely limited. About 60% of area irrigated by sprinkler system is under field crops like cereals, pulses, oil seeds, cotton, sugarcane and vegetables, the rest being under tea, coffee, cardamom gardens in the western ghats and north-eastern states.

The developmental work on sprinkler irrigation system has been mainly aimed at reducing the cost of the equipment and introducing smaller low pressure sprinklers. Till recently, the most popular piping material has been aluminum but to reduce the cost of the systems 'High Density Polyethylene' (HDPE) pipes with suitable modifications have been introduced. The spread and popularization of sprinkler irrigation method among farmers has received significant support under the various schemes, involving subsidy of the central and state governments in the recent years.

13.2.2. Drip Irrigation

Drip irrigation involves application of water only at the roots of the plant where it is required and thereby saving more water and brings more area under irrigation. The crop yields by this method of irrigation are higher with reduction in cost of fertilizers, pesticides and power-use for irrigation. In drip irrigation, water is directly applied into the root zone of plants and it permits the irrigator to limit the watering closely to the consumptive use of the plants. This method is well-suited for wide-spaced high-value crops like coconut, grapes, banana, lime, etc. The required quantity of water is provided to each plant daily or at short intervals at the root zone through a network of piping system. Hence, there is no loss of water in either conveyance or application. Evaporation losses from the soil surface are also minimal, since water is applied only to the root zone and crop canopy provides the shade to prevent evaporation. Research studies have indicated that the water savings can range from 40% to 70% or higher with enhanced yield in various crops, if the drip method is used.

Considering the irrigation needs in Indian agriculture, emphasis is being laid on promoting drip irrigation which helps conserve water, reduces fertilizer inputs and ensures higher productivity. This technology though was introduced in early-1970s in India, its adoption took off in the 1980s, particularly in fruits, vegetable and cash crops. Drip irrigation is becoming increasingly popular in water-scarce and salt-affected areas. In India, almost 1.5 lakh hectares of fruit crops and 2.5 lakh hectares of vegetable crops are being added annually under drip irrigation. As per the estimates of Ministry of Agriculture, Government of India, about 27 Mha area in the country can be brought under drip irrigation (Table 17).

To tap this potential, the Government of India has initiated various schemes for motivation of farmers, ensuring availability of material, technical support and credit availability. This has helped in promoting the technology in a big way.

13.2.3. Rainguns

Rainguns are the modified form of sprinklers and are less capital-intensive compared to conventional sprinkler systems and can be modified to cater to many other crops. They support the highly recommended practice of trash mulching in sugarcane, enriching the soil with nutrients. They also hold a lot of promise as an efficient tool for application of irrigation water.

13.2.4. Micro Sprinklers

The micro sprinkler method of irrigation has the advantages of both the sprinkler and drip irrigation methods. Water is sprinkled or sprayed around the root zone of the trees with small sprinklers that work under low pressure. This unit is fixed in a network of tubings, but can be shifted from place to place. The exact quantity of water can be delivered to each plant daily at the root zone. Water is given only to the root zone area as in the case of drip irrigation unlike to the entire ground surface as in case of sprinkler irrigation method. This method is highly suitable for tree/orchard crops and vegetable crops and is suited for protected agriculture.

13.2.5. Fertigation

It is also possible to supply the nutrients to the crops through the pressurized system of irrigation popularly called as fertigation. Supplying liquid-soluble fertilizers through a drip system can lead to savings in fertilizer applied to the extent of 40% to 60% without affecting the yield and a much higher application efficiency compared to the conventional methods. This is in addition to the savings in water applied which could be anywhere from 40% to 70%. The quality of produce is also superior. These aspects will play a decisive role in the adoption of these technologies when the subsidies to the two vital inputs to agriculture, water and fertilizer, are reduced which is inevitable in future.

13.2.6.Use of Poor and Marginal Quality Waters for Irrigation

Due to intense competition with domestic, industrial and power sectors, agriculture will have to depend more on the use of marginal/ poor quality of water, as it is projected that the share of water allocated to agriculture is likely to decrease by 10-15% in the next 20-30 years. One of the advantages of pressurized irrigation systems is that they can be used with poor quality waters also. In addition to the specific maintenance requirement of the system, it is important to ensure removal of salts or other chemicals periodically to prevent their critical concentrations, which will ultimately affect production. If the field can be saturated or flooded at the time of seed germination, much of the salinity can be leached out, if only for a brief period. Most species are more susceptible to salt effects during germination and are more resistant during other portions of their life-cycle. With this system, the soil salinity will remain below the crop's damage threshold throughout the year, although during sometimes in the season salt levels may become quite high. In case of drip/sprinkler irrigation systems, it is also essential to follow an application schedule that encourages deeper root system in trees (horticulture/orchard crops) for anchorage in regions which are prone to high wind speeds.

14. Scheduling of Irrigation to Crops

Irrigation scheduling — the timing and amount of irrigation to crops — plays a significant role in optimizing crop production with a given amount of water and avoiding the adverse effects of either over-irrigation or under-irrigation on soil environment. These objectives are achieved through optimum scheduling of irrigations based on sound scientific principles of soil-water-plant-atmosphere relationships. Approaches to irrigation scheduling vary depending on situations, e.g. (i) where adequate irrigation water is available on demand to secure potential yield, and (ii) where available supplies fall short of the full irrigation water requirement of crops over the entire command area.

14.1. Plentiful Water

Where ample water supplies are available for meeting the water requirements of crops over the entire command area, growers' aim is to secure potential yields without wasting water. For such a situation, an optimum irrigation schedule is that time sequence of the amounts of needed minimum irrigation which may eliminate under- and over-irrigation and ensure high water-use efficiency. For securing potential yield, the following points need consideration:

(i) Irrigation must be repeated while soil water content is sufficiently high, and
(ii) Soil can and does supply water sufficiently fast to meet the atmospheric evaporative demand without putting the plants under stress that will reduce yield or quality of produce.

In any irrigation practice we have to answer two questions, viz.

(i) When to irrigate? and
(ii) How much to irrigate?

For optimum irrigation scheduling, we have to first look for the point on the yield versus soil water status curve at which yield starts declining so that the irrigation is applied before this point is reached. Also, we should have means of knowing the soil-water deficit in the root zone when this point is reached. Several criteria that are used for finding out this point can be classified into three categories. These are based on:

(i) Soil-water regime,
(ii) Meteorological parameters, and
(iii) Plant indices.

14.1.1. Soil-Water Regime

The criteria for scheduling irrigation based on soil-water regime use soil water content, available soil water depletion and soil water tension to find the threshold point for irrigation.

Soil-Water Content — Yields are related to soil-water contents in assumed root zone at which irrigations are applied and the soil-water content is identified where yields begin to decline. Alternatively, fixed time intervals between successive irrigations are tried experimentally and the most suitable interval is identified. These criteria suffer from lack of generalization because both threshold soil-water content and fixed time intervals for a given crop vary widely with soil-water characteristics. It has been suggested that the soil-water content at which the ratio of actual to potential evapotranspiration (ET) falls below a threshold value for potential yield is the point before which irrigation must be applied. This soil-water content has been termed as lower limit for potential ET (LLE_p). Once the threshold limit for application of irrigation is established, irrigation could be repeated any time before this point is reached. Water addition should be only sufficient to make up the soil water deficit and keep the salts below the root zone.

Available Soil-Water Depletion — Irrigation is advocated on the basis of the fraction of plant available soil-water depletion from a fixed root zone usually the upper 15 cm. In view of the ever-deepening characteristics of the root system, at least till flowering, the assumption of fixed shallow root zone is not valid.

Soil-Water Suction — The soil-water suction measured by tensiometers has also been used as a criterion for scheduling irrigation to crops. Tensiometers are installed at specified depths to monitor soil water suction (or tension) for irrigation purpose. Soil water suction integrated across the root zone depth is a more dependable criterion than soil water suction of a shallow layer. The traditional tensiometers have a narrow working range of 0-80 kPa which render this method unsuitable for fine-textured soils, but now new devices with higher range are also available. It may be noted that this technique does not measure the amount of water needed at each irrigation. This has to be assessed through other means.

The response of crops to irrigation is better correlated with soil moisture tension than with soil-water content. Crop yield is related to soil moisture tension at which irrigation is repeated and the period is chosen when yield starts declining. This approach is applicable to fruit and vegetable crops which require more frequent irrigations and on sandy soils in which most of the available water is depleted at tensions well below 80 kPa.

14.1.2. Meteorological Parameters

Irrigation can be scheduled on the basis of water lost from the canopy and the permissible soil water depletion for potential yield. The former is determined by the ambient environment and the degree of crop cover and is obtained from the potential ET and the crop cover factor. Now computers can be used to advise the farmers about the timing of irrigation based on water budgets of individual fields prepared from the potential evapotranspiration (PET) rates and initial soil water storage. A serious problem with this approach is one of establishing the suitable crop factors. Moreover, it is now being increasingly recognised that under field conditions soil-water depletion does not necessarily equal ET by the crop. Water is also lost by deep percolation between irrigations. Another approach based on pan evaporation has been suggested. In this approach the ratio of fixed amount of water (in each irrigation) and evaporation from open pan is used as the basis for irrigation. It is mathematically denoted as:

$$\frac{\mathrm{IW}}{\mathrm{Pan-E}}$$

where, IW denotes the fixed amount of water in each irrigation (based on the permissible profile water depletion) and Pan–E is the net cumulative evaporation from open pan since previous irrigation. It is particularly convenient during low rainfall *rabi* season. Wheat sown on deep alluvial soils after pre-sowing irrigation and given first irrigation after four weeks could be irrigated on the basis of IW/(Pan–E) ratio of 0.75 to 0.90 in northern India. This approach could save irrigation water without reducing the yield (Table 18).

Intervals between successive irrigations, based on the optimum IW/(Pan-E) ratio, can be worked out using average pan-evaporation

Table 18. A comparison of irrigation schedules in wheat (average of two years)

Irrigation		Mean irrigation water (cm)	Mean profile water depletion (cm)	Mean grain yield (kg / ha)
Amount	Basis of timing			
7 cm	5 growth stages	35	15.5	4260
0.9 PAN-E	5 growth stages	27	14.6	4060
0.10 PAN-E	5 growth stages	23	15.4	4145
7 cm	IW / (PAN–E) = 0.9	23	18.5	4080
7 cm	IW / (PAN–E) = 0.75	22	15.6	4125
L.S.D. (0.05)		–	–	NS

The mean rainfall during these seasons was 800 mm for a few years and time tables of the type given in Table 19 can be recommended for different crops for various locations (Prihar and Sandhu, 1987).

14.1.3. Plant Indices

Several plant indices ranging from simple appearance of plants, their rates of growth, plant water stress measured as relative leaf water content, xylem water potential, leaf diffusion resistance and canopy temperature in relation to ambient air temperature have been investigated as the criteria for scheduling irrigations to crops. It is generally agreed that the plant integrates the effects of soil-water availability and the evaporative demand and the growth and yield are also found to relate with certain plant-water stress. However, it has not been possible so far to exploit these indices for the practical irrigation scheduling. These parameters only indicate when to irrigate but not how much to irrigate.

Canopy temperature appears to be the most promising index which could be exploited for scheduling irrigations (Jackson, 1982). Certain instruments like infrared thermometers are easy to operate and permit rapid determination of canopy temperature.

14.2. Limited Water Supplies

Irrigation scheduling for optimizing production with limited water supplies is a bigger challenge than that with adequate water supplies. As stated earlier, the first step in scheduling irrigations with limited water is to assess the relative sensitivity of different growth stages of a crop to water stress.

15. Special Purpose Irrigation

Irrigations are often applied to save crops from frost damage, even if the soil is sufficiently moist to meet transpirational requirements of crops. Similarly, irrigations are applied to soften the crust that develops on seedbeds before the emergence of seedlings. Early post-seeding irrigations which do not conform to the given criteria of scheduling irrigation are reported to enhance availability of surface applied nutrients and promote root extension to deeper layers.

16. Adverse Effects of Irrigation

While irrigation does help in increasing and stabilizing the crop production in water deficient areas, it may have short-term and long-term adverse effects on crop yields and soil productivity. Heavy irrigation (or rain) on soils of low permeability may cause anaerobic conditions resulting in aeration stress to plant roots for variable lengths of time which reduces crop yields. Similarly, over irrigation on excessively permeable and low retentivity soils may leach down the native or applied plant nutrients out

Table 19. Irrigation time table for wheat sown after a heavy pre-sowing irrigation on deep alluvial soil in Punjab

Date of sowing	First irrigation	Second irrigation 7.5 cm	Third irrigation 7.5 cm	Fourth irrigation 7.5 cm
	Days after sowing/previous irrigation			
Nov. 1-7	28	36	43	26
Nov. 8-21	28	41	38	25
Nov.22-Dec.5	28	44	30	15
Dec.6-20	28	35	25	15
Dec.21-Jan.4	28	30	20	14
Jan.5-Jan.15	28	28	18	12

* First irrigation should be comparatively lighter

Notes: This time-table is not applicable if any one or more of the following conditions prevail:

(a) A crop has been sown without giving a heavy pre-sowing irrigation

(b) A hard impervious layer of lime concretions (*kankar*) or clay lies in soil profile within one metre from the surface

(c) The soil is excessively sandy or saline of the root zone and, thus, make them physically inaccessible to crop. This, too, results in reduced yield.

The long-term adverse effect of irrigation is one of raising the groundwater table. The rising water, if not intercepted, ultimately causes waterlogging, often accompanied by salinity. The development of waterlogging and salinity in some areas of Punjab, Indira Gandhi Canal Command Areas of Rajasthan, several parts of Haryana and low-lying areas of Tungabhadra canal region are examples of such an effect (Bhumbla, 1984) (Table 20). Provision of canals is essential for agriculture in arid and semi-arid areas, but permanent irrigated agriculture is not possible unless effective steps are taken to control the rise of water table. The surface drainage is of limited avail in this regard and the water table can be controlled through sub-surface drainage. Where quality of ground water is good, shallow tubewells may be sunk to augment irrigation water supplies and prevent rise of water table. Shallow water tables with good quality ground water can be exploited *in situ* to partly or fully meet the water requirements of crops.

17. Irrigation with Poor Quality Water

It is estimated that in India about 25% of ground water is of poor quality. The use of saline and sodic water has till recently been guided by the standards laid down in the U.S. Salinity Laboratory Handbook 60. However, these standards are not applicable under Indian monsoonal type of climate. We have evolved our own classification of saline and sodic water for arid and semi-arid areas and these are given in Table 21.

The poor quality waters can be used conjunctively with good quality water by either mixing the two or irrigating with poor and good quality water in suitable succession. The quality of water resulting from the blending of good and poor quality waters can be computed from the quality of the respective waters by Equations (10) and (11):

$$EC_{iw} = (EC_{pw} * f_{pw}) + (EC_{gw} * f_{gw}) \qquad \ldots(10)$$

$$RSC_{iw} = (RSC_{pw} * f_{pw}) + (RSC_{gw} * f_{gw}) \qquad \ldots(11)$$

where, EC denotes the electrical conductivity, RSC denote residual sodium carbonate, and the subscript 'iw' stands for irrigation water and 'pw' and 'gw' denote poor quality and good quality waters, respectively and 'f' is the fraction of water. The fractions of poor and good quality water can be adjusted to have an irrigation water within permissible limits.

18. Drainage

The purpose of drainage is to remove the excess water from soil to ensure proper soil aeration. When soil gets saturated with water, the entire pore space is occupied by water and soil aeration is cut off. Also, water ponded at soil surface cuts off gaseous exchange between atmospheric air and soil air even if there are enough air-filled pores in the soil below. Draining out the excess water to create air porosity creates soil conditions favourable for soil aeration. Drainage is of two types, viz. (i) Surface drainage, and (ii) Subsurface drainage.

Table 20. Rate of water table rise in some irrigation commands in India

Name of irrigation command	Water table rise (cm/yr)
Mahi Right Bank Canal Command, Gujarat	28
Rajasthan Canal Command (IGNP), Rajasthan	29-88
Western Jamuna and Bhakra Canal Command, Haryana	30-100
Sirhand Canal Command, Punjab	10-100
Sharda Sahayak Canal Command, Uttar Pradesh	68
Malprabha Canal Command, Karnataka	60-120
Nagarjuna Sagar Irrigation Project, Andhra Pradesh	32

Source: Gupta and Tyagi (1996)

Table 21. Classification of saline waters (RSC <2.5) for irrigation of soils of different textures for crops sensitive, semi-tolerant and tolerant to salinity

Soil texture and percentage of clay	Crop tolerance to salinity	Upper limits of EC_{iw} (dS/m) in different regions, mm		
		< 350	350-500	500-750
Fine > 30%	Sensitive	1.0	1.0	1.5
	Semi-tolerant	1.5	2.0	3.0
	Tolerant	2.0	3.0	4.5
Moderately fine 20-30%	Sensitive	1.5	2.0	2.5
	Semi-tolerant	2.0	3.0	4.5
	Tolerant	4.0	6.0	8.0
Moderately coarse 10-20%	Sensitive	2.0	2.5	3.0
	Semi-tolerant	4.0	6.0	8.0
	Tolerant	6.0	8.0	10.0
Coarse < 10%	Sensitive	-	3.0	3.0
	Semi-tolerant	6.0	7.5	9.0
	Tolerant	8.0	10.0	12.5

Source: Gupta *et al.* (1994)

18.1. Surface Drainage

For the removal of the surface-accumulated water, surface drainage methods are adopted. On a sloping land, water does not accumulate over the land and most of it flows out by gravity. Therefore, surface drainage methods are more applicable in flat lands or lands with mild (<0.5%) slopes. For flat land drainage, a network of drains in a hierarchical order, from field drains to main drain are established. The lowest order drains are constructed in the field or at the lower boundary of the field. These drains are shallow about 15 cm deep and are triangular or trapezoidal in cross-section. They are referred to as "field drains". Their requirement exists in the monsoon season in providing a smooth access of the overland flow into it and carrying this water towards the hierarchically higher order drain. Hence, they need to be constructed or renovated before the monsoon arrives. The distance between adjacent field drains may be between 100 m and 200 m and may have a longitudinal slope of 0.1% to 0.2%. A number of such field drains convey their water to a "collector drain", which further conveys the water towards the main drain. Collector drains are trapezoidal in cross-section, 40 to 50 cm deep with 1:1 side slope and 0.2% to 0.3% longitudinal slope. Generally, the collector drain capacity is between 5 to 10 Lps/ha. A number of collector drains lead their water into a main drain, which takes the excess surface water out of the drained agricultural land. The capacity of the main drain depends on the number of collector drains joining it. The above-mentioned depth, spacing and slope are only general guidelines and they vary depending on local topography, rainfall, field boundary and size, soil type, crops grown and outlet condition. Regular maintenance of surface drainage network is essential to keep them functional at the time of maximum need, mostly in the monsoon season.

18.2. Sub-surface Drainage

Sub-surface drainage is accomplished through a network of pipe drains buried 1 to 2 m below the land surface. This is used for controlling shallow water table and for leaching of harmful salt solutions in the soil profile. The drainpipes are usually of baked clay or plastic material. Baked clay pipes are normally 30 cm to 60 cm long and 10 cm to 15 cm in diameter. The most common plastic drainpipes are made in continuous length of 200 m to 500 m and are corrugated PVC pipes with small perforations on their surface. For commissioning a sub-surface drainage system, a trench is dug to the desired depth, its bottom is smoothened and is given a uniform slope of 0.05% to 0.2%. A layer of filter is placed at the bottom of the trench. A common filter is a mixture of sand and small size gravel. The drainpipe is laid over the filter bed and is covered from sides and top by filter. The trench is then backfilled. Organic filters such as paddy straw or coir fibre may also be used at the top and at the sides of the drainpipe. Synthetic filter of nylon fabric is also used, particularly for the corrugated PVC drainpipes. If clay pipes are used, the fine gaps between the adjacent pipes convey the excess water in the soil profile into the drainpipe.

Like in the case of surface drainage, a sub-surface drainage network also has a hierarchical order of drains, from the lowest level lateral drains to the highest-level main drain. The final outlet drain may be an open drain if topography permits gravity flow into it from a drainage network buried 1 to 2 m below the ground. Otherwise, the sub-surface drained water is collected in a sump well from where it is pumped and disposed in any available shallow surface drain nearby.

The problem of waterlogging and soil salinization is increasing in the commands of many irrigation projects, especially in the arid and semi-arid areas. It is estimated that almost 2.5 Mha are has become waterlogged and 3.3 Mha has been affected by salinity in the irrigation commands. Inefficient use of irrigation water is the primary cause for converting these highly productive lands into wastelands. To improve the soil environment and make it more conducive and favourable for crop growth, surface or subsurface drainage is the only effective alternative (Table 22). Investigations have revealed that the cost of restoring the productivity of these affected lands is very high and can totally upset the economy and durability of the projects.

In many coastal areas, excessive pumping of groundwater through tubewells leads to deterioration of water quality due to seawater intrusion. For skimming fresh water, Dorouv technology/infiltration gallery methods have been found beneficial to prevent seawater intrusion. Similarly, in many island areas overdrawal of good quality groundwater leads to lowering of water table and acute water shortage. Scientific regulation of the intensity of tubewells according to local aquifer characteristics, is highly essential to ensure a proper water balance.

Table 22. Effect of subsurface drainage (SSD) on crop yields in Chambal Command (Rajasthan)

Crop	Yield (t/ha)		
	Without SSD	With SSD	Percent Increase
Wheat	2.2	3.4	55
Mustard	0.8	1.5	84
Berseem	36.0	52.0	44
Paddy Rice	1.9	2.9	53
Soybean	1.4	2.2	56

19. References

Abrol, I.P. (1988) Dryland farming — The Indian experience. In: *Challenges in Dryland Agriculture — A Global Perspective*. Proceedings of International Conference on Dryland Farming, August 15-19, Amarillo/Bushland, Texas, USA, pp 50-53.

Aggarwal, M.C. and S.S. Khanna (1983) Efficient soil and water management in Haryana. *Technical Bulletin of Haryana Agricultural University, Hissar.*

Annonymous (1986) *Annual Report*, ICRISAT. Patancheru.

Annual Progress Report (1993-94) Integrated Research Project for Soil Conservation and Watershed Management, College of Agriculture, Indore.

Bhumbla, D.R. (1984) *Small Reservoirs – A Programme for Improved Rainfed Agriculture.* Policy paper series. Society for Promotion of Wasteland Development. New Delhi 110 057.

CWC (Central Water Commission) (2008) *Water and Related Statistics*, Ministry of Water Resources, Govt. of India. New Delhi.

Chaudhary, T.N. (1997). *Water Management in Rice for Efficient Production. Research Bulletin No.* 1. Directorate of Water Management Research. Patna, Bihar.

Soil Erosion and Soil Conservation

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1. Introduction

Soil is a dynamic natural body on the surface of the earth and is a critical resource for supporting plant growth. It is composed of minerals, organic materials and living forms and provides the necessary nutrients to the growing plants. Water acts as a nutrient carrier as well as an agent of soil erosion. Mismanagement and indiscriminate use of soil and water resources result in land and environmental degradation and may prove disastrous for humankind as well as animals. Hence, judicious management and conservation of soil and water are essential for realizing sustainable productivity and environmental benefits.

2. Soil Erosion

Soil erosion is defined as the detachment of soil particles, their transportation from one place to another and deposition elsewhere through water, wind, coastal waves, snow, gravity and other forces. Soil erosion involves three steps, viz. (i) detachment of soil particles from the main soil body, (ii) their transportation by splashing, floating, rolling, dragging, and (iii) their deposition at another place. The major factors which encourage detachment of soil particles are: rainfall impact, vegetation destruction, freezing and thawing, flowing water, and wind velocity. Rain drop splash, runoff water and blowing wind facilitate the carrying away of soil to distant places. In gullies, most of the loosening and cutting of soil are due to water flow, while on flat soil surfaces, beating action of rain drops causes the detachment of soil particle.

Soil erosion due to runoff is more on slopping and denuded lands. In high rainfall areas, loss of water may be as high as 50 to 60% of the annual rainfall. The loss of water through runoff could be reduced by suitable methods. The soil erosion accompanying runoff is usually more serious. The surface soil is taken away by the running water causing loss of soil along with nutrients, both native and applied. Since erosion starts with the loss of finer fraction of soil, the soil fertility is depleted. The loss of soil fertility and top soil makes the land unsuitable and unproductive for biomass production. In certain cases, the nutrient losses exceed the gains through renewal by the growing crops.

In deserts, the formation of sand dunes and their erosion due to high wind velocity are common phenomena. Under normal physical, hydrological and biotic equilibria, the erosion or removal of soil should be balanced with soil formation processes. If the balance gets disturbed due to biotic interference or natural factors, the eroding agencies become more active and cause accelerated soil erosion.

2.1. Geological Erosion

It refers to natural or normal erosion. It represents the erosion when the land is in natural equilibrium under the cover of vegetation.

Geological erosion takes place as a result of the action of water, wind, gravity and glaciers and its rate is so low that the loss of soil is compensated by the formation of new soil under natural weathering processes. Geological erosion may be considered a part of the natural soil forming processes which result in the existing form and distribution of soils. This kind of erosion does not pose any problem.

2.2. *Accelerated Soil Erosion*

When soil erosion exceeds the normal rate and becomes unusually destructive and unproductive, it is called accelerated erosion. When man or animal interferes with the existing balance in nature, by way of removal of vegetative cover, indiscriminate cutting of trees, overgrazing, faulty agricultural practices, etc., there is deterioration and loss of soil. This signifies accelerated erosion. Accelerated soil erosion has been a major process in degradation of land resources and has destroyed about 430 million hectares (Mha) of area in different countries or about 30% of the cultivated land of the world or 50% of the geographical area of India. This erosion often results in vast deposits from sedimentary rocks.

2.3. *Extent of Soil Erosion in India*

A precise estimation of the extent of land affected by soil erosion is still lacking. As per earlier estimates, out of 305.9 Mha of the reported area, about 50% (145 Mha) needs conservation measures and about 25% of the reported utilizable land is precariously degraded, needing immediate soil-conservation measures. This includes sub-humid and per-humid regions, Western Ghats, coastal and north-eastern hill regions afflicted with severe erosion, and facing soil loss of 20-40 tonnes/ha/year.

Agricultural land in major parts of the country suffers from the decline in fertility due to loss of soil matrix which produces runoff. These areas need conservation for sustaining production. For example, in Maharashtra, over 70% of the cultivated land has been affected by erosion in varying degrees and 32% of the highly eroded land is no longer cultivable. In the Sholapur district, nearly 17% of the land of medium depth (> 45 cm) was converted into shallow soils (< 45 cm) during 1870 to 1945. Similarly, in the districts of Akola, Buldhana and Yeotmal, the number of fields with less than 37.5 cm soil depth increased during the same period by 54%, 16% and 8%, respectively. As much as 2.68 Mha land is already under ravines scattered all over India and is a constant threat to the adjoining fertile cultivated lands.

In the Himalayan region, landslides and landslips pose a very serious problem due to improper land management. The landslide disasters in the winter of 1969 in the Darjeeling district of West Bengal, and in 1998 in the Pithoragarh district of Uttarakhand are eye-openers of future potential hazards. The erosion problem along the mountain roads is assuming alarming proportions. The Border Road Organization spends huge resources to stabilize the slopes along the highways for keeping these important lines of communication serviceable.

2.4. *Consequences of Soil Erosion*

Soil erosion has both on-site and off-site effects. It is the severest form of land degradation, which has assumed national as well as global importance.

2.4.1. *Land Degradation*

Soil degradation is a major threat to our food and environmental security. Many thriving civilizations have vanished in the past because of inadequate attention to land care. Various agencies have been constantly inventorizing and characterizing soil resources for drawing up land use and land developmental plans (Table 1). The estimates are at great variance due to differences in approaches and criteria for assessment. The National Bureau of Soil Survey and Land Use Planning, Nagpur (ICAR) has furnished the latest data on soil degradation on 1:250,000 scale for different states of the country. About 45% of the total geographical area of the country is degraded due to different degrading agents, in which

Table 1. Estimates of soil degradation in India by different agencies

Agency	Estimated degraded area (Mha)	Criteria for delineation
National Commission on Agriculture (1976)	148.09	Based on secondary data
Ministry of Agriculture (1978)	175.00	Based on National Commission on Agriculture (NCA) estimates
Society for Promotion of Wastelands Development (1984)	129.58	Based on secondary collected data
National Remote Sensing Agency (1985)	53.28	Mapping on 1:1 million scale based on remote-sensing techniques
Ministry of Agriculture (1985)	173.64	Land degradation statistics for the states
Ministry of Agriculture (1994)	107.43	Elimination of duplication of area
National Bureau of Soil Survey and Land Use Planning (1994)	187.70	Mapping 1:4.4 million scale based on Global Assessment of Soil Degradation (GLASOD) guidelines
National Rainfed Area Authority (NRAA) (2008)	120-72	Harmonization of various maps

more than 65% of the land degradation is due to water and wind erosion (NBSSLUP, 2004).

2.4.2. *Soil Degradation and Loss in Soil Productivity*

Soil degradation through the loss of topsoil is one of the major factors of low and unstable crop yields in the rainfed semi-arid to subhumid subtropics of India. The study on the effect of surface/topsoil depth on the yield of rainfed crops grown in an Alfisol soil has revealed that yield responses are up to 2.5-times higher in soils with deeper topsoils when rainfall in the critical period exceeds evapotranspiration than under drier conditions (Vittal *et al.*, 1990).

Various agronomic experiments have shown significant decline in yield of different crops with increasing degrees of erosion by water, slope per cent and decreasing soil depth. The estimated loss in crop productivity on soils showing different degrees of erosion by water has been shown in Table 2. The reduction in yield is significant when the soil depth decreases. The erosion classes may be modified with availability of more data on kind, degree and extent of problem vis-à-vis loss of productivity of different soils. It may also be noted that a moderate water erosion in alluvium-derived deep soils may show significantly less reduction in soil productivity as compared with the deep Red and Black soils (Table 2). Similarly, in deep alluvial silty loam soils at Dehradun, each centimetre desurfacing of soil caused 76 kg/ha decline in maize yield. This reduction is likely to be more severe in shallow soils.

2.4.3. *Siltation of Reservoirs*

The off-site effects of water erosion include siltation of reservoirs. Multipurpose reservoirs constructed at huge costs under the river-valley projects are getting silted at an alarming rate due to excessive soil erosion. It has been estimated that 5,334 million tonnes of soil (16.35 tonnes/ha/annum) is lost annually, which is 4-times higher than the permissible soil loss (4-5 tonnes/ha/annum). About 29% of the eroded material is permanently lost to the sea, 61% is dislocated from one place to another, and 10% is deposited in the reservoirs, resulting in loss of storage capacity by 1-2% per annum. The data on river valley projects on 17 medium and small reservoirs in India has shown that the rate of inflow of annual sediment rate is 3-fold (9.17 ha-m/100 km^2) in comparison to the designed rate of 2.93 ha-m/100 km^2/annum; thus reducing their life expectancy and hydro-electric power generation capacity to only one-third.

Table 2. Estimated loss of soil productivity due to erosion by water in different soils

Soil erosion class	Soil loss (t/ha)	Loss in productivity (%) in		
		Alluvium-derived soils (Inceptisols)	Black soils (Vertisols)	Red soils (Alfisols)
Nil to very slight	< 5	Nil	< 5	< 10
Slight	5 - 10	< 5	5-10	10 - 25
Moderate	10 - 20	5 – 10	10 – 25	15 – 50
Strong	20 – 40	10 – 25	25 – 50	> 50
Severe	> 40	25 – 50	> 50	N.A.

2.4.4. Floods and Flood Plains

The soil erosion, siltation and reduced transport capacity of rivers, *chos* and drains, and serious catastrophies, viz. floods and landslides cause large-scale damage to animal as well as human life and property. Productive agricultural land might go out of cultivation due to soil and nutrient erosion or silt deposition, in the flood plains. Flood water from the Brahmaputra river has been reported to carry a sediment load of 1500-30000 mg/L of soil, 6.4-25.8 mg/L of NO_3-N, 2.3-8.5 mg/L of PO_4-P, 15.4-33.8 mg/L of K_2O, 2.5-6.4 mg/L of Ca, 6.5-14.0 mg/L of Mg, 5.0-8.5 mg/L of SO_4^{2-} and 7.5-22.7 mg/L of micronutrients (Sharma and Sharma, 2004)

2.4.5. Environmental Pollution

Sediment is one of the major pollutants. In India, the situation is still worse, since the sediment load from agricultural lands not only continues unabated, but is also on the increase due to the fast rate of our developmental activities. If erosion continues unchecked at its present rate, we shall be left with the reclamation of soil rather than its conservation and management.

2.4.6. Changing Forest Cover

India has 2% land, 10% forests and 0.5% grazing area of the world, but has the burden of supporting 18% human and 15% cattle population. The recorded forest area of the country is 67.7 Mha. Out of this, 1.66% is very dense (more than 70% canopy), 10.12% is moderate dense (40-70% canopy) and 8.82% is open forest (10-40% canopy). Apart from that, 3.8 Mha is scrub or non-forest cover (canopy <10%) (FSI, 2005). About 29 Mha area is under degraded forests. The forest cover is declining due to the degradation of permanent pastures and open forests in the desert and arid regions.

2.4.7. Loss of Biodiversity

Loss of biodiversity is one of the major global concerns today. India's biodiversity is unique in the world. As many as 45,000 species of wild plants and over 77,000 of wild animals have been recorded, which comprise 6.5% of the world's known wildlife. An assessment of the loss of wildlife habitat in tropical Asia in 1986 showed that the country had only 615,095 km² out of its original wildlife habitat of 3,017,009 km², i.e. loss of 20%. Loss of seeds and propagules in the erosion fluxes, washing away of nutrients, run-offs, etc. depletes the biodiversity. During the past few decades, India has lost at least half of its forest, has polluted over 70% of its water-bodies and has degraded most of its coasts.

3. Factors Affecting Soil Erosion

Numerous factors affect soil erosion depending upon the local conditions with regard to physical, chemical and biological nature of the soil. The major factors affecting soil erosion are :

(i) Climate, especially rainfall and wind velocity – Their intensity in time and space

(ii) Topography – Slope of land, its nature, degree and length

(iii) Physical, chemical and biological characteristics of the soil

(iv) Nature of ground cover and land use

(v) Biotic interference due to human beings and animals

(vii) Conservation practices, and

(viii) Mass erosion, such as earthquakes, land slides, torrents (flash) floods, etc.

Mathematically, soil erosion may be depicted by functional Equation (1):

$$\text{Erosion} = f(cl, v, t, s, h, \ldots) \qquad \ldots(1)$$

where, f, is the function of climate (cl), vegetation (v), topography (t), soil (s), human and animal activities (h).

3.1. Climate

Rainfall is by far the most important factor causing soil erosion by water. The soil erosion depends on the intensity, kinetic energy, amount, duration and frequency of rainfall. The intensity, duration and frequency of rainfall govern the rate and the volume of runoff. A light rain that can be easily absorbed in the soil causes no runoff and soil loss. When the intensity of rain is more than infiltration rate, it causes runoff and soil loss. The principal effect of rain drops is to detach soil (splash) and the principal effect of surface flow is to transport the detached particles. The interaction of kinetic energy (KE) [$\frac{1}{2}mv^2$] and rainfall intensity (say I_{30}) expressed as erosion index (EI_{30}) could be best related with rainfall erosion as per Equation (2):

$$EI_{30} = KE \times I_{30}/100 \qquad \ldots(2)$$

where, EI_{30} is the erosion index; KE is the kinetic energy of storm, 210.3 + 89 log I; I_{30} is the maximum 30-min intensity of storm; and I is the rainfall intensity (cm/hr), which can be computed through automatic raingauge charts. The sequence of storms and rainfall distribution also determines the extent of soil erosion. Wischmeier and Smith (1958) had used precipitation and soil loss data from fallow plots to determine the best characteristics of rainfall for estimating storm soil loss. The results indicated that the rainfall characteristic best for estimating single storm soil erosion was the product of the total kinetic energy of the rainfall and the maximum rainfall intensity over a continuous 30-minute period during the rainstorm. This was also known as EI or the R factor (Laflen and Moldenhauer, 2003).

3.2. Topography

The land slope, its degree and length, is important in determining the extent of soil erosion. Higher the degree of slope and longer is its length, more is the soil erosion due to increased velocity of water running down the slope. If the land slope is increased 4-fold, the velocity of water flowing down the slope becomes approximately doubled. If the velocity of the runoff water is doubled, its energy, i.e. erosive power, increases 4-times, because the latter varies as the square of the velocity. Similarly, the quantity of material of a given size that can be carried is increased 32-fold (varies as the fifth power of the velocity), and the size of the particles that can be transported by pushing or rolling is increased 64-fold (varies as the sixth power of the velocity). Similarly, the velocity of winds and storms also affects soil erosion in the arid regions. Large sand dunes are transported from one place to another within no time due to wind action. Smith and Wischmeier (1957) evaluated the effect of slope and length on soil erosion for several locations. Data evaluated included slopes ranging from 1 % to 25%. The derived relationship was a quadratic relationship expressing the effect of slope on soil loss (Laflen and Moldenhauer, 2003) as:

$$S = (0.43 + 0.30s + 0.043s^2)/6.613 \qquad \ldots(3)$$

where, S is the slope factor and s is per cent slope. Smith and Wischmeier (1957) also evaluated the effect of slope length on soil erosion. They defined slope length as 'the distance from the point of origin of overland flow to either where the slope decreases to the point that deposition begins, or to the point where runoff enters a well-defined channel'. They have also found that soil loss per unit area varies as the

m^{th} power of slope length, expressed as Equation (4):

$$L = (\lambda/72.6)^{m} \qquad \ldots(4)$$

where, L is the slope length factor, λ is the slope length and m is the slope length exponent. It was observed that the value of m varies widely from year to year, even at times becoming negative. The effect of slope length on soil erosion is influenced by slope, vegetation, soil and management. The Universal Soil Loss Equation (USLE) used a value of 0.5, but values of 0.6 have been recommended for slopes steeper than 10%, and 0.3 for very long slopes under furrow irrigated conditions. For most situations, the length and slope factors have been combined into Equation (5):

$$LS = \lambda^{0.5}(0.0076 + 0.0053\,s + 0.00076\,s^{2}) \qquad \ldots(5)$$

3.3. *Soil Characteristics*

Erodibility of soil is influenced by its physical and chemical properties, including soil texture, structure, organic matter, nature of clay and amounts and kinds of salts present. Soil structure, texture, organic matter, infiltration and permeability influence the runoff and soil loss. Fine soils are more susceptible to erosion than coarse soils. The organic matter helps in binding the soil particles and improves the water holding capacity of soil. The three most significant soil characteristics which influence soil erosion are: (i) infiltration capacity, (ii) structural stability, and (iii) antecedent soil moisture. Various empirical equations like dispersion ratio and erosion ratio based on the ratio of sand + silt to clay and nomographs, have been used to estimate erodibility of soil. In India, the deep lateritic soils of Tamil Nadu and red soils of Bihar have the lowest rate of runoff, whereas, the alluvial soils of Gujarat and Uttarakhand have a very high rate of runoff.

3.4. *Nature of Ground Cover and Land Use*

The nature of vegetative cover on the soil surface such as crops, cropping systems and the accompanying vegetation, is a major factor influencing soil erosion. A thick mantle of plants reduces erosivity of rainfall and most of the water either quickly percolates through the soil or flows over the surface with non-erosive velocity. The areas which are not protected with thick cover of plants, are not able to absorb water effectively, because the falling rains shatter the soil surface. The fine soil particles go into a suspension and other particles form a thick mixture of water and soil, which quickly fills and clogs the tiny interspaces in the soil, reducing water infiltration and consequently increasing the runoff and soil loss. Moreover, the rooting system of vegetation also improves the binding capacity of the soil particles as well as soil infiltration. The effect of canopy is more when it is closer to the land surface. Hence, mulches being directly in touch with land surface are most efficient in reducing the runoff and soil loss, followed by grasses and crops.

The cropping and management factor (C) for the USLE is defined as the ratio of soil loss from a particular cropping and management to the soil loss from a continuously tilled fallow land.

If the land is left undisturbed under a natural cover, the runoff and soil loss are the least, which increase steeply when the vegetation is removed and the land is cultivated (Table 3).

Along with the loss of runoff water and soil, considerable amounts of plant nutrients are also lost.

3.5. *Mass Landslides and Biotic Interference*

Soil erosion due to loosening and cutting of soil on the river banks is accelerated during floods. Earthquake shakes and de-stabilizes the soil surface, thereby rendering it prone to erosion in the form of mass landslides. Faulty agricultural practices without proper care for resource conservation like shifting cultivation and free range grazing by cattle cause disturbance and removal of top fertile soil. This results in soil fertility depletion and makes land unsuitable for cultivation. Construction of roads and railway tracks in the hills produce mass wasting.

Table 3. Effect of land use on runoff and soil erosion

Treatments	Runoff (mm)	Soil loss (tonnes/ha)
Alluvial soils : 8% slope (Uttarakhand)		
Bare fallow	339	42.4
Cultivated fallow	889	156.0
Natural grasses	265	1.0
Maize-wheat (up-and-down cultivation)	870	28.5
Black soils : 0.5% slope (Rajasthan)		
Natural cover	33	0.3
Cultivated fallow	111	3.5
Sorghum (*kharif*)	79	2.9
Red soils : 2% slope (Bihar)		
Natural fallow	105	0.6
Overgrazed fallow	222	3.3
Maize (contour cultivation)	64	3.3

3.6. *Conservation Practices*

The Conservation Practice Factor (P), also called as the Erosion Control Practice Factor, is the ratio of soil loss for a specific practice to the soil loss with up-and-down hill culture. The initial practices considered for the USLE were contouring, strip cropping, contour strip cropping and terraces. These were expanded to include contour listing, controlled-row grade ridge planting contoured residue strips and terraces of various types.

3.7. *K- Soil Erodibility*

Wischmeier and Mannering (1969), using a rainfall simulator, measured soil loss and computed soil erodibility on the data adjusted to the unit plot as the slope of the linear relationship between the rainfall factor and soil erosion. They related soil erodibility to a number of variables using multiple regression techniques. A major finding was that very fine sand behaves more like silt than like sand. Using these results, Wischmeier *et al.* (1971) developed a soil erodibility monograph that has proven to be easily usable for estimating soil erodibility for most soils.

4. Water Erosion

It is the removal of soil from the land surface by water, including runoff from melted snow and ice and is one of the major causes of soil degradation (Table 4) (Velayutham and Bhattacharyya, 2000). Water erosion has been sub-divided into various types in relation to progressive concentration of surface runoff. Different types of water erosion are described below:

4.1. *Splash Erosion*

A falling rain drop is capable of generat-ing a force equals to almost 14-times of its weight. A rain drop produces three important effects during splash erosion: (a) its beating action breaks down the soil aggregates, (b) it detaches soil particles, and (c) its splash after touching the ground causes displacement of soil. Under low intensity and low rainfall conditions, if the dispersed material is not removed by the runoff, it forms a hard crust after drying. Under such conditions, germination of seeds is hampered and sometimes, if this condition develops just after sowing, the crop stand remains poor due to very low germination. The rain drops beat the soil with force and splash may take away soil even up to 2 metres.

4.2. *Sheet Erosion*

The erosion in which soil matrix is lost but remains undetected for a long period and a thin venier of soil is removed from large areas uniformly during every rain, even producing a runoff, is called sheet erosion. The eroding and

Table 4. Land degradation status in India

Type of degradation (million ha)	Area
1. *Water erosion*	148.9
(i) Loss of top soil	132.5
(ii) Terrain deformation	16.4
2. *Wind eorosion*	13.5
(i) Loss of top soil	6.2
(ii) Loss of top soil/terrain deformation	4.6
(iii) Terrain formation/overblowing	2.7
3. *Chemical deterioration*	13.8
(i) Loss of nutrients	3.7
(ii) Salinization	10.1
4. *Physical deterioration (waterlogging)*	11.6
Total degraded area	187.8

transporting power of sheet flow is a function of the depth and velocity of runoff for a given size, shape and quantity of soil particles or aggregates.

Sheet erosion can result from: (a) surface creeping of the soil in uniform layers and its deposition at another place or washing away along with runoff in the drainage system, (b) saltation, which results when water causes soil particles to hop or slip during their downward movement, and (c) suspension of the soil particles in runoff as they do not touch the soil surface. Sheet erosion continuously makes the soil shallower with a progressive decline in crop yields. It is generally neglected, although the soil deteriorates slowly and imperceptibly. Its existence, however, can be detected by the muddy colour of the runoff arising from the fields/areas. The most serious sheet erosion occurs in red and black soils, occupying 69 Mha and 67 Mha of land, respectively in the country.

4.3. Rill Erosion

It is the removal of soil by water from small channels which develop due to concentration of surface flow along the slope. It generally develops on bare and unprotected land where soil conservation measures are not adopted. In advanced stage, it may take the shape of gullies. The rill erosion is an intermediary stage between sheet erosion and gully erosion. The rill can be removed by normal tillage operations. Generally, these rills vanish after cultivation, but may reappear with ensuing rains.

4.4. Gully Erosion

It is the advanced stage of rill erosion which cannot be obliterated by normal tillage operations. Rills with more than 30 cm depth are generally called gullies. Gullies obstruct the operation of farm machinery and tends to deepen and widen with every heavy rainfall. They cut up large fields into small fragments and in course of time, make them unfit for cultivation. The rate of gully erosion depends on several factors, viz. runoff producing characteristics of the watershed, drainage area, soil characteristics, channel alignment, slope and vegetative cover on the ground, etc. Gully erosion indicates misuse, neglect and mismanagement of land over a long period of time.

5. Special Forms of Erosion

Besides the four main water erosions described above, there are some special forms of erosion also. These are discussed below:

5.1. Landslides or Slip Erosion

During monsoon season or heavy down-pour, the hills get saturated with water, resulting in instability of the land mass. The complete surface portion of soil may slips down from its actual position causing landslides. Landslides are very common in hills and huge quantities of soil are washed away in streams and rivers. Hills devoid of vegetation are more prone to slip erosion, causing damage to agricultural lands, orchards, communication systems and human lives. Minor landslips occur during the drying phase in the months of May and June, especially in the sandy hills of Shiwaliks.

5.2. Stream Bank Erosion

It is caused by streams and torrents, also called *chos*, cutting their banks and adjoining fields. It is mainly due to runoff flowing over the side of the stream banks or by scouring and undercutting below the water surface, the latter process being more serious. It is influenced by deforestation and overgrazing. Scour erosion is influenced by the velocity and direction of water flow, depth and width of channel and soil texture.

5.3. River Bank Erosion

This is a common phenomenon in the rivers of the eastern and western Himalayan regions. During floods, the rivers undermine their banks and big chunks of land mass are engulfed by flood water. This leads to broadening of river beds and changes in their courses.

5.4. Slumping

It is a geological process and is one of the most common types of slippage. It may be defined as the downward slipping of a mass of unconsolidated matter or rock, moving as a unit or as several subsidiary units. It is an important process in widening of the stream channels and gullies under high rainfall areas.

5.5. Coastal Erosion

The strong waves of sea strike against the seashore causing large-scale soil erosion. In this case, erosion is a combined effect of wind and water.

6. Estimation of Soil and Water Erosions

Universal Soil Loss Equation (USLE) has been designed to predict soil loss from agricultural lands under various agro-climatic and topographical situations. The essence of the USLE is to isolate each variable and reduce its effect to a number so that when the numbers are multiplied together, the answer is the amount of soil loss.

The Universal Soil Loss Equation is presented by expression (6) (Wischmeier and Smith, 1978):

$$A = R \times K \times L \times S \times C \times P \qquad \ldots(6)$$

where,

A = Computed soil loss per unit area,

R = Rainfall factor, is the number of erosion-index units in a normal year's rain. The erosion index is a measure of the erosive force of specific rainfall,

K = Soil-erodibility factor, is the erosion rate per unit of erosion index for a specific soil in cultivated continuous fallow, on a 9% slope, 22 m long,

L = Slope-length factor, is the ratio of soil loss from the field slope length to that from a 22-m length on the same soil type and gradient,

S = Slope-gradient factor, is the ratio of soil loss from the field gradient to that from a 9% slope,

C = Cropping-management factor, is the ratio of soil loss from a field with specified cropping and management to that from the fallow condition on which the factor K is evaluated, and

P = Erosion-control practice factor, is the ratio of soil loss with contouring, strip-cropping or terracing to that with straight-row farming, up and down slope.

6.1. Soil Erodibility Factor

The soil erodibility factor (K) in the USLE is related to the rates at which different soils erode. The soil erodibility is different from soil erosion in the sense that the total erosion may be influenced by other factors like rainfall, crops, management of the lands, etc. However, a soil may erode relatively more than the other, although the rainfall, crop condition, etc. may be the same in these two soils. This difference which is caused by the inherent soil properties is referred to as "soil erodibility". The soil properties that influence erodibility are texture, structure, permeability, infiltration, organic matter and mineralogy. Usually, the deep, permeable, coarse sands are the least erosive.

To obtain the value of K, soil loss is required to be measured from a unit plot of 22.13 m × 1.83 m size having uniform slope of 9% in a continuous fallow, tilled up and down the slope. The soil loss obtained from above unit plot is divided by yearly EI_{30}, where EI_{30} is the cross-product of the total kinetic energy and the maximum 30-minuie intensity (I_{30})

(Wischmeier, 1959) to obtain the soil erodibility factor (K).

In case the plot gradient and length are different than the defined unit plot, the soil loss may be adjusted by dividing it with the topographic factor (LS).

The value of K can be estimated from Equation (7):

$$K = \frac{\text{Total adjusted soil loss}}{\text{Total } EI_{30}} \qquad \ldots(7)$$

6.2. Estimation of Sedimentation Yield

For estimation of sedimentation yield in runoff plots/ watersheds, the first runoff is measured through the flumes with the help of stage level recorders. A portion of the total runoff can then be collected with the help of devices like Coshocton wheel and this runoff sample is then used for the determination of soil loss from the plots/ watershed. The runoff is calculated using Equation (8):

$$\text{Runoff (mm)} = \frac{\text{Runoff (litres)}}{\text{Area of plot } (m^2)} \qquad \ldots(8)$$

For estimating the soil loss, the runoff collected in the tank is churned thoroughly and a small runoff sample of about 500 mL to 1000 mL is drawn. The sample is then evaporated in the laboratory and the amount of soil is measured gravimetrically to give soil loss in g/litre. This quantity is multiplied by the total runoff volume and divided by 10^6 to give total soil loss in tonnes per plot. It is further multiplied by the suitable factor as per size of the plot to get the soil loss in tonnes/ha, for that particular storm. The soil losses from all the storms during one year are added to get the annual soil loss in tonnes/ha/year (Singh *el al.*,1990).

7. Development of Universal Soil Loss Equation (USLE)

The evolution of Universal Soil Loss Equation (USLE) is one of the most significant developments in soil and water conservation. It is an empirical technology that has been developed in an evolutionary manner during about past 60 years and is still undergoing evolution with the development of various revisions (Laflen and Moldenhauer, 2003). The USLE is the contribution of many individuals over a very long period of time. The period from 1930 to 1942, was the golden period of conservation research. Most of the work related to erosion prediction was conducted by USDA scientists (Soil Conservation Service). Duley and Ackerman (1934) had mentioned about the effect of slope and length on soil erosion but no relationship was described. Zingg (1940) evaluated data from field experiments under natural rainfall and from rainfall simulation experiments and proposed the relationship (9):

$$X = C\,S^m L^n \qquad \ldots(9)$$

where, X is the total soil loss from a land slope of unit width, C is a constant of variation, S is the land slope (%), L is the horizontal length of land slope and m and n are exponents. Soil loss per unit area from a land slope of unit width may be expressed as per Equation (10):

$$A = C\,S^m L^{n-1} \qquad \ldots(10)$$

The Zingg's equation was expanded by Smith (1941) to Equation (11):

$$A = C\,S^{1.4} L^{0.6} P \qquad \ldots(11)$$

where, P is the ratio of soil loss with a mechanical conservation practice to soil loss without the practice. Smith (1941) used Equation (11) with measured annual values of A, S and L from individual plots to compute C values for various treatments. Browning *et al.* (1947) developed the "erodibility factor". Musgrave (1947) used vegetal cover factor that expressed relative soil erosions for different vegetative covers.

$$\text{Soil loss} = (R)^{1.75} (K) (L)^{0.35} (S)^{1.35} (C) \qquad \ldots(12)$$

where, R is 30-minute maximum rainfall, K is soil erodibility, L is the length, S is the slope and C is the vegetal factor.

In an example presented for the soil (erodibility of 0.33 inches/year), where 30 minute maximum rainfall was 1.35 inches, slope was 5%, length was 150 feet and the crop was wheat (vegetal factor 0.2), the average predicted soil loss was :

$$\text{Soil loss} = (1.35)^{1.75}\,(0.33)\,(150\ \text{ft})^{0.35}\,(5)^{1.35}\,(0.2) \quad \ldots(13)$$

$$\text{Soil loss} = 5.67\ \text{tonnes/acre/year} \quad \ldots(14)$$

Gavrilovic (1962; 1970; 1972; as described by Globevnik *et al.*, 2003) proposed an analytical relationship for determining the annual volume of detached soil due to surface erosion, as Equation (15):

$$W_p = {}^{\wedge}P\,F_w\,K_t\,\sqrt{K_z^3} \quad \ldots(15)$$

where, P is the average yearly precipitation, F_w is the drainage area (km^2), K_t is the temperature coefficient, and K_z is the erosion coefficient. The equation for K_t is :

$$K_t = \sqrt{(0.1 + T/10)} \quad \ldots(16)$$

where, T is the average yearly temperature. The K_z is given by Equation (17):

$$K_z = K_y\,K_s\,(K_\varphi + \sqrt{F_a}) \quad \ldots(17)$$

where, $\sqrt{F_a}$ is the average slope of the basin (%), K_y is the soil erodibility coefficient, K_s is the soil protection coefficient, and K_φ is the erosion and stream network development coefficient. Various soil loss equations developed by different workers between 1940 and 1997 are presented in Table 5.

Table 5. Equations in the development of soil erosion prediction

Scientist/Acronymous	Year	Equation
Zingg	1940	$A = C'\,L^{0.6}\,S^{1.4}$
Smith	1941	$A = C''\,L^{0.6}\,S^{1.4}\,P$
Browning *et al.*	1947	$A = C'''\,L^{0.6}\,S^{1.4}\,P$
Musgrave	1947	$A' = (P_{30}/1.25)^{1.75}\,K'\,(L/72)^{0.35}\,(S/10)^{1.35}\,C^*$
USLE	1965	$A = EI_{30}\,K\,(L/72.6)^{0.5}\,(0.065+0.045+0.0065\,S^2)\,C\,P$
USLE	1978	$A = EI_{30}\,K\,(L/72.6)^{0.5}\,(65.4\sin^2\Theta + 4.56\sin\Theta + 0.065)\,C\,P$
RUSLE	1997	$A = EI_{30}\,(L/72.6)^m\,(a\sin^2\Theta + b)\,C\,P$

A = Soil loss in tonnes/acre; A′ = Soil loss in inches/year; C′, C″, C‴ = Coefficients; C* = Vegetal cover factor; P_{30} = Maximum precipitation amount (inches) falling in 30 minutes in a storm; K′, K = Soil erodibility factors; L = Slope length in feet; S = Slope in per cent; Θ = Slope angle in degrees; C = Cropping management factor; E = Storm rainfall energy in hundreds of foot-tonnes per acre; I_{30} = Maximum rainfall intensity in a 30-minute period within a storm in inches per hour; P = Conservation practice factor; M = Exponent on length term-values depend on slope or slope and rill/inter-rill ratio; a, b = Coefficients in function making up slope term-values depend on slope.

Adopted from Laflen and Moldenhauer (2003).

8. New Developments in Predicting Soil Erosion

8.1. *USLE, RUSLE and WEPP*

By 1980, there was considerable interest in updating USLE. There was a successful development of a field-scale model for Chemicals, Runoff, Erosion from Agricultural Management Systems (CREAMS) (Kinsel, 1980). Consensus was arrived to revise the USLE [eventually named as Revised Universal Soil Loss Equation (RUSLE)] and the other to begin planning for a technology [which became Water Erosion Prediction Project (WEPP)] to replace the empirical technology for erosion prediction.

8.2. *Developments in Predicting Erosion from Ungauged Areas*

Simulating erosion, transport and deposition of sediment involve complex interactions of precipitation, surface and sub-surface hydrological processes, soil properties, land cover and topography. For this kind of complex spatial process, conventional rainfall-runoff models are of little use (Xu *et al.*, 2003). Although much has been achieved in the field of hydrological research, the issue of how to reliably predict the erosion and deposition of sediment in ungauged river basins is still a big challenge. Simulating erosion and sedimentation usually requires prediction of spatially variable hydrological processes at a fine resolution (Xu *et al.*, 2003). Overland flow generally detaches and transports large amounts of sediment, depending on ground slope, soil characteristics, vegetation cover and conservation practices. In a spatially distributed model, the spatial variability within the basin may be properly accounted for (Wigmosta *et al.*, 1994). An economical and reasonably efficient alternative would be to monitor the major basins with a reliable gauging system and to create a network of representative experimental basins for predictive purposes in medium and small sized basins (Toebes, 1965).

Many countries or certain areas in some countries face shortage of water, and at the same time, lack scientific support for water management. 'Prediction in Ungauged Basins' (PUB) is a recent initiative undertaken by the International Association of Hydrological Sciences (IAHS). Many countries and regions do not have sufficient hydrological stations to construct hydrological models with a high degree of accuracy. The International Hydrological Programme (IHP), World Climate Research Programme (WCRP), and International Geosphere-Biosphere Programme/ Biospheric Aspects of the Hydrological Cycle (IGBP/ BAHC) are all concerned, at some level, with research in ungauged basins (Jun and Ge, 2002; Zhu, 2002). PUB is a unique scientific endeavour that IAHS has conceived, initiated and is implementing. The aim of PUB is to develop hydrological sciences so that basin hydrology can be predicted using the limited information available for any basin, and to reduce uncertainty associated with such predictions (IAHS, 2003). At the same time, PUB has a social mission to provide the necessary hydrological information, especially discharge data, for ungauged or poorly-gauged basins, where those data are urgently needed.

The PUB initiative aims to increase our understanding of how hydrological systems work and can best be described by mathematical models. Success is to be measured by the reduction of uncertainty in the predictions from such models. Mismanagement of the rainwater has caused enormous soil erosion from the basins and silting of river beds, resulting in floods in many parts of the world. We are faced with a major water crisis that has emerged as a result of population growth, increase in water demand, environmental degradation and climate change. Increasing water shortages, flood hazards and groundwater contamination are problems intensifying during the recent past. There are numerous ungauged or poorly-gauged basins. It is especially true in developing countries where data are urgently needed to enable water resources to be developed and managed (IAHS, 2003). Numerical simulation models will be the new way for PUB (Jun *et al.*, 2003).

PUB Science Agenda

1. Exploration of ways to characterize the heterogeneity of landscape properties and climate inputs;
2. Development of general model framework for heterogeneity and predictive uncertainty;
3. Conducting model inter-comparisons in selected gauged basins to investigate uncertainty resulting from the choice of model structure;
4. Investigation of methods for efficient assimilation of gauged and/or remotely sensed data to constrain predictive uncertainty;
5. Promotion of process studies and field experiments to advance process understanding and conceptualizations with the aim of reducing uncertainty;
6. Advancing our understanding and predictions of the effects of hydro-climatological variability and change; and
7. Transforming the available global research products into hydrological information useful for local and regional water resources management.

Objectives of PUB

PUB is a policy-relevant hydrological science activity with the following objectives (IAHS, 2003):

- To advance our ability to predict with confidence the fluxes of water out of drainage basins worldwide in the areas of societal and ecological relevance in different biomes of hydro-climatic regions, and to verify predictions with data from selected basins;
- To advance the scientific foundations of hydrology, including understanding the climatic and landscape controls on the natural variability of hydrological processes, and the impact of human induced alterations to climate and landscape;
- To increase awareness about the value of data, especially of the measurement of hydrological variables, for the management of water resources around the world, and to demonstrate the importance of existing gauging systems and the need for targeted gauging to enhance currently inadequate or non-existent data sources;
- To advance technological capability around the world, so that predictions in ungauged basins are firmly based on local knowledge of the climatic and landscape controls on hydrological processes, along with access to the latest data sources; and
- To promote capacity building activities in the development of appropriate scientific knowledge and technology for areas and communities where they are needed.

Sharma and Sharma (2003) evolved a mathematical model to predict soil erosion from ungauged basins; based on the slope, rainfall, vegetation cover and soil texture. The soil erosion prediction model involves a partial regression equation of the form (18):

$$S_e = a + b_1X_1 + b_2X_2 + b_3X_3 + b_4X_4 \qquad \ldots(18)$$

where, S_e is the soil erosion (t/ha) and b_1, b_2, b_3 and b_4 are the partial regression coefficients for slope (X_1) in %, annual rainfall (X_2) in cm, clay content (X_3) in %, and vegetation cover (X_4), respectively.

Equation (18) was developed using the Dolittle method, which gave a reasonably high degree of accuracy in prediction or estimation of soil erosion. The vegetation is the only independent variable for which the values (1-5) have to be based on visual estimates. For uniformity, the classification used for the vegetation values was: 1, bare soil surface with soil stirred or ploughed; 2, scrubs or effective covered area below 25%; 3, cropped soil surface or effective covered area of 25-50%; 4, open forest vegetation or effective covered area of 50-75%; and 5, dense forest with trees and bushes. The model given by Equation (19):

$$S_e = 41.73 + 0.181 \times \text{slope (\%)} + 0.046 \times \text{rainfall (cm)} - 0.387 \times \text{clay (\%)} - 8.125 \times \text{vegetation (1 to 5 scale)} \qquad \ldots(19)$$

More available data for calculating the values of the coefficients may result in an increase in the predictive power of the equation. In principle, the model can be used for predicting soil loss by flowing water, under a wide range of climatic and physical conditions.

An explicit finite difference scheme for erosion and sediment transport on upland area of a watershed was derived by Aksoy (2003). The derivation ends up with a linear partial differential equation. Calibration of parameters is of great importance for ungauged basins, where data do not exist. Erosion by rainfall is usually related to the rainfall intensity (R) as:

$$E = \lambda R^{\delta} \qquad \ldots(20)$$

where, λ and δ are parameters to be determined at the calibration stage of the model, λ is the soil detachability coefficient, which has a dimension that depends on the value of δ, which generally equals to 1 or 2 (Aksoy, 2003). The model derived is given by Equation (21):

$$E = \lambda R^{\delta} + \sigma[\eta(\gamma h S)^{\varepsilon} - \rho_s C_s \alpha h] \qquad \ldots(21)$$

where, E_r is erosion; λ and δ are dimensional rainfall erosion coefficient and rainfall erosion exponent, respectively; σ is transfer rate coefficient (m^{-1}); η is dimensional coefficient; γ is the specific weight of water; h is the flow depth; S is the topographical slope (% slope/100); ε is the exponent; ρ_s is the specific mass of sediment (kg/m^3); C_s is the volumetric sediment concentration, and α is the coefficient. The performance of the model on the hill slope indicates that it has potential for application on watershed scale.

8.3. *Adjusted Tolerable Soil Loss for Conservation Planning*

The tolerable soil loss (T) has been defined in several ways (McCormack *et al.*, 1982; Lal, 1988; ISSS, 1996), but it is generally considered as the upper limit of soil erosion that can be allowed without degrading the long-term productivity of a specific soil. The T-value is also sometimes called "permissible soil loss" (Kok *et al.*, 1995). If erosion rates are higher than T, they are considered unacceptable and should be reduced. In India, a generalized T-value of 11.2 Mg /ha/yr is used for different sets of conditions to develop specific best resource management practices for each specific area. Mandal *et al.* (2006) have applied a quantitative model to sum up the overall soil performance to define the current state of soil resource for the north-west Himalayas of India in which scaling functions were used to convert soil parameters to a 0 to 1 scale. The normalized values were then multiplied by the appropriate weighting factors, based on relative importance and sensitivity analysis of each indictor. A categorical ranking of soil as 1, 2 or 3 was given to a soil group and general guideline was developed by following USDA Natural Resource Conservation Service (NRCS) to calculate soil loss tolerance for each soil group by using effective soil depth. Adjusted T values for the area ranged between 5.0 and 12.5 Mg/ha/yr compared to 11.2 Mg/ha/yr. Hence, the use of adjusted T-values is useful in improving conservation planning, meeting erosion control regulations for the development of sustainable farm operations, and improving watershed management.

8.4. *Assessment of Erosion Hazards for Conservation Planning*

Globally, vast areas face serious erosion and land degradation problems. Various erosion hazards occurring in the form of rills, gullies and torrents in arable and non-arable areas can be assessed and mapped for the purpose of resource conservation. Remote sensing techniques (Lillesand and Kiefer, 1994) are operationally useful for soil resource inventory and natural resource mapping at a scale of 1:50,000. They provide real lime and accurate information related to landuse/soil (Manchanda and Kudrat, 1999). For mapping the existing land uses and erosion hazards of a hilly watershed from Uttarakhand Himalaya, Dadhwal *et al.* (2007) employed two approaches: (i) visual interpretation of the merged FCC, and (ii) the digital classification of LISS-III data. Erosion hazard sites in the watershed were identified covering topographical variations from top to bottom. Index-based method was used for soil, landuse, slope and proximity to drain for the hazard assessment. Stream bank, rill and gully erosions were found as the dominant form of degradation associated with the drainage net work.

Different land uses, viz. open scrub, mixed forest and agriculture were considered to assess and understand their role towards degradation. Erosion hazard mapped by digital classification showed the maximum area under moderate rill erosion (e_2), followed by high (e_3) and very high (e_4) hazards. The average accuracy of erosion hazards mapped by remote sensing data was > 80%, inclusive of all the available land uses in the watershed. Thus, index based method for assessing erosion hazards has been found satisfactory. The problems in assessment with remote sensing data are mainly due to infrared reflectance of the prevailing vegetation which masks the effect of other reflectance, while interpreting the erosion hazards.

9. Wind Erosion

The south-western desertic zone of India covers about 28.6 Mha of area (about 8.7% of the geographical area of the country). The annual rainfall in the area ranges from 100 to 500 mm against the annual potential evapotranspiration of 1600 to 2000 mm. Wind erosion is mainly due to a strong wind regime, low atmospheric humidity, high solar radiation and single grain structure of sandy soil. About 14% and 28% of the area of the arid zone of Rajasthan suffers from severe and moderately severe form of wind erosion, respectively. The wind erosion under different land-use situations is shown in Table 6 (Dhir, 1989).

It occurs in arid and semi-arid areas which are devoid of vegetation, and where wind velocity is high. The single-grained sandy soil particles on the land surface are lifted and blown off as dust-storms. When the velocity of dust-bearing wind retards, the coarser soil particles are deposited in the form of dunes, sometimes rendering fertile lands unfit for cultivation. The finer particles are carried away to farther distances. The fertile soil is blown away by winds and the subsoil gets exposed, as a result the productive capacity of the soil is considerably reduced. Areas susceptible to wind erosion were delineated on the global as well as country levels. The Thar desert of the western Rajasthan shows high soil losses due to wind erosion. Deep ploughing of sand plains lost more than 3,000 tonnes soil/ha during a sand storm in 1987, whereas similar areas with 10-12% plant cover or with higher cloddy surface, suffered negligible soil loss.

9.1. *Wind Erosion Processes*

The abrasive action of the wind causes detachment of soil particles, followed by their transportation to distant places. The impact of high speed soil particles dislodges even other soil particles from the soil aggregates. The dislodged soil particles move by three phenomena: saltation, suspension and surface creep, usually occurring simultaneously.

9.1.1. *Saltation*

The movement of soil by wind in a series of short bounces along the surface of land carrying fine particles of 0.1 to 0.5 mm diameter is called 'saltation'. It is caused by the direct pressure of wind on soil particles and their collisions with other particles. After being pushed along the ground surface by the wind, the particles leap almost vertically in the first stage of saltation. Some grains rise only to small heights and, others leap up to 30 cm or even higher, depending directly on the wind velocity and the rise from the ground. In this process, the particles drop down within the eroded area or its immediate vicinity. Depending on the nature of the soil, wind speed and vegetative cover, saltation accounts for 50 to 75% of the total soil movement through wind erosion.

9.1.2. *Suspension*

The movement of very fine soil particles of less than 0.1 mm in diameter into the air and remaining suspended there is called suspension. The particles are kicked up into the air by the action of particles in saltation. The movement of fine dust in suspension is completely governed by the characteristic movement of the wind. The suspended material is carried to long distances from its original location and thus there is a considerable loss to the eroded area, accounting for about 40% of total soil movement.

9.1.3. *Surface Creep*

The soil particles varying between 0.5 mm and 1.0 mm in diameter, are too heavy and move along the surface resulting in the formation of surface creep. The movement through

Table 6. Soil loss due to wind erosion on flat sandy plains under different situations

Land use	Mean soil loss (tonnes/ha)
Cropped and left uncultivated a year before	80
Cropped and ploughed	1000
Long fallow	50
Degraded pasture	90
Dunes	200

surface creep is the outcome of the kinetic energy derived from the impact of particles moved in saltation. The particles raised from the ground move by rolling and sliding along the surface of soil drop down within the eroded area or its immediate vicinity. It accounts for about 5% to 25% of the soil erosion through wind.

9.3. Factors Affecting Wind Erosion

The major factors affecting wind erosion are: (i) climate, (ii) soil, and (iii) vegetation. The climatic parameters influencing wind erosion are: rainfall, Temperature, wind, humidity, viscosity and intensity of the air. The amount and distribution of rainfall and its effect on soil moisture regime are the most important factors. The principal characteristics of wind affecting erosion are: velocity, direction, duration and turbulence. Soil characteristics affecting wind erosion are: texture, structure, density of soil particles, organic matter, and surface roughness. The vegetative factors that affect wind erosion are: kind of vegetation, height, density, etc. Living plant roots and tops are more effective in retarding wind erosion.

9.4. Control of Wind Erosion

The basic principles in the control of wind erosion are breaking of wind velocity at the soil-atmosphere interface and the reduction in soil erodibility. This can be achieved through vegetation or cultural practices. The wind erosion control measures include: (i) creation of permanent grass strips across the prevailing wind direction, (ii) roughening of the soil surface to trap some of the moving particles, (iii) provision of stubble mulch, (iv) tillage to provide cloddy surface condition which is at right angles to the wind direction, (v) strip cropping which can prove useful in reducing the effect of wind, and (vi) wind-breaks with tree linings which are effective in reducing wind velocity for short distances.

9.5. Estimation of Wind Erosion

Wind erosion can be estimated by using Equation (22), given by Chepil and Woodruff (1963), which predicts the relationship between annual soil loss by wind erosion from a given field and the five factors influencing wind erosion:

$$E = f(I \times K \times C \times L \times V) \qquad \ldots(22)$$

where,

E = Average annual soil loss in tonnes/acre,

I = Soil erodibility index indicated by soil aggregates greater than 0.84 mm in diameter and percentage of land slope,

C = Climatic factor indicated by wind velocity and surface soil moisture,

K = Soil surface roughness,

L = Unsheltered field width measured along the direction of the prevailing wind, and

V = Vegetative cover.

Equation (22) is useful in determining (i) potential wind erosion on any field under existing conditions, and (ii) conditions of surface roughness, soil cloddiness, vegetative cover, sheltering or width and orientation of field necessary to reduce wind erosion to tolerable level.

10. Extent of Soil and Nutrient Erosion due to Shifting Cultivation (*Jhuming*)

Major physical manifestations of the processes causing degradation are the removal of top soil through water and wind erosions, reduced capacity of soil to hold water, increased susceptibility to runoff and gradual increase in the loss of nutrients. The process of soil erosion begins when a farmer enters his field either for selection of site for cultivation or cutting of the forest vegetation. There is a considerable loss of soil even due to the movement of farmers on hill slopes, earthworm castings, dibblings, etc. The problems of soil erosion in shifting cultivation are mainly of splash and wash. Soil erosion from hill slopes (60 to 70%) under first year, second year, abandoned *Jhum* (shifting cultivation) and bamboo forest was 147, 170, 30 and 8 tonnes/ha/yr, respectively. The second year of shifting cultivation is more hazardous than that of the first year. It has been estimated that about 88.3 million tonnes of soil is lost annually in the North-Eastern states of India alongwith 10.7 thousand tonnes of N, 0.4 thousand tonnes of P and 6.0 thousand tonnes of K (Table 7) (Sharma and Prasad, 1995). However, wide variations in soil and nutrients losses are found, depending on the slope gradient, nature of the soil, crop canopy, agricultural activities, etc. Improper land management in hills can cause annual soil loss of about 80 t/ha/yr as compared to the national average of 16.35 t//ha/yr. The permissible soil loss limit is from 4.5 to 11.2 t/ha/yr under different soil types. About 5334 million tonnes of soil is being eroded annually due to agricultural and associated activities in the country and 29% of the eroded material is permanently lost into the sea (Dhruva Narayana, 1993). The loss of soil nutrients through water erosion is estimated to be 5.4 to 8.4 million tonnes.

11. Soil Erosion *vis-à-vis* Carbon Sequestration

Growing of trees and crops as well as rearing of animals on the same piece of land simultaneously or sequentially is viewed an optimum strategy for resource conservation and environmental security. Sustainable development and management of land-use system through such innovations not only meet the multiple needs for food, fodder, fuelwood, fibre, fertilizer, timber, gum, etc. but guarantees economic and environmental security also for safe and protected habitats for human beings and other living organisms. Indian soils are largely carbon depleted but can be brought back to their native carrying capacity by such interventions. Hence, sustainable forest/tree-crop/animal combinations are very important for the environmental conservation, as they help in regulating water cycle, protect against erosion, recharge groundwater, improve soil fertility, conserve soil and water, moderate climate, sequester carbon, conserve biodiversity, combat desertification and degradation, provide perpetual flow of clean water in streams, improve soil health, etc. They play an important role in carbon sequestration and may hold even up to 100-times more carbon per unit area as compared to purely agricultural lands (Houghton and Hackler, 1994). Though the major part of the carbon stored is held in the above-ground biomass, soils are also important reservoirs of carbon. Thus, forest/tree- crop systems serve as carbon sink/source, depending on the management system and also protect the soil from erosion losses (Dadhwal and Kumar, 2004).

11.1. Carbon Emissions

Tropical deforestation is responsible in part for the increasing concentration of CO_2 in the atmosphere (Houghton, 1990). Estimates of net release of carbon at the global level depict wide variations, ranging from 0.4-1.6 GtC/ year (Detwiler and Hall, 1988) to 1.1-3.6 Gt C/ year (Houghton, 1991). Carbon emissions can be from deforested as well as degraded lands also (Ravindranath *et al.*, 1997).

Table 7. Annual soil and nutrient losses in different states of North-Eastern region due to shifting cultivation

State	Soil loss (million tonnes)	Nutrients loss ('000 tonnes)		
		N	P	K
Arunachal Pradesh	14.49	2.04	0.058	1.01
Assam	12.32	0.98	0.061	0.64
Manipur	20.43	2.68	0.082	1.53
Meghalaya	14.15	1.90	0.042	1.05
Mizoram	13.04	1.69	0.065	0.98
Nagaland	7.60	0.87	0.040	0.52
Tripura	5.95	0.51	0.024	0.31
NE Region	88.34	10.67	0.372	6.05

Source: Sharma and Prasad (1995)

As per estimates, 13.5×10^6 t C is released from combustion. There is no balance between the total carbon stored in the felled vegetation and the total carbon released from deforestation. In addition to carbon release from deforestation, degradation of the existing forests also contributes to the carbon release. This degradation may be due to grazing, fire, death due to diseases and pests, illegal removal of timber, non-sustainable harvest of firewood or timber, etc. (Houghton, 1991; Flint and Richards, 1991). The analysis carried out by Flint and Richards (1991) and Houghton (1991) for southern Asia (including northern parts of India) for the period 1950-1980 has indicated a degradation ratio of 1.54, which means that for every tonne of carbon released into the atmosphere through deforestation, an additional 0.54t of C is released from the degradation of standing forests which amounts to 15.4×10^6 t C. Of this, about 50% is assumed to be used as firewood, 25% for long-term purposes, and 25% decomposes.

In India, about 0.99×10^6 ha area is annually subjected to shifting cultivation (FSI, 2005). The fallow period has also been shortened and considering this, a net emission of 1.56×10^6 t C is estimated after accounting for the cutting of secondary forests in fallow areas, uptake in the fallow, and degradation or burning of new areas.

11.2. Carbon Storage and Uptake

Carbon storage is estimated by taking 50% of the biomass as carbon and is based on the soil organic content in the top 30-cm soil. Total carbon stored in Indian forests is estimated at 9585×10^6 t C, of which vegetation accounts for 44% and soil, the remaining 56% (Ravindranath *et al.*, 1997).

While considering the net carbon release, carbon uptake and storage in forests, and tree plantations, e.g. *eucalyptus, casuarina*, etc. are also considered. But, large tracts of forests in India have been subjected to anthropogenic pressures over centuries and thus are at various stages of succession. The C-accumulation is estimated using the data on standing biomass.

11.3. Balance between Carbon Emission and Sequestration

It is established that the carbon flux is nearly balanced in India because of the reason that the Forest Conservation Act, 1980, prescribes that no land should be alienated without the consent of the Central Government. Moreover, in the late-1970s, the high level of subsidies enjoyed by the forest industries were gradually reduced and farmers have been encouraged to plant tree crops. The states like Haryana and Gujarat have initiated measures to create village-based management systems involving local people in sustainable use and restoration of degraded lands (Ravindranath *et al.*, 1997).

11.4. Limitations of Carbon Emission Estimation in India

The carbon emission estimates in India are based on the standing biomass determined using the crown cover estimates and a few published studies on the standing biomass. Similarly, the C-uptake is estimated to be in the range of 1.25-2.84% of the standing biomass. Therefore, any changes in the estimates of crown cover, standing biomass and annual productivity will also change the net emission rates (Ravindranath *et al.*, 1997).

11.5. Agroforestry and Carbon Sequestration

Trees growing on non-forest areas meet a major part of the demand for fuelwood and timber in the rural areas which envisages that agroforestry has great potential in carbon sequestration and resource conservation in India. But, the constraints in agroforestry on marginal lands include low income of farmers, low productivity of land, heavy grazing, insecure tenurial rights, etc. (Singh, 2001). Despite problems, agroforestry is picking up and about 50% of the planting done in India after 1980 is reported under agroforetry (MOEF, 1989). Thus, agroforestry is picking up an important opportunity for C-sequestration. Plantations on marginal and degraded lands, on bunds and boundaries in agricultural lands and growing of trees in homesteads in all parts of the country would be helpful in carbon sequestration (Ravindranath and Somashekhar, 1995).

12. Soil and Water Conservation

Soil and water conservation is a combination of all management and land use practices which protect the soil against depletion or deterioration by natural or man-induced factors and improve the productivity of the natural resources on sustainable basis. It is said that nature takes centuries to produce the soil but man may loose it in a few years due to its indiscriminate exploitation for his need and greed. Soil once lost, is difficult and expensive to replace. The conservation of soil and water is an anti-erosion and anti-runoff approach, and is also a comprehensive and integrated approach for the judicious use of these resources rather than their negligent and wasteful use.

12.1. Soil Conservation Measures

The following soil conservation measures may be adopted for controlling soil erosion, depending on the degree and length of the slope and physical configuration of the land.

12.1.1. Mechanical Measures

(i) Contour Bunds — Contour bunds are mechanical barriers built across the slope for safe diversion of excess runoff and retention of eroded soil. The land area in between the two bunds gets levelled in due course of time. Due to deposition of eroded soil along the bund, the latter takes the shape of a riser. These risers should be planted with grasses to check their erosion.

(ii) Graded Bunds — The graded bund is a small earthen bund with slight grade constructed across the slope for safe disposal of runoff. The graded bunds are recommended up to 10% slope for areas where annual rainfall exceeds 800 mm, particularly on clayey and black soils with poor drainage. However, efficacy of graded bunds gets reduced gradually beyond 4% slope. The purpose of graded bunds is to reduce the velocity of runoff water, for *in-situ* conservation of rain water and to minimize soil erosion.

(iii)Bench Terraces — Bench terraces are flat beds constructed on the hills across the slope. The height of the riser should not be more than one metre and the width of bench terrace depends on the degree of slope. The bench terraces are important because they promote uniform distribution of soil moisture, irrigation water, etc. and control soil erosion. The bench terraces may be table top (level), outward sloping or inward sloping, with or without mild longitudinal grades. On steep slopes, it is better to construct terraces on foot hills for agricultural crops when soil depth is more than one metre.

(iv) Half Moon Terraces — Half moon terraces are semi-circular beds of appropriate diameter with the shape resembling a half moon. These terraces are recommended for fruit trees or other plantation crops on steep slopes.

(v) Grassed Waterways — Grasses are well-known for their soil binding characteristics. They are most effective in moderating the flow and reducing the erosive velocity of runoff water, particularly on the rolling topography. The runoff water moves with high velocity down the slope, carrying with it soil and nutrients. If some suitable grasses are planted on the runoff route or natural channels, the soil and nutrient losses can be reduced. These grassed waterways are laid on the natural drainage lines in the watershed. Stilling basins or water ponds are constructed *en route* at appropriate locations, with earthen and boulder pitched bunds for the retention of runoff water. By reducing the velocity of runoff water, the erosion losses can be minimized.

(vi) Water Harvesting Ponds — Water harvesting structures can be dug out for retaining runoff water on seasonal or perennial basis. These are generally constructed down the slope. Earthen dams should be used for retaining silt loads at appropriate locations on the slope of a watershed. The water thus harvested

or stored can be used for pisciculture and supplemental irrigation.

(vii) Conservation Bench Terraces (CBT) — These are used to stabilize the yield of rainfed crops by inter-field water harvesting. A part of the field is levelled to retain the runoff originating from rest of the field.

(viii) Gully Control Structures — Gully control structures are provided to (i) reduce the erosive velocity of runoff water, (ii) facilitate establishment of vegetation, and (iii) provide protection at points that cannot be adequately protected by other methods. Loose boulder check-dams perform well in gullies which do not carry much runoff and it also helps in silt deposition, thereby helping the stabilization of gully beds. Permanent gully control structures are constructed to control the overfalls either at gully head or in gully bed. Erosion from the extending heads and sides of gully and main channel are the major sources of sediment. There is also a need to construct diversion bunds to divert surplus water to water harvesting structures or to the grassed water ways.

(ix) Contour Trenches — Contour trenches are dug-out, piling up the dug-out earth on lower side of the trench, for trapping sediment and runoff at the early stage of their movement. These trenches also improve soil moisture and favour quick growth of trees and grasses.

(x) *Stream Bank and Torrents Control* — The vulnerable stream banks should be protected by providing spurs and retaining walls, etc. To control torrents, structures like barrages, paved channels, etc. need to be provided.

12.1.2. Agronomical Measures

For preventing soil erosion on cultivated lands, proper choice of crops and cropping patterns is necessary, particularly on hill slopes. The protection through vegetative shield, forest cover, grasses, crops and mulches, etc. are some important measures to prevent soil erosion. Such a protection by absorbing the energy of rain impact prevents the loss of both water and soil. The following crop management practices can be useful in minimizing the erosion of soil and nutrients.

(i) *Cropping Systems* — Crops with the ability to develop canopy quickly provide an early protection to the soil. Inter-planting of ero-sion-resistant crops like cowpea, soybean, etc. are also useful. Strip cropping of erosion-resistant legumes along with cereals can conserve rain water and reduce the velocity of runoff.

(ii) Crop Geometry — It is essential to manipulate the crop layout in the field in a manner which may prevent soil erosion. A closer spacing of rows across the slope can help in this regard.

(iii) Contour Cultivation — Contour cultivation reduces the runoff to a large extent, thereby reducing the soil and nutrient losses. Contour cultivation as well as furrows and ridges have been found useful.

(iv) Tillage — Low intensity tillage favours consolidation of soil through better structure, infiltration and pore distribution. This imparts erosion resistance. A study of conventional method of cultivation of maize with zero tillage, with or without live mulch, has shown that runoff and soil losses are greatly reduced with low intensity tillage (Table 8) (Bharadwaj, 1998).

(v) Grasses — Grasses are perhaps the best friend of soil conservationists. Low and evenly distributed canopy and fibrous root systems with much soil binding capacity make grasses highly effective in controlling soil erosion. The performance of various grass species in controlling soil erosion and runoff losses at 9% and 11% slopes are given in Table 9. The selection of grasses should be based on their production potential considering edaphic conditions and local preferences.

Table 8. Effect of tillage and living mulch on runoff and soil losses

Conservation measure	Runoff (%)	Soil loss (tonnes/ha)
Conventional maize	61.2	15.76
Zero tillage without live mulch	41.8	12.00
Normal tillage with live mulch	31.2	7.00
Zero tillage with live mulch	21.6	3.30

Source: Bharadwaj (1998)

12.1.3. Agronomical-cum-mechanical Measures

Depending upon the needs of land and intensity of erosion, bio-engineering measures are sometimes preferred as compared to either mechanical or agronomical measures in isolation. In black soils with 45-50% clay, both agronomical and mechanical measures have been found to reduce runoff and soil losses. These conserve soil moisture and thereby increase crop yield. The treatment of vertical mulch and surface mulch has been found to be most effective in controlling runoff and soil losses, and consequently, could produce maximum sorghum grain yield. Under mechanical measures, the graded bunding has been found to be superior to conservation ditch and contour bunds (Table 10).

12.1.4. Agroforestry Measures

Agroforestry has become popular as a useful land-use system on slopes in the recent past. Growing of trees along with agricultural crops satisfies the multifarious needs of farmers. The data presented in Table 11 provide an idea of the usefulness of agroforestry system as a soil conservation measure (Narain *et al.*, 1994). Growing of multipurpose tree species (MPTs) has been recommended alongwith crops.

(a) Soil Health Improvement

Ameliorative effects of agroforestry landuse in marginal and degraded lands have been observed in Doon Valley (Uttarakhand). A decrease in soil pH by 0.4 units and in bulk density (BD) by 0.1 mg/m^3, improvement in organic matter content by 0.16% and in water holding capacity by 1-5% have been observed (Dadhwal and Tomar, 1999). Low pH values and high organic carbon (OC) contents beneath the tree cover than bare soil have also been observed by Aggarwal *et al.* (1975). In another study in the Garhwal Himalaya, improvement in soil health has been recorded over time in minespoil degraded lands. After 14 years with agroforestry interventions, pH of the minespoil decreased to 7.5 from 8.1, OC

Table 9. Runoff and soil losses under different grass species at 9% and 11 % slopes at Dehradun

Land slope	Treatment	Runoff (%)	Soil loss (tonnes/ha)	Reduction (%)		Biomass (kg/ha)
				Runoff	Soil loss	
9%	Bare and ploughed	59.6	155.95	-	-	-
	Bare fallow	71.1	92.42	19.3	40.7	-
	Cynodon dactylon	27.1	2.10	54.5	98.7	-
	Natural grass	21.2	1.02	64.4	99.3	4355
11%	Cultivated fallow	16.2	18.45	-	-	
	Puerara hirsuta	1.8	0.11	88.2	99.4	14962
	Dicanthium annulatum	1.9	0.23	88.1	98.7	6805
	Chrysopogon fulvus	2.5	0.30	84.2	98.4	19170
	Eulaliopsis binata	5.2	0.29	67.1	98.4	16290

Source: Tejwani *et al.* (1975)

Table 10. Effect of conservation measures on runoff and soil losses and yield of sorghum under black soil of Bellary (Karnataka) at 1 % slope (Average of 8 years)

Treatments	Runoff (% of rainfall)	Soil loss (kg/ha)	Sorghum grain yield (kg/ha)
Agronomical Measures			
Up and down cultivation	54.8	2167	685
Vegetative barrier	39.9	1372	848
Contour cultivation	38.6	1640	897
Vertical mulch + surface mulch	23.3	172	1204
Mechanical Measures			
Contour bund	33.6	606	859
Graded bund	26.7	601	943
Conservation ditch	18.4	618	931

Source Rama Mohan Rao *et al.* (1997-98)

Table 11. Runoff and soil loss under different agroforestry systems

Land use	Runoff (%)	Soil loss (tonne/ha)
Maize	18.3	17.70
Maize + *Subabul*	8.9	5.00
Maize + Eucalyptus	3.6	0.91
Chrysopogon fulvus	1.6	0.33
Grass + *Subabul*	0.6	0.13
Subabul	0.4	0.04
Grass + Eucalyptus	0.1	0.02
Eucalyptus	0.1	0.01

Source: Narain *et al.* (1994)

Table 12. Change in minespoil characteristics over time (average of 14 locations) from Uttarakhand Himalayas

Characteristics	Year		
	1985	1991	1997
pH (1:2.5)	8.10	7.90	7.50
Organic C (%)	0.13	0.18	0.42
Total N (%)	0.01	0.01	0.05
$CaCO_3$ (%)	54.60	34.60	31.00
Bulk density (Mg/m^3)	1.63	1.53	1.47

Source: Dadhwal (1999)

increased from 0.13% to 0.42%, whereas $CaCO_3$-content decreased from 54.6% to 31.0% and BD (mg/m^3) decreased from 1.63 to 1.47 (Dadhwal, 1999); it shows that planting of MPTs, shrubs and grasses help in improving soil health and arresting soil erosion (Table 12).

Further, bio-engineering measures applied in the mined watershed at Sahastradhara (Uttarakhand), checked the heavy soil erosion, drastically reduced the surface runoff the surface runoff with attenuation in the flood peaks, increased the lean period flow due to groundwater recharge and improved the biodiversity over time (Table 13) (Juyal *et al.*, 2007).

Leguminous shade trees benefit through shade and nitrogen they provide to tea plants. A shade tree may add as much as 2500-5000 kg leaf, twigs and pods and 63-126 kg N, 18-36 kg P_2O_5, 22-44 kg K_2O, 32-64 CaO and 16-32 kg MgO per ha per annum to the soil.

The accumulation of soil organic matter under trees is the most commonly reported effect of trees on soils (Odum, 1960). Quantity and quality of soil organic matter bring about important changes in physical and chemical properties of soil. Reduction in soil bulk density and concomitant increases in soil aeration and soil hydraulic conductivity are caused by the incorporation of low specific gravity organic matter and metabolic activity of decomposer organisms. Rolfe and Boggess (1973) had found that 30-35 years after pine establishment in Southern Illinois (USA), the soil bulk density had decreased while soil hydraulic conductivity and exchangeable calcium and magnesium contents had increased. It is worth noting that these improvements in soil

Table 13. Impact of rehabilitation measures at minespoil watershed, Sahastradhara, Derhadun (Uttarakhand) (treated since 1984)

Particulars	Pre-treatment (1984)	Post-treatment (1996)
Debris outflow (t/ha/yr)	550	6
Monsoon runoff (%)	57	37
Lean period flow (days)	60	240
Water quality	Not potable	Potable
Vegetation cover (%)	10	80
Cost of debris clearance (Rs/yr)	1 lakh	Nil

conditions under trees have occurred on young soils, which contain weatherable minerals and are dominated by fixed charged clays. In many soils, a large portion of cation exchange capacity (CEC) is associated with carboxyl group in the organic matter. This can have a significant effect on soils that have very low mineral based CEC (Fisher, 1990).

The age-old practice of shifting agriculture takes the advantage of ameliorating effects of a period of forest fallow. This causes organic matter additions greatly in excess of decompositional losses and restores good physical and chemical conditions to the site (Ewel, 1986; Venkataramanan *et al.*, 1983). Singh *et al.* (1994) have reported 6-10 fold increase in organic carbon status of a sodic soil when occupied by trees such as *Prosopis juliflora*, *Acacia nilotica*, *Eucalyptus tereticornis*, *Terminalia arjuna* and *Albizia lebbek* for more than 20 years. Similarly, Shankaranarayan (1984) had reported greater increase in the amount of organic matter, nitrogen and phosphorus under *Prosopis cineraria* than under *Prosopis juliflora* and bare field under arid conditions.

(b) Carbon Sequestration and Soil Fertility in Tree-Crop Interaction

Tree legumes are often important component of arid and semi-arid ecosystems. Many of these trees and shrubs have the ability to develop a deep root system and symbiotically fix atmospheric nitrogen. Martin (1977) had noticed significant formation of soil organic matter during active root growth, much of it coming directly from the root tissue without the intervention of soil microfauna. Woody legume trees, which have potential to fix significant quantities of nitrogen, may have an influence on the yield in the nitrogen-deficient soils. Some nitrogen-fixing trees have proved to be very efficient in phosphorus uptake and can effectively serve as nutrient pumps for phosphorus and other mineral nutrients (Balkrishan and Toky, 1993).

Nair (1984) had predicted the possibility of nitrogen fixation in woody legumes to the order of 100 kg N/ha/year in areas receiving 500 mm annual rainfall. Nitrogen fixing trees can achieve the rates of fixation comparable with those of leguminous crops (Young, 1990). The major recognized avenue for addition of organic matter to the soil from the tree standing on it, is through the fall of litter, twigs, branches and fruits (Brinsen *et al.*, 1980).

Kellman (1979) had emphasized that trees also show preferential enrichment of soil below them in terms of N, P, K, Ca, Mg and Na. In some cases, the levels of these nutrients approach or exceed those found in the nearby rain forest. He concluded that gradual accumulation of mineral nutrients by perennial, slow growing trees, and incorporation of these into an enlarged plant-litter-soil-nutrient cycle was the mechanism responsible for this soil enrichment. Similar results of an increase in nutrient content of soils under species *P. cineraria*, growing in the arid and *F. albida* trees in semi-arid environment, were reported by Singh and Lal (1969) and Felker (1978), respectively. By adopting various agroforestry/silvipasture models, degraded soils can be rehabilitated (Pathak and Dagar, 1998).

In the semi-arid conditions at Jhansi, in the silvipastoral systems of *Albizia lebbeck* – *Cenchrus* – *Stylo* and *A. procera* – *Cenchrus* – *Stylo*, a significant improvement in soil organic matter builds-up was recorded; it was 38% in the former and with 25% in the latter system (Pathak, 1992).

In a nutrient management study conducted at Meghalaya in the north-eastern India, the contents of exchangeable Al and Mg decreased, while those of Ca, K and Bray's P increased in the silvipastoral system (Prasad, 1992). Similarly, in the north-west Himalayas, a study on the dynamics of herbage in *Acacia catechu* and *Punus roxburghii* based natural silvipastoral systems has revealed that as the distance from tree trunk increases, pH and EC enhance and organic carbon, N, P, K, Ca and Mg decrease. Recent studies on soil fertility improvement in *Grevillea robusta, Gmelina arborea* and *Tectona grandis* based silvipastoral systems under semi-arid condition at Mettupalayam, Tamil Nadu, have revealed that due to inclusion of either tree alone or tree + grass, the organic carbon, available N, P and K in the soils increase appreciably as compared to those in open (barren) land and pure grass, except N in *G. robusta* + grass; P with pure trees of *G. arborea* and *T. grandis* + grass, and K with *T. grandis* (Thiyageswari *et al.*, 1998).

In arid condition, Singh and Lal (1969) investigated the profile characteristics and level of fertility under *P. cineraria* and *A. nilotica* canopies and concluded that silvipasture systems enhance organic matter, total N, available P, soluble C and brings down soil pH. Better mechanical composition of soil up to 120 cm depth due to silvipasture systems promote the growth and yield of crops grown in the vicinity of *P. cineraria*. Aggarwal *et al.* (1975) concluded that the status of available micronutrients generally improves under plantation especially under *P. cineraria*.

The soil properties, in general improve with tree-cropping as compared to non-tree situation. Hazara (1990) had observed that field capacity, wilting point, organic carbon, cation exchange capacity and available N and P contents of soils were greatly improved, whereas bulk density, pH and EC values were appreciably decreased under *Albizia lebbeck* plantation as compared to normal cropping. The canopy structure and density of tree species had influenced the grass production underneath. The leguminous shrubs and trees have great influence in building up soil organic matter (0.32-0.91%), available soil N (131-293 kg/ha), available P (6.2-18.5 kg/ha) and field capacity (11-15.8%).

In one of the studies conducted at IGFRI, Jhansi, under seventeen years old silvipastoral systems, it was found that the organic carbon was maximum at the top layer (0-15 cm), followed by 30-45 cm soil depths in pure grasses as well as in *C. ciliaris* + *A. tortilis* and *P. maximum* + *L. leucocephala* based silvipastoral systems (Mishra, 2002). The increase in organic carbon by 10 to 40 per cent under the silvipasture system as compared to the sole pastures may be because of more litter decomposition and available root biomass. The increase in organic carbon under the silvipasture systems has indicated the improvement in soil health, resulting in higher system productivity and sustainability. *L. leucocephala* + *P. maximum* based silvipasture system maintains higher organic carbon content and therefore, the sustainability of production system is maintained over the years. Maximum increase has been estimated in the contents of available N, P and K under the tree pasture canopies than the sole pastures.

13. Watershed Approach

The soil and water conservation measures in the lower catchment areas are prone to dam-age due to uncontrolled runoff from the upper areas. It is therefore necessary to protect the upper catchment areas through the watershed approach.

Watershed can be defined as a topographic hydrological unit draining runoff water at a common or a particular point by a network of channels and streams. Thus, a watershed represents a hydrological unit of area but can also be described as a bio-physical and socio-economic and sometimes a political unit for planning and management of natural resources.

Small watersheds are those where the overland flow is the chief contributor to the peak flows and the overland flows are not affected by the channel characteristics. On large watersheds, the channel flows are highly pronounced and peak flows are influenced by the channel characteristics. The area to be managed for soil and water conservation programme is also important in categorizing a watershed. Thus, these are sub-divided into different levels called as sub-, milli-, micro- and mini-watersheds managing the area of 10000-50000 ha, 1000-10000 ha, 100-1000 ha and 10-100 ha, respectively.

The main objectives of watershed approach for the management of land, water and vegetation resources are: (i) to establish a watershed as a basis to conserve and upgrade crop and degraded lands, (ii) to develop and demonstrate location-specific technologies for soil and water conservation and for crop production stabilization, and (iii) to augment the fodder, fruit and fuel resources through alternate land-use systems. For steep hill slopes, the following aspects are to be considered (Samra, 1997):

(i) Ensuring adequate protection of land against soil erosion with the use of local resources and skill-based soil conservation measures.

(ii) Maximum retention of rainfall within the area without affecting the crops.

(iii) Storage of runoff water, its use for pisciculture and irrigation.

Based on the micro-watershed approach, different land uses such as agriculture, horticulture, agri-horti, agri-horti-silvi-pastoral and livestock based systems, etc. may be adopted for a judicious resource conservation. In the watershed approach, proper land-use systems may be adopted to minimize runoff and soil losses. The data presented in Table 14 show the average runoff and nutrient loss from seven land-use systems under different slopes and soil conservation measures.

13.1. Hydrology of Watershed

A thorough knowledge of the different hydrological phenomena is required in all aspects of soil and water conservation in the watershed. Information about rainfall is needed for in estimating runoff, planning erosion-control measures, planning for irrigation, removal of excess water and conserving water in low rainfall regions. Knowledge of runoff is required in designing structures and channels that will handle natural flows of water. Data on infiltration, evaporation and transpiration are required in planning irrigation and drainage systems, moisture conservation practices, etc.

13.2. Watershed Management

A good watershed management must consider the social, economic, and environmental sustainability and institutional factors operating within and outside the watershed area. The essentials features of watershed-management programmes involve newly emerging paradigms — priority areas should be identified and integrated development approaches should be followed; community-driven involving interaction between users, agencies and developmental departments; bottom-up and flexible approach with the capacity for making interim corrections; execution through village cooperative societies in perfect harmony; peoples' participation is necessary right from planning, execution and monitoring of the programme; judicious and equitable sharing of costs and benefits to landowners and the landless people in the watershed; gender neutrality, women, who play pivotal role in agriculture should be involved in all the decisions without any gender bias; involving non-governmental organizations (NGOs) in watershed management; and the common property resources should be managed by the people with their rights. Joint forest management needs serious thinking/rethinking. Village societies are required to be established for resource management with peoples' empowerment for their effective management.

13.3. Participatory Integrated Watershed Management

Watershed management is an integrated approach, considering the holistic development for all the users of the watershed. It is necessary to protect, conserve and improve the land resources for an efficient and sustained production; to protect and enhance water resource, moderate floods and reduce silting up of tanks, increase irrigation and conserve rainwater for crops and thus mitigate droughts; and to utilize the natural local resources for improving agriculture and allied occupations or industries (small and cottage industries) to improve socio-economic status of the local residents.

It is universally accepted that conservation should be essentially an area approach and the unit of its adoption should be a watershed. Therefore, a combination of few approaches is required to be adopted in a given watershed. Now-a-days, integrated watershed management of the resources with peoples' participation is a well-accepted concept for the management of natural resources.

14. Future Scenario as Affected by Resource Conservation Technologies

Adoption of resource conserving technologies for efficient utilization of land, water, vegetation and human resources following bottom-up integrated watershed management approach is the only solution to meet the ever increasing demand of foodgrains, fodder, fuelwood and fibre for the growing population. Apart from increasing production, integrated watershed management (IWSM) is quite beneficial in generating employment opportunities. The Participatory IWSM approach exhibites long-lasting, people-oriented and environment-friendly impact. Application of multi-objectives decision support system (MODSS) technique to the Himalayan watersheds has revealed great potential in watershed planning, monitoring and evaluation through adoption of optimal watershed management plans. For example, at the Fakot watershed in Uttarakhand, income employment and food sufficiency increased by 18%, 8% and 46%, respectively, with 53% reduction in soil loss. Similarly, based on the experiences of model watersheds developed by CSWCRTI, Dehradun (India), potential of IWSM in increasing production of crops, horticulture, livestock, etc. in terms of rice equivalent yields in the rainfed areas has been analyzed in 11 agro-climatic zones. The average productivity of the predominant rainfed cropping systems and allied agriculture sectors in the model watersheds has been compared with the average productivity at the district level and the concerned agro-climatic region for extrapolation of production potentials based upon the yield gaps. The analysis has revealed that production can be enhanced by 152 million

Table 14. Runoff and soil losses under different land-use systems on micro-watersheds

(Average of 7 years)

Watershed No.	Dominant land-use system	Conservation measures	Average slope (%)	Average runoff (mm)	Soil loss (tonnes/ha)
1	Livestock based	Contour bunds, trenches and grassed water-ways	32.0	12.7	0.02
2	Forestry	-	38.0	80.1	0.11
3	Agroforestry	Contour bunds	32.2	39.7	0.06
4	Agriculture	Contour bunds, bench terraces and grassed waterways	32.4	23.2	0.08
5	Agri-horti-silvi-pastoral	As in watershed No. 4 also half-moon terraces	41.8	12.1	0.19
6	Horticulture	As in watershed No. 5	53.2	29.2	1.89
7	Natural fallow	-	45.8	11.5	0.01

Source: Singh and Singh (1978)

Table 15. Potential of additional production and income generation through integrated watershed management in different agro-climatic regions of India

Sl No.	Agro-climatic region	Additional regional production ('000 tonne)	Additional regional gross return advantage (in lakh Rs)
1.	Western Himalayan Region	13132.7	735433.2
2.	North-eastern Himalayan Region	16234.1	909111.6
3.	Upper Gangetic Plain	1933.8	108293.9
4.	Trans-Gangetic Plain	25791.5	1444321.9
5.	Eastern Plateau and Hill Region	29350.6	1643634.7
6.	Central Plateau and Hill Region	20487.7	1147310.1
7.	Western Plateau and Hill Region	8254.1	462226.8
8.	Southern Plateau and Hill Region	18967.2	1062163.2
9.	East Coast Plains and Ghat Region	10858.4	608068.2
10.	West Coast Plains and Ghat Region	1077.1	60317.0
11.	Gujarat Plains and Hill Region	5790.9	324289.3
12.	All India	151878.1	8505170.0

tonnes, if efficiency of technology transfer is assumed as 100%, similar to the model watersheds (Table 15). Even at 70% efficiency, a significant increase in production (106 Mt) can be achieved, which would be sufficient to meet the demands of growing population by 2025 (CSWCRTI, 2007).

Better inventorization/characterization of natural resources and prioritization of critical areas for erosion control on watershed basis (through application of advanced RS/GIS/GPS techniques), development of area-specific soil loss tolerance limits along with estimation of erosion-fertility-productivity relationships under primary production systems of the country and site-specific land suitability criteria can lead to efficient management of natural resources.

14. References

Aggarwal, R.K., Gupta, J.P., Saxena, S.K. and Muthana, K.D. (1975) Studies on soil physico-chemical and ecological changes under twelve years old five desert species of western Rajasthan. *Indian Forester* **202**, 863-872.

Aksoy, H. (2003) Physically-based mathematical formulation for hill-slope scale prediction of Erosion in ungauged basins. In: *Erosion Predictions in Ungauged Basins: Integrating Methods and Techniques* (D. Boer, F. Wojciech, T. Mizuyama and A. Pietroni, Eds), IARS Publ. 279. IAHS Press, Wallingford, UK. pp 79-83.

Balkrishan and Toky, O.P. (1993) Significance of nitrogen fixing woody legume trees in forestry. *Indian Forester* 126-132.

Bharadwaj, S.P. (1998) Zero tillage and weed mulch for erosion control on slopy farmland in Doon valley. *Indian Journal of Soil Conservation* **26**(2), 81-85.

Brinsen, M., Brandshaw, B.D., Holmes, R.N. and Elkins, J.B. (1980) Litter fall stremflow and through fall nutrient fluxes in alluvial swamp forest. *Ecology* **61**, 827-835.

Browning, G.M., Parish, C.L. and Glass, J. (1947) A method for determining the use and limitations of rotation and conservation practices in the control of soil erosion in Iowa. *Journal of American Society of Agronomy* **39**, 65-73.

Chepil, W.S. and Woodruff, N.P. (1963) The physics of wind erosion and its control. *Advnnces in Agronomy* **15**, 211-301.

CSWCRTI (2007) *Perspective Plan, Vision 2025*, Central Soil and Water Conservation Research and Training Institute, Dehradun, India, 87 p.

Dadhwal, K.S. (1999) Arresting environmental degradation of limestone mined lands from north-west Himalaya through agroforestry. In: *Short Course on Sustainable Rehabilitation of Degraded Lands through Agroforestry* (K.R. Solanki, A.K. Bisarria and A.K. Handa, Eds), National Research Centre for Agroforestry, Jhansi, (UP), India. pp 95-101.

Chemical Composition of Soils

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1. Introduction

Soils are the products of many forces and agents. They exist in the lithosphere where geological deposits get affected by the forces of wind, water and climate, and occasionally by earthquakes, volcanic eruptions and glaciation. The climate of a region plays a dominant role in altering the lithospheric materials into soils. But, soils are produced and will continue to support us even in places where climate has little apparent influence, as in very dry regions or in the Arctic. It is because of the past history of the Earth, existing plants and animals, groundwater, and man's multitudinous activities which all contribute to the production of the more or less loose material at the earth's surface that we call 'soil' in which plants grow and on which we live.

Soil composition concerns the geochemistry of that portion of the mantle of the earth which is currently under exposure to the atmosphere and is subjected to biotic influences. Considering soil as a material, its essential ingredients are:

(a) Mineral matter, which is originated from a rock by the action of a series of weathering processes,
(b) Organic matter, which is generated from the residues of organisms (plant and animal),
(c) Soil water, containing substances in colloidal or true solution, and
(d) Soil air.

The solid phases (mineral and organic) generally make up only about 50% of the soil volume, the other half being constituted by the soil solution and soil air (Figure 1).

Since the density of organic matter in a soil is only about 0.5 Mg/m^3, while that of mineral matter is approximately 2.7 Mg/m^3, 5% (by weight) of organic matter occupies nearly 25% of the volume of solids. The liquid and gas phases together often make up as mentioned above, about half of the soil volume, in the relative proportions of each varying rapidly as the liquid phase (water) is drained away or used by the plants.

The gas phase of soils, which often occupies about one-fourth of the soil volume, contains approximately 80% nitrogen, as in the ordinary air, and about 20% (by volume) oxygen. In well-drained soils, the CO_2-content of soil air is less than 0.1% by volume; in poorly-drained soils, it rises to 5-10%, and in extreme cases like frozen soil horizons, it may be as high as 20% and thus can replace all oxygen. Rare gases like Ne, Ar, Kr and Xe also occur in the soil atmosphere, though in trace amounts.

Soils are the products of several chemical, biochemical and physical processes. These processes affect the earth material under various topographic and climatic conditions and the soils formed reflect landforms and climatic and geomorphic history of the region in which they are evolved. Under the control of these factors, chemical compounds are translocated and deposited in deeper soil horizons; sometimes these are returned to the upper horizons by biocycling, and sometimes they are removed from the soil by leaching. In spite of this, most

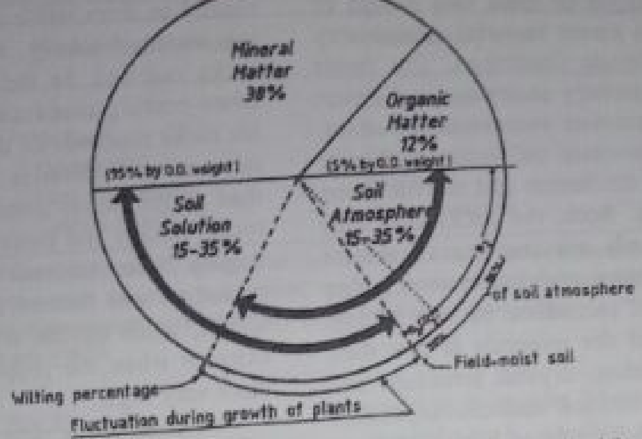


Figure 1. Percentage composition of soil (volume basis)

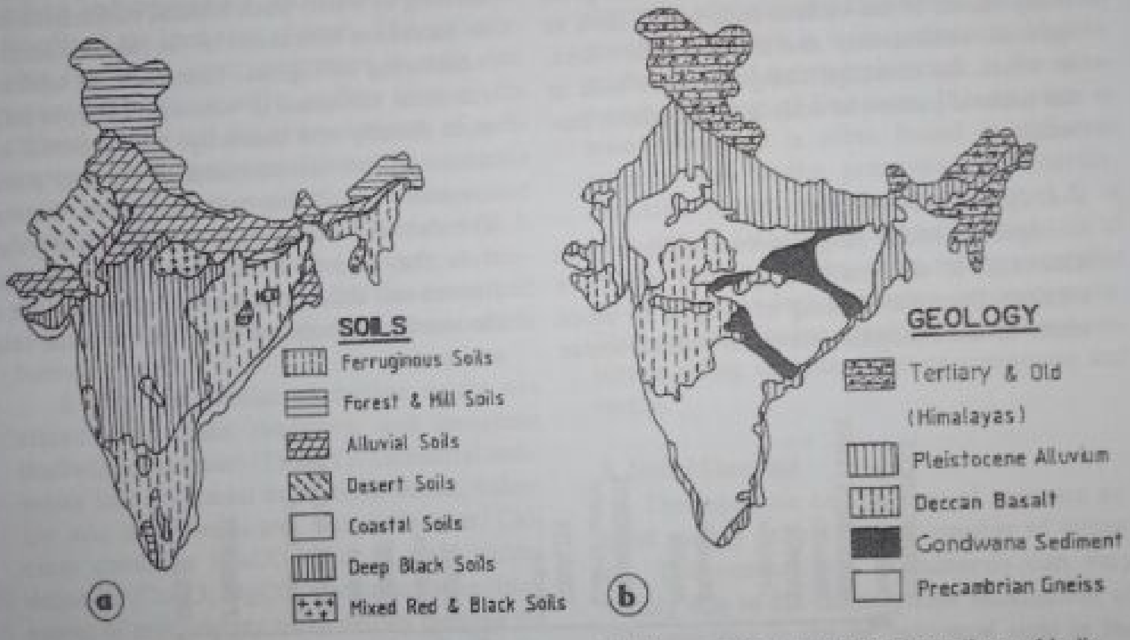


Figure 2. (a) Distribution of different kinds of soil, and (b) the spread of geological formations of India

soils reveal through the chemical composition of their constituent minerals, the nature of their original parent rock.

2. Composition of Earth's Crust

Coarse-grained rocks tend to weather faster than fine-grained ones, and basic rocks weather faster than acid ones. Soils formed from the basic rocks are usually more fertile than those formed from the acid rocks. Distribution of different kinds of soils of India (Figure 2a) and the different geological rock formations (Figure 2b) demonstrate the effect of parent rock on soil formation. The fine-textured Deccan basalt (*see* later) has been responsible for the formation of clayey black soils, whereas coarse-grained metamorphic rocks (*see* later) have given rise to ferruginous soils. However, the occurrence of ferruginous soils in black soil areas, and of black soils in the ferruginous soil areas reflect two contrasting chemical environments that were conducive to the formation of these two groups of soils on the same parent material, presumably under similar climatic conditions. The forest and hill soils on tertiary sediments and desert soils in the Pleistocene sediments, however, indicate the predominant influence of climate and vegetation on the former and climate alone on the latter soils. Both, the rocks and their mineral components are important because their structure, texture and hardness together determine physical resistance to weathering. The significance of the minerals is that their chemical composition, crystal structure and susceptibility to alteration and decomposition are all important factors from both pedologic and agricultural standpoints (Mohr *et al.*, 1972). Therefore, it is useful to note the mineralogical composition of the various rock types, such as igneous, sedimentary and metamorphic rocks, in which the chemical compounds are held in the mineral portion of soils and soil parent materials.

2.1. Igneous Rocks

Igneous rocks range in composition from ultrabasic to acid rocks (Figure 3), and have resulted from the cooling of magma or silica melts of the lithophile elements (which ionize readily or form stable oxyanions). When cooling occurred slowly, coarse-grained, plutonic rocks resulted. In the event of an eruption, when cooling was rapid, finer-grained, volcanic rocks resulted. Of the plutonic rocks, granites and granodiorites are far more plentiful than all the others combined. Among volcanic rocks, basalts and basic andesites predominate largely. When the melt was deficient in silica, nepheline was formed from melts high in Na and K, while olivine was formed at the other extreme when the proportions of Fe and Mg were very high.

Volcanic rocks range in texture from massive to ash, depending on the degree of fracturing that occurred by either gas expansion or chilling in water. Such textural differences lead to enormous alterations in the rate of chemical weathering during soil formation. In addition, chemical variations in various rock types influence directly and indirectly, the elemental and mineralogical compositions of the soil parent material and of the soils developed thereon. The elemental composition of rocks mainly controls the nature of the secondary minerals formed and influences the ionic composition of the weathering solution in contact with the mineral grains.

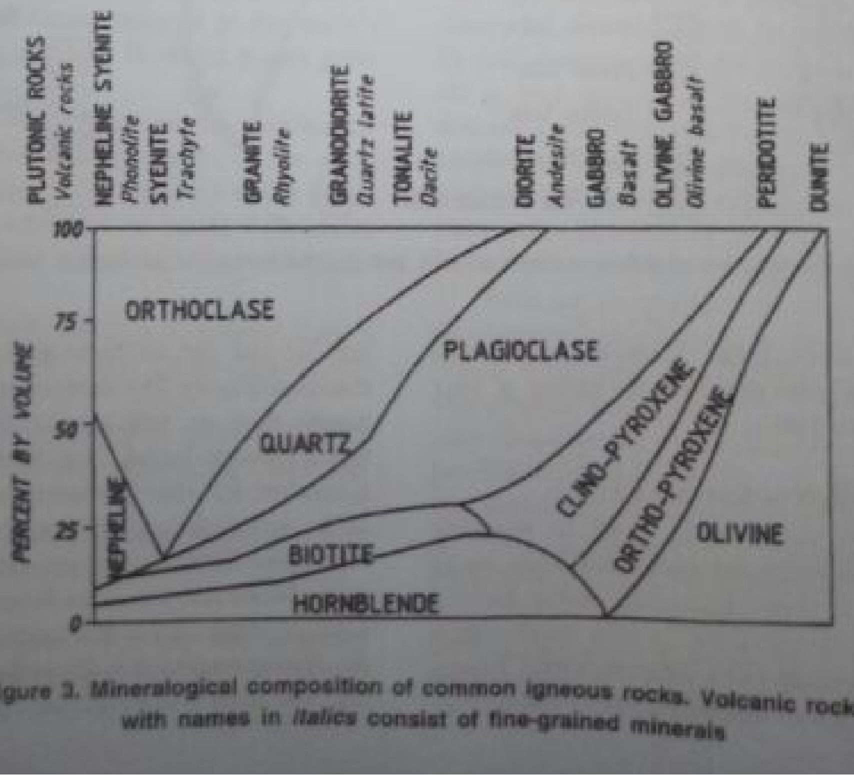


Figure 3. Mineralogical composition of common igneous rocks. Volcanic rocks with names in *italics* consist of fine-grained minerals

Table 1. Some important metamorphic and sedimentary rocks and their predominant minerals

Metamorphic rock	Predominant mineral	Sedimentary rock	Predominant mineral
Gneiss	Varies*	Limestone	Calcite ($CaCO_3$)
Schist	Varies*	Dolomite	Dolomite [$CaMg(CO_3)_2$]
Quartzite	Quartz (SiO_2)	Sandstone	Quartz (SiO_2)
Slate	Clay minerals	Shale	Clay minerals
Marble	Calcite ($CaCO_3$)	Conglomerate	Varies**

* The minerals present are determined by the original rock which has been metamorphosed. Primary minerals present in the igneous rocks commonly dominate these rocks; however, some secondary minerals are also present.

** Small stones of varied mineralogical composition are cemented into conglomerate.

Source: Buckman and Brady (1971)

2.2. *Sedimentary Rocks*

Sedimentary rocks are formed near the interface of the earth's crust with the hydrosphere and atmosphere, and cover about three-fourths of the land area of earth. This geomorphic fact is of great importance to soils, because the higher specific surface area of the mineral grains and the general nature of sedimentary rocks (less consolidated than igneous rocks) permit more rapid biotic invasion and development of deep, highly productive soils. It is probable that accumulation of hydrous minerals in the sedimentary muds favoured and conditioned the evolutionary development of life forms on earth.

Sedimentary rocks of marine origin are classified as shale, sandstone, and limestone (including dolomite) (Table 1). Terrestrial sediments include glacial drift, loess, dunes, volcanic ash, and stream and lake sediments. Calcium carbonate ($CaCO_3$) and to some extent dolomite ($CaCO_3.MgCO_3$), accumulate in limestones of sedimentary rocks mainly through the chemical and biochemical precipitation in the ocean waters, and also as chemical precipitates (evaporites) in lakes and soils.

2.3. *Metamorphic Rocks*

Metamorphic rocks are formed by the effect of heat and pressure on sedimentary and igneous rocks. The mineralogical composition and texture of the rocks strongly affect the composition of soils formed. Details about minerals in the metamorphic rocks are given in Table 1.

A gneissic rock is a laminated or banded metamorphic rock similar in mineralogical composition to granite. Mica schists, another metamorphic rock, weather rapidly to give rise to soil high in vermiculite and quartz. The third form, slates are produced by compression and alteration of shales and mudstones, giving fine-grained rocks high in mica and quartz. Marble, another form of metamorphic rock, is the metamorphosed form of limestone. It is resistant to weathering and is often found in different colours due to the presence of minerals. Quartzite is metamorphosed sandstone. It is very rich in SiO_2 and is extremely resistant to weathering. In addition, there are metamorphic rocks that are less extensive in occurrence. They are amphibolite, amphibole schist or hornblende schist, chlorite schist or greenstone and tactite.

3. Soil Minerals

The inorganic components of soils are derived mainly from a limited number of minerals. The composition of minerals can vary widely due to the isomorphous substitution of different ions in similar structural sites in the crystal; each mineral structure may have an isomorphous series of minerals. Thus, many of the essential trace elements are present in host minerals, and constitute sources of the plant available pool of trace metal ions, in addition to that of metal ions present in larger amounts.

A small portion of various ions in soil minerals is readily available in exchangeable or adsorbed form. This form assumes importance in the nutrition of crop plants grown on different soils. However, the renewal of ions in the exchangeable form through weathering of minerals is a slow process. The mineral species in soils are many and they influence the release of structural ions and the chemical activity of exchangeable ions. Therefore, the mineral composition of soils has a strong relevance to soil fertility.

Soil minerals are classified as: (a) primary minerals formed at elevated temperatures and inherited from igneous and metamorphic rocks, and (b) secondary minerals formed by low-temperature reactions and inherited by soils from some sedimentary rocks or formed in soils by weathering (Jackson, 1964). The primary and secondary mineral classes are not mutually exclusive, and some minerals occur in both.

The silicates comprise a bulk of most of the soils. These occur as both primary minerals inherited from igneous or metamorphic rocks, and secondary minerals formed from the weathering products of primary minerals. The fundamental unit of all silicate structures is the SiO_4^{4-} – tetrahedron. Several different arrangements of the SiO_4^{4-} – tetrahedron occur, partly accounting for the large number of silicate minerals and providing a basis for their classification. The tetrahedra may be present as single tetrahedra (nesosilicates), double tetrahedra (sorosilicates), rings (cyclosilicates), single or double chains (inosilicates) or frameworks (tectosilicates) (Table 2). In all these arrangements, adjacent SiO_4^{4-} – tetrahedra share only corners. It means they share only corners. The minerals from each silicate class, commonly found in soils, are listed in Table 2.

3.1. *Primary Minerals in Soils*

Primary minerals are inherited from the parent rock and constitute a major part of the sand and silt fractions of most of the soils. For a given soil, the sedimentation history of the parent material determines the sand content (2000-50 μm) of the soils, whereas the combined effect of weathering and sedimentation determines the silt content (50-2 μm). The combination of sand and silt determines the percentages of primary minerals in a soil. The primary minerals may also occur in the clay fraction (< 2 μm) of less-weathered soils.

Throughout the world, the most abundant minerals and mineral groups in sand and silt of soils are quartz–SiO_2 (silica), and feldspars–$MAlSi_3O_8$ (silicates of basic cations, M^+). There are other minerals like pyroxene, amphibole, olivine and accessory minerals of primary origin that are present in the soils. Some carbonates, sulphur–bearing minerals and zeolites are also present. Though present only in smaller amounts, they are important to the chemistry of soils.

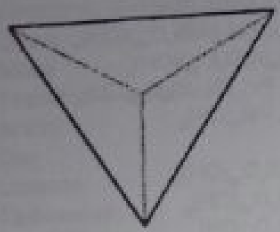
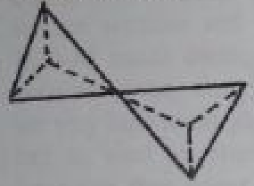
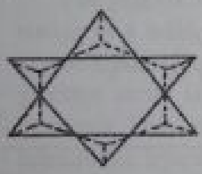
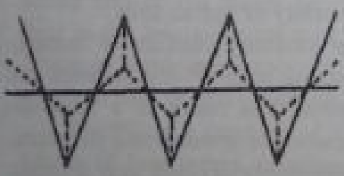
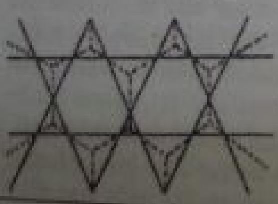
3.1.1. Quartz

Free silica occurs in soils mainly in the form of quartz (SiO_2). It consists of a continuous framework of silica tetrahedra. It is almost ubiquitous in soils and it is only a rare soil that does not contain some quartz. It is often present in coarse clays (2-0.2 μm). Free silica also occurs in some soils as *opal* ($SiO_2.nH_2O$), including opal pseudomorphic after plant cells (phytoliths), chalcedony, agate, flint and chert. The other quartz polymorphs (cristobalite and tridymite) are rare in soils. However, cristobalite has been found in the sand and silt fractions in ferruginous and black soils developed in alluvium of weathering Deccan basalt (Pal and Deshpande, 1987; Bhattacharyya *et al.*, 1993).

Quartz is inherited mostly from the soil parent material. But, aeolian contributions are believed to be widespread and are usually considered to be the source of quartz in soils derived from basalt. However, its presence in soils derived from Deccan basalt, cannot be attributed entirely to the aeolian origin because Deccan basalts contain quartz (Figure 4a) as a secondary mineral.

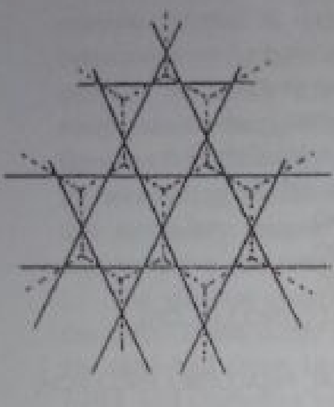
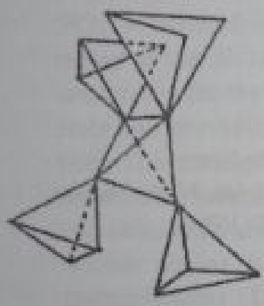
The main process of concentration of free silica is podzolization, under which organic matter chelates the sesquioxidic cations and reduces ferric iron so that these ions are leached out of the bleached A2 (or E) horizons. Despite its stability, quartz may be subject to dissolution under intense weathering, especially when the grains are very small. The tendency for SiO_2 is, therefore, to accumulate as coarser grains of sand, silt, and even coarse clay in highly podzolized horizons. The most common and conspicuous light grey or white E horizons of many podzolic soils consist of sand-sized SiO_2 and other aluminosilicate particles, but a high content of bleached silt and rarely of coarse clay, do occasionally occur in the E

Table 2. Classification of silicate minerals

Silicate class and unit composition arrangement of $(SiO_4)^{4-}$-tetrahedra	Mineral	Ideal formula
Nesosilicates $(SiO_4)^{4-}$ 	Olivine	$(Mg, Fe)_2SiO_4$
	Forsterite	Mg_2SiO_4
	Fayalite	Fe_2SiO_4
	Zircon	$ZrSiO_4$
	Sphene	$CaTiO(SiO_4)$
	Topaz	$Al_2SiO_4(F, OH)_2$
	Garnets	$X_3Y_2(SiO_4)_3$, where $X = Ca, Mg, Fe^{2+}, Mn^{2+}$ $Y = Al, Fe^{3+}, Cr^{3+}$
	Andalusite	Al_2SiO_5
	Sillimanite	Al_2SiO_5
	Kyanite	Al_2SiO_5
	Staurolite	$Fe_2Al_9O_6(SiO_4)_4(O, OH)_2$
Sorosilicates $(Si_2O_7)^{6-}$ 	Epidote	$Ca_2(Al, Fe)Al_2O(SiO_4)(Si_2O_7)(OH)$
Cyclosilicates $(Si_6O_{18})^{12-}$ 	Beryl	$Be_3Al_2(Si_6O_{18})$
	Tourmaline	$(Na, Ca)(Li, Mg, Al)(Al, Fe, Mn)_6$- $(BO_3)_3(Si_6O_{18})(OH)_4$
Inosilicates (single chains) $(SiO_3)^{2-}$ 	Pyroxenes	
	Augite	$(Ca, Na)(Mg, Fe, Al)(Si, Al)_2O_6$
	Enstatite	$MgSiO_3$
	Hypersthene	$(Mg, Fe)SiO_3$
	Diopside	$CaMgSi_2O_6$
	Hedenbergite	$CaFeSi_2O_6$
	Pyroxenoids	
	Wollastonite	$CaSiO_3$
	Rhodonite	$MnSiO_3$
Inosilicates (double chains) $(Si_4O_{11})^{6-}$ 	Amphiboles	
	Hornblende	$(Ca, Na)_{2-3}(Mg, Fe, Al)_5Si_6$- $(Si, Al)_2O_{22}(OH)_2$
	Tremolite	$Ca_2Mg_5Si_8O_{22}(OH)_2$
	Actinolite	$Ca_2(Mg, Fe)_5Si_8O_{22}(OH)_2$
	Cummingtonite	$(Mg, Fe)_7Si_8O_{22}(OH)_2$
	Grunerite	$Fe_7Si_8O_{22}(OH)_2$

Contd

Table 2. Classification of silicate minerals — Contd

Silicate class and unit composition arrangement of $(SiO_4)^{4-}$-tetrahedra	Mineral	Ideal formula
Phyllosilicates $(Si_2O_5)^{2-}$ 	Micas	
	Muscovite	$KAl_2(AlSi_3O_{10})(OH)_2$
	Biotite	$K(Mg, Fe)_3(AlSi_3O_{10})(OH)_2$
	Phlogopite	$KMg_3(AlSi_3O_{10})(OH)_2$
	Chlorites	$(Mg, Fe)_3(Si, Al)_4O_{10}(OH)_2$- $(Mg, Fe)_3(OH)_6$
	Clay minerals (selected)	
	Talc	$Mg_3Si_4O_{10}(OH)_2$
	Pyrophyllite	$Al_2Si_4O_{10}(OH)_2$
	Kaolinite	$Al_2Si_2O_5(OH)_4$
	Smectite	Variable
	Vermiculite	Variable
	Serpentines	
	Antigorite	$Mg_2Si_2O_5(OH)_4$
	Chrysotile	$Mg_3Si_2O_5(OH)_4$
Tectosilicates $(SiO_2)^0$ 	Feldspars	
	Orthoclase	$KAlSi_3O_8$
	Albite	$NaAlSi_3O_8$
	Anorthite	$CaAl_2Si_2O_8$
	SiO_2 Group	
	Quartz	SiO_2
	Tridymite	SiO_2
	Cristobalite	SiO_2
	Zeolites	
	Analcime	$NaAlSi_2O_6 \cdot H_2O$
	Clinoptilolite	$(Na_3K_3)(Al_6Si_{30}O_{72}) \cdot 24H_2O$
	Feldspathoids	
	Nephelene	$(Na, K)AlSiO_4$

Source: Adapted from Schulze (1989)

horizons. Neoformed quartz has also been detected in petroplinthite of ferruginous soils of southern India.

3.1.2. Feldspars

Feldspars are anhydrous aluminosilicates of K, Na and Ca and occasionally of other large cations such as Ba. The feldspars make up an average of about 60% (by weight) of igneous rocks and large fractions of sedimentary rocks (Table 3).

Next to quartz, feldspars are the most widespread primary minerals in soils. They may constitute most of the sand and silt in some slightly weathered soils developed from highly feldspathic parent materials. There is usually more feldspar in the coarser soil fractions than in the clay, but it is often reported as a minor component in coarse clay (2-0.2 µm). Feldspars occur as components of the clay fractions in a large variety of Indian soils.

The feldspar structure consists of tetrahedra linked by sharing each O^{2-} ion between the adjacent tetrahedra. The result is a three-dimensional framework or tectosilicate structure. Feldspars are commonly divided into potassic species, or K-feldspar ($KAlSi_3O_8$, polymorphs orthoclase and microcline), and plagioclase series (ranging from albite, $NaAlSi_3O_8$ to anorthite, $CaAl_2Si_2O_8$). Orthoclase and microcline are common in plutonic and metamorphic rocks such as granites, syenites, gneisses, and some pegmatites. The potassium feldspars occur commonly in the silt and sand fractions of soils, but may be abundant in the clay-size fraction of soils which have undergone less weather-

Table 3. An average mineral composition of igneous and sedimentary rocks

Minerals	Average percentage		
	Igneous rock	Shale	Sandstone
Feldspar	59.5	30.0	11.5
Amphibole and pyroxenes	16.8	—	Trace
Quartz	12.0	22.3	66.8
Micas	3.8	—	Trace
Titaniferous minerals	1.5	—	Trace
Apatite	0.6	—	Trace
Clay	—	25.0	6.6
Limonite	—	5.6	1.8
Carbonates	—	5.7	11.1
Other minerals	5.8	11.4	2.2

Source: Jackson (1964)

ing. In the finer silt and clay sizes, they indicate a reserve of K, but the release of K by surface reaction and also by the rupture of Si-O-Al bonds during breakdown of mineral structure is very slow and K-feldspars contribute very little to the soil pool of potassium that can be used by plants. Thus, they are less important than micas as a mineral source of potassium in soils.

The plagioclase feldspars are a series consisting of a solid solution of albite–$NaAlSi_3O_8$ and anorthite–$CaAl_2Si_2O_8$. Classically, the plagioclase series is named by compositional increments. They are albite, oligoclase, andesine, labradorite, bytownite and anorthite. Albite and oligoclase occur commonly in pegmatites and in metamorphic rocks, but do occur in some igneous rocks too. The bytownite and labradorite end of the series becomes more abundant in the less-siliceous rocks such as basalts. These minerals are important sources of available calcium of less-weathered soils.

Anorthite, a product of contact metamorphism of limestones, weathers rapidly in soils even in temperate climates. Thus, the chemistry and dissolution of plagioclase series of feldspars provide an interesting and important relationship between the degree of weathering (Figures 4b and 4c) and the fertility status of soils. The exchange complex of many of the ferruginous soils of southern and central India is dominated by Ca^{2+} ions and the soils are moderately fertile, indicating the importance of the presence of plagioclase. Thus, in many highly-weathered tropical and subtropical soils, the presence or absence of plagioclases in the parent rock makes a clear difference between productive and unproductive soils.

Many minerals have been reported as alteration products of feldspars, depending on the feldspar composition and the weathering environment. The clay-size dioctahedral smectite (*see* later) was the first weathering product of plagioclase of the Peninsular gneiss of southern India during the tropical humid climate of the pre-Pliocene period (Figure 4d) (Pal *et al.*, 1989).

3.1.3. Accessory Minerals of Soils

A wide variety of accessory minerals in small but significant amounts is found in the soils. These minerals occur in the coarser fractions of soils, but are sometimes found in the clay fraction as well. Most of these minerals are heavy minerals falling in the fractions with density greater than 2.95 Mg/m^3 and are used profusely in soil genesis studies to infer the origin of the parent material (Pal and Roy, 1978; Rengasamy *et al.*, 1978).

3.1.4. Pyroxenes and Amphiboles

These two minerals are two groups of ferro-magnesian minerals and their structures consist of long chains of linked silica tetrahedra. The pyroxenes consist of a single chain (two O^{2-} ions shared in each tetrahedron) and the amphiboles of double chains (two and three O^{2-} ions shared alternatively between succes-

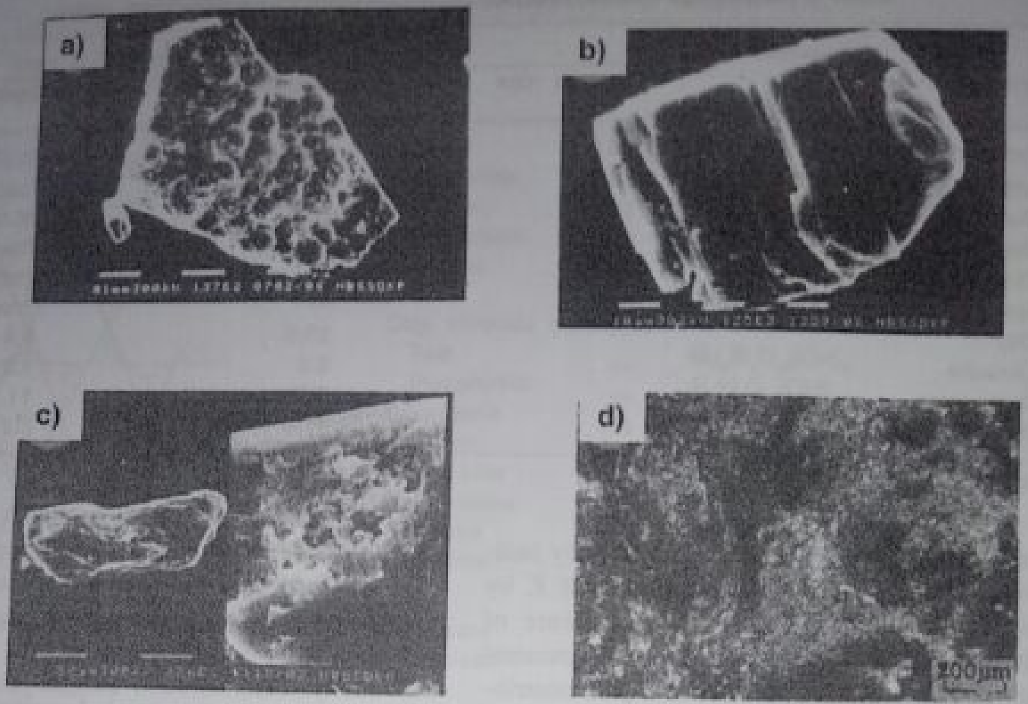


Figure 4. Scanning electron microscope photographs of (a) quartz particle from ferruginous soil, (b) relatively unweathered plagioclase from alluvial soils of the Indo-Gangetic Plains, (c) weathered plagioclase from alluvial soils of the Indo-Gangetic Plains, (d) photomicrographs of smectitic clay mineral pseudomorphs after plagioclase in shrink-swell soils

sive tetrahedra). These chain silicates are classified as inosilicates. Due to a variety of substitutions, these minerals are excellent host minerals for trace elements in soils and also for main constituent cations like Ca, Mg and Fe. Their weathering rate is quite favourable to provide sufficient amounts of these ions in available form for plant nutrition.

The compositions of pyroxene are represented by enstatite — $MgSiO_3$, hypersthene — $(Mg, Fe)SiO_3$, diopside — $CaMgSi_2O_6$, and augite — $(Ca,Na)(Mg,Fe,Al)\ (Si,Al)_2O_6$. Among these, augite is the most important in amounts and compositional significance to soils. This dark-colour mineral is common in the Deccan basalt.

The compositions of amphibole may be represented by the isomorphous series between the end members tremolite — $Ca_2Mg_5Si_8O_{22}.(OH)_2$ and actinolite — $Ca_2Fe_5Si_8O_{22}(OH)_2$ or $Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2$. Tremolite is characteristic of metamorphosed dolomites; actinolite is also characteristic of metamorphic rocks.

The amphibole group is widely represented in the hornblende series— $(Na,Ca)_2(Mg,Fe,Al)_5(Si,Al)_8O_{22}(OH)_2$ and it is widely distributed in igneous rocks from syenite and granite to gabbro (Figure 3), and in metamorphic rocks such as gneiss, hornblende-schist, and amphibolites.

3.1.5. Olivines

Olivines $[(Mg,Fe)_2\ SiO_4]$ are olive-green minerals forming an isomorphous series between the end members forsterite — Mg_2SiO_4, and fayalite — Fe_2SiO_4. The structure consists of independent silica tetrahedra in which oxygen ions are not shared and Al ions are not substituted for Si ions, while alkali ions are excluded. The magnesium and ferrous ions are both surrounded by six oxide ions of two tetrahedra. These minerals are classified as nesosilicates. Large crystals or phenocrysts of olivine are frequently found in the basic rocks (basalt) elsewhere but not in basalt of Deccan (Pal and Deshpande, 1987). On weathering in soils, these crystals yield smectite (nontronite) and free iron oxides. Olivines are relatively easily weathered.

3.1.6. Other Accessory Minerals

The mineral apatite [$Ca_{10}(PO_4)_6F_2$ and its variant], is the most common primary mineral carrier of phosphorus in rocks and parent materials. In basalt, apatites occur commonly as fine mineral grains. They also occur in calcareous sediments. Fluorapatite—$Ca_{10}F_2(PO_4)_6$, is a common form in the igneous and metamorphic rocks. In a few deposits, F has been partially substituted by either Cl^- or OH^-. In most of the sedimentary deposits, PO_4 has been partially substituted by CO_3, causing subsequently a series of inclusion of Na, Mg, and F, leading to the formation of a group of minerals called 'francolites' (also called carbonate apatite). The occurrence of phosphate minerals in the Indian soils is vary rare. The presence of the plumbogummite group of phosphate minerals has been reported in some Indian soils, and therefore, these are considered to be pedogenetic.

Tourmaline—$Na(Mg,Fe)_3Al_6(BO_3)_3.Si_6O_{18}(OH)_4$ has the cyclosilicate (closed ring) structure represented by the $[Si_6O_{18}]^{12-}$ part of the formula. This mineral has relevance to soil chemistry because it contains boron, a plant essential element. Although the mineral is highly insoluble in normal soils, its solubility is enhanced in alkali (sodic) soils to the extent that boron toxicity may occur.

Epidote — $Ca_2(Al,Fe)Al_2O(OH)(SiO_4).(Si_2O_7)$ represents a mineral group in which some of the Ca may be substituted by rare earths and some Mg may be substituted for Fe. Minerals of the epidote group characteristically occur in metamorphic rocks.

Other accessory minerals, used to establish the origin of soil parent material, are: zircon ($ZrSiO_4$), topaz $[Al_2SiO_4(OH,F]_2$, garnets $[X_3Y_2(SiO_4)_3]$, in which X may be Ca^{2+}, Mg^{2+}, Fe^{2+} or Mn^{2+} and Y may be Al^{3+}, Fe^{3+} or Cr^{3+}, and sometimes Ti or Mn^{3+}, magnetite (Fe_3O_4), ilmenite ($FeTiO_3$), sphene ($CaTiSiO_5$), rutile and anatase (TiO_2), kyanite, sillimanite, and andalusite (Al_2SiO_5), and similarly structured staurolite $[2Al_2SiO_5.Fe(OH)_2]$. The presence of some of these minerals has been helpful in establishing a relationship between the parent material and ferruginous soils of the eastern and southern India (Pal and Roy, 1978; Rengasamy *et al.*, 1978).

3.2. Secondary Minerals

3.2.1. Carbonates, Sulphur-bearing Minerals and Zeolites

Carbonates – Calcite ($CaCO_3$) is the most abundant carbonate mineral in dry (subhumid to arid) regions. It may be inherited (Figure 5a) or neoformed (Figure 5b). It often makes up much of the soil mass in thick calcic or petrocalcic horizons (calcretes). In some instances, the isomorph, aragonite ($CaCO_3$), occurs in the sediments and mantle rocks and may, therefore, be found in some soils. Arid and semi-arid climate is the prime factor responsible for the depletion of Ca^{2+} ions from the soil solution in the form of pedogenic carbonates (neoformed calcites) and also in the simultaneous increase of both sodium adsorption ratio (SAR) and exchangeable sodium percentage (ESP) with pedon depth (Balpande *et al.*, 1996). The impairment of percolative moisture regime due to subsoil sodicity provides an example of a soil where gains exceeds losses. This self-terminating process will lead to not only extreme sodicity but also formation of a calcic horizon.

With the establishment of vegetation on soils, some $CaCO_3$ of soils is dissolved as $Ca(HCO_3)_2$ and this dissolution provides a source of available calcium in soils. This reaction has relevance to reclamation of alkali (sodic) soils of the Indo-Gangetic Plains with or without addition of gypsum (Gupta and Abrol, 1990). Only in rare environments magnesite ($MgCO_3$) can persist in surficial soil and rock layers, because of its relatively high solubility in water.

Dolomite — $CaCO_3.MgCO_3$, in soil parent materials is inherited from dolomitic limestone and mantle rocks derived therefrom. Its dissolution by H_2CO_3 supplies both calcium and magnesium ions that could become available in soils.

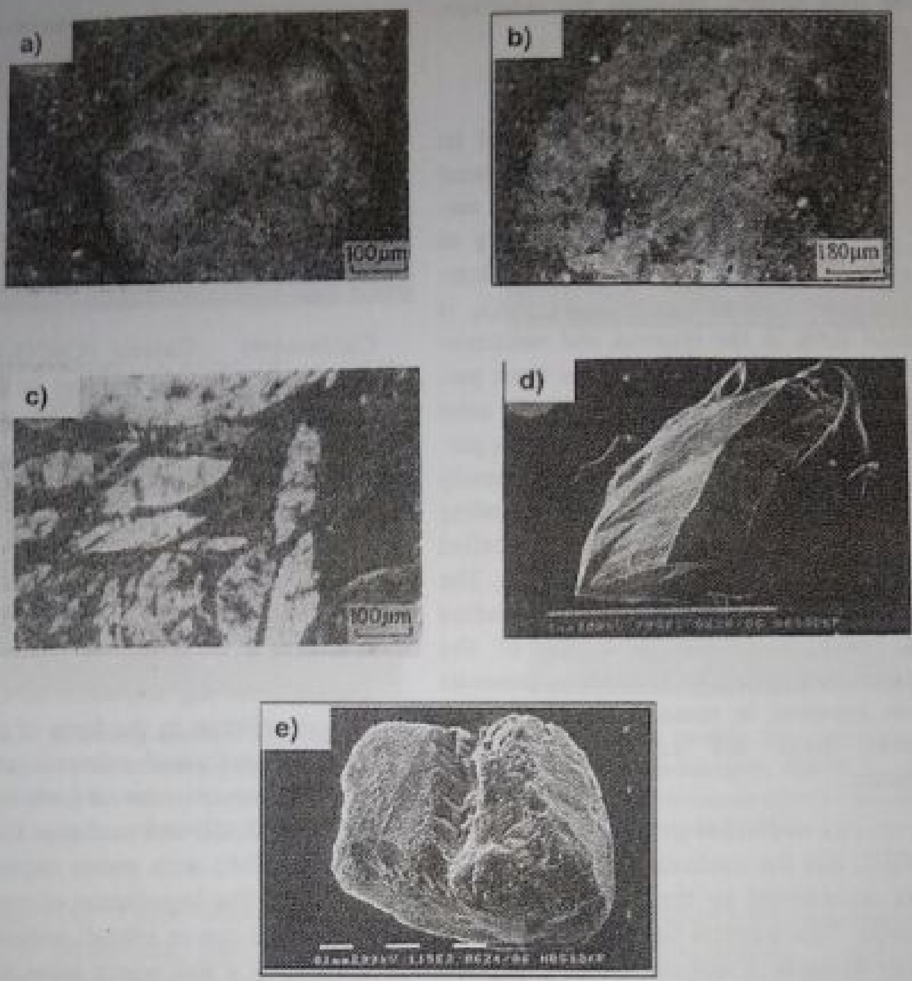


Figure 5. Photomicrographs of (a) lithogenic (parent material) calcium carbonate of shrink-swell soils, (b) pedogenic calcium carbonate of shrink-swell soils, (c) lenticular gypsum crystal of shrink-swell soils; and scanning electron microscope photographs of (d) unweathered heulandite of shrink-swell soils, and (e) weathered heulandite of ferruginous soils

Siderite — $FeCO_3$, is formed by the decomposition of iron-bearing minerals in environments suitable for carbonate accumulation. It is more resistant to the action of H_2CO_3 than $CaCO_3$.

Gypsum — $CaSO_4.2H_2O$ is a relatively common component in soils of arid climates, found frequently in a horizon situated somewhat beneath the horizon in which $CaCO_3$ is accumulated. It occurs sometimes as indurated horizons and as crystals disseminated in surface soil horizons. Gypsum is found in certain subsoils as a weathering product of pyrites (FeS_2), even in soils of humid regions. In arid regions, epsomite ($MgSO_4.7H_2O$) and other soluble sulphates may occur in soils. Minor amounts of the insoluble minerals, namely sphalerite (ZnS), chalcopyrite ($CuFeS_2$), and cobaltite (CoAsS), may also occur in soils. The presence of gypsum in the Indian soils is found very rarely. One such example is its presence in shrink-swell soils of Tamil Nadu (Figure 5c). The presence of gypsum in these soils may have positive implications in water-management practices.

Pyrite and Marcasite — FeS_2 ('fool's gold'), are formed during the reduction phase in the geomorphic cycle of sedimentation when iron of ferric oxides and sulphur of sulphates get reduced [Equation (1)].

$$2Fe_2O_3 + 4Na_2SO_4 + 4H_2O + 9/n\ Cn \longrightarrow 4FeS + 8NaHCO_3 + CO_2 \quad ...(1)$$

where, Cn represents organic matter.

The occurrence of sulphides is frequent in shales, underclays of coal seams, and some sedimentary rocks, including limestones. Pyrite is also found in igneous and metamorphic rocks.

The sulphate reduction reaction is extremely fundamental in as much as it utilizes iron of ferric oxide as a cation to replace Na of Na_2SO_4 and to liberate free alkali into the soil solution, lake, or ocean. This reaction occurs under anaerobic conditions. The sulphate ion is reduced in the presence of sulphur reducing bacteria and this reaction is responsible for the development of alkalinity in poorly-drained saline soils. However, in the development of alkali (sodic) soils of the Indo-Gangetic Plains (Pal *et al.*, 2003a) and also of the Purna Valley (Balpande *et al.*, 1996), there has been no role of this reaction.

With time FeS transforms into FeS_2 which on oxidation under good drainage conditions, produces H_2SO_4, resulting in extreme acidity in soils (pH 2.5). Calcium is removed from Ca-bearing minerals by H_2SO_4, and gypsum crystals [$CaSO_4.2H_2O$] are formed. The reaction of sulphuric acid with soil minerals yields jarosite [$KFe_3(OH)_6(SO_4)_2$] and alunite [$KAl_3(OH)_6(SO_4)_2$]. The presence of these minerals in soils has rarely been reported though favourable conditions prevail in some soils of the coastal belt of the country.

Zeolites — Zeolites are crystalline, hydrated aluminosilicates of alkali and alkaline earth cations. They are further characterized by their ability to hydrate and dehydrate reversibly and to exchange some of their constituent cations, without a major change in structure. Zeolites were discovered in 1756 by the Swedish mineralogist, Cronstedt. Since that time, about 50 natural species have been recognized, and at least 150 species having no natural counterparts have been synthesized in the laboratory (Ming and Mumpton, 1989).

During the past three decades, zeolite minerals have been recognized with increasing regularity as common constituents of Caenozoic volcanogenic sedimentary rocks and altered pyroclastic rocks. Zeolites have also been found as secondary minerals in the Deccan basalt of the Western Ghats in Maharashtra (Sabale and Vishwakarma, 1996).

Along with quartz and feldspars, zeolites are tectosilicates. They consist of three-dimensional frameworks of SiO_4^{4-}-tetrahedra wherein all O^{2-} ions of each tetrahedron are shared with the adjacent tetrahedron. The zeolites may be represented by the following general formula:

$$[M_x^{+}, M_y^{2+}]\ [\{Al_{(x+2y)}Si_{n-(x+2y)}O_{2n}\}.mH_2O]$$

where, M^+ and M^{2+} are monovalent and divalent cations, respectively. Cations within square brackets are the exchangeable cations, and the others are known as structural cations because with oxygen they make up the framework of the structure.

Zeolites were not believed to be significant in soils until the 1960s, when analcime ($NaAlSi_2O_6.H_2O$) was found in several saline, high-pH soils formed in alluvium in the San Joaquin Valley of California. The analcime in high-pH soils was considered to be authigenic (Baldar and Whittig, 1968). Authigenic analcime has also been found in sodic soils of the Indo-Gangetic Plains. The occurrence of inherited zeolites in soils has been rare. The presence of heulandite [$Ca_4(Al_8Si_{28}O_{72}).24H_2O$] has been found in the sand, silt and clay of red and black soils of the Western Ghats of Maharashtra (Bhattacharyya *et al.*, 1993) (Figures 5d and 5e). It is believed that zeolites can provide adequate amount of bases to prevent the transformation of smectite to kaolinite, and thus enable formation of shrink-swell soils in microdepressions, even in the humid tropical climate of the Western Ghats. The presence of zeolites in acid ferruginous soils of the Western Ghats may be the reason as to why crops do not show response to liming.

3.3. Layer-Silicate Minerals of Soils

Layer silicates or phyllosilicates (Table 4) strongly influence both the physical and chemical properties of soils because of their generally small particle sizes, high specific surface areas, and unique cation exchange properties. Comprehension about phyllosilicate minerals is very basic to soil clay mineralogy. In fact, many clays contain other hydrous silicate and oxide components that may greatly affect soil properties. Layer silicates are also important components of silt and sand and even gravel of soils. Therefore, the layer silicate names need not be restricted in meaning to clays.

3.3.1. Structural Principles

The principal building elements of the layer silicates are two-dimensional arrays of silicon-oxygen tetrahedra and two-dimensional arrays of aluminum or magnesium-oxygen-hydroxyl octahedra. Such sheets of tetrahedra and of octahedra are layered in different fashions in various layered silicates.

In the silicon-oxygen sheets, silicon atoms are coordinated with four oxygen atoms. The oxygen atoms are located on the four corners of a regular tetrahedron with silicon atom in the centre. Three of the four oxygen atoms of each tetrahedron are shared by three neighbouring tetrahedra. The fourth oxygen atom of each tetrahedron is not shared. This arrangement gives hexagonal symmetry to the sheet, in which rings of six oxygen atoms appear. The silicon-oxygen sheet is called the 'tetrahedral sheet' or the 'silica sheet'.

In the Al-, Mg-O-OH sheets, the Al or Mg atoms are coordinated by oxygen atoms or OH groups, forming octahedra. The sharing of oxygen atoms by edges of the neighbouring octahedra results in a sheet. The oxygen atoms and hydroxyl groups lie in two parallel planes with Al or Mg atoms between these planes. The projection of the sheet indicates that the oxygen atoms and hydroxyl groups form a hexagonal close packing. This sheet is called the octahedral sheet or the alumina or magnesia sheet, and is also called as gibbsite layer or brucite layer, respectively.

Phyllosilicates are divided into two groups: 1:1 and 2:1 type minerals, based on the number of tetrahedral and octahedral sheets in each unit layer. Different combinations of these two general structural units, the octahedral and tetrahedral sheets, yield the structures of various layer silicates of importance in soils, including mica, vermiculite, smectite, attapulgite, chlorite, kaolinite, and halloysite, as well as various interstratified and intergradient forms of these layer silicates.

3.3.2. 1:1 Type Minerals

The 1:1 Layer Structure — The 1:1 layer structure consists of a unit made up of one octahedral sheet and one tetrahedral sheet, with the apical oxide (O^{2-}) ions of the tetrahedral sheet being shared with (and becoming part of) the octahedral sheet (Figure 6a). There are three planes of anions. One plane consists of the basal O^{2-} ions of the tetrahedral sheet, and the second consists of O^{2-} ions common to both the tetrahedral and octahedral sheets plus OH^- belonging to the octahedral sheet, and the third consists of only OH^- belonging to the octahedral sheet.

Kaolinite Group — This group includes a number of clay minerals: kaolinite, halloysite, anauxite, dickite, etc. Of these, kaolinite is of maximum importance in soils. The structure of kaolinite minerals consists of 1:1 layers stacked one above the other. Kaolinite is dioctahedral mineral and contains Al^{3+} ions in the octahedral sites and Si^{4+} in the tetrahedral sites (Figure 6b). The 1:1 layer is electrically neutral, and the adjacent layers are held together by hydrogen bonds between the basal oxygens of the tetrahedral sheet of one unit and the hydroxyls of the surface plane of the octahedral sheet of the adjacent unit. Consequently, the structure is fixed and no expansion ordinarily occurs between units when the clay is wetted. Cations and water do not enter between the structural units of the mineral. The effective surface area of kaolinite is thus restricted to its external faces. This is one reason for its low adsorptive capacity for cations. Soils high in kaolinite are generally less fertile than soils in which 2:1 clay minerals predominate.

Kaolinite crystals usually are hexagonal with clear-cut edges (Figure 7a). In comparison with smectite particles, kaolinites are large in size, the diameter generally ranging from 0.1 µm to 5 µm. In contrast to other minerals layer silicate groups, the plasticity, cohesion, shrinkage and swelling properties of kaolinite are very low (Table 4). Kaolinite is abundant in more weathered soils such as Ultisols and Oxisols. It is also commonly present in many ferruginous soils (Alfisols) of southern India. Kaolinites have very little isomorphous

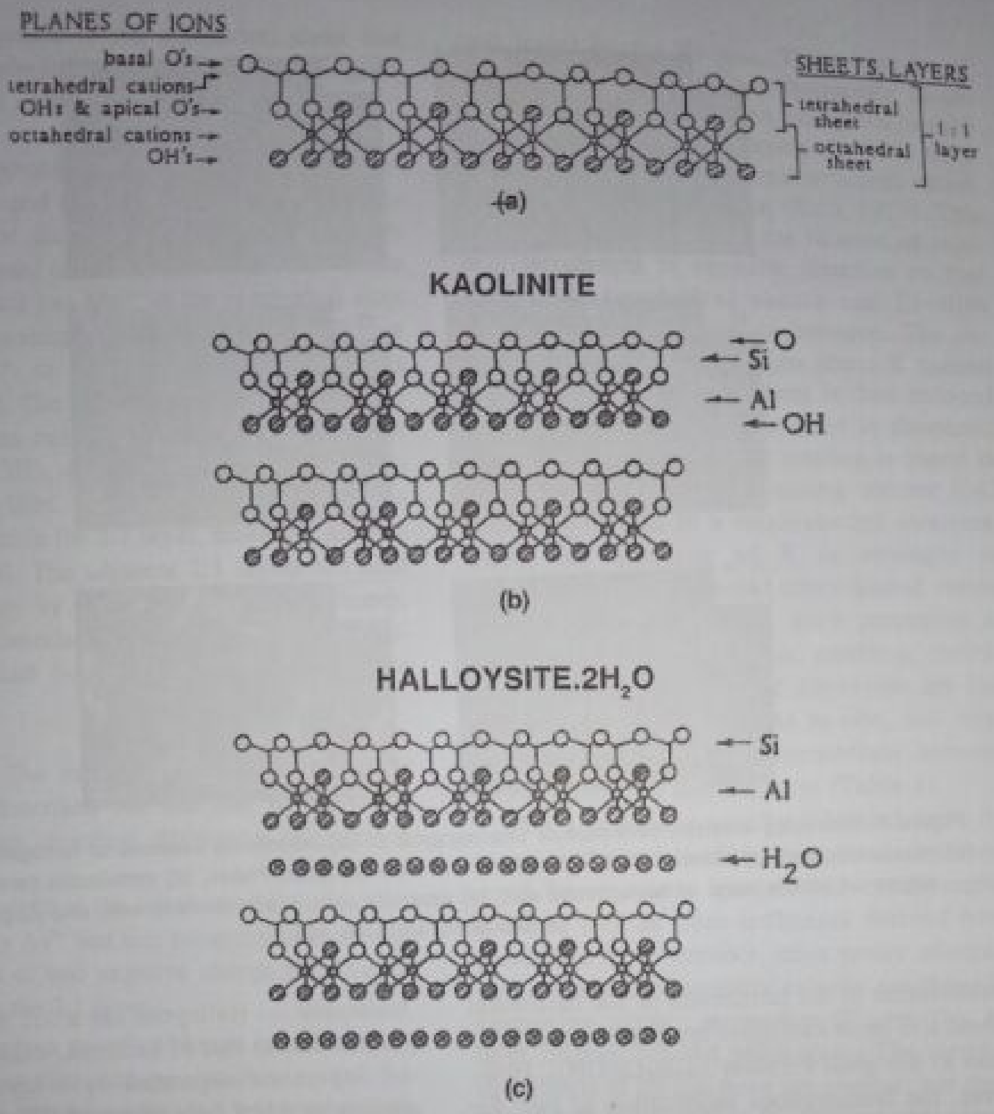


Figure 6. Structural scheme of soil minerals based on tetrahedral and octahedral sheets (a) 1:1 type minerals, (b) kaolinite, and (c) halloysite

Table 4. Properties of three major groups of clay minerals – A comparison

Property	Mineral group		
	Smectite	Mica	Kaolinite
Particle size, µm	0.01-1.0	0.1-2.0	0.1-5.0
Shape	Irregular flakes	Irregular flakes	Hexagonal crystals
Specific surface, m^2/g	700-800	100–120	5–20
External surface area	High	Medium	Low
Internal surface area	Very high	Medium	None
Cohesion and plasticity	High	Medium	Low
Swelling capacity	High	Medium	Low
CEC, $cmol(p^+)kg^{-1}$	80–100	15–40	3–15

Source: Buckman and Brady (1971)

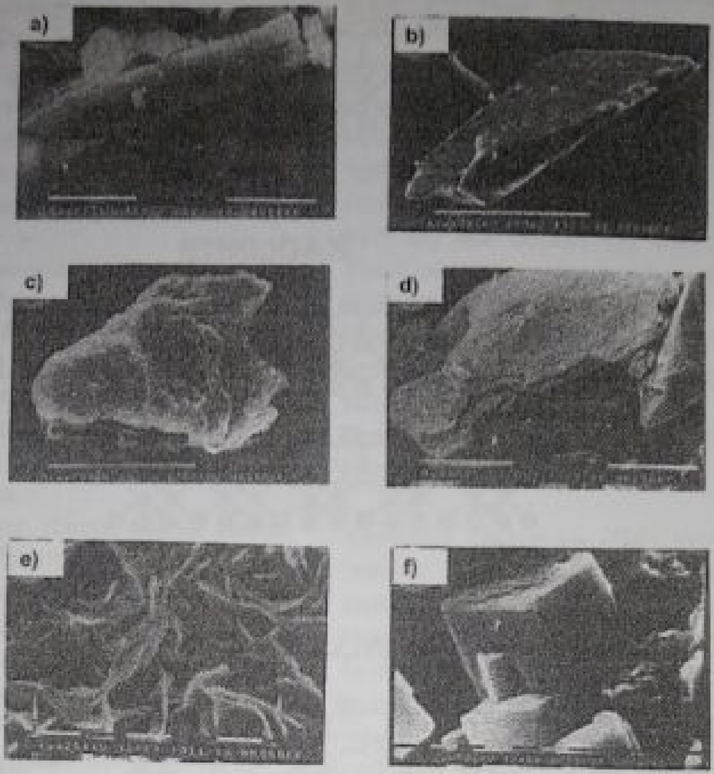


Figure 7. Scanning electron microscope photographs of (a) vermiform kaolinite of ferruginous soils, (b) muscovite, and (c) biotite of alluvial soils of the Indo-Gangetic Plains, (d) vermiculite forming at the edges of biotite mica of ferruginous soil, (e) smectite clay of shrink-swell soil, and (f) gibbsite crystal of ferruginous soil

substitution in the tetrahedral or the octahedral sheet and most kaolinites are close in composition to the ideal formula $Al_2Si_2O_5(OH)_4$. However, the isomorphous substitution of Fe^{3+} for Al^{3+} in the octahedral sheet does occur in ferruginous Alfisols of southern India (Rengasamy *et al.*, 1975).

Kaolinite may be formed in the soils by neosynthesis from the products of hydrolytic decomposition of feldspars and other primary minerals (Rengasamy *et al.*, 1978) and by the conversion of smectite to kaolin by hydroxy Al-interlayering in the expandable minerals and/or mixed layering between 2:1 and 1:1 layers (Pal *et al.*, 1989; Bhattacharyya *et al.*, 1993). The latter process, however, has to provide for the inversion of tetrahedra of alternate tetrahedral sheets so that all the tetrahedra in the crystal are oriented in the same direction. Kaolinite also can be inherited from soil parent materials such as clayey sediments.

Halloysite — Halloysite has a 1:1 layer structure similar to that of kaolinite, except that the 1:1 layers are separated by a layer of H_2O molecules when fully hydrated (Figure 6c), but halloysite often occurs as tubular or spheroidal particles. Halloysite forms early in the weathering process, but it is generally less stable than kaolinite and gives way to kaolinite with time.

3.3.3. 2:1 Type Minerals

In contrast to the 1:1 minerals, the 2:1 minerals are structurally more diverse and are represented by several mineral species.

The 2:1 Layer Structure — The 2:1 layer structure consists of two tetrahedral sheets bound to either side of an octahedral sheet (Figure 8a). There are four planes of anions. The outer two planes consist of the basal oxygens (O^{2-} ions) of the two tetrahedral sheets, while the two inner planes consist of oxygens common to the octahedral sheet and the two tetrahedral sheets, plus the hydroxyls (OH^-) of the octahedral sheet.

Talc and Pyrophyllite — The simple structures of talc and pyrophyllite are a good starting point for discussing 2:1 structures. Both minerals consist of 2:1 layers stacked one above the other. Talc has Mg^{2+} in the octahedral sites and is trioctahedral, while pyrophyllite (Figure 8b) has Al^{3+} in the octahedral sites and is dioctahedral. The tetrahedral sheet in both minerals contains only Si^{4+}, giving ideal formulae $Mg_3Si_4O_{10}(OH)_2$ for talc and $Al_2Si_4O_{10}(OH)_2$ for pyrophyllite. In each case, the charge is balanced within the 2:1 layer, making it electrically neutral. The adjacent 2:1 layers are held together only by weak van der Waal's forces. Both the minerals when found in soils, are usually inherited from low-grade metamorphic rocks.

Micas — The mica minerals have 2:1 layer structure described for talc and pyrophyllite, but with two important differences (Figure 8c), viz.

1. One-fourth of the tetrahedral sites is occupied by Al^{3+} and this substitution causes an excess of one negative charge per formula unit in the 2:1 layer.
2. The excess negative charge is balanced by K^+ ions that occupy interlayer sites between the two adjacent 2:1 layers. This gives an ideal formula of $K(Al_2)(AlSi_3)O_{10}(OH)_2$ for a dioctahedral mica mineral (muscovite) and $KMg_3(AlSi_3)O_{10}(OH)_2$ for phlogophite.

The octahedral sheet of mica may be either dioctahedral or trioctahedral. There are several different mica species because Fe^{2+} and Fe^{3+} can be substituted for Mg^{2+} and Al^{3+} in the octahedral sheet, and Na^+ and Ca^{2+} can be substituted for K^+ in the interlayer. Since the unit layers do not part on addition of water, the K^+ ions between the unit layers are not available for exchange; they are fixed. Only the K^+-ions on the external surfaces can be exchanged for other cations.

The non-swelling character of the micas is attributed to the specific linking effect of the unit layers by the K^+ ions. These ions are of the right size to establish a favourable 12-coordination with the opposite hexagonal oxygen rings of adjoining unit layers. In fact, these rings are not hexagonal in real structure, rather, they are ditrigonal in shape (Rich, 1972). This arrangement is caused by the rotation of adjacent tetrahedral in opposite direction so that the larger tetrahedral sheets can fit with smaller octahedral sheet in between. The coordination number of oxygen about K caused by the ditrigonal arrangement is thus reduced from 12 to 6 and the K-O bond is shortened and strengthened. A larger rotation is found in a dioctahedral structure causing shorter K-O bond length than in a trioctrahedral structure. Thus, the bonding of K is stronger in dioctahedral micas than in trioctahedral varieties of this mineral group. Such properties as hydration, cation adsorption, swelling, shrinkage, plasticity, and ease of dispersion are less developed in soil micas. As to size, soil mica crystals are apparently intermediate between those of smectite and kaolinite (Table 4).

Mica in soils is usually inherited from the parent rocks and is likely to occur in soils derived from various igneous and metamorphic rocks as well as from sediments derived from them. The most common mica group minerals in rocks and consequently in soils are three K-containing micas : muscovite (Figure 7b), biotite (Figure 7c) and phlogopite. The tetrahedral sheets of all the three are similar, but their octahedral sheets differ. Biotite and phlogopite are trioctahedral, with phlogopite containing Mg^{2+} in the octahedral sites, while biotite contains Mg^{2+} and Fe^{2+}. Muscovite is dioctahedral and contains Al in the octahedral sheet.

Soil clay mica often has poorer crystallinity, lower K-content, and higher H_2O-content than the macroscopic muscovite which most closely resembles. This clay-sized mica is often referred to as illite. However, illite is not to be used as a specific mineral name but as a term indicating that it is a mixed assemblage of minerals with mica as the main component. Unless one is using the mineral name in the strict crystallographic sense, for common understanding, soil clay minerals with a dominance of mica can still be called illite (Raman,1984).

Micas weather to other minerals, particularly to vermiculites and smectites, and the K^+ ion released during weathering is an important source of K for plants. It is known that muscovite micas are more resistant to weathering than biotite and phlogopite and in the presence of latter two, muscovite tends to retain its K still more firmly and is a useless K-reserve (Pal, 2003). As a result, muscovite is more likely to persist than biotite or phlogopite in the highly weathered soils (Pal , 2003).

Vermiculite — Vermiculite consists of the 2:1 layer silicate structure described for mica, but instead of having a layer charge of -1 per formula unit and K^+ ion in the interlayer positions, vermiculite has a layer charge of 0.9 to 0.6 and contains Ca^{2+} and Mg^{2+} ions as exchangeable cations in the interlayer (Figures 7d and 8d). The high charge per formula unit gives vermiculites a high CEC and causes them to have a high affinity for weakly hydrated cations such as K^+, NH_4^+, and Cs^+. Fixation of K^+ by vermiculites can be significant in soils high in vermiculite. Reported fixation of K^+ ions in black, alluvial and ferruginous soils of India is due only to the presence of vermiculite and not any other mineral like illite, hydrous micas and smectites (Pal, 2003).

Vermiculites in soils are believed to be formed almost exclusively from the weathering of micas and chlorites. It is often reported that dioctahedral vermiculites are most common in soils. However, trioctahedral vermiculite is very common in soils of arid and semi-arid regions of India and is the weathering product of biotite mica (Figure 7d).

Smectites — The smectite group contains several minerals, such as montmorillonite, beidellite, nontronite, and saponite. Of these, montmorillonite is the most dominant, although others also occur in certain soils. The flake-like crystals of montmorillonite (Figure 7e) are composed of crystal units. The smectite group consists of minerals with the 2:1 structure like mica and vermiculite, but with a still lower charge per formula weight (0.6-0.25). As in vermiculite, the interlayer contains exchangeable cations (Figure 8d).

Dioctahedral smectites are the most common in soils. Their compositions range between the three end members like montmorillonite, beidellite and nontronite. In montmorillonite, Mg^{2+} is substituted for Al^{3+} in the octahedral sheet to produce the layer charge. In beidellite and nontronite, the layer charge is generated in the tetrahedral sheet. Beidellite contains Al^{3+} in the octahedral sheet, while in the nontronite, the octahedral sheet contains Fe^{3+}. Smectites do not fix K^+ ions as readily as vermiculites do because their layer charge is lower.

Smectite dominates in the clays of Indian shrink-swell soils developed in alluvium of Deccan basalt and is nearer to montmorillonite of the montmorillonite-nontronite series and does not have any K-selectivity (Pal, 2003). Its layer charge ranges from 0.28 to 0.78 mol(-)/$(SiAl)_4O_{10}(OH)_2$ (Ray *et al.*, 2003). The observed K-adsorption by these soils is due to the occurrence of small amount of vermiculite mainly concentrated in the coarse clay (2-0.2 μm) fractions (Pal and Durge, 1987).

Smectite crystals range in diameter from 0.01 μm to 1.00 μm. They are thus much smaller than the average kaolinite particles (Table 4). Cations and water molecules are able to move in between the crystal units of smectites. The movement of water between crystal units hydrates the cations and the large hydration energy involved is able to overcome the attractive forces between the unit layers for the interlayer cations, causing smectite to swell more than other layer silicates (van Olphen, 1966).

Smectites are also noted for their high plasticity and cohesion, marked shrinkage on drying and ready dispersion of their fine, and flaky particles (Table 4). The shrink-swell behaviour is most pronounced in the Vertisol order and in Vertic subgroups. The shrink-swell properties lead to engineering problems when houses, roads, and other structures are built on smectitic soils.

Chlorites — Chlorites are a group of clay minerals structurally related to the three-layer minerals. In chlorites, the charge compensation between smectite-type unit layers is by a positively charged interlayer octahedral hydroxide sheet (Figure 8e). A possible formula for such a hydroxide sheet is $[Mg_2Al(OH)_6]^+$. Both octahedral sheets (the one that is part of the 2:1 layer or the interlayer hydroxide sheet) may be either dioctahedral or trioctahedral, and may contain Mg^{2+}, Fe^{2+}, Mn^{2+}, Ni^{2+}, Al^{3+}, Fe^{3+}, and Cr^{3+} ions, giving a large number of different mineral species.

Chlorite minerals in soils are generally primary mineral, inherited from the parent rock and are found only in small amounts in soils. Under suitable (possibly acidic) environments, chlorites weather to vermiculite and smectite minerals (Kapoor, 1972). Chlorites in soils of the Indo-Gangetic Plains are rich in iron, as is evident from the ease of dissolution in HCl.

Hydroxy-interlayered Vermiculite and Hydroxy-interlayered Smectite — Hydroxy-interlayered vermiculite (HIV) and hydroxy-interlayered smectite (HIS) can be considered to form a solid-solution series, with the "pure end-numbers" consisting of smectite or vermiculite at one and pedogenic or aluminous chlorite at the other. Hydroxy-interlayered minerals are formed as Al^{3+} ions are released during weathering, which hydrolyse and polymerize to form large polycations with a postulated formula $[Al_6(OH)_{15}]^{3+}$ (or similar) in the interlayers of vermiculite and smectite. These polycations balance some of the charge of the 2:1 layer. The combination of a 2:1 layer with hydroxy-Al in the interlayer gives a structure similar to that of chlorite (Figure 8f). Thus, these minerals are called secondary chlorites. The degree of infilling of the interlayer with hydroxy-Al can vary from zero to almost complete. The interlayer hydroxy-Al is not exchangeable and, therefore, it lowers the CEC of smectite or vermiculite.

The interlayer hydroxy-Al impairs the shrinking and swelling characteristic of smectite (Kalbande *et al.*, 1992). In vermiculite, it reduces K-adsorption capacity by lowering the exchange capacity and by preventing the layers from collapsing around the K^+ ions. The positively charged hydroxy-Al interlayers may also provide active sites for adsorption of anions (Schulze, 1989).

Both, HIV and HIS are most common in Alfisols and Ultisols of India (Pal and Deshpande, 1987; Bhattacharyya *et al.*, 1997).

Mixed-layer or Interstratified Minerals — Phyllosilicates in soils do not always occur as discrete particles of mica, vermiculite, smectite, chlorite or kaolinite. One physically discrete particle may consist of a mixture of both mica and vermiculite layers. Such minerals are referred to as mixed layer or interstratified minerals.

Different types of interstratified minerals have been identified. Two component systems include mica-vermiculite, mica-smectite, mica-chlorite, chlorite-vermiculite, chlorite-smectite, chlorite-swelling chlorite and kaolinite-smectite. Three component mixed-layer systems can also occur. The sequence of layers can be regular or random. The random interstratification of layer silicates is common in Indian soils (Kapoor *et al.*, 1981; Tomar, 1985; Bhattacharyya *et al.*, 1993; Chandran *et al.*, 2005).

Partial loss of interlayer K^+ ions from micas or of interlayer hydroxides from chlorite is one way for the formation of interstratified minerals in soils. Other possibilities include (i) fixation of adsorbed K^+ ions by some vermiculite layers to yield mica-like layers, and (ii) formation of hydroxide interlayers to provide chlorite-like layers.

Palygorskite and Sepiolite — Structurally important element in these minerals is the amphibole double silica chain oriented with its long direction parallel to the c axis. The apexes of the tetrahedrons in successive chains point in opposite directions. Thus, they are structurally distinct from the typical 1:1 and 2:1 layer structures. The linked chains form a kind of double-ribbed sheet, with two rows of tetrahedral apexes at alternate intervals in the top and bottom of the sheets. The ribbed sheets are arranged so that the apexes of successive sheets point together, and the sheets are held together by aluminium and/or magnesium in octahedral co-ordination between the apex oxygens of successive sheets (Figure 8g).

Palygorskite and sepiolite are often found in soils of arid and semi-arid environments and have fibrous morphology in contrast to the platy morphology of most other clay minerals. Palygorskite has been identified in some desert soils of Rajasthan (Sarma and Sidhu,1982) and

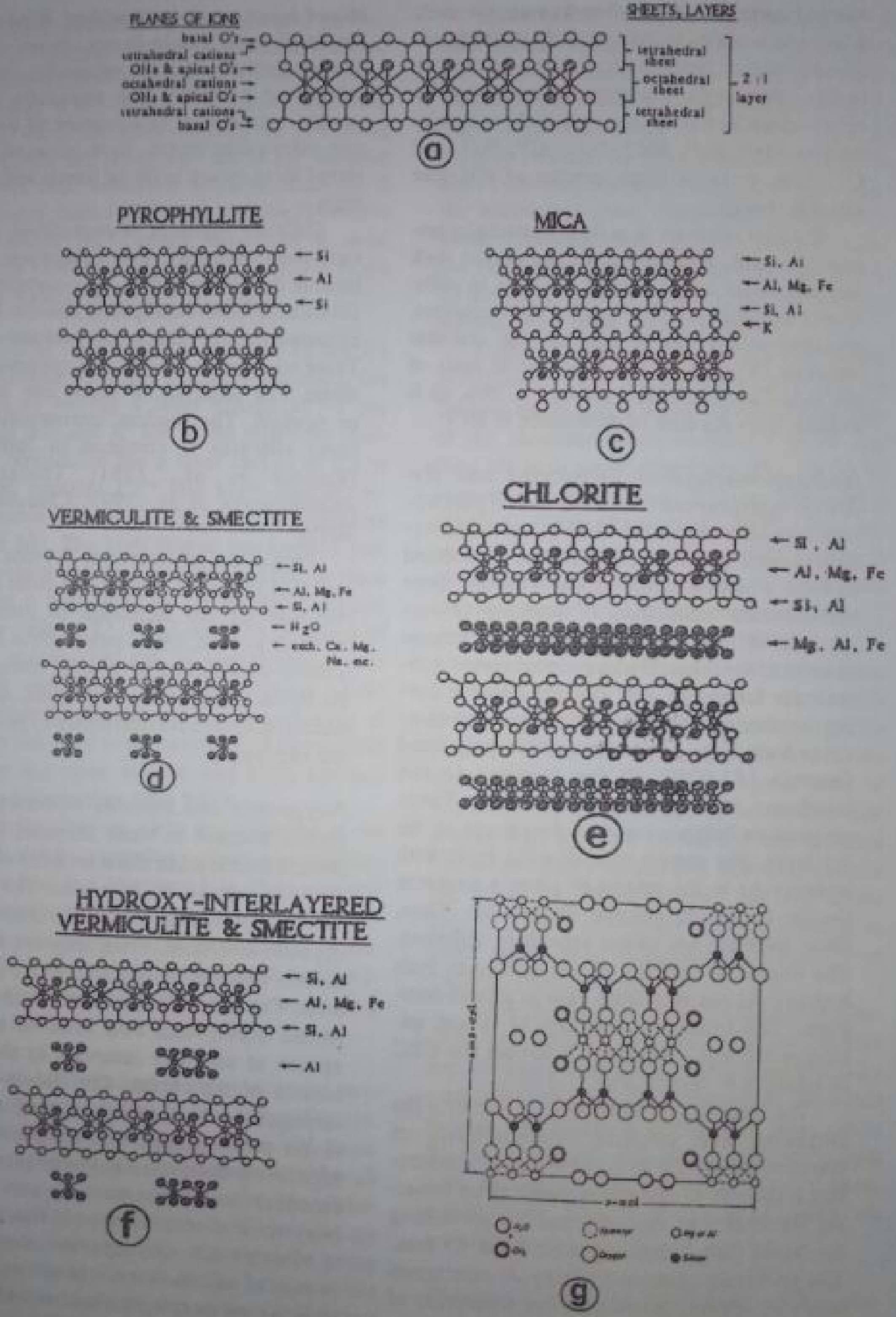


Figure 8. Structural scheme of soil minerals based on tetrahedral and octahedral sheets (a) 2:1 type minerals, (b) pyrophyllite, (c) mica, (d) vermiculite and smectite, (e) chlorite, (f) hydroxy-interlayered vermiculite and smectite, and (g) palygorskite

also in Vertisols of Gujarat (Pal *et al.*, 2003b). It releases Mg^{2+} ions when soils containing it are irrigated and high exchangeable Mg^{2+} ions cause dispersibility of clay particles, and in this way soils become poorly drained (Neaman and Singer, 2004).

Allophane and Imogolite — These two aluminosilicates exhibit short-range (or local) order, and are essentially poorly-crystalline or the structure is disordered. Structures with short-range order exhibit order over several nanometres, but on a larger scale, the structure is disordered.

Allophane is a mineral consisting chemically of variable amounts of O^{2-}, OH^-, Al^{3+}, and Si^{4+}, and is characterized by short-range order and a predominance of Si-O-Al bonds. It consists of small (dia: 3.5-5.0 nm) spheres, the structure of which has not been determined.

Imogolite consists of tubes several micrometres long with outer diameter of 2.3 nm to 2.7 nm and an inner diameter of ~1.0 nm. The tubes consist of a single dioctahedral sheet with the inner surface of OH^- replaced by O_3SiOH groups (Farmer *et al.*, 1983). The tubes are arranged in bundles 10-30 nm across and several micrometres long.

Soils containing large amounts of allophane and imogolite exhibit unique physical properties. These soils often have a low bulk density (0.25 - 0.3 Mg/m^3), high water-holding capacity (80-180% on oven dry basis), high liquid and plastic limits, and thixotropic tendency.

3.4. *Oxide Minerals in Soils*

In humid tropical climate, accumulation of soil colloids with compositions enriched in hydrous oxides of Al, Fe and Ti occurs because of de-silication. These inorganic colloidal substances range from amorphous to crystalline states in degree of organization.

3.4.1. Aluminium Oxides

A number of crystalline hydroxides, oxyhydroxides, and oxides of aluminium are found in nature. Gibbsite, one of the $Al(OH)_3$ polymorphs, is common in soils. It is one of the major minerals in Ultisols and Oxisols of tropical and subtropical regions. It is common in the silt and clay fractions of Ultisols of Meghalaya (Figure 7f). The gibbsite layer consists of a pair of hydroxyl sheets held together by octahedrally coordinated aluminium ions. The layers are held together by hydrogen bonding between hydroxyls of adjacent layers.

The oxyhydroxide of aluminium, boehmite (AlOOH), in which aluminium ions are octahedrally coordinated by oxygen and hydroxyl ions, occurs in the intensely leached, highly weathered soils, frequently in association with gibbsite. Geological materials that have undergone weathering for periods of several million years may contain appreciable amounts of boehmite along with large amounts of gibbsite.

3.4.2. Iron Oxides

Iron oxides, oxyhydroxides, and hydrated oxides are the most abundant among the metallic oxides in soils. These are present in most of the soils of different climatic regions as very fine particles in one or more of their mineral forms in varying concentrations. These different mineral phases have been referred to as "free iron oxides" previously in an attempt to differentiate between true oxides and the Fe bound in other minerals such as the silicates. This distinction is, however, unnecessary, because Fe in other minerals is not really present as an oxide. Therefore, it will be more meaningful to refer to these minerals collectively as "iron oxides" and subsequently distinguish between primary and secondary or pedogenic forms (Schwertmann and Taylor, 1989).

The most common iron oxides in soils are haematite (Fe_2O_3), which imparts pink to bright red colour to soils, and goethite (α-FeOOH or α-$Fe_2O_3.H_2O$), which gives brown and dark reddish brown colours to soils. Iron oxides provide an extremely important reflection of the chemical properties of soils and the genetic processes that have governed soil formation.

Haematite, purple in colour, is of coarsely crystalline form, and may occur in silt and sand fractions of soils (Rengasamy *et al.*, 1978). Martite (Fe_2O_3) has a structure similar to that of haematite. Dark iron oxide coatings are often associated with coatings of pyrolusite (MnO_2).

Goethite is abundant in concretions in certain Oxisols and Podzols. Finely divided yellowish goethite is sometimes designated as limonite which absorbs much water. Thin films of goethite and haematite occur on the surfaces of grains in soils to give reddish and brownish tints. Lepidocrocite (FeOOH), an isomer of goethite, has often been reported in poorly-drained soils rich in organic matter (Jackson, 1964), particularly as bog ores. It imparts bright orange colour to soils.

The mineral magnetite (Fe_3O_4) is a magnetic iron oxide of spinel structure, inherited from the parent rock. It occurs in soils usually as a sand-size mineral of specific gravity 5.17. Oxidation of iron in Fe_3O_4 yields maghemite (Fe_2O_3), which is also magnetic.

3.4.3. Titanium Oxides

Rutile (TiO_2) and anatase (TiO_2) commonly occur in soils and clay sediments. Rutile in soils is usually inherited from the parent rocks, while anatase may be pedogenic (Jackson, 1964). These minerals are commonly found in soils developed from the metamorphic rocks and also from basalt. Leucoxene ($TiO_2.nH_2O$) is an amorphous hydrous titanium oxide.

3.4.4. Ilmenite

Ilmenite ($TiFeO_3$ or $TiO_2.FeO$) is generally found in soils developed from the gneissic rocks. The substitution of Ti for Fe in magnetite gives the titanomagnetite ($TiFe_2O_4$) to titanomaghemite ($TiFe_2O_5$) series which occurs in soils of Hawaii (Jackson, 1964).

4. Elemental Composition of Soils

Just as in the earth's crust (Table 5), a large proportion of the inorganic material of soils is composed of four elements: oxygen, silicon, aluminium and iron. Nearly 90% of the mineral matter of most of the soils consists of the combined oxides of silicon, aluminum and iron. During chemical weathering and leaching, the proportions of these and other constituents get changed largely. This is comprehended in the composition of the fine fractions of the subsoil, especially in the illuvial Bt horizons (Table 6).

In general, the oxides of calcium, magnesium, sodium and potassium each make up about 1-2%, and together constitute about 5-7% of many highly-weathered soils. However, in soils of arid and sub-humid climates, the percentages of these oxides are higher. The oxides of manganese, phosphorus, and sulphur generally constitute only a fraction of a per cent of the mineral soil components.

Table 5. Average chemical composition of the earth's crust

Element	Percentage	Oxide	Percentage
O	46.5	SiO_2	59.07
Si	27.6	Al_2O_3	15.22
Al	8.1	Fe_2O_3	3.10
Fe	5.1	FeO	3.71
Ca	3.6	CaO	5.10
Mg	2.1	MgO	3.45
Na	2.8	Na_2O	3.71
K	2.6	K_2O	3.11
Ti	0.6	TiO_2	1.03
P	0.12	P_2O_5	0.30
S	0.06	MnO	0.11
Cl	0.05	H_2O	1.30
C	0.04		

Source: Jackson (1964)

4.1. Silicon

The silicon content (as SiO_2) makes up a major portion of the mineral soils if not highly weathered. Due to de-silication in soils of humid tropical climates, silica is highly depleted. Therefore, laterites contain very small amount of SiO_2 (Table 6).

4.2. Aluminium

Next to oxygen and silicon, aluminium is the most abundant element in the earth's crust and in the majority of rocks and soils. As Al_2O_3, it ranges from 20 to 60% in highly-weathered soils and laterites (Jackson, 1964). The higher percentage of Al_2O_3 is generally associated with high percentage of gibbsite, as in bauxite ores. As SiO_2 decreases and Al_2O_3 is enriched, the molar ratio of SiO_2/Al_2O_3 in soil clays decreases from about 4 in crystalline clays rich in layer silicate minerals to less than 1 in clays high in allophane (amorphous).

4.3. Iron

The iron content (as Fe_2O_3) makes up 3-15% of soils in general. However, it can ex-

Table 6. Chemical composition of some major soils of India

Horizon	Depth (cm)	Chemical composition, %								
		SiO_2	Al_2O_3	Fe_2O_3	TiO_2	CaO	MgO	MnO_2	Na_2O	K_2O
Alluvial soil of the Indo-Gangetic Plain in Sitapur district, Uttar Pradesh (Typic Haplustalf)[a]										
Ap	0-18	76.8	14.2	3.1	0.6	0.4	0.5	0.1	1.2	3.0
A/B	18-36	73.8	16.2	3.6	1.2	0.3	0.3	0.1	0.8	3.1
Bt1	36-88	70.1	19.7	3.9	1.5	0.3	0.3	0.1	0.7	3.1
Bt2	88-133	69.1	20.4	4.2	1.6	0.3	0.3	0.1	0.7	3.1
Bt3	133-175	67.4	20.6	4.9	1.4	0.3	0.7	0.1	0.8	3.2
B/C	175-200	68.9	19.3	4.3	1.3	0.5	0.8	0.1	1.0	3.6
Ferruginous soil on granite-gneiss in Bankura district, West Bengal (Typic Rhodustalf)[b]										
Ap	0-13	72.0	13.0	5.1	1.6	2.3	0.7	—	—	7.8
Bt1	13-32	63.0	27.2	6.9	1.7	2.0	0.7	—	—	7.1
Bt2	32-49	65.3	19.0	8.8	0.4	1.0	0.5	—	—	7.8
Bt3	49+	57.0	23.1	12.9	0.3	1.2	0.4	—	—	7.7
Laterite on serpentine rock in Kukusan mountains, S.E. Kalimantan, Indonesia (Oxisol)[c]										
A	0-100	7.6	10.9	65.1	0.3	—	0.3	0.3	—	—
B1	100-200	3.7	13.8	65.4	0.3	—	0.4	0.8	—	—
B2	200-500	2.5	12.1	68.7	0.2	—	0.9	0.8	—	—
B3	500-650	7.3	9.8	60.4	0.1	—	6.4	0.9	—	—
C1	650-700	19.0	8.4	50.0	0.1	—	7.6	0.7	—	—
C2	700-750	34.0	5.6	34.9	0.1	—	11.4.	0.5	—	—
R	750+	39.4	2.5	8.4	Tr	—	33.2'	0.1	—	—
Black soil of Deccan trap in Bijapur district, Karnataka (Calcic Pellustert)[d]										
Ap	0-20	44.2	14.8	12.9	1.2	7.6	4.6	—	0.8	1.1
A12	20-45	42.6	13.9	13.1	1.2	9.0	4.8	—	0.9	1.0
A13Ca	45-62	40.2	13.4	10.1	1.0	14.0	4.4	—	0.8	0.5
AC	62-95	42.1	13.3	11.6	1.2	11.8	5.3	—	0.7	0.7

Sources: [a]Srivastava *et al.* (1994); [b] Pal and Roy (1978); [c] Mohr *et al.* (1972); [d] Dasog and Hadimani (1980).

ceed 50% in laterites (Table 6). The higher content of iron in soils than in the earth's crust (Table 5), is due to soil development processes. Because of relatively higher iron oxide content, high degree of stability of soil aggregates and high porosity are usually associated with ferruginous soils.

The principal form of iron in soils is generally iron oxides, but iron is also an integral part of layer silicates of soil, like biotite, vermiculite and smectite. Smaller quantitites of iron occur in soils in pyroxenes and amphiboles.

4.4. Titanium

The titanium content (TiO_2) makes up not more than 2% in many soils of temperate, humid and arid climates. In soils, titanum occurs primarily as fine-grained crystals of free TiO_2 (anatase and rutile) and $FeTiO_3$ (ilmenite). Since these minerals are highly resistant to weathering, the titanium content of soils tends to increase as other elements are leached out. The TiO_2 content of soils other than laterites, is very low (< 2%, Table 6). In laterites, TiO_2 content ranges from less than 1% (Table 6) to 20% or more (Jackson, 1964).

4.5. Calcium

Except where calcium carbonate is present, the calcium content (CaO) is generally in the low range (<3%) in soils (Table 6). The CaO content of soils is lower than that of igneous and sedimentary rocks (Table 6), because of leaching of calcium from the upper horizons of well-drained soils, except under arid climates.

The calcium of igneous and metamorphic rocks occurs mainly in the plagioclase series of feldspars. Non-calcareous soils contain calcium in the form of plagioclase, augite, hornblende, and epidote. Because of the presence of these minerals, even the moderately-weathered ferruginous soils (Ustalfs) are dominated by exchangeable Ca^{2+} ions (Pal *et al.*, 1989). However, in black soils of semi-arid climates, the concentration of Mg^{2+} ions is higher than that of Ca^{2+} ions in the subsoil (Dasog and Hadimani, 1980; Balpande *et al.*, 1996).

Calcium carbonate occurs in many soils in lower horizons. However, $CaCO_3$ is found at increasingly lower depths as one proceeds from more humid to more arid regions. Calcareousness also increases in black soils of semi-arid climates from high rainfall (>900 mm) to low rainfall area (<900 mm) (Balpande *et al.*, 1996). Pedogenic $CaCO_3$ is mainly the mineral calcite, precipitated when the soil solution gets concentrated in the arid climatic conditions. This precipitation triggers the development of sodicity in the subsoils of black soils (Balpande *et al.*, 1996). Crystalline gypsum ($CaSO_4.2H_2O$) is also accumulated in soils of semi-arid and arid climates, as observed in black soils of southern India (Pal, 2003).

4.6. Magnesium

In soils with percolative moisture regime, magnesium is retained primarily in mafic minerals, such as biotite, augite, hornblende, and smectites. The magnesium content of soils (as MgO), is generally less than 1% (Table 6) in non-calcareous soils. Dolomite ($CaCO_3.MgCO_3$), which occurs in substantial quantities in some soils that are inherited from the parent rocks or are formed pedogenically (Jackson, 1964). In black soils of semi-arid parts of India, MgO content exceeds 2% (Table 6). As exchangeable cations, Mg^{2+} ions along with Ca^{2+} ions dominate the exchange complex of soils of non-percolative moisture regime, especially in the black soils of semi-arid climates.

4.7. Potassium

In the alluvial and ferruginous soils, the K_2O content is higher than 3%, whereas in black soils, it is lower (Table 6). The former two soils are rich in K-bearing minerals, whereas the latter are not.

The proportion of the total potassium of soils held in soluble and exchangeable forms is usually relatively small. The high potassium primary silicate minerals are: muscovite, biotite, orthoclase and microcline. The potassium availability of major soils of India is related to the K release from biotite mica contained in sand, silt and clay fractions (Pal *et al.*, 2001). Of the layer-silicates, only micas have a substantial potassium content. However, the K release from clay mica depends on the nature of mica, it being slow when the clay mica has more muscovitic character. Alluvial and ferruginous soils of the northwestern and southern parts of India are rich in K-bearing minerals, especially biotite mica and therefore, no crop response to chemical K-fertilizers is noticed even after long periods of cropping. Application of water-soluble forms of fertilizer, such KCl or K_2SO_4 is, however, necessary for intensive agriculture in soils of humid climates where biotite mica is fast depleted during weathering.

4.8. Sodium

The sodium-content (as Na_2O), in general, does not exceed 1% in most soils (Table 6). This is much less than the average of 3.7% in the earth's crust (Table 5). The lower Na_2O content in soils reflects weathering away of sodium-bearing minerals, primarily plagioclase feldspars. The high sodium content of ocean waters indicates the depletion of sodium through leaching from soil minerals. The release of Na^+ ions from plagioclase may be sufficient to cause the sodicity in soils of semi-arid climate (Pal *et al.*, 2003a).

Sodium in well-leached soils occurs in high-albite plagioclases and in small amounts in micas, pyroxenes, and amphiboles, primarily in the fine sand and silt fractions. Sodium occurs as NaCl, Na_2SO_4, $NaHCO_3$, and Na_2CO_3 in saline and saline-alkali soils.

4.9. Phosphorus

The P-content of most mineral soils is between 0.02% and 0.5%. About half of the soil phosphorus occurs in combination with organic matter of surface soils and the remainder resides in mineral or inorganic combination.

Phosphorus in inorganic form occurs mainly as calcium phosphate in alkaline and calcareous soils (Gupta and Abrol, 1990). The main calcium phosphate in soils is a variant of fluorapatite [$Ca_{10}(PO_4)_6F_2$]. With progressive weathering, soils develop acidity and phosphate becomes increasingly bonded to aluminium and iron. Calcium phosphate is insoluble at high soil pH conditions. However, on lowering the pH by amendment (through gypsum, for instance), it becomes available to the extent that application of phosphatic fertilizers may be avoided for the first few years of cropping. Iron and aluminium phosphates of acid soils become available only when pH of the soil is raised by amendments (lime, for example).

4.10. Molybdenum

The molybdenum-content of soils generally ranges from 1 to 10 ppm, but in certain soils it may be as high as 30 ppm or even more. Molybdenum occurs in soil mainly as MoO_4^{2-} ions and like phosphate, it undergoes fixation as basic iron and aluminum molybdates. With increase in soil pH by liming, the availability of soil molybdenum is enhanced.

4.11. Nitrogen

The amount of nitrogen in surface soils ranges from 0.02% to 0.25% in general, and is closely related to the amount of soil organic matter of which nitrogen makes up approximately 5%. Nitrates, nitrites, and exchangeable ammonium usually account for less than 1% of the total nitrogen content of mineral soils. When NH_4^+ ions substitute K^+ ions in micas of subsoil and rocks, these provide an additional amount of nitrogen. The mineral-humus complexes also account for much additional nitrogen of soils.

4.12. Sulphur

Sulphur is present in soils in both organic and inorganic forms. In well-leached soils much of the sulphur gets associated with organic matter. Soils of humid temperate region contain 50 to 500 ppm of water-soluble sulphate and 100 to 1500 ppm of total sulphate. Soils high in free iron oxides tend to contain sulphate ions substituting for hydroxyl ions in basic iron sulphate complexes. This form of sulphate, which is slowly extractable in alkali (NaOH) solutions, constitutes a reserve of slowly available sulphate.

Sulphur is added to soils through rain and snowfall. It is also added to soils through many commercial fertilizers and by irrigation waters used for crops and greenhouse culture.

4.13. Selenium

Like other trace elements, selenium is found in traces in most soils and rocks. Selenium-content from 2 to 10 ppm is considered very high in soils. High Se-content is likely to produce vegetation toxic to livestock. Certain native plants are "accumulators" of Se and their chemical analysis may be used as a guide in areas of excess selenium supply. During weathering, this element oxidizes to selenate which is subject to leaching, and combines with free iron oxides to form basic selenates.

4.14. Boron

The total boron-content of soils, in general, is in the range 4-98 ppm , compared to 10 ppm in igneous rocks and 4.5 ppm in sea water. The boron soluble in 85% phosphoric acid at 100 °C represents the amount that is present in organic matter and layer silicates. This acid extraction excludes boron present in the mineral tourmaline. Acid-soluble boron is associated with the mica of mature sediments and mica-derived clays of soils (Jackson, 1964).

Plant-available soil boron is well correlated with boiling water soluble boron, which frequently ranges from 0.2 ppm to 1.5 ppm in soils. Soils of semi-arid climates contain more boron (10-40 ppm) than soils of humid climates. Alkali soils carry excessive boron. Hence, leaching of undesirable boron to non-toxic levels before cropping should be an essential step in land management practices.

4.15. Zinc and Copper

The total content of zinc and copper of soils is generally lower (40-60 mg/kg for Zn and 20 mg/kg for Cu) than in the material from which the soils are formed (Hodgson, 1963). These cations are held in soils principally on organic or inorganic surfaces or substituted as accessory constituents in common soil minerals. They are strongly held by the soil surface. But, when these cations are present in bivalent states in highly insoluble oxides and phosphates, they become much less liable to leaching. This explains their low concentration during the formation of soils from rocks of the earth's crust. This also explains the lower concentration of Zn and Cu in soils of the Indo-Gangetic Plains.

These elements form a host of naturally occurring minerals, but only rarely these minerals represent an important source of zinc and copper from an agronomic point of view (Hodgson, 1963).

5. References

Baldar, N.A. and Whittig, L.D. (1968) Occurrence and synthesis of soil zeolites. *Soil Science Society of America Proceedings* **32**, 235-238.

Balpande, S.S., Deshpande, S.B. and Pal, D.K. (1996) Factors and processes of soil degradation in Vertisols of Purna Valley, Maharashtra, India. *Land Degradation and Development* **7**, 313-324.

Bhattacharyya, T., Pal, D.K. and Deshpande, S.B. (1993) Genesis and transformation of minerals in the formation of red (Alfisols) and black (Inceptisols and Vertisols) soils on Deccan basalt in the Western Ghats, India. *Journal of Soil Science* **44**, 159-171.

Bhattacharyya, T., Pal, D.K. and Deshpande, S.B. (1997) On kaolinitic and mixed mineralogy classes of shrink-swell soils. *Australian Journal of Soil Research* **35**, 1245-1252.

Buckman, H.O. and Brady, N.C. (1971) *The Nature and Properties of Soils*, 3rd Indian Reprint, Eurasia Publishing House (Pvt.) Ltd., New Delhi, 653 p.

Chandran, P., Ray, S.K., Bhattacharyya,T., Srivastava, P., Krishnan,P. and Pal, D.K. (2005) Lateritic soils of Kerala, India: Their mineralogy, genesis and taxonomy. *Australian Journal of Soil Research* **43**, 839-852.

Dasog, G.S. and Hadimani, A.S. (1980) Genesis and chemical properties of some Vertisols. *Journal of the Indian Society of Soil Science* **28**, 49-56.

Farmer, V.C., Adams, M.J., Fraser, A.R. and Palmieri, F. (1983) Synthetic imogolite: Properties, synthesis and possible applications. *Clay Minerals* **18**, 459-472.

Gupta, R.K. and Abrol, I.P. (1990) Salt-affected soils: Their reclamation and management for crop production. In: *Advances in Soil Science, Vol. 11*, (B.A. Stewart, Ed), Springer-Verlag, Berlin, pp 223-288.

Hodgson, J.F. (1963) Chemistry of the micronutrient elements in soils. *Advances in Agronomy* **15**, 119-159.

Jackson, M.L. (1964) Chemical composition of soils. In: *Chemistry of Soil*, 2nd edition (F.E. Bear, Ed), Oxford and IBH Publishing Company, Calcutta, pp 71-141.

Kapoor, B.S. (1972) Weathering of micaceous clays in some Norwegian Podzols. *Clay Minerals* **9**, 383-394.

Kapoor, B.S., Singh, H.B., Goswami, S.C., Abrol, I.P., Bhargava, G.P. and Pal, D.K. (1981) Weathering of micaceous minerals in some salt-affected soils. *Journal of the Indian Society of Soil Science* **29**, 486-492.

Ming, D.W. and Mumpton, F.A. (1989) Zeolites in soils. In: *Minerals in Soil Environments*, 2nd edition (J.B. Dixon and S.B. Weed, Eds), Soil Science Society of America, Madison, USA, pp 873-911.

Mohr, E.C.J., van Baren, F.A. and van Schuylenborgh, J. (1972) *Tropical Soils — A Comprehensive Study of their Genesis*. Mouton-Ichtiarbaru-van Hoeve, The Hague, 481p.

Neaman, A. and Singer, A. (2004) The effects of palygorskite on chemical and physico-chemical properties of soils: A review. *Geoderma* **123**, 297-303.

Pal, D.K. (2003) Significance of clays, clay and other minerals in the formation of and management of Indian soils. *Journal of the Indian Society of Soil Science* **51**, 338-364.

Pal, D.K. and Roy, B.B. (1978) Characteristics and genesis of some red and lateritic soils occurring in toposequence in eastern part of India. *Indian Agriculturist* **22**, 9-28.

Pal, D.K. and Deshpande, S.B. (1987) Characteristics and genesis of minerals in some benchmark Vertisols of India. *Pedologie (Ghent)* **37**, 259-275.

Chapter 12

Soil Colloids and Ion Exchange in Soil

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1. Soil Colloids: Introduction

A colloidal system is a heterogeneous mixture of at least two phases* in which one phase is dispersed in a state of fine subdivision (particle size 1µm to 1nm) in another continuous medium. This continuous medium is termed as the dispersion medium. The size limits of 1µm to 1nm are not very rigid and there may be variations on either side of the range. When the dispersed phase is a solid (e.g., soil colloidal particles) and the dispersion medium is water (e.g., soil water), the colloidal system is referred to as a 'sol'. The properties peculiar to a colloidal system, which distinguish them from suspensions (particle size > 1µm), and true uni-phase solutions (particle size < 1nm), are primarily those arising from the large interfacial area, which is charged, characterizing the colloidal system. In other words, most of the important properties of colloidal systems in soils may be attributed to the large specific surface (or interface) charge density.

The soil colloidal fraction is the site of important processes in soil, governing ion-exchange, nutrient availability and fixation, soil physical properties, especially soil structure, hydraulic conductivity, infiltration, and also soil management. Soil colloids can be broadly classified into two types, depending on the nature of the linkages present and the types of compounds formed. These are:

(i) Inorganic colloids, and
(ii) Organic colloids.

The inorganic colloids cover aluminosilicate layer-lattice clay minerals; hydrous oxides and hydroxides of Fe, Al, Si and Mn in clay-size dimensions; amorphous silicates; and clay-sized primary minerals. The organic colloidal fraction in soil is constituted by soil humus.

2. Inorganic Soil Colloids

The aluminosilicate layer-lattice clays, the chief inorganic colloidal fraction in soil, are derived from the silicate minerals. Silicates are the most common minerals in the earth's crust, accounting for more than 90% of the latter. Most of the soils are also dominated by these silicates which occur as primary minerals, inherited from igneous or metamorphic rocks, and as secondary minerals formed from the weathering products of primary minerals. These silicates consist of the SiO_4^{4-}- tetrahedron (*see* later) as the basic structural unit, present as either a single, or in several joint combinations. Depending on the arrangement of SiO_4^{4-}- tetrahedron in the structure, the silicate minerals

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* In Physical Chemistry, a *phase* is defined as any homogeneous and physically distinct part of a system which is separated from other such parts of system by definite bounding surfaces.

are classified into different groups, as already discussed in the Chapter on 'Chemical Composition of Soils'.

2.1. *Basics of Phyllosilicate Minerals*

In common with colloids of diverse origin, the layer-lattice aluminosilicate clays (phyllosilicates) are characterized by large specific surface area and charge density. Because the crystal structure of phyllosilicates largely governs their properties, it is necessary to understand these structures. These structures have been extensively studied by employing various techniques, particularly since the discovery in the early-1920s that most of the clays are crystalline in nature, and subsequent application of the x-ray diffraction technique to elucidate the above referred crystal structures.

2.1.1. *Structural Units in Phyllosilicates*

The convenient way of treating the atoms in the aluminosilicate structures as rigid spheres is oversimplified, but it provides a useful starting point.

Closest Packing of Spheres — The central concept of stability of crystal structure involves packing of anions around a central cation in such a way that each of these ionic spheres touches six of its nearest surrounding neighbours. Such an arrangement is known as the hexagonal closest packing, as shown in Figure 1. This forms a sheet leaving two kinds of central voids (A-type and B-type) between the spheres (Figure 1). With another sheet of such hexagonal packing, when stacked on the top of the first sheet, then a central cation placed in A-type of voids will have four surrounding spheres (such as anions) arranged in a tetrahedral manner around the cation, while a cation occupying a B-type of void will have six such surrounding spheres, above and below the central cation, pointing towards the apices of a regular octahedron (Figure 2). A cation occupying a tetrahedral site (the A-site) is said to be in a four-fold or tetrahedral coordination state (*see* later), being bonded to four oxide ions (O^{2-} ions), which form the apices of a regular tetrahedron (Figure 2). A cation occupying an octahedral site (B-site) is in six-fold (or octahedral) coordination state, being bonded

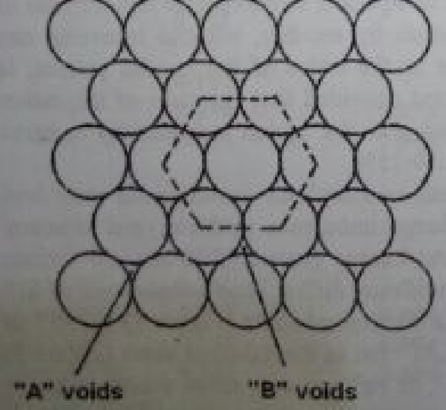


Figure 1. Hexagonal closest packing of spheres in a plane
Source: Schulze (1989)

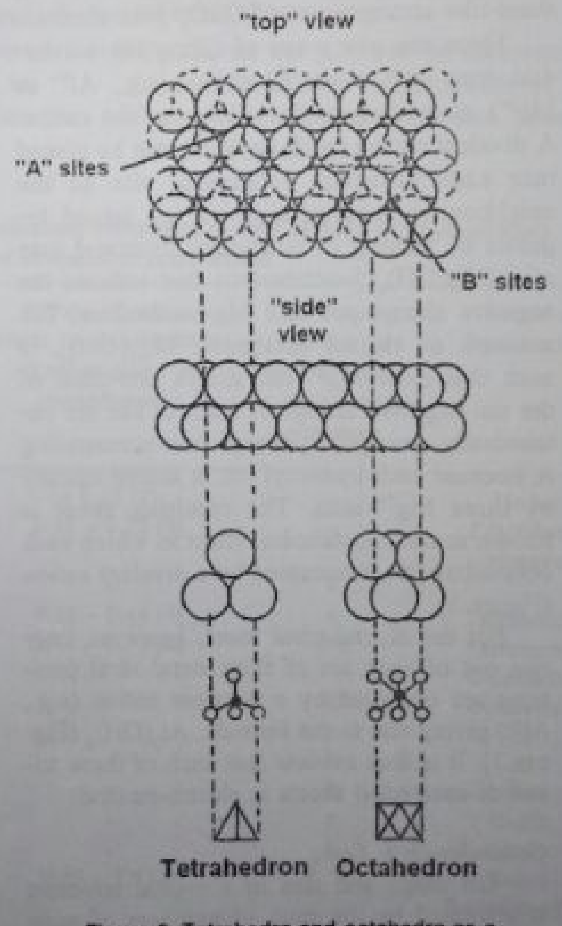


Figure 2. Tetrahedra and octahedra as a consequence of two planes of close-packed spheres
Source : Schulze (1989)

to six hydroxyl (OH^-) or oxide ions, arranged in the shape of an octahedron (Figure 2).

Silica Tetrahedral and Alumina Octahedral Sheets — A silica tetrahedron (SiO_4^{4-}) and an aluminium octahedron [$Al(OH)_6^{3-}$] are schematically depicted in Figure 2. A silica tetrahedral sheet is formed by sharing of three basal oxide ions of each (SiO_4^{4-})-tetrahedron with adjacent such tetrahedra in a and b directions (i.e., two dimensional) which will bring down the negative charge per unit tetrahedron (Figure 3). This leaves one unshared apical oxide ion in each tetrahedron, pointing upward from the basal plane accommodating the shared oxide ions. The apical oxide ion is free to form bond with other polyhedral atoms/ions. Indeed the term, 'Phyllosilicate' means "sheet silicates", which are thus minerals containing sheet-like arrangement of (SiO_4^{4-})-tetrahedra.

There are two ways of filling the octahedral sites with a central cation (e.g., Al^{3+} or Mg^{2+}), depending on the valency of the cation. A divalent cation (such as Mg^{2+}) can be placed into each adjacent octahedral site of the neighbouring octahedra which are joined together by sharing of octahedral hydroxyl ions of a [$Mg(OH)_6^{4-}$]-octahedron that reduces the negative charge per unit Mg-octahedron. The network of shared octahedra, $Mg_3(OH)_6$, is such that each Mg^{2+} ion shares one-third of the uni-negative charge of each of the six octahedrally arranged hydroxyl ions surrounding it, because each hydroxyl ion is shared equally by three Mg^{2+} ions. The resulting sheet is known as the trioctahedral sheet in which each octahedral site is occupied by a divalent cation (Figure 3).

For the dioctahedral sheet, however, only two out of each set of three octahedral positions are occupied by a trivalent cation (e.g., Al^{3+}) giving rise to the formula, $Al_2(OH)_6$ (Figure 3). It is thus evident that each of these tri- and di-octahedral sheets is electro-neutral.

Goldschmidt's Laws

The shape and size of a crystal structure is governed by the ratio of numbers of units (e.g., cations and anions), ratio of ionic sizes and properties of polarization of its structural units. Besides, allowance must be made for repulsive forces in ionic compounds, especially when the central cation is small, e.g., Li^+, and the surrounding anions have virtual contact with one another.

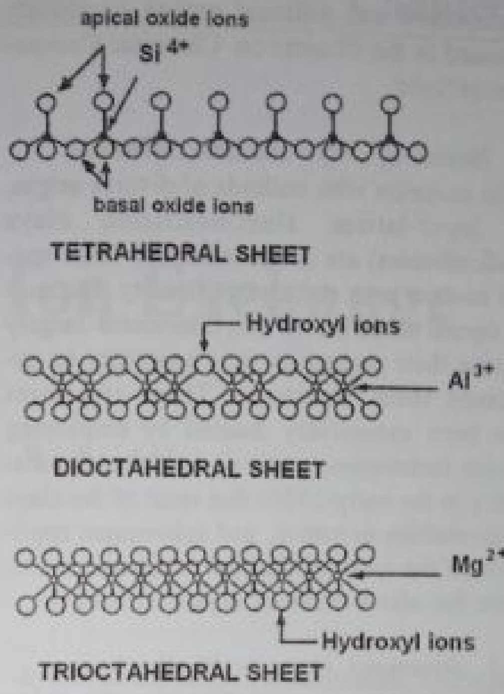


Figure 3. The tetrahedral, dioctahedral and trioctahedral sheets
Source: Schulze (1989)

The laws governing structural characteristics of the phyllosilicates are known as Goldschmidt's Laws – Law I and Law II.

(i) Goldschmidt's Law I — In an ionic crystalline compound, isomorphous replacement of one cation by another, without incurring any change in the order of the crystal pattern, is permitted provided that the radii of the cation replaced and the cation substituting it agree within 10-15%.

Such isomorphous substitutions may lead to a charge imbalance in the crystal structure since the charge (valency) of the two cations concerned may differ. Thus, substitution of Si^{4+} ion in a tetrahedral sheet (Figure 3) by Al^{3+} or that of Al^{3+} ion in dioctahedral sheet (Figure 3) by Mg^{2+} or Fe^{2+} ion will cause excess negative charge in sheet structures. This charge imbalance may be balanced by an appropriate and simultaneous substitution elsewhere, and/or by the retention of oppositely charged ions in the overall mineral structure (*see* later) so that the electroneutrality is maintained.

(ii) Goldschmidt's Law II: Radius Ratio Law — When the number of anions surround a central cation (with the closest packing of spheres) such that they satisfy the charge (of the cation) completely, the number of anions that can be so accommodated around the cation depends on the ratio of the radius of the central cation to that of the surrounding anions. Thus, this ratio of ionic radii governs what is known as the 'coordination number' of the central cation. The latter is thus basic to the shape of crystal forms. The ionic radii of a number of cations and anions pertaining to phyllosilicates, the cationic to anionic radius ratio ($r_+ : r_-$) [based on the radius of oxide ion, taken as the anionic radius (r_-) of relevance to phyllosilicates], the range of the values of these ratios ($r_+ : r_-$) corresponding to different coordination number of central cations, as well as the shape of the crystal structure or lattice are given in Table 1.

It can be shown by simple calculations that the tetrahedral site has a radius of 0.035 nm, while an octahedral site possesses a radius of 0.050 nm. By considering these sizes and data presented in Table 1, as well as by taking due cognizance of the Goldschmidt's laws, one arrives at the following important conclusions:

(i) A sphere of radius up to 0.41-times the radius of oxide ion can be accommodated in a tetrahedral site, while a sphere of radius up to 0.73-times the radius of oxide ion can be fitted into the octahedral site.

(ii) The Si^{4+} ion can remain in only four-fold coordination state in a tetrahedral sheet, while Al^{3+} ion can have both four- and six-fold coordination states in the tetrahedral and the octahedral sheets of phyllosilicates.

(iii) The cations such as Mg^{2+}, Fe^{3+} and Fe^{2+} ions can be accommodated only in the octahedral sites.

(iv) Potassium ion (K^+) can be accommodated in a dodecahedral arrangement of oxide ions, i.e. in between the two hexagonal arrays of oxide ions.

Table 1. Ionic radius, radius ratio, coordination number of central cations of phyllosilicate structures and shape of crystal lattice

Ion	Ionic radius (nm)	Radius ratio* (r^+/r^-)	Range of radius (ratio for coordination number given in parentheses)	Coordination number of central cation in phyllosilicate structure	Shape of crystal lattice
O^{2-}	0.140	-	< 0.15 (2)	-	Linear
F^-	0.133	-	0.15 – 0.22 (3)	-	Equilateral triangle
Cl^-	0.181	-	-	-	-
Si^{4+}	0.039	0.278	0.22 – 0.41 (4)	4	Tetrahedral
Al^{3+}	0.051	0.364	0.41 – 0.73 (6)	4, 6	Tetrahedral Octahedral
Fe^{3+}	0.064	0.457	-	6	Octahedral
Mg^{2+}	0.066	0.471	-	6	-do-
Fe^{2+}	0.074	0.529	-	6	-do-
Mn^{2+}	0.080	0.571	-	6	-do-
Na^+	0.097	0.693	-	8	Cubic
Ca^{2+}	0.099	0.707	-	8	-do-
K^+	0.133	0.950	0.73 – 1.0 (8)	8, 12	Cubic Dodecahedral
Ba^{2+}	0.134	0.957	> 1.0 (12)	8, 12	- do-
Rb^+	0.149	1.050	-	8,12	-do-

*Ratio of cation radius (r^+) to O^{2-} ion radius (r^-)

2.2. *Common Phyllosilicate Minerals in Soil*

Among the common layer-lattice minerals, the important ones are the 1:1 and 2:1 type minerals. The details of the chemical composition and structure of these phyllosilicates have already been discussed in the Chapter on 'Chemical Composition of Soils'.

3. Organic Soil Colloids

These refer to the humic colloids, the main fraction of the soil organic matter, which are widely distributed all over the earth's surface. Nearly 70–80% of the soil organic matter consists of humic substances, viz. humic acid, fulvic acid and humin. These are believed to be produced by continuous decomposition of plant, animal and microbial bodies, encompassing alteration of carbohydrates, proteins, fats, resins, wax, and so on, followed by synthesis of complex humic products. The humic colloids are dark in colour, partly aromatic and amorphous with very high specific surface area and charge density and are essentially hydrophilic in nature. These substances are characterized by flexible polyelectrolyte behaviour, having polyfunctional groups, and very high molecular weights, ranging from a few hundreds to several thousands. A typical value of CEC (*see* later) of soil humic colloids (which is pH-dependent) may be of the order of 200–250 $cmol(p^+)kg^{-1}$ or even higher. The typical functional groups of the organic colloids are carboxylic, phenolic hydroxyl, amino, amide, etc. and as a result, these act as proton donors or proton acceptors, depending on soil pH.

The humic colloids in soil often impregnate the soil mineral matter, e.g. the inorganic phyllosilicate clays (of soil) with which they remain in intimate association, forming clay-humic complexes.

4. Sources of Charges on Soil Colloids: Their Charge Characteristics

4.1. *Isomorphous Substitution*

Isomorphous substitution of cations by those of different charges (but having similar sizes within permissible limits) within the tetrahedral and the octahedral layer-lattice clay minerals leads to the development of net unbalanced charge within the lattice. Such a substitution takes place following the Goldschmidt's laws of crystal chemistry (*see* Section 2). Indeed an isomorphous substitution within Al-octahedral and Si-tetrahedral layers of higher-valent cations by lower-valent cations having similar sizes leads to permanent (pH-independent) negative charge. Examples are the substitution of Si^{4+} in tetrahedral layer by Al^{3+} in mica, vermiculite, illite, etc. and of Al^{3+} in the octahedral layer by Mg^{2+}/Fe^{2+} in montmorillonite. Positive charges on clay lattice may also arise through isomorphous substitution, e.g. by substitution of a lower-valent cation by a higher-valent one, such as occurs in the trioctahedral sheet forming substituted $[Mg_2AlO_2(OH)_2^+]$ sheet. The latter is sometimes found in the trioctahedral sheet in some vermiculites and chlorites. The net permanent charge on the clay lattice obviously depends on the extent of these two types of isomorphous substitutions, although it is nearly always negative.

The charge of soil clays arising out of isomorphous substitution remains constant.

4.2. *pH-dependent Charge*

There is yet another source of charge of soil colloids, especially for kaolinite, the aluminosilicate clay, and also for the hydrous oxides and hydroxides of Fe and Al in clay-size dimensions, allophones and non-crystalline clays, as well as the organic humic colloids of soil. The negative charge here arises through the dissociation of proton from the exposed hydroxyl group or bound water of constitution at the edge and surfaces of the inorganic colloids, and is naturally pH-dependent. The carboxylic and phenolic (OH) groups of soil humic substances also behave as weak acids, dissociation of which (thereby contributing towards soil acidity) would depend on the dissociation constant of the acid, and the pH of the surrounding medium. The carboxylic group ionizes at relatively lower pH (close to pH 5.5-6.0) than the phenolic (OH) group (which ionizes at pH 9.0), and contributes to the pool of pH-dependent negative charges. However, the substituted phenols, such as nitrophenols, are stronger acids than phenols (due to electron-withdrawing character of nitro group) and ionize, contributing to negative charge of humic colloids, at intermediate pH between 6.0 and 9.0.

No soil organic fraction with a net positive charge has been reported so far at normal soil pH values. However, organic species (R, *see* later), bonded to (NH_2) functional groups, may protonate at low pH to yield positively charged ($R\text{-}NH_3^+$) units, that hold anions, capable of anion exchange.

From the above discussion on pH-dependent charge of soil colloids, e.g., Fe and Al hydroxides, it is clear that the exposed edge/surface hydroxyl groups can act as both an acid (donating a proton to the surrounding OH group) or a base (accepting a proton from the surrounding soil solution), depending on pH of the soil system, i.e., relative preponderance of H^+/OH^- ions in the vicinity of exposed edge and surface OH group of soil colloid. Such amphoteric nature of these soil colloids warrants that these exposed hydroxyls would behave as an "uncharged" group (neither a proton donor nor a proton acceptor) at an intermediate pH. The pH at which the net surface charge of the soil colloids is zero is known as the 'zero-point of charge' or ZPC. The latter is thus a characteristic of soil colloids (such as in acid soils) which are mixtures of permanent and variable-charge minerals. The concept of ZPC is illustrated in Scheme 1 with hydrous oxides of Fe and Al (in clay-size dimensions).

The ZPC of a soil component depends upon its relative acidity with respect to water. Thus, ZPC of silicic acid is 2.0, of goethite (α - $Fe^{III}OOH$) is 8.5 and of gibbsite [$Al(OH)_3$] is more than 9.0. However, even at ZPC, the permanent charge of soil colloid (such as the aluminosilicate clay) persists unchanged.

On the other hand, the modification of surface/edge charge characteristics of soil colloids (especially the inorganic colloids) leads to the modification of ZPC. Thus, phosphate-fixation by ligand-exchange (e.g., exchange of neutral OH_2 ligand to Fe or Al by the applied $H_2PO_4^-$ ion; *see* later) process raises the surface negative charge, and pushes the ZPC towards lower pH, i.e. a stronger H^+ ion concentration is required in the surrounding soil solution to render the surface of soil colloid neutral. Furthermore, intense weathering of a soil shifts its ZPC towards higher pH owing to greater accumulation of hydrous oxides/hydroxides of Fe and Al and also enrichment of soil layer-lattice clays with kaolinite (possessing mostly pH-dependent edge charge). Indeed, the ZPC of Ultisol or Alfisol is generally higher than that of an Inceptisol. Organic matter, on the other hand, pushes the ZPC towards lower pH.

The development of charges on soil components is schematically shown in Scheme 2.

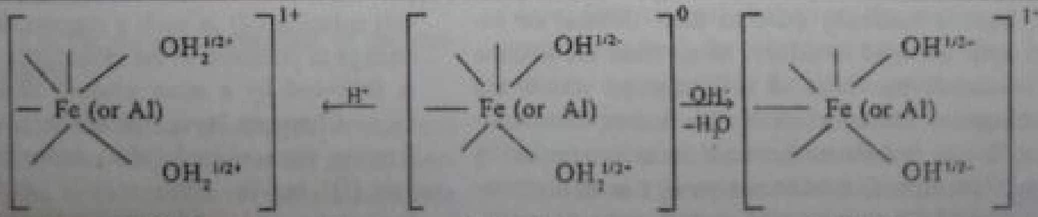
(i) On clay minerals (at edge/surface; pH-dependent negative charge)

(ii) On clay minerals (within lattice; permanent negative charge)

Mg^{2+}, Fe^{2+} replacing Al^{3+} in octahedral layer

Al^{3+} replacing Si^{4+} in tetrahedral layer

(iii) On hydrous oxides of Fe and Al (pH-dependent negative edge/ surface charge) Shown in Scheme 1.

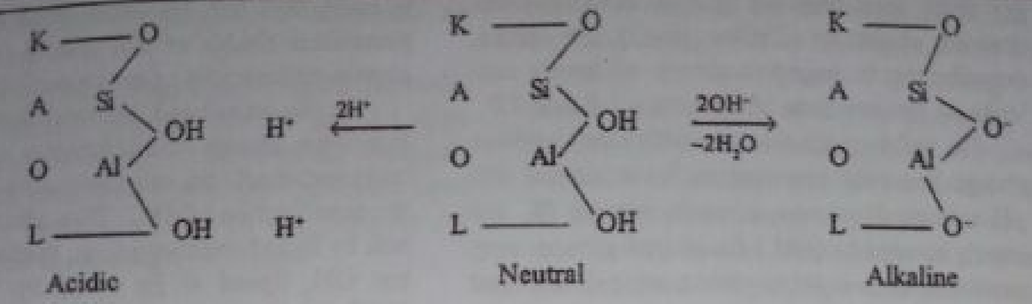


Acidic
(at pH less than ZPC,
contributes towards positive
charge and anion exchange)

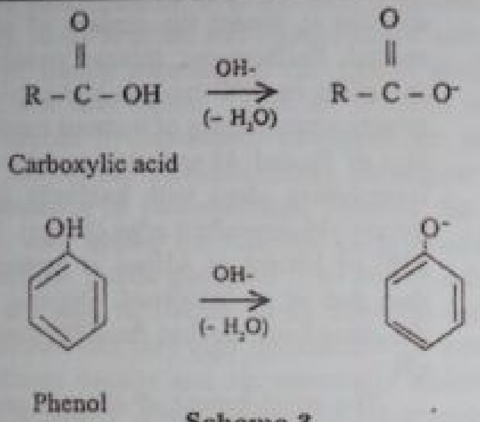
Neutral
(at the pH of ZPC)

Alkaline
(at pH greater than ZPC,
contributes towards negative
charge and cation exchange)

Scheme 1



Scheme 2



Scheme 3

(iv) On soil organic matter (pH-dependent negative charge)

Shown in Scheme 3.

4.3. *Electrical Double Layer of Soil Colloids*

The net charge on soil colloids is balanced by counter ions, such as interlayer cations of clay minerals, forming an electrical double layer. These counter ions, electrostatically attracted by the oppositely charged soil colloids, have a tendency (due to their thermal or kinetic energy) to diffuse away from the surface towards the bulk of soil solution where the counter ion concentration is lower. The equilibrium or balance between these two opposing (competitive) tendencies results in a distribution of the counter ions such that their concentration gradually decreases with increasing distance from the colloidal surface. In fact, it is postulated that the solution side of the aforesaid double layer consists of counter ions distributed into two broad regions, namely (i) a more or less fixed layer at a uni-molecular or uni-ionic (hydrated) distance from the colloidal particle surface (the so-called Helmholtz layer), trailed by (ii) a diffuse layer (the so-called Goüy-Chapman layer) of counter ions (in which some kinetic movement of the counter ions is possible), till the bulk solution is reached where the distribution of positive and negative ions (unaffected by the charge on the colloid which is far away) becomes uniform. Evidently, there is excess of cations and a deficit of anions in the diffuse double layer compared to their respective concentrations in the bulk solution. The above model or description of the electrical double layer at the colloid/solution interface is known as the 'Stern model'.

4.4. *Stability of Soil Colloids and their Coagulation/Flocculation*

The soil colloids are negatively charged, and as mentioned above, the counter ions (cations) are present in the fixed and the diffuse layer on the solution side of the electrical double layer at the soil colloid/soil solution interface. This causes a change of electrostatic potential from the solid surface (P) to the bulk soil solution (R) in such a manner that a sharp change in potential in the Helmholtz layer (PQ) is followed by a more gradual change in the Goüy-Chapman layer (QR). The potential difference, denoted by ξ (zeta), between the fixed part (Q) and the freely mobile portion (R) of the solution side of the double layer (Figure 4) governs the stability of the colloids. The higher the value of this potential (which is also known as the electrokinetic potential), the greater is the repulsion between two approaching soil col-

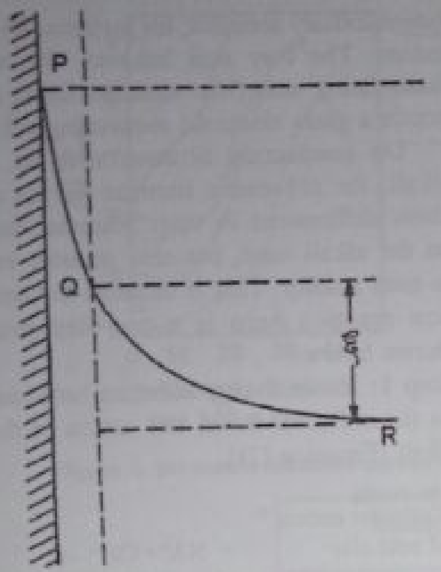


Figure 4. Electrokinetic (zeta) potential in an electrical double layer

loidal particles, swarmed by the respective counter ions of like charge in the neighbouring soil solution (e.g., the so-called 'clay micelles'), and hence less is the likelihood of these approaching particles to coalesce so that the larger will be the buoyancy force to keep these colloidal particles afloat and dispersed. This will lead to the stability of the soil colloidal sol.

By considering, as a first approximation, the electrical double layer at the soil colloid/soil solution interface to be equivalent to a parallel plate electrical condenser, with each plate carrying a charge of numerical value e per cm^2, the zeta potential (ξ) can be approximated by the potential difference between the plates, at a distance d apart [Equation (1)]:

$$\xi = \frac{4\pi ed}{D} \qquad ...(1)$$

where, D is the dielectric constant of the medium (water).

On addition of an electrolyte to the soil colloidal system, the cations of the added electrolyte tend to accumulate on the solution side of the electrical double layer, in the vicinity of the plane PQ (Figure 4). Upon gradual increase in the concentration of the added electrolyte, this cationic accumulation may reach such a proportion that there will be a virtual neutralization of some of the charge on the soil colloidal surface through a phenomenon which is termed as 'ion-pair' formation. It is facilitated by an increase in the cationic charge (i.e. valency) of the added electrolyte due to a higher bonding energy of the added cation for the given colloidal surface. This will bring down the value of 'e' [Equation (1)], and also of 'd', owing to release of some of the counter ions from the diffuse portion of the double layer to the bulk solution. The latter arises from the fall in the effective charge on the solid (colloid) surface, and hence its electrostatic attraction for the counter ions. In other words, the electrical double layer will be compressed with reduced charge density. Both these factors will bring down the ξ potential (Equation 1), enabling the approaching colloidal particles to coalesce with a concomitant reduction in buoyancy. The particles will thus settle under gravity, leading to coagulation or flocculation of the disperse phase (i.e. soil colloids) out of the dispersion medium, namely soil solution.

When the zeta potential attains the value of zero, the given colloidal system is said to correspond to the *isoelectric state* of the system concerned. It is apparent that for the aluminosilicate layer-lattice clay minerals (and soil, in general), possessing permanent (pH-independent) charge, the ξ potential is the combined function of the latter as well as the pH-dependent surface charge, while at the corresponding ZPC, only the net surface charge is zero, not necessarily the corresponding ξ potential. This underlines the difference between the ZPC and the *isoelectric point*. For the soil organic colloids and the hydrous oxides of iron and aluminium (possessing only pH-dependent charge), however, the two may be quite identical for obvious reasons. For instance, a polypeptide molecule (an amphoteric electrolyte), in its isoelectric state, is regarded as a neutral molecule even though it still has the ionized groups in a distribution of equal number of acidic and basic ionic groups.

The flocculation-dispersion behaviour of soil colloids has a strong bearing on the maintenance of favourable soil structure, thereby aiding soil aeration and drainage.

5. Electrometric Properties of Soil Colloids

5.1. Potentiometric and Conductometric Titrations

It is advantageous to study the electrometric (or electrochemical) properties of soil colloids such as soil clays when the latter are rendered homoionic with respect to a single cation that satisfies the entire cation-exchange capacity (*see* later) of the given clay. An acid clay, for instance, may be prepared by repeated leaching of the clay with a dilute mineral acid (e.g., 0.1 *N* HCl), followed by electrodialysis of the resulting acid clay to remove the excess electrolytes. A better way of obtaining the H-clay is to pass the clay suspension finally through a column of H^+ ion-exchange resin. The titration of such clay acids with alkalis provides important information regarding not only the exchange capacity, but also the nature and properties of the acid clay. Before proceeding further, let us refresh our knowledge of simple acid-base neutralization in aqueous solutions.

The gradual neutralization of an aqueous acid (e.g., acetic acid or HCl) by an alkali leads to a pH-metric (potentiometric) titration curve. An n-basic acid may show n-inflexions provided that the dissociation constants (K) of the $(n-1)^{th}$ and the n^{th} stages of acidity are widely separated. Typically, the ratio (K_{n-1}/K_n) should be of the order of 10^4. In actual practice, however, the dissociation constants of a polybasic acid (such as orthophosphoric acid, H_3PO_4) are often closer together, and neutralization at one stage may overlap partially with the preceding or succeeding stages, giving rise to statistically monobasic acid behaviour with an average dissociation constant (Mukherjee, 1974). The total acidity of such an acid solution is independent of the alkali chosen to measure the former.

If, however, the anion of an acid is of colloidal dimension (such as the negatively charged clay lattice of an acid clay), the protons form altogether a different phase from the former, giving rise to a two-phase acid system. The H^+ ions (i.e., the exchangeable cations) cannot be separated out into the intermicellary solution, for instance, by ultrafiltration. The clay acid behaves like a weakly dissociating acid (like aqueous acetic acid) towards a glass electrode measuring pH.

On conducting titration with an aqueous alkali, the pH-metric titration curves not only show differences in their features, depending on the alkali used, but also register variations in total acidity. This is because the neutralization reaction here is a two-step process as shown below.

Step 1: Ion-exchange reaction between H^+ ion on the colloidal anion and cation of the added alkali [Equation (2)]:

$$\boxed{\begin{matrix}\text{Colloidal antion}\\ \text{of acid clay}\end{matrix}}^{n-} + \underbrace{NA^+ + OH^-}_{\text{Alkali}} \rightarrow \boxed{\begin{matrix}\text{Colloidal anion}\\ \text{of acid clay}\end{matrix}}^{Na\,n-} + H^+ + OH^- \qquad ...(2)$$

Step 2: Neutralization of H^+ ion by OH^- ion in the solution phase [Equation (3)]:

$$H^+ + OH^- \rightarrow H_2O \qquad ...(3)$$

Some typical pH-metric titration curves of acid montmorillonite are shown in Figure 5.

The divalent Ca^{2+} ions, forming a stronger association with the colloidal anion of acid clay than the monovalent Na^+ ions, possess a greater H^+ ion replacing power (from the acid clay) than that of Na^+ ion. This leads to higher acidity (Figure 5) of the acid clay when titrated with $Ca(OH)_2$ than with NaOH. With $Ca(OH)_2$, the titrant, buffering (which amounts to flattening of the pH-metric titration curve; Figure 5) is observed at lower pH (Figure 5) which is the characteristic of a strong acid. With NaOH, such buffering is observed at relatively higher pH, a characteristic of weak acid.

Furthermore, the acidity of a colloidal acid increases (unlike a true acid) when titrated in the presence of neutral salts, so also the features of the titration curve change from those of a relatively weak acid to those of a stronger acid (Figure 5). This, again, results from an exchange of H^+ ion (also Al^{3+} ion, *see* later) for cations of the added salt preceding neutralization.

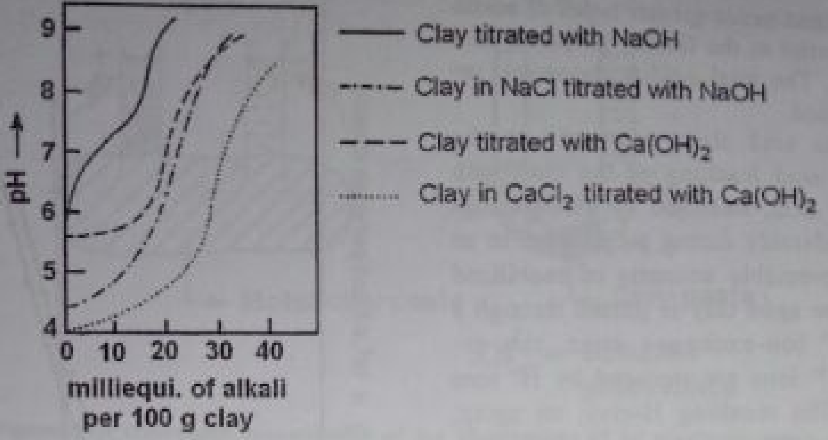


Figure 5. pH-metric titration curves of a clay separated from a Bengal soil showing the effect of alkali used and presence of salts
Source: Mitra and Kapoor (1969)

5.2. Exchangeable Aluminium Ions in Acid Soils

An acid soil is actually a mixed H-Al system, i.e. such a soil has both H^+ and Al^{3+} ions as exchangeable ions. As to the origin of exchangeable Al^{+3} ions, the H-soil or H-clay, when titrated immediately after preparation (on passage of the suspension through a column of H^+ ion-exchange resin), shows strong acid characteristics and a buffering at a low pH (Figure 6). The neutral salt extract contains little or no Al^{3+} ions. However, on aging (or on heating the suspension to, say, 95 °C), the titration curve changes markedly to those of weak acids, buffering occurring at higher pH with the total acidity remaining unchanged. The neutral salt extract, however, shows the presence of an increasing amount of Al^{3+} ions with the time of aging (or heating) (Figure 6). The number of inflexions in the titration curve also increases on aging. Such a conversion of H-soil or H-clay to mixed H-Al system can be arrested by keeping the suspension in a non-aqueous solvent such as acetone or methanol.

The exchangeable Al^{3+} ions in H-clay or H-soil arise from the lateral mobilization of the octahedral Al^{3+} ions in the clay lattice (such as acid montmorillonite) by the exchangeable H^+ ions with which they exchange the position. This is a slow process, but it gets accelerated on heating. Such ionic migration is arrested in a non-aqueous solvent of relatively low dielectric constant (e.g., acetone/methanol). The exchangeable Al^{3+} ions being weaker acids (Lewis acid) than H^+ ions, it is no wonder that freshly prepared acid clays, showing strong acid behaviour, change over to weaker acid type on aging (Figure 6), exhibiting a greater number

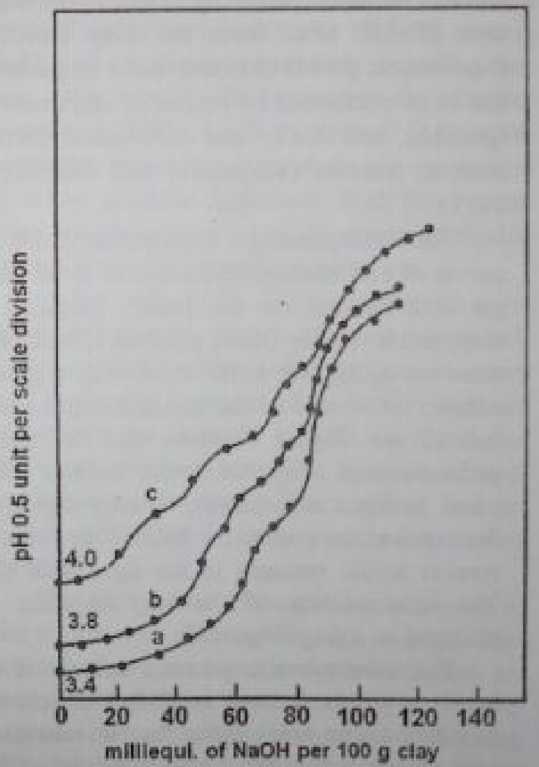


Figure 6. pH-metric titration curves of acid montmorillonite clay: (a) freshly prepared, (b) aged for one week, and (c) aged for nine weeks
Source: Mitra and Kapoor (1969)

of inflexions (and hence greater types of acidic groups) compared to the freshly prepared acid (i.e. H-) clay. The total acidity, however, remains unaffected.

Homoionic acid clays (prepared in the usual way of acid leaching of the clay with dilute mineral acid, followed by electrodialysis), age considerably during preparation so as to contain appreciable amounts of mobilized Al^{3+} ions. If the aged clay is passed through a column of H^+ ion-exchange resin, the exchangeable Al^{3+} ions are replaced by H^+ ions of the resin. The resulting H-clay, on aging, would again mobilize octahedral Al^{3+} ions. Thus, if the processes of aging and recovering the H-clay are repeated, the resulting H-clay would reach a stage when the clay would show signs of degradation owing to continued depletion of Al^{3+} ions from the clay structure (Mukherjee, 1974). Indeed, such degradation has been confirmed by means of electrometric titrations, and X-ray and differential thermal analysis studies (Majumdar and Mukherjee, 1979).

The corresponding conductometric titration curves of acid montmorillonite also show stronger acid features for the freshly prepared H-clay, and a weaker (more polybasic) acid character on aging, the latter exhibiting a greater number of breaks in the conductometric titration curves. Figure 7 shows that the freshly prepared acid clay has fewer breaks (compared to aged acid clays), thereby signifying the appearance of new acidic species of weaker acidic strength in the aged acid clay. The total acidity of the clay remains unchanged on aging (Figure 7).

The electrometric titration curves of soil humic colloids (humic acid, fulvic acid and hymatomelanic acid) show features characteristic of weak polybasic acid with the cation-exchange capacity being variable and a function of the alkali used for titration.

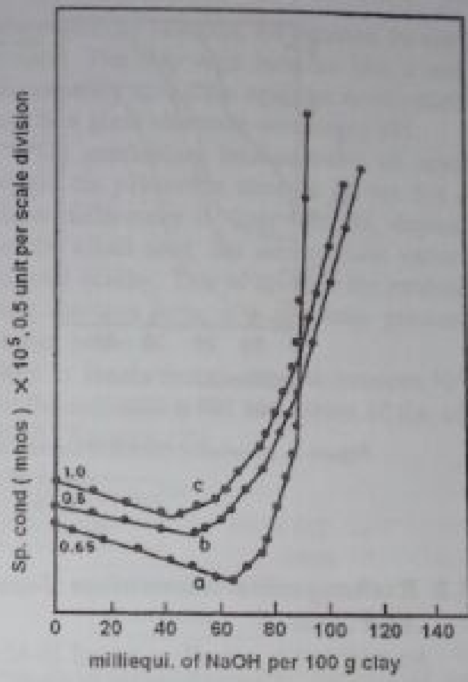


Figure 7. Conductometric titration curves of acid montmorillonite clay: (a) freshly prepared, (b) aged for one week, and (c) aged for four weeks
Source: Mitra and Kapoor (1969)

6. Electrokinetic Phenomena in Soil Colloids

It is well known that when an electric field is applied by sticking two water-filled glass tubes, encasing two platinum electrodes, into moist clay, and connecting these to an external source of electromotive force (E.M.F.) such as a battery, the liquid in the tube serving as a positive electrode (Figure 8a) becomes turbid while the water in the other tube remains clear. This indicates that clay particles move under the applied electric field towards the positive electrode. In another experiment, a clay plug is placed at the central part of a U-tube (Figure 8b) and is enclosed on both sides by sintered glass discs. On filling the U-tube with water, and applying an electric field through two electrodes inserted into water in the two arms of the U-tube, the water level in the arm having the negative electrode rises until the difference in water level between the two arms attains a constant value.

These two phenomena are known as electrophoresis and electro-osmosis, respectively, and are the direct consequences of the existence of electrical double layer at the clay/water interface, and slipping past of one layer of the said double layer over the other, the two layers possessing opposite (and equal) charges.

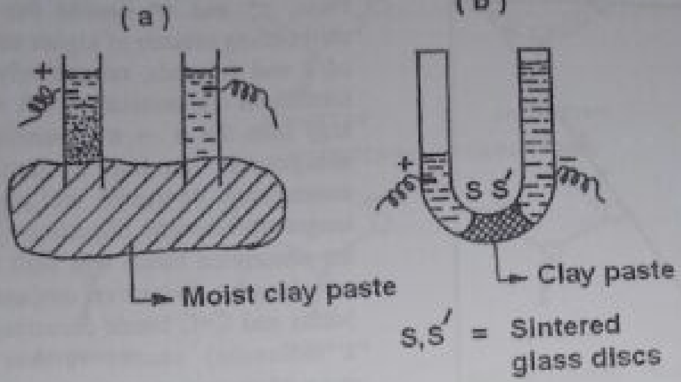


Figure 8. A schematic presentation of the phenomena of (a) electrophoresis of clay, and (b) electro-osmosis of clay

In electrophoresis, the negatively charged clay particles are free to move, and therefore migrate towards the positive electrode, while in electro-osmosis, the clay particles having been restricted from movement by the glass discs, the positively charged solution side of the double layer would move under an electric field towards the negative electrode.

A closer examination of electro-osmosis reveals that although the liquid flow is generally associated with the application of pressure difference (ΔP), the application of an electrical potential difference (i.e. an electric field) is doing this job in electro-osmosis. By analogy, one would then expect a pressure difference to induce an electric current which is generally the result of application of an electric field. Indeed, this is found to be so when the clay plug (Figure 8b) is subjected to a pressure difference. The ensuing current is termed as the streaming current, while the corresponding potential difference developed across the clay plug is known as the streaming potential. Here again, a mechanical separation of the solution side of the double layer (from the stationary clay particles) under the applied ΔP leads to development of streaming potential and the consequent current. It can be shown by Equation (4):

$$\left[\frac{v}{X}\right]_{\Delta P=0} = \left[\frac{i}{\Delta P}\right]_{X=0} \quad \ldots(4)$$

where, X is the applied electric field across the clay plug (in electro-osmosis; Figure 8b), v is the corresponding velocity of electro-osmotic flow of water, ΔP is the pressure difference applied under zero (externally applied) electric field, and i is the corresponding streaming current. The term (v/X) represents the electro-osmotic mobility, while the term ($i/\Delta P$) represents the streaming current density generated by a unit pressure difference. Both these quantities are characteristics of the electrical double layer at the colloid/ solution interface.

In the year 1878, Dorn discovered a related phenomenon (Shaw, 1970) in which colloidal particles under sedimentation in water lead to the development of an E.M.F. between two electrodes inserted at different depths of the liquid column. This effect, opposite of electrophoresis, is known as Dorn effect, and the E.M.F. generated is known as sedimentation potential.

The above four phenomena are known as electrokinetic phenomena for reasons explained above as regards the mechanism responsible for these processes. Figure 9 demonstrates the streaming current as a function of hydraulic pressure difference applied across kaolinite/water system.

7. Visible and Infrared Spectroscopic Studies of Humic Colloids

7.1. *Basic Principles*

The electronic transitions in humic molecules lead to characteristic spectral absorption bands in the ultraviolet (UV) and visible

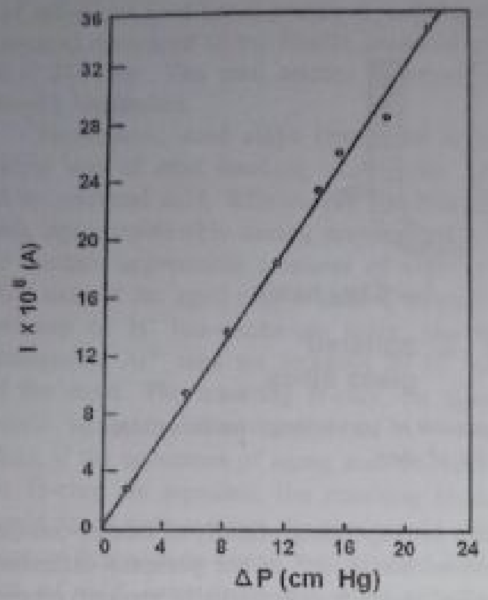


Figure 9. Relationship between streaming current (I) and hydraulic pressure difference (ΔP)

Source: Srivastava and Pal (1973)

region (together covering the wavelengths from about 200 nm to 800 nm) of the electromagnetic spectrum. The electrons involved are those in σ-bond (formed through face-to-face overlapping of atomic orbitals, as found in single bonds between C and H in CH_4), π-bond (formed through lateral partial overlapping of atomic orbitals such as found in olefinic bonds; the electrons are more mobile than those in σ-bonds), and the non-bonding (n) electrons (on oxygen, sulphur, nitrogen and halogens). The absorption of electromagnetic (e.g. light) energy (E) is quantized and is governed by the well known Planck's equation (Equation 5):

$$E = h\nu = h\frac{c}{\lambda} \qquad \text{...(5)}$$

where, ν is the frequency; λ is the wavelength of the absorbed radiation; c, the velocity of light, and h is the Planck's constant.

The absorption of light energy in visible (and UV) regions by humic colloids is accompanied with promotion of the aforesaid electrons from the ground state to a higher energy state. It is frequently described as $\sigma \rightarrow \sigma^*$, $n \rightarrow \sigma^*$, $n \rightarrow \pi^*$ and $\pi \rightarrow \pi^*$ transitions. Here, σ^* and π^* denote the corresponding antibonding orbitals of higher energy than those of σ and π bonds, respectively. The $\pi \rightarrow \pi^*$ transitions are associated with much lower energy than the $\sigma \rightarrow \sigma^*$ transitions. Therefore, compounds having a large degree of unsaturation (π-bonds) will absorb energy at longer wavelength [Equation (5)], and hence the absorption bands will shift towards longer wavelengths. Extensive conjugation (i.e. C-C bonds and C=C bonds occurring alternately in a molecule) causes further lowering of $\pi \rightarrow \pi^*$ transition energy, owing to extensive delocalization of π-electrons. It leads to a greater shift of the absorption bands to longer wavelengths. This is particularly true in aromatic compounds, compared to their aliphatic analogues having the same number of C-atoms. Indeed, as the number of fused (coplanar) benzene rings increases, as in, say, anthracene, the extensive conjugative delocalization of π-electrons takes place, pushing the absorption band far in the visible spectrum of electromagnetic radiation. Such an absorption spectrum is given as a plot of absorbance (A) or optical density at various wavelengths (λ) of incident light as function of λ. The absorbance, following the Lambert-Beer's law, is given by Equation (6):

$$A = \log\frac{I_0}{I} = abc \qquad \text{...(6)}$$

where, I_0 and I are the intensities of the incident and the transmitted radiations, respectively, at λ corresponding to maximum absorption by a given substance, 'a' is the molar extinction coefficient of the light-absorbing substance (e.g., humic colloid), 'c', its concentration and 'b' is the path length traversed by light through the given substance.

However, as compared to the absorption bands observed in the visible (and UV) regions (200-800 nm), the humic molecules (in common with organic molecules) show much larger number of peaks in the infrared (IR) spectrum (λ ranging from 2.5 to 15 μm, with the region 0.8 to 2.5 μm being known as near infrared and the one from 15 to 200 μm, known as far infrared region of electromagnetic spectrum).

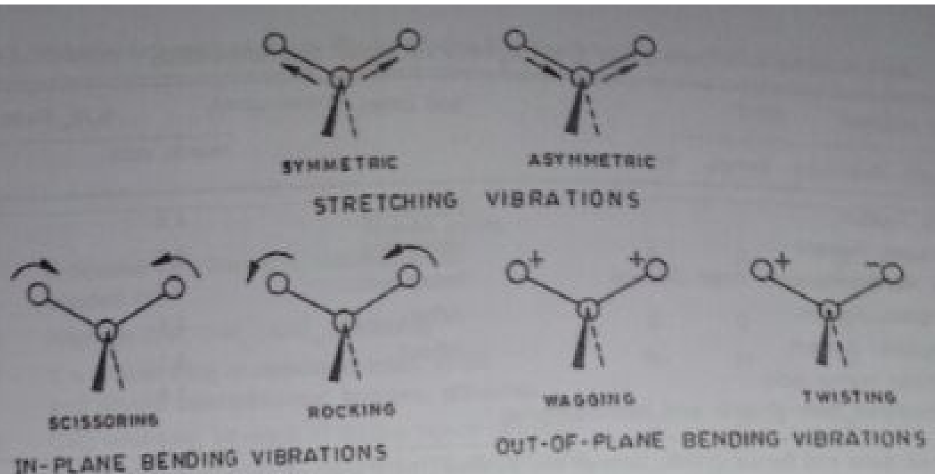


Figure 10. Vibrations of a group of atoms (+ and – signify vibrations perpendicular to the plane of the paper)

Source: Dyer (1971)

In the IR region, absorptions are often shown as a function of reciprocal wavelength or what is known as wave number. The wave number for instance for the near IR region, ranges from 12,500 to 4000 cm^{-1}. The energy (E) of absorption in the IR region is smaller than that in UV or visible region, the corresponding wavelength being longer (Equation 5), and is associated with the vibration of molecules, including stretching (in which the distance between two bonding atoms changes, but the atoms remain in the same position) and bending (or deformation in which the position of the atoms changes relative to the original bond axis). These modes of vibrations are illustrated in Figure 10.

As before, the energy of the various stretching and the bending vibrations of a bond is quantized, and the incidence of IR radiation of identical frequency causes absorption of energy by the organic molecule, leading to an increase in the amplitude of the given vibration. This, in turn, leads to an absorption peak in the IR spectrum. Perhaps the most significant application of the IR spectrum lies in the identification of the presence of a number of functional groups from the occurrence of the highly characteristic absorption peaks in the so-called "finger-print" region (7 to 11 μm).

7.2. *Absorption of Energy in the Visible Region*

The absorption of radiation energy in the visible region by soil humic and fulvic colloids often indicates the degree of aromatization and relative preponderance of aromatic and aliphatic nature of a colloid. Thus, the comparative absorption at 445 nm (denoted by E_4) and that at the longer wavelength of 665 nm (denoted by E_6) is taken to show the relative preponderance of aliphatic and aromatic groups, respectively, in the given humic molecule. The higher degree of aromatization of an organic molecule causes stronger absorption at longer λ for reasons explained earlier. More important parameter in this context is neither E_4 nor E_6, but rather the ratio, (E_4/E_6), which provides a better index of the above-mentioned preponderance. The fulvic colloids (the lower polymer and an earlier fraction than humic colloid of the soil organic matter, the humic colloid being of a higher polycondensate nature) possess typically higher E_4/E_6 ratio than their corresponding humic fraction. It is obviously linked with a lower extent of aromatization of the former, and the presence of a relatively larger proportion of aliphatic structures in the fulvic acids. Some representative values of the E_4/E_6 ratio for fulvic and humic colloids are given in Table 2.

7.3. *Absorption of Energy in the Infrared Region*

The functional group composition of humic colloids (e.g., humic acid and fulvic acid) can be inferred from their infrared (IR) spectrum. Thus, the presence of phenolic (OH), amide, methyl, free-NH-, carboxylic, carbonyl and

Soil location	Soil order	E_4/E_6 Ratio	
		Humic acid	Fulvic acid
Janji, Assam	Inceptisol	4.4	6.2
Kokrajan, Assam	Inceptisol	5.0	6.3
Raja Rammohanpur, West Bengal	Inceptisol	3.4	4.7
Golaghat, Assam	Alfisol	4.5	5.9
Deragaon, Assam	Alfisol	4.1	6.2
Synthetic humic acid (synthesized from glycine and hydroquinone)	–	2.9	–

Sources: Saha and Sanyal (1988); Sarmah and Bordoloi (1993); and Lahiri and Chakravarti (1995)

other important characteristics can be inferred from the finger-print region of the IR spectrum. In particular, the degree of aromatization can be ascertained from the presence of absorption bands/peaks in the region 1500-1600 cm^{-1} wave number. The different functional groups, present in the humic and fulvic acids of soil humic colloids, as revealed by a representative infrared spectral study of soil humic colloids, are summarized in Table 3.

Further, the IR spectrum of two synthetic humic acid samples (synthesized by alkaline oxidation of mixtures of the simple constituents of soil humic acids, namely the amino acids and phenolic compounds) is shown in Figure 11, in which per cent transmittance (bearing inverse relationship with absorption) is shown as a function of wave number.

To provide a comparison with soil humic colloids, the major IR bands of some representative clay minerals (the inorganic soil colloids) are listed in Table 4.

8. Coiling in Humic Substances: Hydrophobic Bond

The humic colloids exhibit hydrophobic and hydrophilic characters. While the various functional groups are generally hydrophilic, the (C-H) bonds are essentially hydrophobic. As a consequence, the humic colloids in aqueous medium (e.g., soil solution) remain in a coiled state. Such coiling is, in general, less marked in fulvic colloids than humic colloids; the latter possess a higher polycondensate nature and hence are more hydrophobic.

The interaction of the hydrophobic backbone of humic colloids with neighbouring water is (free) energetically (i.e. thermodynamically) unfavourable. As a consequence, the water molecules are drawn together, as by surface tension, and the non-polar groups tend to aggregate (much like oil droplets in water), leading to coiling, thereby minimizing their contact surface area with water, and hence the mutual interaction. The term, 'hydrophobic bond', has been coined to represent the association (aggregation as referred above) of the hydrophobic parts, excluding neighbouring water. But the term 'bond' may well be considered a misnomer here in that there exists no particular bond localized between two atoms of the hydrophobic residue, the association of which results under 'thermodynamic compulsion'.

9. Rheological Properties of Soil Clays

The science of flow of fluids, or in general, the science of deformation processes, developing with time, is known as rheology. The rheological or flow behaviour of clay suspensions is largely governed by the mode and the extent of interaction between the clay particles. These interactions may be of three types: edge-to-edge (EE), edge-to-face (EF), and face-to-face (FF) associations. This presumes that a well developed positively charged double layer exists at the edges of the clay particles, opposite in charge characteristics to that on the faces, which are predominantly negatively charged. As a result, there is an EF attraction causing "internal mutual flocculation", leading

Table 3. Relative intensity of major IR absorption bands of humic and fulvic acids of soils

Frequency (cm⁻¹)	Assignment	Soils			
		Janji	Jorhat	Kakrajan	Golaghat
	Humic acids				
3450-3300	H-bonded OH, Free OH, Intermolecular bonded OH	B	B	B	B
2950 – 2850	Aliphatic C-H, C-H_2, C-H_3 stretching	S	S	S	S
1725 – 1640	C = O stretching of carboxylic acids, cyclic and acyclic aldehydes and ketones, quinones	Sh	W	M	W
1640 – 1585	C=C stretching vibration of double bonds of cyclic and acyclic compounds, benzene ring substitution	S	S	S	S
1540	NO_2 vibration of nitro groups	W	W	W	W
1515	C=C stretching vibration of benezene, pyridine, etc., benezene ring substitution, secondary amines	A	M	W	W
1470 – 1420	Aliphatic C-H deformation	W	W	S	S
1025	Si – O – Si vibration of silicates	S	S	S	S
	Fulvic Acids				
3450-3300	H-bonded OH, Free OH, Intermolecular bonded OH	SB	SB	SB	SB
2950 – 2850	Aliphatic C-H, C-H_2, C-H_3 stretching	W	A	W	W
2850 – 2500	Carboxylate ion	Sh	Sh	A	A
1725 – 1640	C = O stretching of carboxylic acids, cyclic and acyclic aldehydes and ketones, quinones	M	M	M	M
1640 – 1585	C=C stretching vibration of double bonds of cyclic and acyclic compounds, benzene ring substitution	S	S	S	S
1515	C=C stretching vibration of benzene, pyridine, etc., benzene ring substitution, secondary amines	W	W	W	W

A= Absent, B = Broad, M = Medium, S = Strong, Sh = Shoulder, W = Weak

Source: Sarmah and Bordoloi (1993)

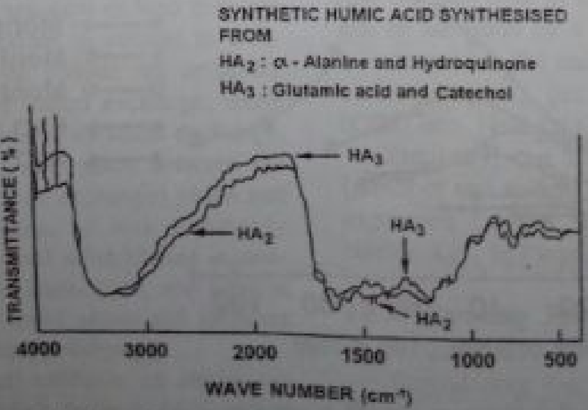


Figure 11. Infrared spectra of two samples of synthetic humic acids

Source : Saha and Sanyal (1988)

Table 4. The major IR absorption bands of some clay minerals

Major IR absorption bands (cm^{-1})		
Montmorillonite	Illite	Kaolinite
3620	3620	3700
3440	3460	3625
1650	2320	3490
1640 - 1630	1630	1430
1560	1460	1110
1505	1030 - 990	1025
1115	910	1000
1050	830	930
925	795	905
890	750	785
855	445	745
805	400	685
630	370	565
530		455
470		400
380		370
350		305

Source: Mondal *et al.* (1997)

to high viscosity (i.e., high resistance to flow). In fact, EF (and to an extent EE) association causes formation of card-house structure throughout the clay suspension. This imparts high viscosity to the system. On the other hand, FF association (through van der Waal's interaction, for instance) causes thickening of clay particles and an effective reduction of EF linkages. This leads to lowering of viscosity.

In the presence of electrolytes, when both the edge and face double layers are compressed, the nature of viscosity variations of clay suspensions is determined by two factors: (i) the balance between the residual EF attraction and FF repulsion, and (ii) the enhanced van der Waal's FF and EE attractions. The viscosity of montmorillonite (in sodium form, for instance) suspension exhibits a high degree of sensitivity towards the presence of extraneous electrolytes. The illite and kaolinite suspensions show progressively less dependence of their viscosity on the concentration of added electrolytes. This provides a relatively simple method (employing the viscosity determinations) for identifying the presence of montmorillonite as the dominating clay in a soil clay fraction. It has been observed experimentally that the viscosity-buffer curves of montmorillonite in particular, and of clay mixtures, in general, in which montmorillonite is dominant, show a distinct peak/hump. This hump gradually becomes less marked and shifts towards lower degree of per cent base saturation as the proportion of montmorillonite in the clay mixtures decreases. This is illustrated by the viscosity-buffer curves of clays and clay mixtures in Figure 12.

The distinct behaviour of montmorillonite, as against that of kaolinite or mica, may be attributed to the nature of its clay lattice to expand in the polar liquids. It causes easy accommodation of the hydrated cation (e.g., Na$^+$)

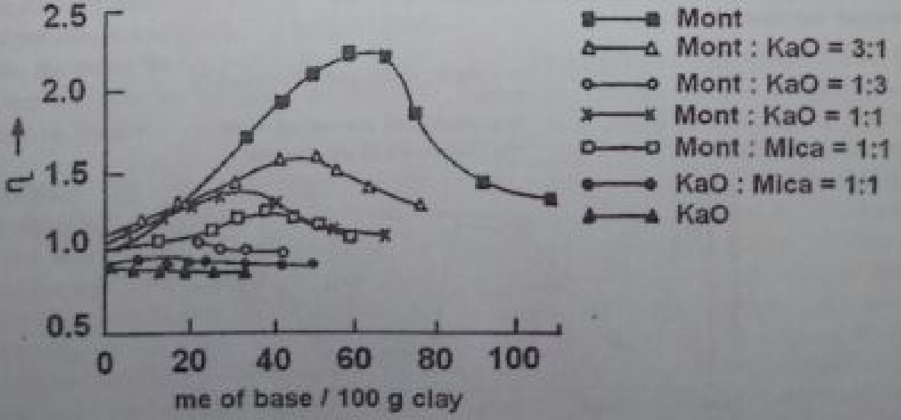


Figure 12. Viscosity buffer-curves of clays and of their mixtures
Source: Sahu and Das (1974)

of the added base largely in the expanding interlayer space up to about 75-80% base saturation, leading to a high value of the volume fraction (ϕ) of the dispersed clay particles (i.e., fraction of total volume occupied by the clay particles). This imparts high viscosity according to the Einstein's equation [Equation (7)].

$$\frac{\eta}{\eta_0} = 1 + 2.5\phi \qquad ...(7)$$

where, η and η_0 are the viscosity coefficients of clay suspension and pure water, respectively, at the same temperature. The numerical factor 2.5 has been assigned to spherical suspended particles, and for other shapes, different values have been assigned. Hence, the value of η increases, relative to the value of η_0, with concomitant increase in the value of ϕ.

In mixtures of montmorillonite with non-expanding clays, e.g., kaolinite, mica, etc., the volume-fraction at the same per cent base saturation will obviously be less, leading to a lower value of η.

Furthermore, this nature of montmorillonite to expand favours EF linkages, leading to high viscosity. However, at sufficiently high pH (e.g., beyond 75-80% base saturation in Figure 12), the positively charged double layer at the edges is significantly reduced so that the viscosity of the clay suspension falls due to snapping of EF linkages.

10. Ion Exchange in Soil

Ion exchange in soil system refers to exchange of equivalent amounts of ions between two phases in equilibrium in contact in a reversible process. When cations are involved, the process is termed cation exchange, while for anions, it is referred to as anion exchange. Such exchanges may take place between the soil solid (soil colloid or exchange) phase and the soil solution phase, or less commonly though, between the soil solid phases in contact (e.g. the Inverse Ratio Law; *see* later in Section 13), or soil solid phase and growing plant in contact (contact exchange).

The experimental evidence of cation exchange was provided around 1850 when Thomson, followed by J. Thomas Way leached soil with ammonium sulphate solution, and upon filtration, calcium, and to a lesser extent magnesium, potassium ions were detected in the filtrate, while less of ammonium ions were recovered compared to what was used for leaching. Initially, this phenomenon was described as base exchange. But on recognition of the fact that hydrogen ions can also be so exchanged, the wider term, namely cation exchange was used.

The composition of soil solution is known to change as a result of (1) dilution of soil solution (e.g., during rain or application of irrigation), (2) ion uptake by plants and leaching, (3) fertilizer application, (4) use of brackish groundwater for irrigation, (5) reclamation of salt-affected soils by leaching or use of chemical amendments, (6) sewage and sludge farming and use of industrial effluents for irrigation, etc. This leads to migration of, for instance, cations from soil colloidal phase to soil solution, and/or *vice versa* (cation release/fixation). The process involved is obviously ion exchange.

10.1. Cation Exchange Capacity

As stated above, interchange of a cation in soil solution phase with another on the surface of any surface-active material such as soil clay or organic matter in equivalent proportions in a reversible process is generally termed as cation exchange.

The cation exchange capacity (CEC) is defined as the capacity of the solid (e.g., soil) to adsorb and exchange cations. In other words, it is the sum total of exchangeable cations that can be adsorbed by a soil colloidal surface. It is expressed in the unit of cmol(p^+)kg^{-1} (in S.I. system) or meq. per 100 g in the classical (CGS) unit. Here (p^+) denotes proton.

Similarly, the anion exchange capacity (AEC) of soil colloid can also be defined. It is expressed (in S.I. system) in the units of cmol(e^-)kg^{-1}, (e^-) representing electron.

10.2. Bonding Energy of Exchangeable Cations

Considering the electrical double layer theory as applied to soil colloid system (see Section 11.2 and Section 4.4), the solid surface

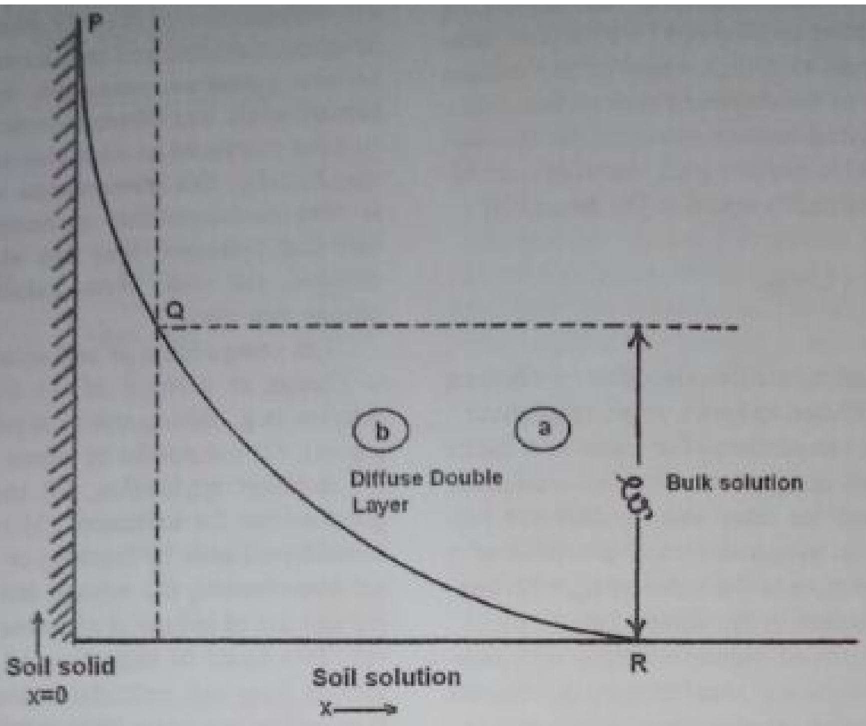


Figure 13. Electrical double layer at soil-soil solution interface

in soil, being predominantly negatively charged, the counter cations on the solution side of the double layer, especially those in the diffuse double layer (Goüy-Chapman layer; Region QR of Figure 4) are the exchangeable cations, amenable to exchange with the externally-added electrolytic cations (e.g., of the added fertilizer/amendments/irrigation water) (Figure 13).

The ionic strength (I) of a solution containing ions, which provides a measure of the intensity of the electric field due to such space change in the solution, is given by Equation (8):

$$I = \frac{1}{2}\sum_i m_i Z_i^2 \qquad \text{...(8)}$$

where, m_i is the molality and Z_i is the valence (including sign) of the ion 'i' in the solution phase, with the summation (Σ) being extended to cover all the ionic species in the solution.

The activity (a_i) of an ion in a solution is related to its molality (m_i) by Equation (9):

$$a_i = \gamma_i m_i \qquad \text{...(9)}$$

where γ_i is the activity coefficient of the ion 'i' in the solution phase while the activity, a_i, is its effective concentration. At infinite dilution, $\gamma_i \rightarrow 1$, and $a_i \rightarrow m_i$.

The activity coefficient (γ_i) of an ion in a solution is given by the Debye-Hückel Extended Law [Equation (10)]:

$$-\log \gamma_i = \frac{AZ_i^2\sqrt{I}}{1 + Bd\sqrt{I}} \qquad \text{...(10)}$$

where, A and B are constants at a given temperature for the solvent (e.g., water for aqueous soil solution), d is the diameter of the hydrated ion, and I is the ionic strength of the solution.

For a soil solution, which is generally dilute (except for the salt-affected soil), the Debye-Hückel Extended Law is approximated by the Debye-Hückel Limiting Law, namely

$$-\log \gamma_i = AZ_i^2\sqrt{I} \qquad \text{...(11)}$$

So that

$$-\log \gamma_i \propto \sqrt{I} \qquad \text{...(12)}$$

for a given ion at a given temperature in aqueous solution.

Considering the position of the same exchangeable cation at either 'a' or 'b' in the diffuse double layer (Figure 13), the bonding energy (B.E.) of the given ion for the present colloidal surface will obviously increase with the fall of distance (i.e. x) of the ion from the colloidal surface (at x = 0 in Figure 13), according to the Coulomb's Inverse Square Law. That is, the B.E. at 'a' will be less than that at 'b'. However, the ionic strength (I) will increase as one moves closer to the surface (i.e., with fall in x) with the concomitant increase in the ionic charge density (i.e. Eq. 8). In other words, both B.E. and I would increase with the fall in x. That is, the bonding energy (B.E.) of the counter ions in the diffuse double layer is directly related to the corresponding ionic strength (I), i.e.

$$\text{B.E.} \propto I \qquad \ldots(13)$$

But, $-\log \gamma_i \propto \sqrt{I}$ (From Eq. 12)

Hence, $(\text{B.E.})_i \propto -\log \gamma_i \qquad \ldots(14)$

Thus, the negative magnitude of the logarithm of the activity coefficient of a counter (exchangeable) ion provides a measure of the corresponding bonding energy of the counter ion for the soil solid surface.

The adsorbability of cations to a charged surface (e.g. soil colloid) has been often discussed in terms of the *Hofmeister series* or *Lyotropic series*. Thus, the monovalent cations can be put in the following sequence according to their adsorbability:

$Li^+ < Na^+ < K^+ < Rb^+ < Cs^+$

Bivalent cations follow the sequence given below:

$Mg^{2+} < Ca^{2+} < Sr^{2+} < Ba^{2+}$

For the cations of varying valencies, the *Lyotropic series* in respect of adsorbability is given as:

$Al^{3+} > Ca^{2+} > Mg^{2+} > K^+ = NH_4^+ > Na^+$, and so on.

Obviously, the guiding factor for such orders of cation binding by a charged surface may be related to the concept of bonding energy that has been derived above [e.g., Eqs (10) and (14)] in that the net bonding energy (or adsorbability) of the counter ions on a colloidal surface is the reflection of the relative weightage of the valency and the size of the hydrated cations concerned.

11. Theories of Ion Exchange

11.1. Adsorption Isotherms

In this sub-section we are primarily concerned with the adsorption of molecules or ions from the soil solution on the surfaces of soil solids. The material adsorbed is called the adsorbate, and the material on which adsorption occurs is called the adsorbent. Indeed, phosphate adsorption by soils is a process which is mainly responsible for rendering the soluble phosphates (e.g., applied phosphatic fertilizer such as single superphosphate or diammonium phosphate) in soil solution unavailable to plants. Desorption of once-sorbed plant nutrients from soils and clays is often slow and irreversible, leading to a large hysteresis effect. Adsorbed ammonium ions are also less susceptible to nitrification in soil. The amount of material adsorbed per unit mass of the adsorbent depends largely on the nature of the adsorbate and the adsorbent; it also depends on the concentration of the adsorbate at equilibrium and temperature. Mathematically, it is expressed by Equation (15):

$$\frac{x}{m} = f(\text{Nature of adsorbate and adsorbent, C, T}) \qquad \ldots(15)$$

where, x is the amount of adsorbate adsorbed on mass m of the adsorbent, C is the equilibrium concentration of the adsorbate and T is the temperature.

For a given mass and surface area of the adsorbent, the amount adsorbed increases with increase in the concentration of the adsorbate. When an adsorbent is placed in contact with a solution, the amount adsorbed gradually increases, and the concentration of the surrounding ions or molecules (adsorbate) decreases. This leads to the release of adsorbate from the adsorbent towards solution. When the rates of such desorption becomes equal to the rate of adsorption, an equilibrium is established. When the concentration of adsorbate in solution increases (or decreases), the adsorbed substance also increases (or decreases) till a new equilibrium is established.

Adsorption data at equilibrium are reported in the form of adsorption isotherms. An adsorption isotherm is a plot between the amount adsorbed per unit mass of the adsorbent and the equilibrium concentration of the adsorbent, at a constant temperature.

Different equations have been developed to predict precisely the shapes of the experimental adsorption isotherms between different adsorbates and adsorbents. The most commonly used ones are due to Langmuir (1918), Freundlich (1926), and Brunauer *et al.* (1938), which are described below.

11.1.1. Langmuir Adsorption Theory

The basic theory of adsorption of gases on solids is due to Langmuir (1918) and it is based on the following assumptions:

(1) The surface of solid is made up of adsorption sites, each of which can adsorb one gas molecule. If another molecule hits the adsorbed molecule, it is deflected back into the gas.

(2) All the adsorption sites are identical in their affinity for that gas molecule.

(3) The presence of a gas molecule on one site does not affect the properties of the neighbouring sites.

(4) The rate of adsorption is proportional to the number of adsorption sites and the pressure of gas. The rate of desorption is proportional to the number of occupied sites.

The system is considered to be at equilibrium when the rate of evaporation of the adsorbed gas is equal to the rate of its condensation.

The linear form of the Langmuir adsorption is given as in Equation (16):

$$\frac{C}{x} = \frac{C}{x_m} + \frac{1}{x_m K} \qquad \text{...(16)}$$

where, C is the concentration of the adsorbate at equilibrium, x is the amount adsorbed per unit mass of adsorbent at any instant and x_m is the maximum amount adsorbed per unit mass of adsorbent as a monolayer, and K is the equilibrium constant related to bonding energy. Experimental adsorption data, when plotted as C/x versus C, should yield a straight line with the slope ($1/x_m$) and the intercept ($1/x_mK$) on the ordinate (Y-axis). It may, however, be remembered that a linear plot between C/x and C is not adequate to prove that adsorption data follow the Langmuir equation. A linear plot of the data is a necessary but not a sufficient condition of obeying the Langmuir equation.

Khasawneh and Copeland (1973) introduced the concept of Supply Parameter (SP) involving the coefficients and constants from the Langmuir equation (Eq. 16). Such SP is taken to provide the buffering capacity of the soil for the adsorbed nutrient due to change of concentration of the nutrient concerned. The SP is given by Equation (16a):

$$SP = \sqrt{(qC/x_m K)} \qquad \text{...(16a)}$$

where, q is identical with x (specific adsorption of the nutrient concerned) of Equation (16), while x_m and K are the Langmuir constants as given by Equation (16).

The adsorption maximum (x_m) of soils for various ions and molecules, such as phosphate, borate, sulphate, K, herbicides, etc., can be estimated from the slope of the linear plot between C/x and C. From such value of x_m, the surface area (A) of soils and/ or clays can be estimated using the relationship (17):

$$A = \frac{x_m\ (g)\ N\ (\text{number of molecules mol}^{-1})\ \pi r^2}{M\ (g\ \text{mol}^{-1})} \qquad \text{...(17)}$$

where, r is the radius of the adsorbate molecule, N is the Avogadro's number and M is the molecular weight expressed in grams of the substance. If the adsorbate happens to be a gas, then x_m and M in Equation (17) should, respectively, be replaced by the volume of the gas required to form a monolayer on the surface (v_m, cm^3), and volume occupied by a mole (molecular weight expressed in grams) of the gas, i.e., 22.4 litre at 273 K, so that Equation (17) becomes Equation (18):

$$A = \frac{v_m\ (\text{cm})^3\ N\ (\text{number of molecules mol}^{-1})\ \pi r^2}{22400\ (\text{cm}^3\ \text{mol}^{-1})} \qquad \ldots(18)$$

Normally, the Langmuir equation is not used to calculate the surface area of solids because it does not define total adsorption process but covers only a certain fraction of adsorption sites.

It is important to emphasize here that soils are mixtures of inorganic and organic substances, consisting of montmorillonite, kaolinite, oxides of iron and aluminium, and decomposed plant material, etc. with each substance reacting towards competing cations differently. In a sense, soil exchangers are essentially polyfunctional ion exchangers. The condition of identical exchange sites (assumption 2) (identical bonding energy) and no change of it during adsorption (assumption 3) thus becomes too restrictive for application of Langmuir equation in a soil system. The experimental exchange isotherm obtained for a soil system is a weighted sum of isotherms for different types/classes of exchange sites rather than for identical exchange sites.

Use of Langmuir equation becomes further restrictive for cation exchange equilibria, as it does not take into account the competition between the cationic species. According to this equation, the amount of a cation on the adsorbed phase is governed only by the concentration of that cation, independent of the concentration of the competing cation. Also, the Langmuir adsorption isotherm refers to gaseous adsorption on non-ionic solid, while adsorption in soil is essentially an ionic process.

11.1.2. Freundlich Isotherm

The Freundlich equation (1926), originally an empirical equation, but subsequently derived on a rigorous mathematical model, implies that the bonding energy of the adsorbate on a given adsorbent decreases with fractional coverage of surface area of the adsorbent. This is closer to reality than the assumption of a constant bonding energy inherent in the Langmuir equation. This is one of the most widely used adsorption equations to describe the experimental data on adsorption of ionic or molecular species in soil. However, the Freundlich isotherm does not predict any adsorption maximum. Mathematically, it is expressed by Equation (19):

$$x = K C^n \qquad (n < 1) \qquad \ldots(19)$$

where, x is the amount adsorbed per unit mass of adsorbent and C is the equilibrium concentration of the adsorbate. K and n are empirical constants, sensitive to the given adsorbent-adsorbate system and temperature. The linear form of this equation is obtained by taking logarithm of its both sides, when we get expression (20):

$$\log x = \log K + n \log C \qquad \ldots(20)$$

A plot of the experimental data of log x *versus* log C should give a straight line with a slope of n and an intercept of log K.

11.1.3. Brunauer, Emmett and Teller (BET) Adsorption Isotherm

The Langmuir equation represents the experimental adsorption data reasonably well in cases where adsorption becomes independent of pressure after the formation of monolayer. For many gases and vapours, however, it fails at higher pressures because more than a monolayer of adsorbed molecules is formed. The theory of Brunauer, Emmett and Teller (BET) extends Langmuir derivation to obtain an equation for multilayer adsorption.

The BET equation is based on the following assumptions (Brunauer *et al.*, 1938):

(i) Heat of adsorption in the second, third, ..., nth layers is the same as the heat of adsorption in the monolayer, i.e. heat of liquefaction or condensation of the gas, and

(ii) If the adsorption takes place on a free surface, then at p° (saturation pressure of the gas), an infinite number of layers can be built up on the adsorbent.

The BET equation used to define multilayer adsorption at any equilibrium pressure, p, is given in its linear form as Equation (21):

$$\frac{p}{v(p^0 - p)} = \frac{(C' - 1)p}{v_m C' p^0} + \frac{1}{v_m C'} \qquad \ldots(21)$$

where, v is the volume of gas adsorbed per unit mass of adsorbent, v_m is the volume of gas required to form a monolayer on the entire adsorbent surface of unit mass, C' is a constant at a given temperature. According to this equation, a plot of the experimental data between $p/v(p^{\circ}-p)$ versus p/p° should give a straight line with a slope of $(C'-1)/v_mC'$ and an intercept of $1/v_mC'$ on the Y-axis. The values of slope and intercept thus yield the values of constant, C', and volume of gas adsorbed when the entire adsorbent surface is covered with a monolayer, v_m. The advantages of BET equation are:

(i) It describes the multilayer adsorption,

(ii) It yields the volume of gas required for the formation of a monolayer, and

(iii) It can be used to calculate the surface area of soils [e.g., by the use of Eq. (18)].

The BET equation is most useful between the relative pressures (p/p^{0}) of 0.05 and 0.45. The surface area can be calculated by multiplying v_m with the cross-sectional area of the adsorbate molecule. To determine surface area of soils and minerals, adsorbate molecules such as nitrogen, ethane, water, ammonia or other gases may be used. Weakly adsorbed molecules like nitrogen, however, may not penetrate into all the interlayer surfaces, resulting into an underestimate of total surface area. Use of strongly adsorbed molecules like ammonia, which penetrates into the internal surfaces also, leads to the measured surface area closer to the total surface area.

11.1.4. Hysteresis

It is generally observed that the ions which are adsorbed by a colloidal surface by ionic interactions, through non-specific retention mechanism, are desorbed during the desorption cycle by the expenditure of the same amount of energy as was released during the adsorption pathway. However, for the ligand exchange processes, such as those involved in the fixation of phosphate, sulphate, borate, etc., by soil colloids, and also the cation, namely K^+ (the latter being specifically sorbed in the interlayer space of illitic clay minerals), desorption generally requires much more energy in view of the fact that the retentive forces in such cases are primarily covalent in nature for the anions concerned, while ionic diffusion of K^+ occurs through a strongly negatively charged electric field in the illitic interlayer space. This leads to irreversibility of the adsorption-desorption processes involved, with the degree of such irreversibility depending on the nature of the soil colloids concerned (e.g., aluminosilicate clay minerals, hydrous oxide of Fe, Al, etc., in clay-size dimensions), and also the ions concerned. This effect is known as 'Hysteresis'. The hysteresis effect in K^+ ionic fixation-release processes depends not only on the amount of illitic minerals present in a given soil, but also on the state of weathering of such minerals. The more weathered the illitic minerals are (e.g., in Alfisols), the less will be the extent of hysteresis, as compared to that in the soils of more recent origin (e.g., Inceptisols, Entisols, etc.).

11.2. Electrical Diffuse Double Layer Theory

As stated earlier (Section 4.2), the layer-lattice silicate minerals have both the permanent and the pH-dependent charges. The oxides and humus components of such complexes, on the other hand, exhibit only the pH-dependent charge. In 2:1 layer-lattice silicate minerals, the pH-dependent charge is invariably less than the permanent negative charge. In majority of the cases, the soil colloidal complex at pH above 6 possesses a net negative charge. As a result, an electrical double layer is formed at the colloidal-solution interface which has already been explained earlier (Section 4.3).

The main assumptions of the theory of diffuse double layer are:

(i) The charged particle has flat surface without edge effect, i.e. the planar surface is effectively infinite in extent.

(ii) The exchange sites are uniformly distributed over the entire planar surface, i.e. the surface charge density is uniform.

(iii) The dielectric constant of water in the aqueous solution phase is constant.

(iv) The ions act as point charges.

(v) The interaction between ions is negligible and the charged surface and the ions interact only electrostatically.

(vi) The charge on the surface is neutralized by the excess of oppositely charged ions in the diffuse double layer, considering the Helmholtz's layer to be too small in extent as compared to the Goüy-Chapman layer.

For a mixture of monovalent and divalent symmetrical salt solutions, Goüy-Chapman's theory of the diffuse double layer has been used to obtain the relative amounts of the monovalent and divalent cations adsorbed on the clay surface. The fractional saturation of the exchange complex by the monovalent cation has been shown to be a function of surface charge density of the exchanger, cationic composition of the equilibrium solution and interaction between the diffuse double layers on the adjacent clay platelets. Such a relationship, based on the diffuse double layer theory, in its simplified form (Bolt, 1955), is given by Equation (22):

$$\frac{\Gamma^{+}}{\Gamma} = \frac{r}{T\sqrt{\beta}}\sinh^{-1}\frac{\Gamma\sqrt{\beta}}{r+4V_{d}\sqrt{C_{2}}} \qquad \ldots(22)$$

where,

Γ = The total surface charge density (mol/ m^2),

Γ^{+}/Γ = The fraction of the total surface charge neutralized by the monovalent cation,

C_1 = Equilibrium concentration of monovalent cation (mol/L)

C_2 = Equilibrium concentration of divalent cation (mol/L)

r = $C_1/\sqrt{C_2}$ ($\sqrt{\text{mol}}/\sqrt{\text{L}}$), is the reduced ratio of monovalent cation,

V_d = cosh Y_d ($Y_d = Ze\Psi_d/kT$), where Ψ_d is the electrical potential at the midway plane between two adjacent clay platelets

β = 1.08×10^{13} m mmol^{-1} at 298 K.

Because of the occurrence of surface charge density (Γ), reduced ratio of Na (r) and square root of the concentration of divalent cation (C_2) in the sine hyperbolic inverse function, the magnitude of their effect on the fractional saturation of the exchange complex by the monovalent cation is not easily discernible. The effect of soil and solution variables on the fractional saturation of the surface charge by the monovalent cation is thus illustrated by taking the following examples.

Example:

Given:

Equilibrium concentration of monovalent cation (C_1) = 0.1 mol/L

Equilibrium concentration of divalent cation (C_2) = 0.01 mol/L

Cation exchange capacity of soil (CEC) = 20 cmol(p$^+$)kg^{-1}

Specific surface area of soil (S) = 2×10^5 m^2/kg

$r = 0.1 / (0.01)^{1/2} = 1$ mol$^{1/2}$ L$^{-1/2}$

$\beta = 1.08 \times 10^{13}$ m mmol^{-1}

Γ = CEC/S = 0.1×10^{-5} mol m^{-2} = 10^{-6} mol m^{-2}

Taking the value of V_d as unity (i.e., non-interacting diffuse double layers) and substituting the values of r, Γ, C_2 and β in Equation (22), one gets Equation (23):

$$\frac{\Gamma^{+}}{\Gamma} = \frac{1}{10^{-6}(1.08\times10^{13})^{1/2}}\sinh^{-1}\frac{10^{-6}(1.08\times10^{13})^{1/2}}{1+4(0.01)^{1/2}} \qquad \ldots(23)$$

$$= 0.304 \sinh^{-1} 2.347$$
$$= 0.304 \times 1.581$$
$$= 0.484$$

which is nearly equivalent to 50% saturation of the exchange capacity by the monovalent cation.

The **thickness of the diffuse double layer**, defined as δ, may be calculated by Equation (24):

$$\delta = \left[\frac{\varepsilon kT}{2n_{0}Z^{2}e^{2}}\right]^{1/2} \qquad \ldots(24)$$

where,

ε = Static permittivity of the medium,

k = Boltzmann constant,

T = Absolute temperature,

e = Electron charge,

n_0 = Number of ions per m^3 in the equilibrium solution, and

Z = Numerical value of the valency of the counter ion (cation in case of negatively charged surface).

The static permittivity of aqueous medium, ε, is equal to ε_0 D, with ε_0 being the permittivity in vacuum (8.85×10^{-12} C^2 J^{-1} mol^{-1}) and D is the dielectric constant of the medium (78.5 at 298 K for water).

Hence, on substituting the values of ε (= $694.7 \times 10^{-12} \times C^2 J^{-1} mol^{-1}$), k (=$1.38 \times 10^{-23}$ J. K^{-1}), T (= 298 K), e (=1.602×10^{-19} C), and n_0 (= $C_0 \times 10^3 \times 6.02 \times 10^{23}$ ions m^{-3}) in Equation (22) at 25 °C (i.e. 298 K), one obtains Equation (25):

$$\delta = \frac{3.0\times10^{-10}}{Z}\left(\frac{1}{C_0}\right)^{1/2} m \qquad ...(25)$$

or

$$\delta = \frac{3}{Z(C_0)^{1/2}} A^0 \qquad ...(26)$$

where, C_0 is the concentration of the equilibrium solution (mol/L).

For a 1:1 electrolyte (i.e., Z =1), and C_0 = 0.1 mol L^{-1}, δ is about 1nm, while at C_0 = 0.001 mol L^{-1}, δ is about 10 nm.

11.2.1. Factors Affecting the Thickness of Diffuse Double Layer (δ)

It is clear from Equations (25) and (26) that δ depends on the total electrolyte concentration, valence of the counter ion, dielectric constant of the medium and temperature.

(i) Total Electrolyte Concentration — The higher is the total electrolyte concentration, the smaller is the thickness of the diffuse double layer. On increasing the electrolyte concentration four-times, the thickness of the diffuse double layer reduces to half.

(ii) Valence of the Counter Ion — The higher is the valence of counter ion, the smaller is the thickness of diffuse double layer. At a fixed electrolyte concentration, the thickness of the diffuse double layer for divalent ions would be half that for monovalent ions.

(iii) Dielectric Constant of the Medium — An increase in the dielectric constant of a medium increases the thickness of the diffuse double layer. An increase in temperature decreases the dielectric constant of medium, causing a fall in the thickness of the diffuse double layer due, obviously, to a higher kinetic energy of the counter ions. These inferences were drawn earlier in a qualitative manner (Section 4.4).

11.3. Donnan Membrane Equilibrium

If an aqueous solution of an electrolyte consisting of two diffusible ions (a cation and an anion), is separated by a membrane from another solution containing a salt with a non-diffusible ion, then at equilibrium (when the flow across the membrane stops at a given temperature), the distribution of the diffusible ions will be unequal on the two sides of the membrane, even though the ionic activity (or concentration in a dilute system) product of equivalent amounts of the oppositely charged ions or ratio of equivalent amounts of the like-charged ions on the two sides of the membrane will be equal. This is known as the 'Donnan Membrane Equilibrium', and the membrane of selective permeability is known as a 'Donnan membrane'.

However, if an aqueous solution of NaCl of concentration C_1 is separated from another such solution of concentration C_2 ($C_1 > C_2$) by an ordinary ("open") membrane of equal permeability to Na^+ and Cl^- ions, there will be a flow from the first solution [say, the left-hand side (L) solution with respect to the membrane (M)] to the second one [say, the right-hand side solution (R)] till the NaCl concentrations on both sides of M become equal and an equilibrium is reached. Under such circumstances, at equilibrium

$$(C_{Na^+})_L = (C_{Na^+})_R \qquad ...27(a)$$

and

$$(C_{Cl^-})_L = (C_{Cl^-})_R \qquad ...27(b)$$

Obviously then,

$$(C_{Na^+})_L (C_{Cl^-})_L = (C_{Na^+})_R (C_{Cl^-})_R \qquad ...27(c)$$

However, if the left-hand side solution (L) with respect to the membrane (M) has a solution (of, say, NaCl) with a non-diffusible ion through the given membrane, i.e., a Donnan membrane, e.g., Na^+R^-, where R^- is a large (non-diffusible) resin anion (Figure 14), then diffusion of Na^+ and Cl^- ions will take place from L to R across M (Figure 14) until an equilibrium (no flow) is attained.

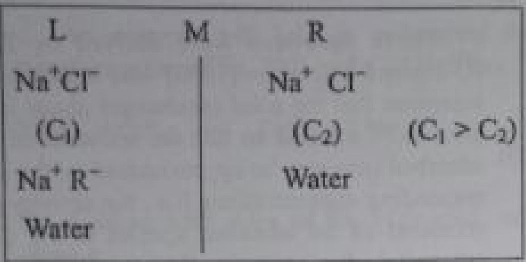


Figure 14. The Donnan membrane system

At equilibrium, for electrical neutrality on each side of the membrane (M):

$$(C_{Na^+})_R = (C_{Cl^-})_R \qquad \ldots(28)$$

And,

$$(C_{Na^+})_L = (C_{Cl^-})_L + (C_{R^-})_L \qquad \ldots(29)$$

where, the concentration terms refer to those at equilibrium.

According to the Donnan Membrane Equilibrium Theory, at equilibrium one obtains:

$$(C_{Na^+})_L\,(C_{Cl^-})_L = (C_{Na^+})_R\,(C_{Cl^-})_R \qquad \ldots(30)$$

Substitution of $(C_{Na^+})_R$ in Eq. (30) from Eq. (28) and $(C_{Na^+})_L$ from Eq. (29), one obtains from Eq. (30):

$$[(C_{Cl^-})_L + (C_{R^-})_L]\,(C_{Cl^-})_L = (C_{Cl^-})^2_R$$

or,

$$(C_{Cl^-})^2_L + (C_{Cl^-})_L\,(C_{R^-})_L = (C_{Cl^-})^2_R \qquad \ldots(31)$$

That is,

$$(C_{Cl^-})_L \neq (C_{Cl^-})_R \qquad \ldots(32)$$

And, in the same manner,

$$(C_{Na^+})_L \neq (C_{Na^+})_R \qquad \ldots(33)$$

Considering a soil colloidal system, the solid (charged) surface is swarmed by the oppositely charged counter ions on the solution side of the electrical double layer such that the counter ion concentration falls off as the distance from the solid surface increases in the diffuse double layer, while the concentration of the ions of the like charges (as on the soil solid or exchange surface) increases. Thus, there will be a region towards the terminal end of the solution side of the diffuse double layer (close to the bulk soil solution of even distribution of cations and anions) which will be dominated by the anions (for a soil colloid of net negative change). This region, intervening between the soil colloidal solid (i.e. exchange) phase and the soil solution phase, will act as if it were a Donnan membrane of selective permeability. Thus, considering the distribution of, say, K^+, Ca^{2+} and Cl^- ions between the soil colloidal (exchange) phase and the solution phase in such a Donnan system, one would have at equilibrium [from Eq. (30)]:

$$(C_{K^+})_{\text{exch}}\,(C_{Cl^-})_{\text{exch}} = (C_{K^+})_{\text{soln}}\,(C_{Cl^-})_{\text{soln}} \qquad \ldots(34)$$

and

$$(C_{Ca^{2+}})_{\text{exch}}\,(C_{Cl^-})^2_{\text{exch}} = (C_{Ca^{2+}})_{\text{soln}}\,(C_{Cl^-})^2_{\text{soln}} \qquad \ldots(35)$$

By combining Eqs (34) and (35), one obtains,

$$\frac{(C_{K^+})_{\text{exch}}}{(C_{K^+})_{\text{soln}}} = \frac{(C_{Cl^-})_{\text{soln}}}{(C_{Cl^-})_{\text{exch}}} = \sqrt{\frac{(C_{Ca^{2+}})_{\text{exch}}}{(C_{Ca^{2+}})_{\text{soln}}}}$$

That is,

$$\left.\frac{C_{K^+}}{\sqrt{C_{Ca^{2+}}}}\right._{\text{exch}} = \left.\frac{C_{K^+}}{\sqrt{C_{Ca^{2+}}}}\right._{\text{soln}} \qquad \ldots(36)$$

This is the essential conclusion, in quantitative terms, in regard to ionic distributions between the exchange and solution phases at equilibrium in soil, behaving like a Donnan system for reasons explained earlier.

In order that the above equilibrium between the exchange and solution phase is maintained, any dilution of the solution phase, for instance, would, therefore, favour release of K^+ ions (monovalent) from the exchange phase to the diluted soil solution phase, and *vice versa* for the divalent Ca^{2+} ions. An increase in the concentration of the soil solution will have a reverse effect on the equilibrium distribution of monovalent and divalent cations between the exchange (or adsorbed) and the soil solution phases.

11.4. Cation Exchange Equations Based on Law of Mass Action

The formulations expressing the relationships between the cations in the solution and adsorbed phases of soil at equilibrium are termed as cation exchange equations. While describing some empirical relationships, it is assumed that (a) all the exchange sites are freely accessible to solution cations such that cation exchange takes place instantaneously, (b) the surface charge (i.e. cation exchange capacity) remains constant over the entire cationic composition of the exchange complex, (c) the adsorbed phase is homogeneous (i.e. the surface charge is distributed uniformly), (d) the anions are effectively excluded from the monolayer of cations, and finally (e) the surface does not form any chelate with cations. The surfaces of different layer-lattice silicate minerals with the net negative charge satisfy the above-stated conditions only to various degrees.

Such cation exchange formulations may involve homovalent ion exchange or heterovalent ion exchange systems.

For homovalent cation exchange equilibria, one has

$$A^{Z+}_{soil} + B^{Z+}_{soln} \rightleftharpoons B^{Z+}_{soil} + A^{Z+}_{soln} \qquad \text{...(37)}$$

For instance, one may consider the Na^+-K^+ ion exchange system in soil. Thus,

$$Na^+_{soil} + K^+_{soln} \rightleftharpoons K^+_{soil} + Na^+_{soln}$$

However, here K^+_{soil} on the product side does not imply that the entire amount of exchangeable Na^+ ion from the homoionic soil exchanger (rendered homoionic with Na^+ ions) will be replaced by K^+ ion at equilibrium so that the exchanger will become K^+ ion-saturated at equilibrium. The exchanger at equilibrium will have both Na^+ ion (the residual Na^+ ion) and K^+ ion (the substituting K^+ ion) as the exchangeable cations, but their proportion will vary, depending primarily on the initial concentration of the bathing electrolyte (e.g., KCl in this case).

The Law of Mass Action leads to the equilibrium constant as:

$$K_m = \frac{[B^{Z+}_{soil}]\,(A^{Z+}_{soln})}{[A^{Z+}_{soil}]\,(B^{Z+}_{soln})} \qquad \text{...(38)}$$

where, [ ] denote the activity of the exchangeable ion concerned on the absorbed phase (e.g., soil), while ( ) denote that in the solution phase at equilibrium.

These equations were derived by Kerr (1928) and Vanselow (1932) who made an assumption that the solid (exchange) phase is an ideal solid solution so that the activities of the adsorbed ions may be approximated by the corresponding concentrations [i.e., the activity coefficients of the adsorbed species in Eq. (38) are unity], for example, the respective mole fractions on the exchange phase at equilibrium.

The activity coefficients of the ions in the soil solution phase may be evaluated from the ionic strength of the solution phase, following the Debye-Hückel formulations [Eqs (10) and (11)], as explained in Section 10.2. For this reason, the Kerr-Vanselow formulations [Eq. (38)] do not lend themselves to the evaluation of the thermodynamic equilibrium constants, and K_m in Eq. (38) is termed as the "Selectivity Constant or Quotient". These formulations are mostly applicable to the homovalent ion exchange systems.

Gapon (1933) also considered the hypothesis of ideal solid solution as above, but dealt with the heterovalent systems which are of more practical value than the homovalent ion exchange systems. Gapon preferred to use the ionic concentrations on the adsorbed (exchange) phase in equivalent basis for taking care of the corresponding ionic activities (in practice, the ionic concentrations for reasons stated above).

For a heterovalent ion exchange equilibrium, namely

$$Z_B\,A^{Z+}_{A\,soil} + Z_A\,B^{Z+}_{B\,soln} \rightleftharpoons Z_A\,B^{Z+}_{B\,soil} + Z_B\,A^{Z+}_{A\,soln}$$
$$(Z_A^+ \neq Z_B^+) \qquad \text{...(39)}$$

Accordingly, the Kerr-Vanselow formulation (Eq. 38) should be modified to Equation (40):

$$K_m = \frac{[B^{Z+}_{B\,soil}]^{Z_A}\,(A^{Z+}_{A\,soln})^{Z_B}}{[A^{Z+}_{A\,soil}]^{Z_B}\,(B^{Z+}_{B\,soln})^{Z_A}} \qquad \text{...(40)}$$

where, square brackets, [ ] and circular brackets, ( ), have the same significance as before. According to Gapon (1933), Eq. (40) reads as:

$$K_m\ (\text{or } K_G) = \frac{[CEC - A^{Z+}_{A\,soil}]\,(A^{Z+}_{A\,soln})^{1/Z_A}}{[A^{Z+}_{A\,soil}]\,(B^{Z+}_{B\,soln})^{1/Z_B}} \qquad \text{...(41)}$$

An example will help to understand the transformation of Eq. (40) to Eq. (41). Thus,

$$2\,Na^+_{soil} + Ca^{2+}_{soln} \rightleftharpoons Ca^{2+}_{soil} + 2\,Na^+_{soln} \quad \text{...(42)}$$

By dividing both sides of Eq. (42) by 2, one obtains:

$$Na^+_{soil} + \tfrac{1}{2}\,Ca^{2+}_{soln} \rightleftharpoons \tfrac{1}{2}\,Ca^{2+}_{soil} + Na^+_{soln} \quad \text{...(43)}$$

In case one starts with a homoionic soil in respect of Na^+ ion as in Eq. (42), then at equilibrium, the soil exchange phase will have only Na^+ and Ca^{2+} ions as exchangeable ions. If the CEC is the cation exchange capacity of the given soil, while the degree of saturation of soil (exchange) phase at equilibrium with respect to Na^+ ion is given as Na^+_{soil}, then obviously, the Ca^{2+} ion saturation of the CEC at equilibrium will be given as:

$$Ca^{2+}_{soil} = CEC - Na^+_{soil}$$

provided that the CEC and the degrees of saturation with respect to the exchangeable cations at equilibrium are expressed on equivalent basis. Substitution of these facts in Eq. (41) in respect of the exchange equilibrium, as given in Eqs (42) and (43), would obviously lead to the form of K_m [Eq. (44)], which is identical with Eq. (41). Thus,

$$K_m\ (\text{or}\ K_G) = \frac{[Ca^{2+}_{soil}]^{1/2}\,(Na^+_{soln})}{[Na^+_{soil}]\,(Ca^{2+}_{soln})^{1/2}}$$

or,

$$K_m\ (\text{or}\ K_G) = \frac{[CEC - Na^+_{soil}]\,(Na^+_{soln})}{[Na^+_{soil}]\,(Ca^{2+}_{soln})^{1/2}} \quad \text{...(44)}$$

A large number of experimental data for homovalent and heterovalent cation exchange equilibria studies have been fitted to the Kerr-Vanselow and the Gapon equations to obtain the corresponding selectivity coefficients (K_m). These values of K_m for, say, a given homovalent ion exchange system (e.g., Na^+-K^+, Na^+-Rb^+, Na^+-Cs^+, Na^+-NH_4^+ exchange equilibria), obtained by equilibrating a given amount of a homovalent soil clay (e.g., Na^+-clay) with varying initial concentrations of a replacing cation (e.g., $K^+/Rb^+/Cs^+/NH_4^+$), when plotted against the corresponding position of equilibrium (e.g., the ionic composition of the solid phase at equilibrium), demonstrate wide variations as shown in Figure 15.

If K_m were a thermodynamic equilibrium constant, its values should have remained constant with the position of equilibrium, unlike what is apparent from Figure 15. When one examines a heterovalent ion exchange equilibrium, such variations of K_m are even wider, as is evident from an examination of the experimental data on K_m for K^+ - Ca^{2+} ion exchange (Figure 16).

Such variations stem from the assumption of ideal solid solution behaviour of the soil exchanger, thereby neglecting the activity coefficients of the adsorbed ions on the solid (exchange) phase, in the Kerr-Vanselow and the

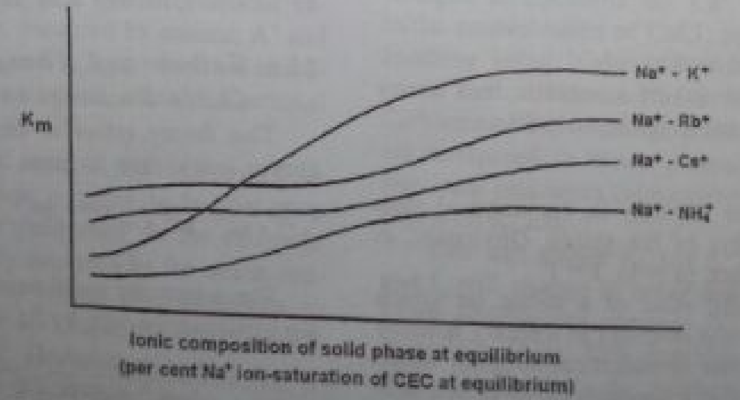


Figure 15. Plot of selectivity coefficient against position of equilibrium in a homovalent cation exchange equilibria

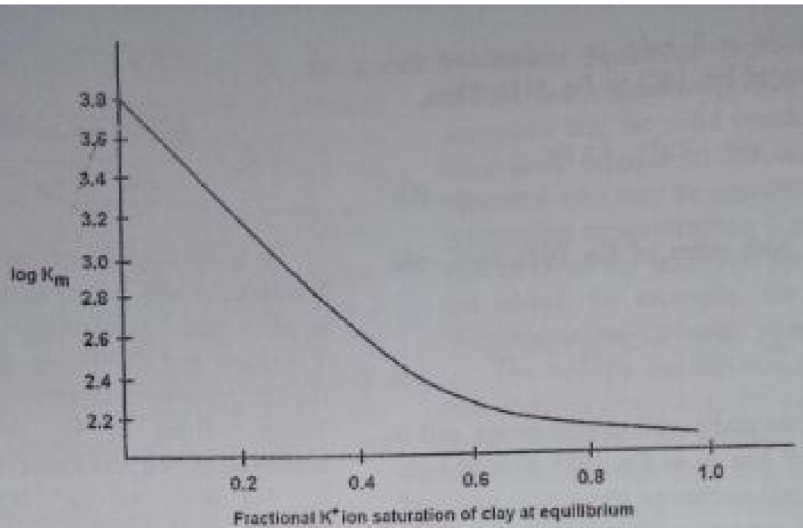


Figure 16. Plot of selectivity coefficient (log scale) against the position of equilibrium in heterovalent cation exchange equilibra

Source: Talibudeen (1972)

Gapon formulations. This is addressed by the thermodynamic theory of ion exchange which will follow.

11.5. *Thermodynamic Theory (due to Gaines and Thomas, 1953)*

Before proceeding further, it will be instructive to introduce, in brief, the concept of fugacity of a component (e.g., the solute) in a solution. The fugacity refers to the "escaping tendency", for instance, of a gas in a mixture. The fugacity (f) of a solute in a binary solution is related to its concentration (c) by the following equation:

$$f/c = 1$$

for all values of 'c' in an ideal solution, while

$$\lim_{c\to 0}\frac{f}{c} = 1$$

for a non-ideal solution.

The activity (a) of a solution in a binary solution is related to its fugacity (f) as,

$$a = f/f^0$$

where, f^0 is the fugacity in the standard state (of unit activity) of the solute. Obviously, in the standard state (a = 1), $f = f^0$.

The standard state of a solute (in which the solute has unit activity) in a binary solution is defined as that hypothetical solution which behaves ideally at a concentration (c) of 1 molar, 1 molal or 1 mole fraction, depending on the scale of concentration used.

However, such a concentration (1 molar, 1 molal, etc.) denotes too concentrated a solution to behave ideally as required in the above-stated definition of the standard state. Hence, the latter has got *no physical existence*, and it is *hypothetical*. In practice, ideal behaviour may be expected only in dilute solutions of concentration, 10^{-4}–10^{-5} *M*, and at further dilution.

For the above-noted standard state of the solute, we have:

$$a = f/f_0 = 1$$

And the solution being ideal, f = c = 1 *M* (say). Hence,

$$f = f^0 = 1M = c^0 \text{ (say)}$$

where, c^0 is the standard state concentration of the solute, namely 1 *M* (say) in this case.

11.6. *Gaines and Thomas Theory of Cation Exchange in Soil*

This theory refers to the general ion exchange equilibrium as given below [Eq. (39)]:

$$z_B A^{z_A+}_{Soil} + z_A B^{z_B+}_{soln} \rightleftharpoons z_A B^{z_B+}_{Soil} + z_B A^{z_A+}_{soln} \quad \text{...(39)}$$

The selectivity coefficient is given, following earlier arguments, by Eq. (45):

$$\zeta_m \text{ (or } K_c) = \frac{(q_B/q_0)^{z_A}(C_A/C_0)^{z_B}}{(q_A/q_0)^{z_B}(C_B/C_0)^{z_A}} \cdot \frac{\gamma_A^{z_B}}{\gamma_B^{z_A}} \cdot C_0^{z_B - z_A} \quad \text{...(45)}$$

where,

q_0 = Total number of equivalent negative sites on the exchange phase,

q_A = Equivalent negative sites occupied by A^+ ions on exchange phase at equilibrium,

q_A/q_0 = Mole fraction on exchange phase at equilibrium as occupied by A^+ ions,

q_B = Equivalent negative sites occupied by B^+ ions on exchange phase at equilibrium,

γ_A = Activity coefficient of A^+ ions in solution phase at equilibrium,

γ_B = Activity coefficient of B^+ ions in solution phase at equilibrium,

C_0 = Total equivalent amount of positive sites in the exchange site (in the diffuse layer). It is equal to q_0, which is in the solid phase ($q_0 = C_0$),

C_A = Equivalent number of cationic sites in the solution at equilibrium occupied by A^+ ion,

C_A/C_0 = Mole fraction of A^+ ions in the solution phase at equilibrium, and

C_B/C_0 = Mole fraction of B^+ ions in the solution phase at equilibrium.

The selectivity quotient (K_m or K_C) is *not* a thermodynamic equilibrium constant in that the activity coefficient of the adsorbed ions in the exchange phase at equilibrium is ignored (is set equal to unity). The Gains and Thomas theory then proceeds to address this issue based on the following two assumptions:

(1) The adsorbed ionic layer is treated as a solid solution (*not* ideal) so that the cationic concentrations are obtained from the fractions of the total ion 'sites' on the exchange phase, occupied by cations, A^+ and B^+ ions (q_A/q_0 and q_B/q_0), which are identical to the parameter, namely fractional saturation of the CEC.

(2) The standard state of unit activity of an adsorbed cation is taken to be that state when all the ionic sites on the exchange phase are fully saturated by the cation concerned and the system behaves ideally so that $a_s = C_s = 1$. That is, the activity coefficient, $f_s = 1$. Hence, the bonding energy parameter in the standard state, namely, $-\log f_{s}$, is given as [Section 10.2, Eq. (14)]:

$$(B.E.)_s \alpha - \log f_s = 0 \qquad \ldots(46)$$

Thus, all the cations in their respective standard states will exchange with equal ease. Referring to Eq. (45), in the standard state for A^+ ions, one has,

$$a_{A^+} = f_{A^+} \frac{q_A}{q_0} = 1$$

But in this state, according to assumption 2, $q_A = q_0$. Hence, $f_{A^+} = 1$

Thus, the value of the activity coefficient of the adsorbed A^+ ion, in its standard state of unit activity (i.e., f_{A^+}), turns out to be unity, as shown above. Thus, the above relationships, based on assumption (2) of the Gains and Thomas theory, are self-consistent and also consistent with the definitions of fugacity, activity, activity coefficient and the standard state.

In general, the thermodynamic equilibrium constant (K) for the given ion exchange equilibrium (Eq. 39) is related to K_m (or K_C) by the relationship:

$$K = K_C (f_B^{ZA} / f_A^{ZB}) \qquad \ldots(47)$$

where, f_A and f_B are the activity coefficients of the adsorbed cations, A^+ and B^+, on the exchange phase at equilibrium.

Because the activity coefficient of an ion varies with the ionic strength (I) of the medium [Eqs (10) and (11) of Section 10.2] for a heterovalent exchange of, say, K^+-Ca^{2+} ion exchange discussed earlier, I [$= \frac{1}{2} \Sigma m_i Z_i^2$; Eq. (8)] will obviously change with the position of equilibrium, i.e., with more of K^+ ion, initially present on the exchange phase, being exchanged at equilibria for Ca^+ ions (with the initial concentration of $CaCl_2$ solution used for leaching being higher), I will increase, and hence the activity coefficient of both K^+ and Ca^{2+} ions on the exchange phase will be different compared to those obtained at a higher fractional saturation (at equilibrium) of the soil clay with K^+ ion.

This will render K_C (or K_m) change for f_A and f_B will change in order that the thermodynamic equilibrium constant (K) remains unaltered [Eq. (47)]. This explains the variations of K_C (or K_m) with the position of equilibrium in a heterovalent ion exchange reaction (Figure 16).

Even for the homovalent cation exchange (Figure 15), the activity coefficients of the monovalent cations involved need not be the same if one examines the Debye-Hückel extended Law [(Eq. (10), giving the activity coefficient of an ion. Further, the relative accessibility on the exchanger (e.g., the clay) of the original and the replacing cations (in a homovalent exchange) need not be the same for a series of replacing cations, depending on the preferential and specific adsorption. Hence, the ionic strength on the exchange (solid) phase will vary with the position of equilibrium for each pair of exchanging cations, so also the corresponding ionic activity coefficients, and hence the value of K_m (Figure 15). However, the variations in K_m have been found much more for the heterovalent ion exchange compared to the homovalent cases (Figures 15 and 16).

Such variations may also be traced to the heterogeneity of the exchange sites in the soil adsorbed phase. These exchange sites have been considered as homogeneous with respect to their specificity for a given cation in the derivation of the exchange equations. These exchange coefficients vary, though in different proportions, with the cationic saturation of the exchange complex. For instance, in Na-Ca exchange, the values of these coefficients increase with Na-saturation of the exchange complex. It has been found that the Gapon coefficient (K_G) can describe Na-Ca exchange isotherm with a fair degree of accuracy in 0-50 ESP range. Studies on illite, montmorillonite and soils indicate that the value of K_G remains fairly constant in the range, 0-50 ESP. Contrary to the increase in the values of selectivity coefficients with Na-saturation in Na-Ca or Na-Mg systems, their values decrease with the increase in K-saturation of the exchange complex in K-Ca or K-Mg systems. Indeed, the presence of K-specific exchange sites on aluminosilicate clay minerals in soils is the main reason for this opposing trend.

11.6. Clay-Membrane Electrode and Ionic Activity

A direct method for ionic activity measurement in suspension was developed by Marshall by way of using clay-membrane electrodes for determining the activities of Na^+, K^+, Mg^{2+}, Ca^{2+} and Ba^{2+} ions with a saturated calomel electrode as a reference electrode. The clay membranes are reversible either to cations, in general, or to monovalent cations only. Because of lack of specificity, the measurement has been limited to homoionic systems and to systems containing monovalent and divalent cations. Marshall and his coworkers (1940) determined the ionic activity as a function of the degree of saturation (titration curves of H-clay, H-Ca and H-Mg clays), the degree of ionization of exchangeable ions, and the mean free bonding energy (ΔG_M). As per this approach, the bonding energy for a monovalent ion, in particular, a monovalent cation (e.g., M^+) is obtained from the formula (48):

$$\Delta G_M = RT \ln \frac{C_M}{a_M} \qquad \ldots(48)$$

where,
C_M = Concentration of total monovalent cation (sum of ionized and non-ionized portions), and
a_M = Measured activity of the given monovalent cation (MV).

Using the relationship introduced earlier [Eq. (9)], we have

$$a_M = f_M C_M \qquad \ldots(49)$$

where, f_M is the activity coefficient of the given MV (M^+), one obtains from Eq. (48),

$$\Delta G_M = RT \ln \frac{C_M}{f_M C_M} = - RT \ln f_M = RT \ln f_M^{-1} \qquad \ldots(50)$$

Marshall termed f_M as "fraction active" in that f_M (= a_M/C_M) is the fraction, accessible to the ionic activity measurement by the given clay membrane electrode, is capable of participating in the ion exchange equilibrium.

One may wish to examine Eqs (48) and (50) little more closely.

Thus, taking a clue from Marshall's idea that the exchangeable ions are only partly ionized, one would envisage the bonding process (and the reverse) of the MV (M^+) on an exchanger, S^- (e.g., soil colloid) as follows:

$$S^- + M^+ \xrightarrow{\text{Bonding process}} \underset{(M^+ \text{ on the exchanger})}{S^- - M^+} \qquad \ldots(51)$$

The reverse of the bonding process is given by expression (52):

$$S^- - M^+ \,(M^+ \text{ in bonded state}) \rightleftharpoons S^- + M^+ \,(M^+ \text{ in soil suspension}) \qquad \ldots(52)$$

Therefore,

$$\Delta G_M \,(\text{bonding}) = RT \ln \frac{a_{M^+} \,(\text{bonded})}{a_{M^+} \,(\text{free in soil suspension})} \qquad \ldots(53)$$

Assuming that the bonded MV (M^+ ion) is only partially ionized in the soil suspension, which is measured by the given clay membrane electrode (reversible to M^+ ions), one has

a_{M^+} (free in soil suspension) = a_M (introduced earlier)

and a_{M^+} (bonded) = Total MV (M^+) concentration with the assumption that a_M is very small compared to total M^+ concentration = C_M

Hence, one has from Eq. (53),

$$\Delta G_M \,(\text{bonding}) = RT \ln \frac{C_M}{a_M} = - RT \ln f_M \qquad \ldots(54)$$

Equations (50) and (54) are identical.

Indeed, it was shown qualitatively earlier (Section 10.2), that - log γ_i (where γ_i was the activity coefficient of a counter ion, 'i') provides a measure of the bonding energy of the given exchangeable ion ('i').

The same conclusion follows here [i.e. Eq. (54)] from an altogether different and a rigorous thermodynamic approach.

11.7. Krishnamoorthy-Davis-Overstreet Equation

Krishnamoorthy *et al.* (1948) introduced another approach, which is based on statistical thermodynamics. For mono-divalent exchange with $Z_A = 1$ and $Z_B = 2$, the general form of the Krishnamoorthy-Davis- Overstreet equation is given by Eq. (55):

$$K_{KDO} = \frac{a_B}{a_A^2} \frac{m_A^2}{m_B \left[\{m_A + 2 - (2/Y)\} m_B \right]} \qquad \ldots(55)$$

where, m_A and m_B are the number of their moles in the adsorbed phase, respectively. Equation (55) can also be written as Eq. (56):

$$K_{KDO} = \frac{a_B}{a_A^2} \frac{M_A^2}{M_B \ \{M_A + 2 - (2/Y)\} M_B} \qquad \ldots(56)$$

where, M_A and M_B denote the mole fractions of cations A and B in the adsorbed phase, respectively.

12. Schofield's Ratio Law

This law states that if a negatively charged surface is in equilibrium with a solution containing certain activities (or concentrations) of monovalent, divalent and trivalent cations, the equilibrium will not be disturbed if the activities of the monovalent cations in the solution phase are altered by a certain factor, of the divalent cations by the square of that factor and of the trivalent cations by the cube of that factor. In other words, in a binary (two cations) system, the ratio of the cations held by a soil depends upon the ratio of the activities of the two cations in the equilibrium solution, in the case of homovalent cations (Na^+ and K^+ or Ca^{2+} and Mg^{2+}); the ratio of the activity of monovalent and square root of the activity of divalent cations, in the case of mono-divalent cations (Na^+ and Ca^{2+} or Na^+ and Mg^{2+}); and the ratio of the activity of monovalent and cube root of the activity of trivalent cations, in the case of mono-trivalent cations (Na^+ and Al^{3+}), and so on. The law thus requires that there is no exchange of cations between the adsorbed and solution phases of the soil so long as a_{Na}/a_K, $a_{Na}/(a_{Ca})^{1/2}$ and $a_{Na}/(a_{Al})^{1/3}$ do not alter with changes in the composition, dilution or concentration of the soil solution for Na^+-K^+, Na^+-Ca^{2+} and Na^+-Al^{3+} binary systems, respectively.

Indeed, for equilibrium in a binary exchange of cations between an adsorbed (e.g., soil colloidal or exchange) phase and a solution (e.g., soil solution) phase, such as one encounters in Eq. (43), namely,

$$K^+_{soil} + \tfrac{1}{2}Ca^{2+}_{soln} \rightleftharpoons \tfrac{1}{2}Ca^{2+}_{soil} + K^+_{soln} \quad \ldots(43a)$$

One would have the corresponding selectivity coefficient [Eq. (44)]:

$$\frac{[Ca^{2+}_{soil}]^{1/2}(K^+_{soln})}{[K^+_{soil}](Ca^{2+}_{soln})^{1/2}} = \text{Constant (say, K)} \quad \ldots(44a)$$

where, [ ] denote the activity (or concentration) in the adsorbed phase, and ( ) indicate the activity (or concentration) in the soil solution phase. A rearrangement of Eq. (44a) leads to Eq. (44b):

$$\frac{[K^+_{soil}]}{[Ca^{2+}_{soil}]^{1/2}} = \frac{(K^+_{soln})}{(Ca^{2+}_{soln})^{1/2}} = K = \tfrac{1}{2}\text{ (say)} \quad \ldots(44b)$$

On irrigation, let us assume that the concentrations in the solution phase decrease 4-fold so that one obtains from Eq. (44b),

$$\frac{[K^+_{soil}]}{[Ca^{2+}_{soil}]^{1/2}} = \frac{(K^+_{soln}/4)}{(Ca^{2+}_{soln}/4)^{1/2}} = \tfrac{1}{2} \times \tfrac{1}{2} = \tfrac{1}{4} \quad \ldots(57)$$

Because equilibrium demands that the value of K (= ½) remains constant, as long as the temperature does not change, it implies that to keep K (= ½), there will be re-distribution of ions and change of position of equilibrium between the adsorbed and the solution phases. Indeed, there will be release of K^+ ions from the adsorbed phase to the solution phase and fresh adsorption of Ca^{2+} ions present in the solution phase so that the value of K (= ½), as in Eq. (44b), is restored. In other words, when such equilibrium between the above-noted two phases is disturbed (e.g., on drying/fertilizer application/plant uptake/irrigation/rainfall, etc.), there will be always a shift of the monovalent cations (MV) from the concentrated to the diluted phase (e.g., K^+ ion moving in the above mentioned example from the adsorbed to the diluted solution phase), while the reverse would occur for the divalent cations (DV).

If, however, with the above-stated dilution of the solution phase on applying irrigation, appropriate steps had been taken to increase the K^+ concentration in the solution phase (e.g., by addition of MOP), in such a way that while Ca^{2+} ion concentration in the solution phase is diluted four times, that of K^+ ions is diluted only two times (i.e., √4 times), then Eq. (44b) would take the form of Eq. (58):

$$\frac{[K^+_{soil}]}{[Ca^{2+}_{soil}]^{1/2}} = \frac{(K^+_{soln}/2)}{(Ca^{2+}_{soln}/4)^{1/2}}$$

$$= \frac{(K^+_{soln})}{(Ca^{2+}_{soln})^{1/2}} = \tfrac{1}{2}\text{ (Constant)}$$

$$\ldots(58)$$

That is, the constant of Eq. (44b) would have remained unaltered and the position of equilibrium between the exchange and the solution phases will not be disturbed as required by the Ratio Law.

Similar conclusions may be derived from the Donnan Membrane Equilibrium Theory, as applied to ion exchange processes occurring in soil. Indeed, Eq. (44b) is identical with Eq. (36) describing the Donnan membrane equilibrium between the exchange and the solution phases of the soil.

Thus, an important consequence of the Ratio Law is the applicability of the Donnan Membrane Equilibrium to the description of the distribution of ions between the exchange phase and the soil solution phase in equilibrium with each other. Thus, for a K^+-Ca^{2+} system, as stated earlier, one has the relationship (59):

$$\left(\frac{a_{K^+}}{\sqrt{a_{Ca^{2+}}}}\right)_{\text{Exchange or adsorbed phase}} = \left(\frac{a_{K^+}}{\sqrt{a_{Ca^{2+}}}}\right)_{\text{Soil solution phase}} \quad \ldots(59)$$

While irrigating a soil with sodic/saline water (brackish water, for instance) having relatively high loading of Na^+ ion concentration, compared to the divalent Ca^{2+} and Mg^{2+} ions, the soil solution phase gets concentrated in respect of Na^+ ions, which, following the above-stated argument, tends to move to the relatively dilute phase (i.e., the exchange phase in this case), while the divalent cations move in the opposite direction. This tends to raise the ESP of the soil, and there is the possibility of dispersion of the exchange phase on continued use of such brackish water. This would cause a poor soil physical tilth.

To counter the above-noted alarming trend, pre-treatment of the irrigation source with gypsum (a standard package of practice) is often recommended. This would obviously ensure an increase of Ca^{2+} ion concentration in the irrigation water so as to prevent such rise of ESP of the exchange phase. That is, if the rise in Ca^{2+} ion concentration in the soil solution phase, upon such irrigation, is such that it is the square of the above-stated increase in Na^+ ion concentration, then according to the Ratio Law, there will be no shift in position of equilibrium between the exchange and the solution phases, and hence, the above-mentioned risk of sodification of the exchange phase is avoided.

13. Inverse Ratio Law

The Ratio Law considers the disturbance of the position of equilibrium between the exchange and the solution phases due to change in concentrations of the monovalent, divalent ions, etc., in the solution phase due to extraneous factors. However, such position may also be disturbed by the changes in concentrations (or activities) of the ions in the exchange phase. The consequences of the latter are dealt with by the Inverse Ratio Law (due to Matson).

Thus, given the possibility of occupying the exchangeable positions on the colloid (exchange) phase of a low or high charge density (i.e., with low or high CEC) clays in a mixture of colloidal dispersions, the monovalent cations are likely to be found in more number on the low charge density (i.e., "dilute" phase in respect of charge density) exchange phase of a colloid (e.g., clay), while the reverse will be true for the divalent cations. This ensues from the fact that the ionic fields of the monovalent cations are more compatible with the similar (but oppositely charged) ionic fields of the low CEC clays, while for the divalent cations, such compatibility is in respect of the ionic fields of high CEC clays. This bears similarity to what has been stated above in regard to consequences of the applicability of the Ratio Law or the Donnan membrane distribution in case of soil exchange phase - soil solution phase equilibrium.

Thus, when aqueous dispersions of kaolinite and montmorillonite clays are mixed and allowed to stand to reach a new equilibrium distribution of the corresponding exchangeable ions, the monovalent cations will be found in more numbers on the exchange phase of the kaolinite clay, while the divalent cations will populate mostly the exchange phase of the montmorillonite clay.

14. Plant Root Exchange in Soil

Plant roots behave as feeble ion exchangers in soil by virtue of possessing feeble net negative charge. Plant roots uptake cations from soil solution in exchange for secretion of equivalent quantity of H^+ ions (and as a result, the rhizosphere pH is generally lower than the bulk soil pH), while the anion uptake occurs in exchange for equivalent amounts of OH^- or HCO_3^- ions. However, because of imbalance in the cation/anion uptake by the plant, the corresponding root system develops a net negative charge, and hence an electrical double layer at the root-soil solution interfaces is formed.

Among the plants, the monocotyledons (e.g., cereals) have a lower root CEC than do the dicotyledons (e.g., legumes). Thus, the former root ion exchangers behave more like kaolinite clay, while the dicotyledons behave like montmorillonite clay. The monocotyledons, therefore, generally exhibit a preference for the monovalent cations over the divalent cations, while the reverse is true for the dicotyledons.

For instance, if rice and maize (cereals) are grown in the absence of extraneous potash application in a plot of land, marginal in potash supplying capacity, while berseem and rice are grown in an adjacent plot with suitable cultural and management practices (with no potash application), then it is more likely that both the crops of rice and maize would almost simultaneously show potash deficiency symptoms in the first plot, while there will be much delayed appearance of the deficiency symptoms in the crops of the second plot. This obviously follows from the preference of both rice and

Table 5. Cation exchange capacity (CEC), surface charge density (SCD) and specific surface area (SSA) of some clay minerals

Property	Kaolinite	Clay mineral fine mica	Smectite
Cation exchange capacity (CEC) [cmol(p^+) kg^{-1}]	3 - 15	15-40	80-120
Specific surface area (SSA) (m^2 g^{-1})	5 - 20	100-120	700-800
Surface charge density (SCD) (mol m^{-2})*	6.7×10^{-6}	1.65×10^{-6}	1.25×10^{-6}

*SCD = CEC/SSA

Table 6. Effect of organic carbon (O.C.) and surface charge density (SCD) on ΔG^0_r and K_G for Na-Ca exchange at 25 °C

Soil source	O.C. (%)	CEC [cmol(p^+)kg^{-1}]	SCD x 10^{-6} (mol m^{-2})	ΔG^0_r (kJ eq^{-1})	K_G ($mol^{-1/2}$ $L^{1/2}$)
1. India	0.19	11.2	1.01	0.25	0.460
2. India	0.85	11.7	1.41	0.97	0.300
3. U.K.	1.20	46.8	0.89	1.48	0.328
4. U.K.	2.30	20.8	1.72	3.19	0.260
5. U.K.	1.10	30.7	1.10	0.15	0.445
6. U.K.	5.20	30.0	2.28	5.00	0.260

Source: Poonia and Talibudeen (1977)

maize for potash uptake as compared to that of berseem, as explained above.

15. Factor Affecting Cation Exchange Capacity

The cation exchange equilibria in soils are affected by two types of factors: (i) soil factors, and (ii) solution factors.

15.1 Soil Factors

15.1.1. Effect of Clay Minerals

Some of the relevant properties of important clay minerals (Table 5) show that CEC and specific surface area (SSA) are in the order: smectite > fine mica > kaolinite. Nevertheless, the trend of the ratio of CEC to specific surface area (i.e. SCD) of these clay minerals is just the reverse. It is important to mention that as per the Diffuse Double Layer (DDL) equation, the main emphasis should be given to SCD rather than to CEC of clay minerals while describing the mono-divalent cation exchange equilibria data.

15.1.2. Effect of Organic Matter

Organic matter in a soil affects the cation selectivity in two opposing directions. Closely spaced carboxyl and phenolic groups present in organic compounds cause an increase in SCD of soil and thus selectivity towards multivalent cations. Contrary to this, an increase in organic matter in the soil results in the increase of internal : external surface area / exchange sites. This increases the preference of the exchanger for those monovalent cations like K^+ and NH_4^+, which are mainly influenced by the geometry of adsorption sites (Poonia and Niederbudde, 1990). The overall effect of organic matter on cation specificity is thus a result of these opposing phenomena (Table 6).

15.1.3. Surface Charge Density

The diffuse double layer (DDL) equation predicts that the higher the SCD of soil, the smaller would be the fractional saturation of the soil-exchange complex by the monovalent cation in a mono-divalent cation system. The SCD of soil, however, fails to predict the exchange behaviour of soils when the monovalent cations involved in the exchange process are strongly affected by the geometry of exchange sites. It is assumed in the DDL equation that charge density on soil surfaces is uniform. This assumption, however, does not hold true in most soils consisting of several types of

Table 7. Effect of total electrolyte concentration (TEC) on ΔG_r^0 and K_c for Na-Ca exchange

Soil	CEC [cmol(p⁺) kg⁻¹]	SCD x 10^6 (mol m⁻²)	TEC (millequi L⁻¹)			
			25	50	75	100
			ΔG^0_r (kJ eq⁻¹)			
1.	5.5	1.39	1.91	2.51	2.35	3.24
2.	17.0	4.91	3.88	4.64	5.94	7.70
			K_G(mol$^{-1/2}$ L$^{1/2}$)			
1.	5.5	1.39	0.48	0.44	0.33	0.24
2.	17.0	4.91	0.66	0.54	0.52	0.50

Source: Poonia *et al.* (1984)

Table 8. Effect of Ca : Mg ratio on ΔG^0_r and K_c for Na-(Ca + Mg) system

Soil	CEC [cmol(p⁺)kg⁻¹]	Organic carbon (%)	Ca : Mg ratio	ΔG^0_r (kJ eq⁻¹)	K_G (mol$^{-1/2}$ L$^{1/2}$)
1.	13.32	1.37	3:1	0.69	0.512
			1:3	0.60	0.528
2.	20.27	0.60	3:1	1.07	0.477
			1:3	0.87	0.498
3.	11.19	0.27	3:1	1.97	0.387
			1:3	0.96	0.447

Source: Poonia *et al.* (1980)

clay minerals, organic matter and oxides. To describe the experimental data on mono-divalent cation exchange equilibria, the SCD thus needs to be multiplied by an empirical factor assuming no interaction between the adjacent clay platelets or by invoking interaction between the clay platelets without using an empirical factor.

15.2. Solution Factors

15.2.1. Effect of Total Electrolyte Concentration

The relative preference of soil for a divalent cation increases with the increase in total electrolyte concentration (TEC) of the equilibrium solution (Poonia *et al*., 1984). The results shown in Table 7 indicate an increase in the relative preference for divalent cation over monovalent cation with the increase in TEC of the equilibrium solution.

15.2.2. Effect of Type of Cations

Cations of the same valency but of different hydrated size may have considerable difference in their specificity for exchange sites. To illustrate this, some results on Na-(Ca and Mg) exchange equilibria for soils having different physicochemical properties, are given in Table 8 (Poonia and Pal, 1979; Poonia *et al*., 1980). The relatively larger values of ΔG_r and smaller values of K_G in Ca-dominated as compared to Mg-dominated system show higher preference of the soils for Ca over Mg.

15.2.3. Effect of Type of Anions

The cation exchange equations do not take into account the nature of anions. The formation of more extensive and stronger undissociated ion-pairs by divalent anions with divalent cations (e.g., $CaSO_4^0$ and $MgSO_4^0$) as compared to monovalent (e.g., $NaSO_4^-$) cations influence mono-divalent cation exchange equilibria in soils (Babcock and Schulz, 1963; Rao *et al*., 1968). To illustrate this, some pertinent experimental results on the role of chloride and sulphate ions in water on Na-saturation of some soils are given in Table 9 (Mehta *et al*., 1983).

16. Negative Adsorption

The interaction between anions and clay minerals, in principle, comprises two different

Table 9. Effect of anionic composition on ΔG^0, and K_s for Na-Ca and Na-Mg systems

Soil No.	Na-Ca			Na-Mg		
	Cl	SO_4	SO_4^*	Cl	SO_4	SO_4^*
			ΔG^0, (kJ eq^{-1})			
1.	3.87	2.57	3.52	3.62	2.04	3.57
2.	1.91	1.07	1.92	1.81	0.36	1.48
3.	2.91	1.83	2.87	1.90	0.73	1.73
			K_s (mol$^{-1/2}$ L$^{1/2}$)			
1.	0.49	0.61	0.50	0.54	0.67	0.55
2.	0.66	0.79	0.58	0.67	0.79	0.62
3.	0.54	0.62	0.54	0.62	0.78	0.62

*After considering ion pairs of $CaSO_4^{\circ}$, $MgSO_4^{\circ}$ and $NaSO_4^-$
Source: Mehta *et al.* (1983)

phenomena, viz. the bonding of the anions by positive charges presumably located on the broken edges of the clay, and the repulsion between the anions and the negative charges mainly on the planar surfaces of the clay particles. Both the electrostatic attraction between the anions and positively charged sites, and the chemical bonding may contribute to adsorption of anions. The repulsion of anions by the negatively charged sites leads to a local deficit of anions near a clay particle in comparison with the equilibrium solution. This deficit has been termed as 'negative adsorption'. As almost all the soils found in nature possess net negative charge, negative adsorption or 'anion exclusion' is exhibited to a certain extent by all soils.

17. Volume of Exclusion and Estimation of Specific Surface Area of Soil

The product of the equivalent distance of exclusion and the specific surface area of colloids yields equivalent volume of exclusion. The volume of exclusion is the volume that appears to be completely free of anions. For a soil of a given specific surface area, the volume of exclusion increases with the decrease of total electrolyte concentration, and conversely for a given electrolyte concentration, the higher the specific surface area of soil, the higher would be the volume of exclusion. The experimental estimation of volume of exclusion in combination with the calculated value of the equivalent distance of exclusion yields the specific surface area of colloids or soils. The determination of anion exclusion is an important method for the estimation of the specific surface area of soil.

When the volume of exclusion is multiplied by the equilibrium concentration of the anion (milliequi mL^{-1}), one gets the amount of anion excluded per unit mass of soil. The fraction of the liquid layer that is free of anions or salts increases with the decrease of moisture content in soil. When the equivalent distance of exclusion (δdl, cm) is a non-negligible fraction of the liquid layer (dl, cm), the following practical aspects are involved:

(1) Anions move faster than the mean flow velocity in soils. When a salt solution flows through soil pores, two phenomena, viz. anion exclusion near the walls (soil colloid surfaces) of the pores and the increase in velocity of liquid with the increase in distance from the colloid surface take pace. Both these phenomena result in faster movement of anions through soil than the mean interstitial flow velocity of the liquid calculated as the Darcy velocity divided by the water-filled porosity.

(2) Because of anion exclusion, i.e., repulsion of anions from the vicinity of the negatively charged surface, the concentration of anions in the easily extractable soil solution becomes higher than that in soil solution obtained by means of high-pressure filtration.

18. Anion Exchange

As has been discussed above, positively charged sites exist on the edges of layer-lattice silicate minerals (1:1 type in particular) and surfaces of oxides (with low specific surface area), mainly under acidic conditions. The total amount of anions held exchangeably by a unit mass of soil, termed as its anion exchange capacity (AEC), is therefore much less than the cation exchange capacity (CEC) of the soil. Further, contrary to cation exchange, the capacity of soil to adsorb anions increases as pH decreases or acidity increases, which increases the pH-dependent positive charge of soil colloids. In addition to pH, the adsorption of anions depends on the concentration of anions. Higher is the concentration of an anion, the greater is its adsorption. As the ratio of cation exchange capacity to anion exchange capacity of layer-lattice silicate mineral increases, the adsorption of anions decreases. Soils containing montmorillonite (CEC mainly due to isomorphous substitution) generally exhibit low AEC than those dominant in kaolinite and hydrous oxides of Al and Fe. Similarly, acid soils in the tropical and sub-tropical regions containing hydrous oxides of Al and Fe exhibit much higher AEC than alkaline and calcareous young soils of arid and semi-arid regions. Because humic colloids always carry negative charge, they can compete with other anions for the adsorption sites. Also, the selectivity of the positive sites with respect to different anions, as a rule, is much larger than that of the negative sites for cations. On the basis of experimental data, the preference of different anions for the positive sites has been found to follow the order:

$$SiO_4^{4-} > PO_4^{3-} >> SO_4^{2-} > NO_3^- = Cl^-$$

Consequently, SO_4^{2-}, NO_3^- and Cl^- ions are not adsorbed even if PO_4^{3-} ions are present at very low concentration in comparison to concentrations of other anions. At low pH values (< 6) and in the absence of PO_4^{3-} from a system, adsorption of SO_4^{2-} and Cl^- may also occur.

So far the adsorption of anions on layer-lattice silicates and oxides has been considered as being due to electrostatic force only. But some anions are bound in natural soils by mechanisms other than electrostatic. These mechanisms pertain to isomorphous displacement or substitution of hydroxyl ions from the lattice of clay minerals or hydrous oxides and the formation of chemical bonds with edge Al-groups of layer silicates. Fluoride ion (F^-), for example, has the same size as OH^- ion and may therefore substitute OH^- ion in an isomorphous manner, as shown below:

$$R\text{-}OH + F^- = R\text{-}F + OH^-$$

There is also evidence that the phosphate ion, despite having a different size from that of OH^- ion, is also bound as a structural non-diffusible unit by displacement of lattice OH, as the so-called 'ligand-exchanged' anion (see later).

The relatively strong preference for phosphate ions by clay minerals and hydrous oxides as compared to other anions of lower valency indicates the existence of chemical bonds between the solid phase and phosphate. This is supported by the fact that at high concentration of phosphate ions, the edges of the solids become negatively charged. This, evidently, would not occur if the binding mechanism between phosphate ions and layer-lattice silicate minerals or oxides was electrostatic in nature. The presence of chemical bonds between phosphate ions and edge Al groups of layer-lattice silicates is also supported by the low solubility of aluminium phosphates. More about ligand exchange will follow in the next section.

During decomposition, soil organic matter liberates organic compounds, which acquire negative charges due to the dissociation of protons from the carboxylic acids and phenolic hydroxyl groups. These negatively charged compounds displace phosphate ions from the adsorption sites and thus increase phosphate ions in soil solution. This phenomenon generally enhances phosphate availability to plants, even though organic matter in acid soils, rich in sesquioxides and amorphous Fe and Al oxides, sometimes reduces the plant availability of phosphate in soil. The humic colloids also chelate Al, Fe, Mn ions into soluble chelates which are thereby deactivated in immobilizing the phosphate ions in the soil solution.

19. Ligand Exchange

Let us first introduce as to what one understands by a ligand. A ligand, which may be an anion or a neutral molecule, satisfies the secondary valency or coordination number (Table 1) of a central cation in a crystal lattice. If it is an anion, then it simultaneously satisfies the primary valency as well. Thus, Fe^{3+} ion has got a primary valency of +3 and a secondary valency of 6 (Table 1). In $FeCl_3$, three Cl^- ions satisfy the primary valency (+3), while in $[Fe^{III}(CN)_6]^{3-}$, six CN^- ions satisfy the primary valency (+3) and the secondary valency (6) of Fe^{3+} ion. However, in $[Fe(H_2O)_6]^{3+}$ ion, six water molecules (neutral species) satisfy the secondary valency (6) of Fe^{3+} cation, while the primary valency (+3) remains yet to be satisfied. In the complex ion, $[Fe^{III}(H_2O)_4(OH)_2]^+$, two OH^- ions simultaneously (and partly) satisfy both the primary valency and the secondary valency of Fe^{3+} ion, while the four neutral water molecules satisfy the remaining (4 of 6) secondary valency of Fe^{3+} cation. In this case, OH^- ions and water molecules are both ligand to Fe^{3+} cation, and are liable to undergo exchange with the extraneous anions which would form covalent bonds with Fe^{III} ion.

The conjugate base of strong mineral acids (e.g., HCl, HNO_3, HBr, $HClO_4$, etc.), namely the monovalent anions derived from these acids, e.g., Cl^-, NO_3^-, Br^-, ClO_4^-, etc., undergo anion exchange and are held at the positively charged soil colloidal sites by virtue of electrostatic forces. Such retention of anions is also referred to as the non-specific anion retention. This may possibly be described as the physical adsorption in that no chemical bond is formed between the soil solid surface and the anions in the same sense that the exchangeable cations (e.g., in the diffuse double layer) swarm around a mica-type surface which has been discussed earlier (Section 4.4).

Several other anions, including the organic anions, are generally specifically adsorbed to the surface of clays or clay-sized primary minerals. These anions may be derived from both strong and weak acids, and are of mono- and polyvalent type. In this case, the anions, held at the specific sites, form a chemical bond with the surface group. This phenomenon is also known as ligand exchange in that the extraneous anion, acting as a ligand to the centrally located cation (in the clay lattice), substitutes an existing surface ligand (e.g., an OH^- ion or a H_2O molecule).

While anion exchange takes place only at the positively charged colloid surface, e.g., for a variable-charge surface, namely hydrous oxides of Fe^{III}, such as goethite [α - $Fe^{III}OOH$], at a pH below the corresponding ZPC (pH 8.5, vide. Section 4.2), such oxides, having only the pH-dependent charge, undergo ligand exchange at pH values below, above and even at the ZPC pH of the given variable-charge colloid surface. The same would be true for other types of soil colloid surfaces of variable-charge as well.

The ligand exchange is distinct from the simple anion exchange in that the phosphate, for example, in acid soil readily gets immobilized and removed from the soil solution, following a generous application of soluble phosphatic fertilizer, even in the presence of a strong (say 1*M*) solution of NaCl. This tends to suggest that phosphate and chloride anions compete for different types of adsorption sites. The former is chemisorbed in soil, while the latter undergoes electrostatic (physical) retention.

19.1. Adsorption Envelope

Recalling the Langmuir adsorption isotherm (Eq. 16), one obtains

$$\frac{C}{x} = \frac{C}{x_m} + \frac{1}{x_m K} = \frac{KC+1}{x_m K} \qquad \ldots(60)$$

or,

$$\frac{x}{x_m} = \frac{KC}{1+KC} \qquad \ldots(61)$$

where,

K = Langmuir constant related to bonding energy,

x_m = Adsorption maximum,

C = Equilibrium concentration in soil solution, and

x = Amount adsorbed per unit mass of the adsorbent.

On studying the adsorption in 0.1*M* NaCl solution of a series of anions such as phosphate, fluoride, silicate, etc., on goethite at pH range from 2.5 to 12.5 and a given temperature, followed by plotting of the corresponding x_m against the pH of the study for each adsorbate, the adsorption envelopes obtained are depicted in Figure 17.

The adsorption maximum (x_m) value for F^- is the highest at pH ~ 3.0, and for silicate at pH ~9.6 (a broad maximum), while for phosphate, two inflexion points are obtained, at pH ~6.7 and pH ~11.3 where the phosphate curve changes its slope.

Incidentally, the pK_a of HF is 3.2 (at 25 °C), that of silicic acid is 9.7, while the pK_2 and pK_3 of phosphoric acid are very close to (and little higher than) 6.7 and 11.3, respec-

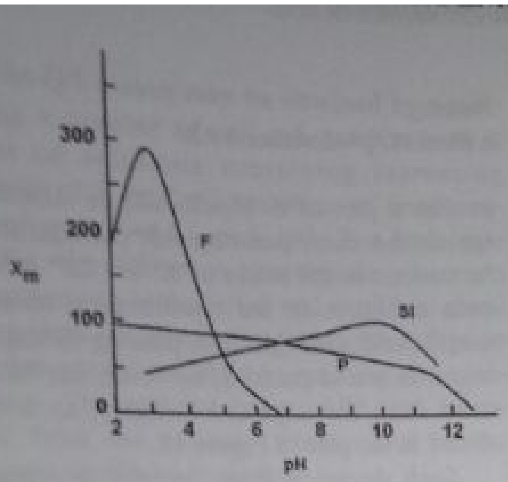


Figure 17. Adsorption envelope of specifically adsorbed anions
Source: Mott (1981)

tively. Presumably, if the experiment could have been continued below pH 2.0 (where goethite starts dissolving and losing its form), the phosphate curve would show a decreasing trend at a pH close to its pK_1, namely 2.1.

One would recall from the Henderson equation, giving the pH of a buffer mixture of a weak acid and its salt, namely Eq. (62):

$$pH = pK_2 + \log (C_{salt} / C_{acid}) \qquad ...(62)$$

Thus, at a pH = pKa,

$$C_{salt} = C_{acid}$$

$$\text{i.e., } C_{anion} = C_{acid} \qquad ...(63)$$

Keeping in view the nature of the adsorption envelopes presented in Figure 17, this tends to suggest that to cause maximum ligand exchange retention on goethite surface, not only the given anion, but the corresponding undissociated acid molecule also has an equally important role to play.

Let us take a look at the possible pathway of such ligand exchange reaction on goethite surface (which can be extended to cover other soil colloidal surfaces as well):

At a pH below the ZPC of goethite (pH < 8.5):

$$\underset{(Fe^{III})}{[\text{Oxide} - \text{M} - OH_2]^{x+}} + O.PO.(OH)_2^- \longrightarrow \underset{(Fe^{III})}{[\text{Oxide} - \text{M} - O.PO.(OH)_2]^{-(1-x)}} + H_2O \qquad ...(64)$$

At the ZPC (pH = 8.5):

$$\underset{(Fe^{III})}{[\text{Oxide}-\text{M} - OH_2]^{x+}} + O.PO.(OH)_2^- \longrightarrow \underset{(Fe^{III})}{[\text{Oxide}-\text{M} - O.PO.(OH)_2]^{-(1-x)}} + H_2O \qquad ...(65)$$

At a pH above the ZPC (pH > 8.5):

$$\underset{(Fe^{III})}{[\text{Oxide} - \text{M} - OH]^{-(1-x)}} + \underset{\text{Undissociated acid}}{HO.PO.(OH)_2} \longrightarrow$$

$$\underset{(Fe^{III})}{[\text{Oxide} - \text{M} - OH]^{-(1-x)}} \ldots H \ldots [O.PO.(OH)_2] \rightleftharpoons$$

$$\underset{(Fe^{III})}{[\text{Oxide} - \text{M} - \underset{H}{\ddot{O}}H]^{x+}} + O.PO.(OH)_2^- \longrightarrow$$

$$\underset{(Fe^{III})}{[\text{Oxide} - \text{M} - O.PO.(OH)_2]^{-(1-x)}} + H_2O \qquad ...(66)$$

Thus, at a pH > ZPC, the undissociated phosphoric acid molecule plays a key role in supplying the proton from one of its dissociable (OH) groups to protonate the OH^- ligand of Fe^{III} in goethite, thereby converting it to a neutral (H_2O) ligand which is then replaced by the negatively charged $H_2PO_4^-$ ion, causing ligand exchange.

It is also apparent that at pH equal to or less than the ZPC, the net negative charge of the soil colloid would increase as a result of ligand exchange, involving, say, phosphate. This will raise the CEC of the soil. Indeed, in the volcanic ash soils of the East Indies, for instance, allophanic clay minerals, being amorphous with high specific surface charge density, have very high phosphate fixation capacity. To these soils, often a massive dose of phosphatic fertilizer (e.g., 1000 kg P_2O_5/ha) is added to satisfy the phosphate hunger of the soil, and in the subsequent crop seasons, moderate dose of such fertilizer is added, which is supplemented by a slow release of small amounts of phosphate from the (earlier) immobilized phosphate in soil. Furthermore, such massive application of phosphatic fertilizer causes a rise in the CEC of the soil due to ligand exchange process as explained above. In such instances, phosphate plays the role of not only a fertilizer, but also that of a soil amendment.

20. Quantity-Intensity Relations

Recalling the Gapon equilibrium [Eq. (39)],

$$\left(\text{Soil}\right)^{-}\begin{matrix}Ca^{2+}/_2\\ Mg^{2+}/_2\end{matrix} + K^{+}_{soln.} \rightleftharpoons \left(\text{Soil}\right)^{-}K^{+} + \tfrac{1}{2}(Ca^{2+} + Mg^{2+})_{soln.} \quad ...(39)$$

One has the corresponding selectivity coefficient, or the Gapon exchange constant, K_G (or K_m), given [vide Eqs (39) and (44)] as Eq. (67):

$$K_G \text{ (or } K_m) = \frac{[K^{+}_{soil}]\,(Ca^{2+} + Mg^{2+})^{1/2}_{soln.}}{[CEC - K^{+}_{soil}]\,(K^{+}_{soln.})} \quad ...(67)$$

where, square brackets [ ], denote the activity (or concentration) of the exchangeable ions concerned on the adsorbed phase and circular brackets ( ), denote the activity in the soil solution phase at equilibrium.

Rewriting Eq. (67), one obtains

$$K_G = \frac{[\text{Exch. } K^{+}]\,(Ca^{2+} + Mg^{2+})^{1/2}_{soln.}}{[\text{Exch. } Ca^{2+} + Mg^{2+}]\,(K^{+}_{soln.})} \quad ...(68)$$

Or,

$$[\text{Exch. } K^{+}] = K_G \cdot [\text{Exch. } Ca^{2+} + Mg^{2+}].\ AR^{K} \quad ...(69)$$

where, AR^K is the ionic activity ratio in the equilibrium soil solution, and is given by Eq. (70) in the equilibrium soil solution.

$$AR^{K} = \frac{a_{K^{+}}}{\sqrt{a_{Ca^{2+} + Mg^{2+}}}} \quad ...(70)$$

Assuming that the change in activity of the exchangeable $[Ca^{2+} + Mg^{2+}]$ in the exchange phase is small due to such ion exchange process, the ionic activity (or concentration) of these ions on the exchange phase may be assumed to remain virtually constant during the process so that it follows from Eq. (69):

$$[\text{Exch. } K^{+}] = \text{Constant} \times AR^{K} \quad ...(71)$$

Hence,

$$\Delta\,[\text{Exch. } K^{+}] = \text{Constant}\ \Delta\, AR^{K} \quad ...(72)$$

so that a plot of Δ [Exch. K^+] or ΔExch. K^+ against the corresponding AR^K at equilibrium for various initial ratios of K^+ to $(Ca^{2+} + Mg^{2+})$ ionic activities in the equilibrating solutions should yield a straight line, passing through the origin. In actual practice, however, deviation is noted from the expected linearity, as demonstrated in the plot in Figure 18.

Such deviation from linearity is generally exhibited by a soil at the lower ranges of the AR^K values. This arises from the release of K^+ ions from the non-exchangeable pool of K^+ ions in the soil when the solution level (and hence the exchangeable K level) is low, causing thereby a small value of AR^K in the soil solution. Such release, in turn, leads to a new position of equilibrium between the exchange and the soil solution phases.

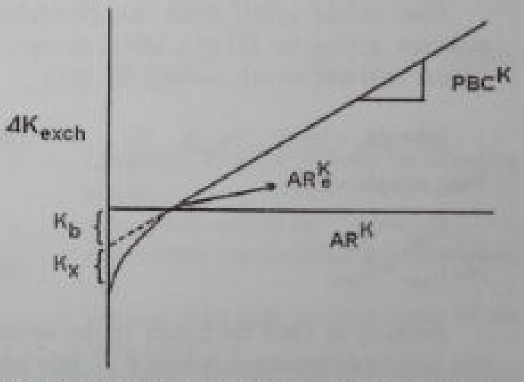


Figure 18. Quantity-Intensity plot of soil potassium

Because the Gapon equation [Eqs (39) and (41)] considers the equilibrium distribution of cations only between the exchange and the soil solution phases, the intervention by the non-exchangeable pool of K^+ ions, thereby disturbing the above noted equilibrium, would cause deviation from the linearity (as observed), demanded by the validity of the Gapon equation.

The plot in Figure 18 is referred to as the Quantity-Intensity (Q/I) relationship (Beckett, 1964), and brings out the changes in the quantity of, for instance, adsorbed potassium, in relation to the corresponding value of the activity ratio, $AR^{K}\left(= a_{K^{+}}/\sqrt{a_{Ca^{2+}+Mg^{2+}}}\right)$ in the equilibrium soil solution.

The Q/I curves may be obtained by equilibrating a number of soil sub-samples with a series of solutions containing increasing amounts of K in $CaCl_2$ solution of fixed concentration (say 0.01 mol L^{-1}). For each suspension, the difference between the potassium concentrations in the initial and equilibrated solution gives the amount by which the soil gains or loses potassium (ΔK) in reaching the equilibrium. The activity ratios, AR^K, are calculated from the cationic composition of the resultant equilibrium solution.

Several (Q/I) parameters are defined from Figure 18. These include the equilibrium activity ratio (AR^K_e) which refers to that value of the activity ratio in the equilibrium soil solution where the exchange phase neither gains nor releases exchangeable K^+ ions from or to the soil solution phase in contact. This value (AR_e^K) provides a measure of the inherent equilibrium soil solution activity ratio of the soil, and hence the inherent K^+ availability (to plants) compared to Ca^{2+} and Mg^{2+} availability.

The parameter, K_x, is related to the number of sites (on the adsorbed phase in soil) contributing towards specifically held K^+ ions. It was estimated from the difference between the intercept made by the interpolated linear portion of the (Q/I) plot and the intercept actually made by the (Q/I) plot on the Y-axis of the curve. Such K_x value is generally much higher for the illitic soil (with interlayer K^+ reserve as the non-exchangeable K pool of soil) as compared to the kaolinitic or the montmorillonitic soils.

The parameter K_L is defined as a measure of the labile K that the soil holds. The linear potential buffering capacity for soil K (PBC^K) is obtained from the slope of the linear portion of the (Q/I) plot (Figure 18). This refers to the inherent buffer capacity of a soil to maintain a given value of AR^K, i.e. the greater is its value [i.e. the steeper is the (Q/I) plot], the more will be the ability/tendency of the given soil to release K^+ from the exchange phase to the soil solution when K^+ is lost (through plant uptake and/or leaching), and *vice-versa* for K^+ gain by the soil solution (e.g., on fertilizer application and/or irrigation). All these (Q/I) parameters are therefore related to the chemistry and the fertility relationships of soil potassium.

21. *Methodologies to Study Ion Exchange Equilibria*

21.1.1. Flowchart Representation

The flowchart representation giving the procedural framework for cation exchange studies in binary cation systems is given in Figure 19. For non-equilibrium, time-dependent slow type of reactions, the Kinetic approach is a viable option for evaluation of thermodynamic parameters of cation exchange in soils with well-defined experimental methodology and theoretical considerations.

21.1.2. Equilibrium Procedure

The determination of cations on the exchange complex is generally carried out by two methods, viz. stripping technique and isotopic dilution technique.

Stripping Technique —In this technique, two approaches are adopted: (a) the soil in equilibrium with the solution is washed free of soluble salts with alcohol and then it is extracted in a salt solution (=1*N*) (free from the cations on the exchanger); or (b) the soil with the equilibrium solution is centrifuged, the supernatant liquid is collected for analysis and the equilibrium solution left in the soil is measured. The total cations extracted in the salt solution minus the cations in the left-over equilibrium solution are designated as the exchangeable cations.

Isotopic Dilution Technique — A given amount of soil (say 1 g) with a known volume of untagged equilibrium solution (say 5 mL) is mixed with an equal volume of tagged solution of the same composition. The contents are fully dispersed and equilibrated for 48 h on a shaker. The equilibrated soil suspension is centrifuged and the radioactivity is measured in the supernatant liquid, using a suitable radiation-measuring device. The amount of adsorbed cation (A_x) is calculated using Eq. (73):

$$A_x = \left(\frac{CPM_{control}}{CPM_{soil}} - 1\right)[A] \qquad \ldots(73)$$

where,

CPM = Counts per minute, and

[A] = Total amount of species x in solution

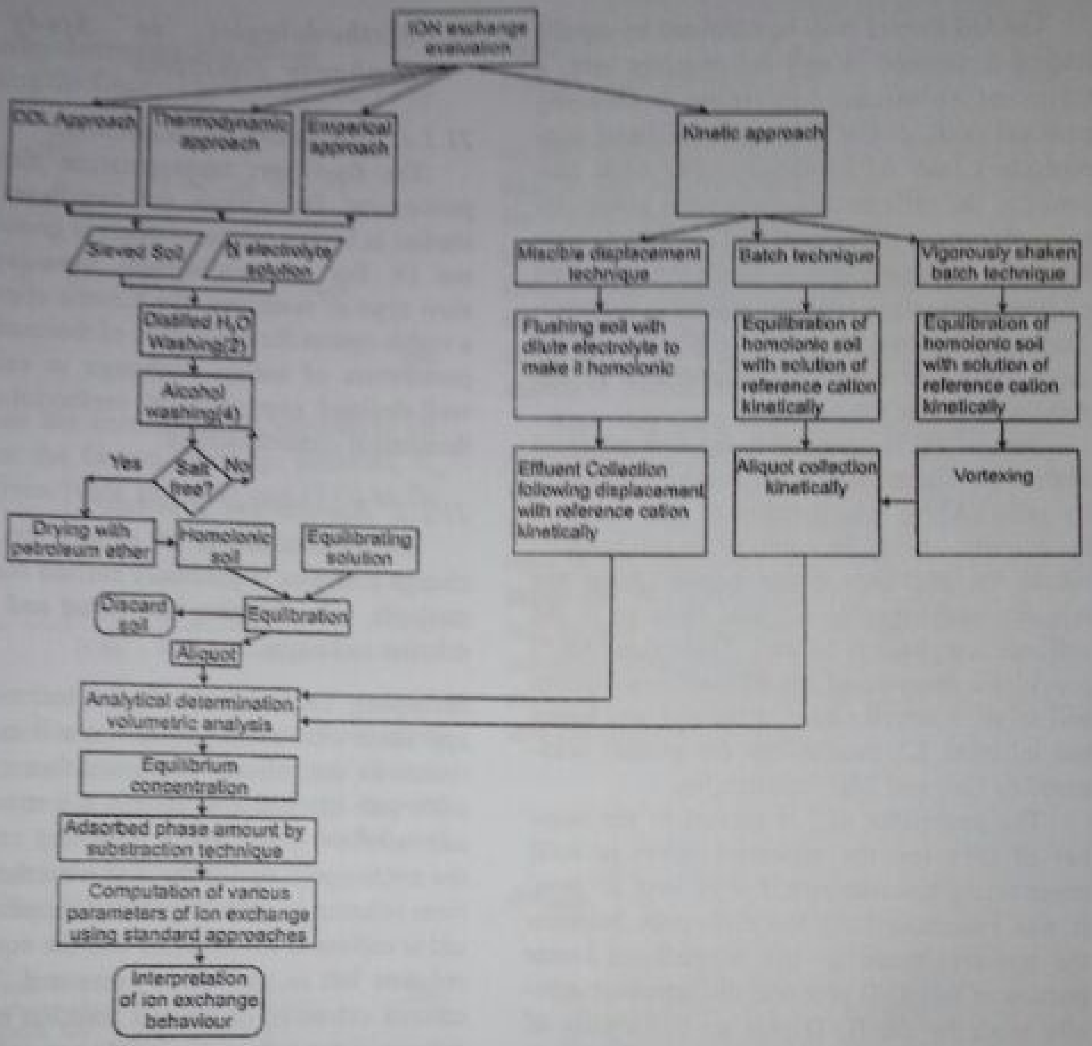


Figure 19. Flowchart representation of cation exchange equilibria studies in binary system

N.B.: Figures within the parentheses are number of washings usually needed

Source: Baruah and Patgiri (1996)

2.1.3. Kinetics of Ion Exchange

Notwithstanding the thermodynamic principles used to characterize exchange equilibria on clay and on soil surfaces, the use of a chemical kinetic approach to determine thermodynamic parameters has received increasing attention recently. It is particularly true for potassium exchange in clays and in soils. Although most studies on the kinetics of cation exchange in soils have been concerned with potassium release, the general principles for cation release are similar.

Techniques of common use for the study of kinetic reaction in clays and soils include Batch, Vigorously shaken batch and Miscible displacement or flow techniques.

For the kinetic model, the reaction written in exact stoichiometric expression is given by Eq. (74):

$$\mathrm{Soil^{-1}Ca^{2+}_{1/2} + K^{+}_{soln} \rightleftharpoons Soil^{-1}K^{+} + \tfrac{1}{2}\,Ca^{2+}_{soln}} \quad \ldots(74)$$

Equilibrium reactions between solution and adsorbed phases of soil profoundly influence K chemistry. The rate and direction of these reactions determine whether applied K will be leached into lower soil horizons or absorbed by plants or converted into unavailable forms, or else released into available forms. Knowledge about rates of reaction between solution and adsorbed phases of soil K is essential in order to predict the fate of the added K fertilizer in soils and make appropriate K fertilizer recommendations.

Miscible Displacement Technique

Miscible displacement technique refers to the mixing of soil solution initially present in both macro (non-capillary) and micro (capillary) pores in mobile and immobile regions with the invading solution, and consequently the displacement of the original solution from the system on the basis of equivalent pore volume. The miscible displacement technique simulates solute movement in soils under field conditions.

Prior to kinetic measurements on adsorption and desorption processes, soil samples should be passed through a 60-mesh stainless steel sieve and saturated with Ca using 1 *N* $CaCl_2$. Excess electrolyte should be removed with distilled water and with a 1:1 acetone-to-water mixture, until a negative test for Cl^- is obtained using $AgNO_3$.

For kinetics of K adsorption, 1 g of Ca-saturated soil is used and mixed with 50 mL of deionized water. The suspension is transferred into a 47-mm nucleopore filter column which is attached to a fraction collector on the terminal side and to a peristalatic pump on the delivery side. The pump is ideally required to deliver the K solution through the soil at a uniform flow rate of 1.0 mL per min. The aliquots are collected, preferably at 10-minute intervals until the concentration of the leachate equals that of the initial K concentration.

For kinetics of K desorption, the tubes, connected to the column, are rinsed thoroughly with distilled water and then the K desorption is initiated using 0.01*M* $CaCl_2$. This solution is passed through the soil until no K appears in the leachate. The quantity of K in solution for both the adsorption and desorption studies is measured. The kinetic experiments should also be carried out ideally at 25±1°C. The apparent adsorption and desorption rate constants are determined using the first-order kinetics as elucidated below.

Batch Technique

The miscible displacement technique suffers from the fact that the diffusion-controlled ion exchange is prominent, due to the prevalence of steady-state condition during the course of fluid dynamics and ion interaction. Thus, ion exchange may be over-estimated under such situations. Batch technique is an alternative to overcome this problem.

Similar to the miscible displacement technique, the soil samples are saturated with Ca using 1 *N* $CaCl_2$ solution; the excess salts are washed out with distilled water first and then with 1:1 acetone-to-water mixture, as long as chloride is present.

The 1-g samples of Ca-saturated soils are equilibrated with 50 mL of 50 μg K/mL of solution in centrifuge tubes at 25±1 °C on a reciprocating shaker. After sampling at 10-minute intervals, the suspension from each tube is centrifuged for 3 minutes at 2,000 rpm; then the supernatant solution is analyzed for K flame-photometrically. From the solution phase measurements of K as a function of time, the adsorption and desorption rate constants are worked out using the equations outlined below.

Vigorously Shaken Batch Technique

This technique is similar to the batch technique except that vortexing is done instead of shaking to achieve a better equilibration, especially in case of slow reactions, and to achieve quicker results. The theoretical consideration and experimental procedure are similar to the batch technique already described.

22. Theoretical Approach for Kinetic Studies

Referring to the $K^+ - Ca^{2+}$ exchange reaction in soil [(Eq. (74)], let us deal with the kinetic frame-work for the first-order K release and fixation reactions in soil (Sanyal and Majumdar, 2001). For K release, this equation assumes that the K concentration at the exchange sites of the soil colloid is the determining factor for the release rate of K into soil solution. In the batch technique, the rate is given by Eq. (75):

$$dK_t/dt = k_d(K_e - K_t) \quad ...(75)$$

where, K_t is the amount of K^+ ion released in time t, K_e the amount of K^+ ion that could be released at equilibrium, while k_d is the desorption rate coefficient.

For a miscible displacement or flow technique, the rate is given by Eq. (76):

$$d(K_t/K_0)/dt = -k_d K_t/K_0 \qquad \text{...(76)}$$

where, K_t and K_0 denote, respectively, the amount of K^+ ion on the exchange sites of the soil colloid at time t and time zero of desorption.

On suitably integrating Eqs (75) and (76) and utilizing the initial and boundary conditions, one arrives at the following integrated forms, namely

For batch equation:

$$\ln(K_0 - K_t) = \ln K_0 - k_d t \qquad \text{...(77)}$$

or,

$$\log(K_0 - K_t) = \log K_0 - (k_d/2.303).t \qquad \text{...(78)}$$

For miscible displacement or flow technique:

$$\ln(K_t/K_0) = -k_d t. \qquad \text{...(79)}$$

or,

$$\log(K_t/K_0) = -(k_d/2.303).t \qquad \text{...(80)}$$

A plot of log $(K_0 - K_t)$ against t leads to a linear plot from the slope of which the specific release reaction rate (k_d) is obtained in a batch technique. For flow technique, a plot of log (K_t/K_0) against t is used.

The first-order kinetics for K-fixation by soil/clays may also be described, e.g., for batch technique, we have:

$$d(K_0 - K_t)/dt = k_a K_t \qquad \text{...(81)}$$

where, K_t and K_0 are the concentrations of K in solution at time t and time zero, while k_a is the adsorption rate coefficient.

For miscible displacement or flow technique, one has

$$d(K_t/K_\infty)/dt = k_a(K_\infty - K_t)/K_\infty = k_a(1 - K_t/K_\infty) \qquad \text{...(82)}$$

where, K_t and K_∞ are the amounts of K at the exchange sites of the colloid at time t and at equilibrium.

The integrated forms are:

$$\ln K_t = \ln K_0 - k_a t \qquad \text{...(83)}$$

or,

$$\log K_t = \log K_0 - (k_a/2.303).t \qquad \text{...(84)}$$

For miscible displacement or flow technique:

$$\ln(1 - K_t/K_\infty) = -k_a t \qquad \text{...(85)}$$

or,

$$\log(1 - K_t/K_\infty) = -(k_a/2.303).t \qquad \text{...(86)}$$

Thus, a plot of log K_t against t leads to k_a from the slope of the resulting linear graph in a batch technique. The adsorption rate coefficient (k_a), in miscible displacement technique, is obtained from the slope of the linear plot of log $(1 - K_t/K_\infty)$ vs. time.

The first-order kinetics have been used by a number of workers to describe the exchange-kinetics and K-release from clay minerals, as well as surface and subsurface horizons of a number of soils (reviewed by Sanyal and Majumdar, 2001).

The equilibrium constant (K) for adsorption-desorption reactions in soil may also be derived from the corresponding adsorption-desorption rate coefficients, namely k_a and k_d introduced above. Thus, it can be shown that for adsorption (fixation) process in soil, the corresponding equilibrium constant (K_a) is given by Eq. (87):

$$K_a = k_a/k_d \qquad \text{...(87)}$$

While for the desorption (release) process in soil, the equilibrium constant (K_d) is obtained as per Eq. (88):

$$K_d = k_d/k_a \qquad \text{...(88)}$$

Many reactions in soil are multi-step ones, each step (elementary reaction) being characterized by the corresponding equilibrium (K) as well as the kinetic (k) constants. In such case, for instance, in a two-step process of adsorption, the overall equilibrium constant (K_a) is given by Eq. (89):

$$K_a = K K' = k_a k'_a / k_d k'_d \qquad \text{...(89)}$$

where, K and K′ are the equilibrium constants for the two elementary reactions, leading to the overall adsorption process, while k_a and k'_a are the corresponding adsorption rate coefficients and k_d and k'_d, the corresponding desorption rate coefficients.

23. Usefulness of Adsorption-Desorption Kinetic Data

The two salient reasons for studying the rate of soil chemical processes are:

- To predict how quickly reactions approach equilibrium or quasi-state equilibrium, and
- To investigate reaction mechanism.

Many reactions in soils are slow, yet they proceed at a measurable rate. Kinetic data on slow reactions may be of importance with regard to plant nutrition. Also, information about the reaction mechanism and processes occurring may be obtained from the kinetic data.

The analysis of adsorption isotherm data allows some prediction to be made on the field dosages of a chemical (e.g., fertilizer) to obtain a particular biological effect. The adsorption-desorption characteristics of a chemical, evaluated under laboratory conditions, and / or even in pot-culture experiments under greenhouse conditions (Hance *et al.*, 1968), merely provide an indication of how that chemical will perform *in situ*. The shift of the field performance from laboratory results is due to the use of excess water to soil ratio, extensive mixing in laboratory studies on one hand, and the effects of climatic variations, biological activity as well as agricultural practices followed (in the field) on the continued changes in the properties of soils, on the other.

In transferring results to field situations, it remains to be understood whether the kinetics of adsorption are comparable and whether the same surfaces are available in natural soils. Generally, the soils in the field are aggregated and the water content seldom exceeds the field capacity during the crop growing season.

References

Babcock, K.L. and Schulz, R.K. (1963) Effect of anions on the sodium-calcium exchange in soils. *Soil Science Society of America Proceedings* **27**, 630-632.

Baruah, T.C. and Patgiri, D.K. (1996) *Physics and Chemistry of Soils*. New Age International Publishers Private Limited, New Delhi.

Beckett, P.H.T. (1964) Studies on soil potassium: 2. The immediate Q/I relations of labile potassium in soils. *Journal of Soil Science* **15**, 9-12.

Bolt, G.H. (1955) Ion-adsorption by clay. *Soil Science* **79**, 267-276.

Brunauer, S., Emmett, P.H. and Teller, E. (1938) Adsorption of gases in multimolecular layers. *Journal of American Chemical Society* **60**, 309-319.

Dyer, J.R. (1971) *Application of Absorption Spectroscopy of Organic Compounds*. Prentice Hall of India Private Limited, New Delhi, p. 23.

Freundlich, H. (1926) *Colloid and Capillary Chemistry*. Translated from the third German edition by H. Stafford Hatfield, Metheun & Co. Ltd, London.

Gaines, G.L. and Thomas, H.C. (1953) Adsorption sites on clay minerals. II. A formulation of the thermodynamics of exchange adsorption. *Journal of Chemical Physics* **21**, 714-718.

Gapon, E.N. (1933) On the theory of exchange adsorption in soils. *Journal of General Chemistry, U.S.S.R.* **3**, 144-152.

Hance, R.J., Hocoombe, S.D. and Holroyd, J. (1968) The phytotoxicity of some herbicides in field and pot experiments in relation to soil properties. *Weed Research* **8**, 136-144.

Kerr, H.W. (1928) Nature of base exchange and soil acidity. *Journal of American Society of Agronomy* **20**, 309-335.

Khasawneh, F.E. and Copeland, J.P. (1973). Cotton root growth and uptake of nutrients: Relation of phosphorus uptake to quantity, intensity, and buffering capacity. *Soil Science Society America Proceedings* **37**, 250-254.

Krishnamoorthy, Ch., Davis, L.E., and Overstreet, R. (1948) Ion-exchange equations derived from statistical thermodynamics. *Science, New York* **108**, 439-440.

Lahiri, T.C. and Chakravarti, S.K. (1995) Distribution and nature of organic matter in some hill soils of West Bengal at various altitudes in the Eastern Himalayan region. *Journal of the Indian Society of Soil Science* **43**, 464-466.

Laugmuir, I. (1918) The adsorption of gases on plane surfaces of glass, mica and platinum. *Journal of American Chemical Society* **40**, 1361-1403.

Majumder, R.N. and Mukherjee, S.K. (1979) Degradation characteristics of hydrogen montmorillonites. *Journal of the Indian Society of Soil Science* **27**, 26-37.

Soil Acidity

N. PANDA, A.K. SARKAR and G.C. CHAMUAH

1. Introduction

Soils differ considerably in their acidity or pH. In aqueous systems, an acid is a substance that donates H^+ to some other substance. Conversely, a base is any substance that accepts H^+ or donates OH^-. Pure water undergoes slight self-ionization [Equation (1)]:

$$H_2O \rightleftharpoons H^+ + OH \qquad \text{...(1)}$$

The H^+ ion actually attaches to another H_2O molecule to give H_3O^+:

$$H_2O + H^+ \rightleftharpoons H_3O^+ \qquad \text{...(2)}$$

Since both H^+ and OH^- are produced by dissociation of a water molecule, H_2O is both a weak acid and a weak base. The product of H^+ and OH^- concentrations, is the dissociation constant of water or Kw:

$$Kw = [H^+]\,[OH^-] = [10^{-7}]\,[10^{-7}] = 10^{-14} \quad \text{(at 25 °C)} \qquad \text{...(3)}$$

Adding of an acid to H_2O will increase $[H^+]$, but $[OH^-]$ would decrease because Kw is a constant, viz. 10^{-14}. For example, in a 0.1 *M* HCl solution, the $[H^+]$ is $10^{-1}\,M$; thus the $[OH^-]$ is:

$$\begin{aligned} Kw = \; & [H^+]\,[OH^-] = 10^{-14} \\ & [10^{-1}]\,[OH^-] = 10^{-14} \\ \therefore \quad & [OH^-] = 10^{-13}\,M \end{aligned}$$

The $[H^+]$ in solution can be conveniently expressed using pH which is defined as :

$$pH = \log 1/[H^+] = -\log\,[H^+]$$

Soil pH measures H^+ activity (Activity = Concentration × Activity coefficient) and is expressed in logarithmic terms. The practical significance of the logarithmic relationship is that each unit change in soil pH means a ten-fold change in the amount of acidity and basicity, as proposed by the Danish Chemist Sorensen in 1909.

Solutions having $pH < 7.0$ are acidic, while those with $pH > 7.0$ are basic or alkaline, and those with pH 7.0 are neutral. Soil pH represents the H^+ activity in soil solution (Active acidity) and does not measure the undissociated or potential acidity.

2. Sources of Soil Acidity

The important sources of soil acidity are: Exchangeable H^+ and Al^{3+}, Fe and Al oxides, soil organic matter and clay minerals.

Two adsorbed cations – Hydrogen and Aluminium — are mainly responsible for the soil acidity. Most of the hydrogen ions are held on the surfaces of clay particles and organic matter.

Active acidity is due to hydrogen ions in the soil solution. The potential acidity is due to hydrogen ions held in various chemical combinations and those adsorbed on the surface of solid particles. The exchangeable hydrogen ions present in soil neutralize the negative charge arising from the isomorphous substitution of cations. The hydrogen ions are thus due to the

Table 1. Relationship between pH and [H^+] concentration

Conc. of H_3O^+ (mol/L)	pH	Conc. of H_3O^+ (mol/L)	pH
10^{-1}	1	10^{-8}	8
10^{-2}	2	10^{-9}	9
10^{-3}	3	10^{-10}	10
10^{-4}	4	10^{-11}	11
10^{-5}	5	10^{-12}	12
10^{-6}	6	10^{-13}	13
10^{-7}	7	10^{-14}	14

permanent charge on the mineral surfaces. The pH-dependent charge may arise from the structural OH^- groups at the corners and edges of soil clay minerals, which dissociate into H^+ ions.

At low pH (< 5.0) aluminium becomes soluble and is adsorbed by silicate clays or is tightly bound by organic matter. The Al^{3+} ions are then hydrolysed in the soil solution releasing H^+ ions which contribute to soil acidity [Equation (5)]:

$$Al^{3+} + 2H_2O^{-} = Al(OH)_2^{+} + 2H^{+} \qquad ...(5)$$

Thus, in strongly acidic soils, H^+ ion activity in soil solution is very high. The influence of adsorbed hydrogen ions is of a lesser magnitude in strongly acidic soils.

In moderately acidic soils, bulk of the aluminium is converted to aluminium hydroxy ions [Equations (6) and (7)]:

$$Al^{3+} + OH^{-} \longrightarrow Al(OH)^{2+} \qquad ...(6)$$

$$Al(OH)^{2+} + OH^{-} \longrightarrow Al(OH)_2^{+} \qquad ...(7)$$

Some of the aluminium hydroxy ions are adsorbed and act as exchangeable ions. They remain in equilibrium in soil solution. In the soil solution, they hydrolyse and produce hydrogen ions and thus contribute to soil acidity [Equation (8) and (9):

$$Al(OH)^{2+} + H_2O \longrightarrow Al(OH)_2^{+} + H^{+} \qquad ...(8)$$

$$Al(OH)_2^{+} + H_2O \longrightarrow Al(OH)_3 + H^{+} \qquad ...(9)$$

At pH below 5.0, iron also behaves like aluminium, but acidity generated by iron is much less than that of aluminium. In acid sulphate soils found in Kuttanad area of Kerala and Sunderbans of West Bengal, dissolved or free acidic substances like H_2SO_4, $Al_2(SO_4)_3$ and $Fe_2(SO_4)_3$ cause acidity. At higher pH (>5.0), amorphous aluminum and iron hydroxyl complexes may undergo hydrolysis and contribute hydrogen ions to the system.

Fine-textured soils containing high amounts of organic matter have more hydrogen ions than lighter soils with lower amounts of clay and organic matter. Moreover, different kinds of clay in soils hold variable amounts of hydrogen.

High aluminium ion concentration at the root surface may prevent the root from taking up phosphate, and the aluminium inside the living cell may interfere with sugar phosphorylation.

Soil organic matter or humus contains reactive carboxylic and phenolic groups that behave as weak acids. They dissociate, releasing H^+ ions. In organic soils and in mineral soils with high organic matter content, organic acids contribute significantly to soil acidity.

3. Classification of Soil Acidity

Soil acidity is of three kinds, namely (a) active acidity, (b) exchangeable acidity, and (c) reserve acidity. The hydrogen ions in the soil solution contribute to active acidity. It may be defined as the acidity developed due to concentration of hydrogen (H^+) and aluminium (Al^{3+}) ions in the soil solution. The concentration of hydrogen ion in soil solution due to active acidity is very small, implying that only a meager amount of lime would be required to neutralize active acidity. Inspite of smaller concentration, active acidity is important since the plant root and the microbes around the rhizosphere are influenced by it and because a dynamic equilibrium exists among active, exchangeable and reserve acidities in the soil.

In strongly acidic soils, the concentrations of exchangeable aluminium and hydrogen ions contribute to exchangeable acidity. It may be defined as 'the acidity developed due to adsorbed hydrogen (H^+) and aluminium (Al^{3+}) ions on soil colloids'. However, this exchangeable aluminium and hydrogen ions concentration is meagre in moderately acidic soils.

Aluminium hydroxy ions, and hydrogen and aluminium ions present in non-exchangeable form with organic matter and clays account for the reserve acidity. It is measured by titrating a soil suspension up to a certain pH, normally about 8.0, the amount of acidity in the soil being equivalent to the amount of NaOH used. When an acid soil is limed, gibbsite is formed from aluminium hydroxy ions with increase in pH. The potential acidity is much higher as compared to active or exchangeable acidity requiring much larger doses of lime to neutralize than what is required for neutralization of active acidity. Therefore, no attempt is made to neutralize reserve acidity. Liming is always limited to neutralize active acidy and part of the exchangeable acidity.

3.1. *Active and Potential Acidity*

The hydrogen ions in soil solution contribute to active acidity which is measured in terms of pH of the soil. Another part of the total acidity of the soil is called potential acidity. This refers to the acidity caused by hydrogen ions held in different chemical combinations and those adsorbed on the surfaces of solid clay particles and organic colloids. These hydrogen ions are in chemical equilibrium with the active hydrogen ions of the soil solution. When active acidity is neutralised progressively, the hydrogen ions from the potential sources are released into soil solution. When hydrogen ions change from the potential to active form, their places on the clay and organic surface are occupied by cations like Ca^{2+}, Mg^{2+}, K^+ and Na^+. Thus, total acidity is the summation of active and exchangeable acidity.

The role of mineral and organic colloids of soil in controlling soil pH has already been discussed. Out of the four factors, the most important is the proportion of cation exchange capacity of the exchange complex satisfied by bases. The other factors are: (a) nature of the clay and organic colloids, (b) kind of adsorbed bases, and (c) concentration of soluble salts in the soil solution.

The per cent base saturation is the ratio of the exchangeable bases and the cation exchange capacity of the soil [Equation (10)] :

$$\text{Percentage base saturation}=\frac{\text{Exchangeable bases }[\text{cmol}(p^+)\text{kg}^{-1}]}{\text{Cation exchange capacity }[\text{cmol}(p^+)\text{kg}^{-1}]}\times 100 \qquad ...(10)$$

Broadly, low per cent base saturation, i.e. below 50%, and a pH around 5.5, indicates acidity. On the other hand, when the base saturation touches 80%, and the pH is around 6.5, then an almost neutral soil exists.

3.2. *Buffering of Soils*

Buffering is defined as the tolerance to changes in pH of a solution. With such tolerance, the rise in soil pH is minimal until enough liming material is applied to effect appreciable depletion of hydrogen and aluminium ions responsible for potential acidity. On the other hand, rapid lowering of pH of soils is prevented by the buffering action, which is ordinarily expected, as organic acids are formed due to organic matter decomposition, leading to the temporary increase in the H^+ concentration. Hydrogen and aluminium ions along with the adsorbed bases largely control buffering of soils. Soils having more of clay and organic matter exhibit higher buffer capacity, resisting the change in soil pH. The higher is the cation exchange capacity (CEC) of a soil, the greater is its buffer capacity, other influencing factors remaining the same.

The degree of buffering is highest between pH 4.5 and 6.0 and drops off below pH 4.5 and above pH 6.0. Soils having equal per cent base saturation will have the highest pH associated with hydrous oxide clay minerals and lowest pH in 2:1 type aluminosilicate clay minerals. However, the pH is intermediate in soils dominant in kaolinite and humus. Soils high in organic matter and clay, particularly montmorillonite or vermienlite, have high buffering capacity. Organic soils and clays have higher CEC values and are more strongly buffered than are sandy soils of similar agroclimatic region. The type of clay affects the pH-per cent base saturation relationship. At low pH, Al^{3+} and hydroxy aluminium ions react and block exchange sites in silicate clays and humus, which reduces the cation exchange capacity.

On liming, obviously the CEC increases and more of Ca^{2+}, Mg^{2+}, etc. are adsorbed. Therefore, to raise the soil pH, additional lime is needed.

The buffering action is due to the influence of weak acids and their salts. Carbonates, bicarbonates and phosphates act as buffering agents. Organic acids are formed continually as a result of microbial activity. These organic acids are weak acids and serve as excellent buffering agents. Both organic and inorganic colloidal complexes present in the soil add to buffering process. The colloidal complex is similar to a mixture of weak acid and its salt and it functions like the acetic acid–sodium acetate buffer. Buffer capacity in soil is caused due to potential hydrogen of the exchange complex and its associated basic cations. Although pH gives some idea regarding the active acidity and to quite an extent base saturation, it hardly gives any idea about the buffering capacity of a soil.

4. Soil Reaction Correlations

The soil chemical properties and microbiological activities are highly influenced by the pH of soil. The concentrations of cations such as Ca^{2+}, Mg^{2+}, K^{+}, Na^{+} as well as Al^{3+} in soil have a direct relationship with soil pH. At low pH, Fe, Mn and Al are highly soluble and attain toxic levels, whereas at high pH, these ions become deficient. Nitrification is slow below pH 5.5. Under acid conditions, the pH-dependent NH_4^+ fixation between lattices of expanding layer silicates decreases with decreasing pH. The availability of phosphorus, primarily $H_2PO_4^-$ and HPO_4^{2-} ions, is highly pH-dependent. Its availability in many soils is highest when the pH is neutral or slightly acidic and it declines as the soil becomes strongly acidic or strongly alkaline. In general, with increase in soil acidity, the availability of micronutrients present in cationic forms such as Fe, Mn, Cu and Zn, increases, whereas the availability of micronutrients present as anions, namely molybdenum and boron, decreases.

Occasionally, the harmful effect of soil acidity on leguminous plants seems to be caused by Mo-deficiency rather than Al-toxicity. In organic soils, the relationship of pH with nutrient availability is different from that in mineral soil. In acid organic soils, Cu may be deficient, whereas it is not so in acid mineral soils.

Generally, bacteria and actinomycetes are active in mild acidic, neutral and high pH soils. Fungi function well in a wide range of pH, predominating in acid soils than in soils of intermediate pH or higher pH. The bacteria such as Rhizobia in legumes, involved in atmospheric nitrogen fixation, are less effective in acid soils. Bacteria responsible for conversion of nitrogen, sulphur, etc. from organic to simple inorganic forms are less active in acid soils.

Degree of tolerance of plants to soil acidity is quite variable. The pH range of 6.0 to 7.2 is suitable for most crop plants. Many leguminous crops are sensitive to soil acidity. Iron requirement of rhododendrons is high, which is sufficiently available at low pH. Advantage is taken to grow many forest trees tolerant to acidity. Many species, particularly conifers, can tolerate soil acidity.

5. Acid Forming Factors

Acid soils are formed mainly due to acidic parent materials and leaching of bases caused by heavy precipitation. Use of acid forming fertilizers also contributes to soil acidity. Acid granite rocks as parent material, in the process of soil formation, render the soil acidic. Nitrogenous fertilizers like ammonium sulphate and ammonium nitrate are responsible for creating soil acidity as calcium and other bases are displaced by ammonium ions, leading to the formation of calcium sulphate which is leachable. Microorganisms are responsible for the decomposition of organic residues and nitrification. Acids formed in the process are neutralized by a base from either free calcium carbonate or an exchange complex. If the base saturation is low, acids generated are not neutralized. Carbonic acid is formed by the reaction of carbon dioxide with water. The solvent action of carbonic acid removes base-forming cations like calcium and magnesium. Stronger organic acids are formed due to microbial decay.

Inorganic acids such as sulphuric and nitric, are the suppliers of hydrogen ions. These acids along with strong organic acids create high acid conditions. Large quantities of sulphuric and nitric acids are formed in the atmosphere from oxides of nitrogen and sulphur emitted from the combustion of coal, gasolene and other fossil fuels in the cities and around large industrial complexes. Rainwater has a pH of about 5.6 due to the presence of carbonic acid. But under the influence of burning fossil fuels or smelting of sulphide ores, the pH of rainwater may be as low as 4.0. However, soils, being highly buffered, tolerate acid rain to a large extent.

6. Determination of Soil pH

The hydrogen ion activity or pH of a soil solution is measured by a standard equipment called the 'pH meter'. It has two electrodes, which are placed in a soil suspension in water. One of these electrodes is the glass electrode which measures the activity of hydrogen ions. A glass electrode consists of a thin bulb of glass separating the soil suspension or paste from an acid solution (HCl) of known H^+ ion activity in which is immersed an internal reference electrode of $AgCl/Cl^-$ (or Hg_2Cl_2/Cl^-). An electrometric potential, proportional to the hydrogen ion activity of the solution into which the electrode is dipped, develops across the glass membrane. The second electrode, which is known as the external reference electrode, is usually a calomel electrode. It is used as a reference electrode for measuring the potential of the glass electrode. Earlier, a hydrogen electrode was being used which has been replaced by a glass electrode (Jackson, 1973).

The description of the electrochemical cell used for the measurement of soil pH is given below:

$Hg, Hg_2Cl_2, KCl \parallel H^+(a_1)|H^+(a_2) \parallel KCl, Hg_2Cl_2, Hg$

E (emf) is given by

$$E = E_{ref} - \frac{RT}{F} \ln a_{H^+} \quad ...(11)$$

$$= E_{ref} + \frac{2.303RT}{F} pH \quad ...(12)$$

or,

$$pH = \frac{E-E_{ref}}{2.303RT} = \frac{E-E_{ref}}{0.059} \text{ at } 25°C \quad ...(13)$$

where, E_{ref} is the emf of the external reference electrode.

The pH of a soil suspension varies considerably with its water content and it is always lower than the pH of the supernatant liquid. For acid and neutral soils, pH is generally low for a large soil : water ratio. In the system of clay lattice-exchangeable cations-solution a complex diffused double layer is formed. This electrical double layer is known as Helmholtz double layer. The surface of the clay lattice (inner layer) carries a negative charge. The outer layer has two parts — a positive layer created due to H-cations held on the clay lattice surface (Stern layer) and a positive layer diffused in the solution close to the lattice surface known as Guoy diffused layer. The pH value of the soil solution decreases through the diffused layer close to the negatively charged surface. The significance of the exchangeable H^+ ions can be appreciated when a soil suspension is allowed to settle and the pH is measured in the supernatant solution and the settled paste; the pH of the paste is generally about 0.2-0.5 pH unit lower.

The pH of a soil having a 0.01 *M* $CaCl_2$ remains unaffected over a wide range of soil : solution ratio. It is based on the relation called Ratio Law. Calcium is the most abundant basic metal cation in most soils and the addition of $CaCl_2$ solution does not usually change the proportions of the exchangeable cations and thus measurement of pH in 0.01 *M* $CaCl_2$ solution is advantageous. It compresses the diffuse double layer and supresses the hydrolysis. A 1 : 1 soil/ water paste may have pH one unit or more lower than a 1 : 5 soil/water suspension, which is a manifestation of the suspension effect. Apart from soil:water ratio, the electrolyte content and the carbon dioxide level also affect the soil pH value. Soils having high pH of more than 8.5 and containing free sodium carbonate and other soluble salts, show increase in pH when soil:water ratio is high. Dilution increases the hydrolysis of sodium clay with the formation of sodium hydroxide, leading to a large hydroxyl ion concentration. For this, 1 : 2.5 soil:water ratio is chosen. Since soil reaction varies with salt content, pH is often measured by making a soil suspension in salt solution of definite concentration. Soil suspension made in dilute calcium chloride solution for measurement of pH is somewhat beneficial. The activity of hydrogen ions divided by the square root of the activity of calcium ions is constant, according to the Ratio Law [Equation (14)]:

$$[H+] / \sqrt{[Ca2+]} = pH - \tfrac{1}{2}\, pCa = \text{Constant} \qquad \ldots(14)$$

The expression, pH – ½pCa, is known as the lime potential (LP) of a soil.

Since the value of lime potential does not depend on soil:water ratio or salt content, measurement of such pH in dilute aqueous $CaCl_2$ is more dependable. Since pH and pCa are measured in a clear solution, suspension effect is avoided. In a neutral soil, salt concentration affects the pH value due to increase in the apparent strength of acidic groups. By increasing salt concentration in an alkaline soil, the pH value is lowered due to reduction in the hydrolysis of the exchangeable cations like sodium. The bicarbonate concentration in the soil solution due to free calcium carbonate and magnesium carbonate influences the pH value.

7. Genesis, Occurrence and Characteristics of Acid Soils

The leaching of bases is the main cause of formation of acid soils, though parent acidic rock is also a contributing factor. In humid tropics, heavy precipitation accelerates the process of acidity. The soil-forming factors like mean annual temperature, vegetation, parent rock and hydrological conditions determine the degree of soil acidity. Acid soils in India have been formed due to drastic weathering associated with hot humid climate and heavy rainfall. Laterization, podzolization and accumulation of undecomposed organic matter under marshy conditions contribute to soil acidity. Estimates (NBSS & LUP, ICAR) show that about 35% of the acid soils of India are strong to moderately acidic (pH < 5.5), while 65% are slightly acidic (pH 5.5 to 6.5) (Table 2).

The acid soils are found in the Himalayan region, the eastern and north-eastern plains,

Table 2. Extent of acid soils in different states of India

(million ha)

States	Strongly acidic (pH < 4.5)	Moderately acidic (pH 4.5-5.5)	Slightly acidic (pH 5.5-6.5)	Total
Anurachal Pradesh	4.78	1.74	0.27	6.79
Assam	0.02	2.31	2.33	4.66
Bihar	—	0.04	2.32	2.36
Chhatisgarh	0.15	6.30	4.39	10.84
Goa	—	0.11	0.19	0.30
Himachal Pradesh	—	0.16	1.62	1.78
Jammu & Kashmir	—	0.09	1.48	1.57
Jharkhand	—	1.00	5.77	6.77
Karnataka	—	0.06	3.25	3.31
Kerala	0.14	2.87	0.75	3.76
Madhya Pradesh	—	1.12	10.60	11.72
Maharashtra	—	0.21	4.33	0.54
Manipur	0.43	1.44	0.32	2.19
Meghalaya	—	1.19	1.05	2.24
Mizoram	—	1.27	0.78	2.05
Nagaland	0.12	1.48	0.05	1.65
Orissa	—	0.26	8.41	8.67
Sikkim	0.28	0.32	—	0.60
Tamil Nadu	0.21	0.35	4.29	4.85
Tripura	0.06	0.75	0.24	1.05
Uttarakhand	—	1.18	2.30	3.48
West Bangal	—	0.56	4.20	4.76
All states	6.19	24.81	58.94	89.94

Source: NPSSLUP, Nagpur

peninsular India and the coastal plains under different agro-climatic situations. The soils occupy about 90 million hectares, constituting over one-fourth of total geographical area of the country (Table 2). About half of the area is under cultivation and rest is under forestry and other uses. About 25 million hectares of cultivated lands with pH value less than 5.5, are critically degraded with very poor physical, chemical and biological characteristics. The states having large areas under degraded acidic soils are: Arunachal Pradesh, Chhatisgarh, Kerala, Assam, Manipur, Nagaland, Mizoram, Meghalaya, Uttrakhand, Madhya Pradesh and Jharkhand. The degraded area in Chhatisgarh and Arunachal Pradesh is predominantly under forests. The alluvial acid soils are found in West Bengal, Bihar, Assam and parts of Orissa. The marshy acid soils are distributed across Assam, Kerala, West Bengal, coastal Orissa, South-east coast of Tamil Nadu and tarai regions of Uttrakhand, Bihar and West Bengal. The acid sulphate soils are unique to the Kuttanad area of Kerala.The classification of acid soils into broad soil groups, types and their distribution is given in Table 3.

Acid soils are base-unsaturated. The extent of base-unsaturation within pH range 5.0-6.0 generally varies from 16% to 67%. The active H^+ and Al^{3+} are bound to the clay. The $Al(OH)^{2+}$ is of minor significance and exists over a narrow pH range. The Al^{3+} is predominant below pH 4.7, $Al(OH)^{2+}$ between pH 4.7 and 6.5, $Al(OH)_3$ between 6.5 and 8.0 and $Al(OH)_4^-$ above pH 8.0. The iron also contributes towards acidity in several acid soils of Himachal Pradesh. Acid sulphate soils of Kerala are influenced by dissolved substances such as H_2SO_4, $Fe_2(SO_4)_3$ and $Al_2(SO_4)_3$. A good correlation between extractable Al and organic carbon content of soil has been found in acid soils of Orissa. A higher amount of exchangeable H^+, compared to Al^{3+}, has been found in the surface layer of Alfisol, which is reversed in the lower layer. The acidity in the soils of north-eastern region is attributed to the presence of exchangeable Al^{3+}, its content [$cmol(p^+)kg^{-1}$] varying from 0.1 to 4.1 in Sikkim, 0.13 to 0.63 in Nagaland, 0.15 to 0.55 in Tripura, 0.04 to 3.53 in Meghalaya and 0.75 to 3.0 in the Brahmaputra valley of Assam.

8. Naturally Occurring Liming Materials including Industrial Wastes

Carbonates, oxides and hydroxides of calcium and magnesium are referred to as agricultural lime. Among the naturally occurring lime sources, calcitic, dolomitic and stromatolitic limestones are important carbonates. The other liming sources are marl, oyster shells and industrial wastes like steel mill slag, blast furnace slag, lime sludge from paper mills, precipitated calcium carbonate, etc. Burnt lime is ordinarily the oxide of lime or quick lime. Quick lime is produced in large kilns. Its reaction in soil is much faster compared to those of carbonates.

$$CaCO_3 + Heat \rightarrow CaO + CO_2\uparrow \quad ...(15)$$

$$[CaMg(CO_3)_2] + Heat \rightarrow CaO + MgO + 2CO_2\uparrow \quad ...(16)$$

Addition of water to burnt lime makes hydroxide or hydrated lime (slaked lime), which is more caustic than burnt lime [Equations (17) and (18)]:

$$CaO + H_2O \rightarrow Ca(OH)_2 \quad ...(17)$$

$$MgO + H_2O \rightarrow Mg(OH)_2 \quad ...(18)$$

The availability of cheap and effective liming materials is crucial for the success of acid soil amelioration programme. The agricultural grade limestone powder and marketable lime are effective, but may not be economical and available in plenty at all the places. A number of industrial by-products like basic slag, lime sludges, phosphogypsum and press mud, etc. are rich sources of calcium and could serve as cheap liming materials in some areas. The availability of basic slag with different steel industries is around 3 million tonnes (Table 4).

8.1. Reactions of Lime with Carbon Dioxide and Soil Colloids

Lime when applied to acidic soils in the form of either oxide, hydroxide or carbonate reacts with carbon dioxide and water to form bicarbonate [Equations (19)-(21)]:

$$CaO + H_2O + 2CO_2 \rightarrow Ca(HCO_3)_2 \quad ...(19)$$

$$Ca(OH)_2 + 2CO_2 \rightarrow Ca(HCO_3)_2 \quad ...(20)$$

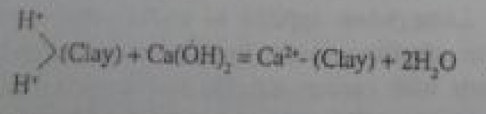
$$CaCO_3 + H_2O + CO_2 \rightarrow Ca(HCO_3)_2 \quad ...(21)$$

Table 3. Classification of acid soils of India and their distribution

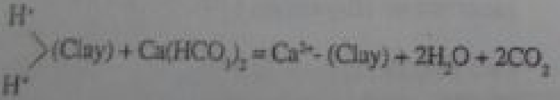
Broad soil group	Type	pH range	Distribution
Laterites	Plinthaquults, Plinthustults and Plinthudults	4.8-7.0	Karnataka, Madhya Pradesh, Orissa, West Bengal, Kerala, Assam, Jharkhand
Lateritic soils	Paleudalfs, Hapludalfs, Hapludults, Trophumults, Dystropepts, Ustropepts, Oxic intergrades	5.0-7.0	Kerala, Orissa, West Bengal, Assam, North-Eastern States, Jharkhand
Mixed red and yellow soils	Dystrochrepts, Udifluvents, Hapludalfs	5.5-6.5	Karnataka, MP, erstwhile UP, Orissa, Chotanagpur and Santhal Praganas of Jharkhand
Ferruginous red soils	Paleudalfs, Udifluvents, Udorthents, Plinthaquepts, Tropaquepts, Ultic intergrades	5.0-6.8	Tamil Nadu, parts of Karnataka, Maharashtra, Andhra Pradesh, Orissa, Jharkhand, Goa
Ferruginous red and gravely soils	Haplustalfs, Rhodustalfs	5.2-6.7	Assam, North-Eastern States, WB, Orissa, erstwhile UP, Andaman Islands
(i) Brown forest soils of Western Himalayas	Hapludolls, Udifluvents, Udorthents, Dystrochrepts, Haplumbrepts	5.5-6.0	Upper regions of Western Himalayas, Jammu and Kashmir, Himachal Pradesh
(ii) Brown forest podzolic soils	Hapludolls, Udorthents, Haplumbrepts, Dystrochrepts	5.1-5.4	Upper regions of Western Himalayas, Jammu and Kashmir and Himachal Pradesh
(iii) Brown forest soils of central Himalayas	Hapludalts, Udorthents, Udifluvents, Dystrochrepts	5.2-5.7	Lower regions of Central Himalayas, erstwhile UP
(iv) Brown forest podzolic soils of central Himalayas	Mollic Hapludalfs, Ultic-Hapludalfs, Haplumbrepts	4.7-6.0	Eastern Himalayas, North-Eastern States of Himalayan region, Arunachal Pradesh
Foot hill soils	Haplaquolls	4.5-6.5	Jammu and Kashmir, West Bengal, erstwhile UP
Peat soils	Histosols	3.5-4.5	Kerala, Assam
Alluvial soils	Haplumbrepts, Udifluvents	4.2-5.8	Assam, West Bengal, Bihar
Coastal alluvial soils	Tropofluvents, Sulfaquepts, Haplaquents, Udifluvents	3.5-5.8	West Bengal (Sunderbans), Kerala, some pockets of Tamil Nadu
Degraded saline and acid saline soils	Acid Fluvaquents	5.2-6.6	West Bengal, Orissa, parts of Tamil Nadu and Kerala

Source: Murthy *et al.* (1976)

These liming materials, on reaction with soil colloid, replace hydrogen and aluminium ions from the colloidal phase to soil solution:



$$\left.\begin{matrix}H^+\\ H^+\end{matrix}\right\rangle(\text{Clay}) + Ca(OH)_2 = Ca^{2+}\text{-}(\text{Clay}) + 2H_2O \quad \ldots(22)$$



$$\left.\begin{matrix}H^+\\ H^+\end{matrix}\right\rangle(\text{Clay}) + Ca(HCO_3)_2 = Ca^{2+}\text{-}(\text{Clay}) + 2H_2O + 2CO_2 \quad \ldots(23)$$

8.2. Lime Requirement

There are four important factors that govern the lime requirement, viz. (a) required change in pH, (b) buffer capacity of the soil to be limed, (c) chemical composition of the liming materials used, and (d) fineness of the liming materials. A fine-textured acid soil requires much larger quantity of lime than does a sandy soil or a loamy soil having the same pH value. Calcitic or dolomitic limestone reacts slowly with soil colloids, whereas burnt lime and hy-

Table 4. Availability of some liming materials in India

Acid soil region/State	Liming material	Estimated quantity available (million tonnes)	Approx. Cost/tonne (Rs)
Assam	Lime stone	15.0	4,000
Himachal Pradesh	Marketable lime	—	4,000
Jharkhand	Lime stone/Basic Slag	1.0	1,000
Kerala	Lime shells	4.0	6,000
Maharashtra	Marketable lime	0.2	2,000
North-Eastern Hill Region	Limestone	14.0	1,000
Orissa	Paper mill sludge	0.2	500
West Bengal	Basic slag	0.3	1,000
Others	Basic Slag*	3.0	1,000

Source: Participatory Resource Appraisal (PRA), ICAR Net Work Project on Acid Soils

*Available with different steel industries in the country.

drated lime react faster and bring about changes in soil pH within a few days.

8.2.1. *Chemical Equivalence of Liming Materials*

One molecule of calcium oxide, magnesium oxide or calcium hydroxide neutralizes the same amount of acidity as does one molecule of calcium carbonate. The $CaCO_3$ equivalent of burnt lime (CaO) can be calculated by the ratio of molecular weights of $CaCO_3$ and CaO [Equation (24)] :

$$\frac{CaCO_3}{CaO} = \frac{100}{56} = 1.786 \qquad \ldots(24)$$

This molecular weight ratio, when multiplied by the weight of pure $CaCO_3$ gives the $CaCO_3$ equivalent; i.e. 100 × 1.786 = 178.6. If the burnt lime has 90% purity, then 100 kg of it will have a $CaCO_3$ equivalent of 90 × 1.786 = 160.7. Likewise, the CaO equivalence of $MgCO_3$ is CaO/ $MgCO_3$ = 56/84 = 0.67, and Mg equivalence of MgO is Mg/MgO = 24/40 = 0.60.

8.2.2. *Fineness of Liming Materials*

Standard sieves having openings of designated size are used to measure fineness of liming materials. A 10-mesh sieve has 10 wires per inch and opening sizes of 2 mm. Similarly, a 60-mesh screen opening is 0.25 mm and 200-mesh screen opening is 0.075 mm. The designated screen size indicates the maximum diameter of particles that can pass along with smaller particles. The coarse particles of liming materials react somewhat slower compared to finer particles in neutralising the soil acidity.

Liming not only causes reduction in soil acidity, it also supplies calcium and magnesium for plant uptake. Frequency of liming varies with climate, soil, and cropping to be undertaken. Soils containing considerable amounts of clay and organic matter need more lime than do sandy and highly-weathered soils of the same pH level. Type of clay is also an important consideration to decide about the dose of lime. A soil having smectitic clay with higher exchange capacity would require more lime than a soil having kaolinite for an equal rise in pH. More lime is needed for a strong acid soil than for weakly acid soils. Liming factor may be defined as the factor by which the actual amount of lime required is calculated from the estimated amount of lime. The former would vary depending on the rate of limestone solubility, plant uptake and leaching during the reaction period. A liming factor of 1.5 to 2.0 is generally used.

8.2.3. *Measurement of Lime Requirement*

Exchangeable Aluminium as Lime Requirement Index — Generally, Al-toxicity is observed in highly-weathered soils of pH below 5.0. Under such condition, plant growth is frequently limited. To manage this, lime application based on the exchangeable Al extracted by neutral unbuffered 1*N* KCl solution is used as an index of lime requirement. Such an index is adopted mostly in soils of low cation-exchange capacity where exchangeable Al equals, and often exceeds the total exchangeable bases. From the values obtained for exchangeable Al, the lime requirement (LR) may be computed by the Equations (25) to (27):

For high Al-tolerant plants:

$$\text{LR} = 1 \times \text{Exchangeable Al} \qquad \ldots(25)$$

For medium Al-tolerant plants:

$$\text{LR} = 1.5 \times \text{Exchangeable Al} \qquad \ldots(26)$$

For low Al-tolerant plants:

$$\text{LR} = 2.0 \times \text{Exchangeable Al} \qquad \ldots(27)$$

where, units for LR in terms of exchangeable Ca and exchangeable Al are $\text{cmol}(p^+)\text{kg}^{-1}$.

For determination of lime requirement, the buffer method of Schoemaker, Mclean and Pratt (SMP) is widely used in India. In this method, 5 mL of distilled water and 10 mL of extractant buffer (1.8 g nitrophenol, 2.5 mL triethanolamine, 3 g potassium chromate, 2 g calcium acetate and 53.2 g calcium chloride dihydrate, are dissolved in one litre of water and the pH of this solution is adjusted to 7.5 with dilute NaOH solution) are added to 5 g of soil. It is then stirred continuously for 10 minutes or intermittently for 20 minutes and the pH of the suspension is determined by a pH meter. The SMP buffer method is based on a generalized relationship between the buffer indicated lime requirement and $CaCO_3$ incubation–measured LR of a group of soils. Complications in the interpretation of the changes in buffer pH, brought about by mixing soil and buffer, arise from the facts that much of the acidity is pH-dependent and only a fraction of the total soil acidity reacts with the buffer. The SMP method has been designed for soils having considerable exchangeable Al and high LR.

$BaCl_2$–Triethanolamine (TEA) Method for Lime Requirement (Mehlich) — In this method, $BaCl_2$–TEA buffer of pH 8.0 is shaken with the soil and the amount of acidity reacting with the buffer is determined by titrating a fresh sample of buffer with acid down to the measured pH of the soil plus buffer. The difference between the two titrations gives a measure from which lime requirement is calculated.

9. Management of Acid Soils

The management of acid soils should aim at improving the production potential of the soil by either addition of amendments or manipulation of agricultural practices to obtain optimum yield under acidic conditions. One of the practices is to grow acid-tolerant plant species and varieties. Application of lime as an amendment to neutralize the exchangeable Al^{3+} to a certain extent is quite effective. Rice has good tolerance to acidity since flooding of rice fields raises the pH to almost neutrality. Short-duration rice grown in the uplands tolerate acidity. Minor millet (*Panicum miliare*) and finger millet (*Eleusine coracana*) are quite tolerant to acidity, showing meagre response to liming. Bengal gram, lentil, groundnut, maize, sorghum and field peas show medium response to liming. Pigeonpea, soyabean and cotton are sensitive to soil acidity and respond well to liming.

Amending acid soil with one time application of full dose lime as per lime requirement should be replaced by small doses of lime application to each crop. Application of 10-20% L.R. dose, mixing with farm yard manure in rows below the seeds or behind the plough at the time of sowing the crop, is quite beneficial. When the soil is moderately acidic, it can be managed with sufficient quantity of organic manure. The leaching loss of basic cations can be minimized by improving the physical conditions of soils through application of organic wastes and lime. Humified organic matter can bind tightly to aluminium ions and prevent them from building toxicity in the soil solution. Microbial decomposition produces low molecular weight organic acids and such materials also come through root exudation. These acids form soluble complexes and can save the plant from toxicity. Organic matter incorporated into soil contains a substantial amount of calcium that leaches and moves down the soil profile and checks aluminium toxicity. By this, calcium and pH levels are raised in both surface and subsoils. Since acid soils have high P-fixing capacity, its inherent acidity and acidity created in the root rhizosphere can dissolve the insoluble rock phosphates and release P. When the pH of soil is less than 5.5, ground rock phosphate as such can be applied. But at pH 5.6 to 6.5, a mixture of rock phosphate and super phosphate in 50:50 ratio is used for economy, since superphosphate will function as a starter.

For a soil with sub-soil acidity, gypsum ($CaSO_4.2H_2O$) or phosphogypsum (a byproduct from phosphate factory) is useful. Calcium from surface applied gypsum moves down the profile much faster than lime. Lime applied to the surface, on dissolution, raises pH and pH-dependent charge on the soil colloids, which hold calcium. Consequently, any anion released (OH^- or CO_3^{2-}) is removed. Gypsum, on the other hand, does not raise the soil pH and CEC. The SO_4^{2-} anion released by the dissolution of gypsum, accompanies calcium in the process of leaching. So gypsum raises the level of calcium and reduces the level of aluminium in both soil solution and exchange complex. Gypsum can be a source of sulphur for surface soil, but it cannot amend the surface acidity.

A large number of experiments on farmers fields (871) have been conducted under ICAR Net Work Project involving different crops and liming materials to evaluate the effect of liming and fertilizer addition on crop yields. Lime @ 2-4 q/ha was applied in furrows along with fertilizers at the time of sowing of crops. Liming alone increased the yields of different crops by about 14-52% over farmers' practice. The increase has been marked on wheat (52%) and mustard (35%) in West Bengal, on groundnut (45%) and pigeonpea (43%) in Orissa, on pigeonpea (34%) and maize (26%) in Jharkhand and on blackgram (26%) in Kerala. The response to the combined application of NPK and lime has been found better than the simple additive effects of fertilizer and lime, indicating a synergy or complementarity between fertilizer-use and liming. The yield of crops with 50% recommended NPK + lime @ 2-4 q/ha has been found equal or slightly higher than the yields obtained with 100% recommended NPK. The liming enhances nutrient-use efficiency, therefore, effects 50% saving of chemical fertilizers (Table 5).

Liming and fertilization have proved to be economical to different crops in acid regions of the country. The benefit – cost ratio increases with (i) liming alone (farmers' practice + lime), (ii) addition of fertilizers (100% NPK), and (iii) conjunctive use of fertilizers and lime (100% NPK + lime). The benefit–

Table 5. Yield of crops in acid soils with recommended fertilizer and half the recommended fertilizer + lime

(q/ha)

State	Crop	100% NPK	50% NPK + lime	Yield change (%)
Assam	Rapessed	9.7	10.1	+4.1
	Summer green gram	4.4	5.2	+18.2
Himachal Pradesh	Maize	34.0	33.1	-2.6
	Wheat	27.9	23.7	-15.0
Jharkhand	Maize + Pigeonpea (Maize equiv. yield)	69.0	65.0	-5.8
	Pea	38.4	50.8	+32.3
Kerala	Cowpea	8.6	10.6	+23.2
	Blackgram	6.4	8.1	+26.6
Meghalaya	Maize	30.5	30.3	-0.7
	Groundnut	14.2	21.3	+50.0
Orissa	Groundnut	22.5	23.6	+4.9
	Pigeonpea	12.0	12.2	+1.7
West Bengal	Mustard	8.2	8.4	+2.4
	Wheat	16.7	17.1	+2.4
All states		21.6	22.8	5.5

8:1

C:N

soil 8:1 < 2/3 respire, 1/3 assimilate

10×2/3 = 20/3 = 6

8:1

(C:1 N)

8×1/3 = 8/3 = 2

100 kg carbon [illegible]

cost ratio has been found to vary from 1.4 to 4.3 with conjunctive use of lime and fertilizers. The net returns are, therefore, Rs 0.4 to Rs 3.3 per rupee invested, depending on the region and crop grown. The application of 50% NPK + lime is more economical than 100% NPK, as has been revealed by higher benefit: cost ratios in the former than the latter.

200 kg

5.0 [illegible]

12 kg [illegible]

10 kg

10. Industrial Wastes as Amendments for Acid Soils

The choice of amendment for acid soils depends highly on the availability of lime source and its cost. Industrial wastes such as steel mill slag, blast furnace slag, lime sludge from paper mills, cement kiln wastes, precipitated calcium carbonate, etc. have been used as amendments. Lime sludge from paper mills contains 65-85% $CaCO_3$, 2% R_2O_3 (sesquioxides), 1% free CaO and 1.5% free alkali. When exposed to rains, the sludge gets freed from alkali. The paper mills located in the acid soil regions of Assam, Nagaland, West Bengal, Orissa, Madhya Pradesh and Andhra Pradesh, produce around 1,87,000 tonnes of lime sludge annually which could be used for amending acid soils. Basic slag is a double silicate and phosphate of lime. It is estimated that for every tonne of hot metal about 500 kg of blast furnace slag is produced. For every tonne of steel, 200-250 kg of steel mill slag is produced. On an average, Indian slag contains 1-7% P_2O_5, 24-50% CaO and 2-10% MgO. Slag being hard is not reactive unless powered to optimum particle size of 60 mesh. Grinding in a ball mill being expensive, ultimate cost becomes high and therefore, slag has not gained popularity among farmers. Moreover, black cement is a potent competitor.

Good grade limestone suitable for industrial use cannot possibly be recommended as agricultural lime. Poor grade lime with comparatively high silica content could be used as an amendment. But, the poor grade limestone quarries are not operative and hence such materials are not marketed. Thus, the only choice is to utilize on the industrial wastes. Since most acid soils are less productive due to low fertility, low water retentivity and poor physico-chemical conditions, the resource-poor farmers of acid soil region are not able to afford commercial limestone. Therefore, it is often said that 'soil acidity and poverty are synonymous'.

11. References

Jackson, M.L. (1973) *Soil Chemical Analysis*. Prentice-Hall of India Pvt. Ltd, New Delhi, p. 40.

Murthy, R.S., Shankaranarayana, H.S. and Hirekerur, L.R. (1976) Distribution, genesis and classification of acid soils of India. *Bulletin of the Indian Society of Soil Science* **11**, 1-17.

Sharma, P.D. and Sarkar, A.K. (2005) Managing acid soils for enhancing productivity. *ICAR Technical Bulletin* p. 1-23.

Sarkar, A.K., Mahapatra, P. and Kumar, A. (2007) *Nutrient Management in Acid Soils* (R.K. Rattan, Ed.) *ISSS Bulletin No.* **25**, pp. 10-27.

1/3 carbon assimilate by micro organism, 2/3 of carbon respired by m.o.

for calculation of N = $\frac{\text{kg carbon}}{8}$ = kg N.

for calculation of C: N × 8 = kg C.