

Karmaveer Kakasaheb Wagh Education Society's
K. K. WAGH COLLEGE OF AGRICULTURE,
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Soil Science And Agriculture
Chemistry Section Theory
Notes of Ist Semester

Course Title:-

Introduction to Soil Science

Course No:- SSAC-111

Course No. : SSAC-111

Course Title : INTRODUCTION TO SOIL SCIENCE (B. SC. (AGRI.))

Course Credit : 2+1=3

Teaching Schedule :

Lecture No.	Topic	Weight age
1 & 2	Soil pedological and edaphological concept. Origin of the earth Earth's crust composition	3
3 & 4	Study of soil forming rocks and minerals	3
5 & 6	Weathering of rocks and minerals	6
7 & 8	Soil forming factors and processes, Components of soils	6
9	Study of soil profile.	4
10 & 11	Soil physical properties: Soil texture, textural classes, particle size analysis	5
12 to 14	Soil structure Classification, soil aggregates, significance of soil consistency, Soil crusting. Bulk density and Particle density. Soil porosity, their significance and manipulation. Soil compaction and soil colour.	6
15, 16 & 17	Soil water: Retention and potentials, soil moisture constants, movement of soil water, infiltration, percolation permeability, Drainage: Methods of determination of soil moisture	6
18	Thermal properties of soils, Soil temperature	3
19	MID TERM	
20 & 21	Soil air: Gaseous exchange. Influence of soil temperature, air on plant growth	5
22 & 23	Soil colloids: Properties, nature, types and significance	5
24 & 25	Layer silicates clays: Genesis and sources of charges	5
26 & 27	Adsorption of ions. Ion exchange. CEC and AEC. Factors influencing ion exchange and its significance	6
28, 29	Soil organic matter: composition, Determination of	6

& 30	organic. Humus. fractionation of organic matter, carbon cycle C:N ratio	
31 to 33	Soil biology: Definition soil Biomass, soil organisms and their beneficial and harmful roles	4
34 & 35	Soil survey and USDA Soil classification. Land Capability classification Soils of India. Soils of Maharashtra.	4
36	Soil erosion. Types, universal soil loss equation & control measures.	3

Text Books :

1. Introductory Soil Science By D.K. Das.
2. Text book of Soil Science by J. A. Daji.
3. Physical properties of soil by C. C. Shah and NK. Narayana (1966)
4. Fundamentals of Soil Science (8th edition) 1990 by Henry. D. Fothk.
5. Text book of Soil Science (Second edition) 1994 by Biswas and Mukharjee
6. Nature and properties of soils (Tenth edition) by N. C. Brady, prentice Hall of India Pvt. Ltd. New Delhi.
7. Fundamentals of Soil Science – A Text Book by V.D. Patil & C.V. Mali
8. Fundamentals of Soil Science by ISSS, New Delhi.

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1 Scope of Soil Science

Soil Science has six well defined and developed disciplines. Scope of soil Science is reflected through these disciplines.

Soil Science : The science dealing with soil as a natural resource on the surface of the earth, including Pedology (soil genesis, classification and mapping) and the physical, chemical and biological and fertility properties of soil and these properties in relation to their management for crop production.

1. **Soil fertility:** Nutrient supplying properties of soil
2. **Soil chemistry:** Chemical constituents, chemical properties and the chemical reactions
3. **Soil physics:** Involves the study of physical properties
4. **Soil microbiology:** deals with micro organisms, its population, classification, its role in transformations
5. **Soil conservation:** Dealing with protection of soil against physical loss by erosion or against chemical deterioration i.e. excessive loss of nutrients either natural or artificial means.
6. **Pedology:** Dealing with the genesis, survey and classification

Soil can be compared to various systems of human body

Digestive - matters decomposition
Respiratory - air circulation & exchange of gases
Circulatory - water movement with in the soil system
Excretory - leaching out of excess salts
Brain - soil clay
Colour - soil colour
Height - soil depth

Components of Soil (Volume basis)

Mineral matter – 45%
Organic matter – 5%
Soil water – 25%
Soil air – 25%

2. Definition of Soil & Approaches of Soil Study

Definition of Soil

Whitney (1882) Hilgard (1892) Dokuchaiev (1900) Joffe (1936): Soil is a natural body of mineral and organic constituents differentiated into horizons usually unconsolidated, of variable depth which differs among themselves as well as from the underlying parent material in morphology, physical makeup, chemical properties and composition and biological characteristics.

SSSA (1970):

(i) The unconsolidated mineral matter on the surface of the earth that has been subjected to and influenced by genetic and environmental factors of parent material, climate (including moisture and temperature effects), macro and microorganisms and topography, all affecting over a period of time and producing a product, that is "SOIL" that differs from the material from which it is derived in many, physical, chemical, biological and morphological properties and characteristics.

(ii) The unconsolidated mineral material on the immediate surface of the earth that serves as a natural medium for the growth of land plants.

Approaches of Soil Study

Two Concepts: One treats soil as a natural body, weathered and synthesized product in nature while other treats soil as a medium for plant growth.

1) Pedological Approach: The origin of the soil, its classification and its description are examined in Pedology. (From Greek word *pedon*, means soil or earth). Pedology is the study of soil as a natural body and does not focus on the soil's immediate practical use. A pedologist studies, examines and classifies soil as they occur in their natural environment.

2) Edaphological Approach: Edaphology (from Greek word *edaphos*, means soil or ground) is the study of soil from the stand point of higher plants. Edaphologists consider the various properties of soil in relation to plant production. They are practical and have the production of food and fiber as their ultimate goal. They must determine the reasons for variation in the productivity of soils and find means for improvement.

3. Weathering of Rocks and Minerals

Rocks and minerals are formed under a very high temperature and pressure, exposed to atmospheric conditions of low pressure and low temperature and they become unstable and weather.

Soils are formed from rocks through the intermediate stage of formation of Regolith which is the resultant of weathering.

The sequence of processes in the formation of soils is:

Weathering of rocks and minerals -> formation of regolith or parent material -> formation of true soil from regolith

Rock ->Weathering ->Regolith ->Soil forming factors and processes ->True soil (otherwise)

Two processes involved in the formation of soil are:

1. Formation of regolith by breaking down (weathering) of the bed rock.
2. The addition of organic matter through the decomposition of plant and animal tissues, and reorganization of these components by soil forming processes to form soil.

Weathering: A process of disintegration and decomposition of rocks and minerals which are brought about by physical agents and chemical processes, leading to the formation of Regolith (unconsolidated residues of the weathering rock on the earth's surface or above the solid rocks).

(OR)

The process by which the earth's crust or lithosphere is broken down by the activities of the atmosphere, with the aid of the hydrosphere and biosphere

(OR)

The process of transformation of solid rocks into parent material or Regolith

Parent material: It is the regolith or at least its upper portion. May be defined as the unconsolidated and more or less chemically weathered mineral materials from which soil are developed

Two basic processes of Weathering:

Physical (or) mechanical - disintegration
Chemical – decomposition

In addition, another process: Biological and all these processes are work hand in hand

Depending up on the agents taking part in weathering processes, it is classified into three types.

Different agents of weathering

Physical/ Mechanical (disintegration)	Chemical (decomposition)	Biological (disint + decomp)
1.Physical condition of rock	1.Hydration	1.Man & animals
2.Change in temperature	2.Hydrolysis	2. higher plants & their roots
3.Action of H ₂ O	3.Solution	3.Micro organisms
-fragment & transport	4.Carbonation	
- action of freezing	5.Oxidation	
- alter. Wet & drying	6.Reduction	
- action of glaciers		
4.Action of wind		
5. Atmosp. electric pheno		

Physical weathering of Rocks

The rocks are disintegrated and are broken down to comparatively smaller pieces, with out producing any new substances

1. Physical condition of rocks: The permeability of rocks is the most important single factor.

1. Coarse textured (porous) sand stone weather more readily than a fine textured (almost solid) basalt.
2. Unconsolidated volcanic ash weather quickly as compared to unconsolidated coarse deposits such as gravels

2. Action of Temperature: The variations in temperature exert great influence on the disintegration of rocks.

1. During day time, the rocks get heated up by the sun and expand. At night, the temperature falls and the rocks get cooled and contract.
2. This alternate expansion and contraction weakens the surface of the rock and crumbles it because the rocks do not conduct heat easily.
3. The minerals with in the rock also vary in their rate of expansion and contraction
4. The cubical expansion of quartz is twice as feldspar

5. Dark coloured rocks are subjected to fast changes in temperature as compared to light coloured rocks
6. The differential expansion of minerals in a rock surface generates stress between the heated surface and cooled unexpanded parts resulting in fragmentation of rocks.
7. This process causes the surface layer to peel off from the parent mass and the rock ultimately disintegrates. This process is called Exfoliation

3. Action of Water: Water acts as a disintegrating, transporting and depositing agent.

i) Fragmentation and transport: Water beats over the surface of the rock when the rain occurs and starts flowing towards the ocean

1. Moving water has the great cutting and carrying force.
2. It forms gullies and ravines and carries with the suspended soil material of variable sizes.
3. Transporting power of water varies. It is estimated that the transporting power of stream varies as the sixth power of its velocity i.e the greater the speed of water, more is the transporting power and carrying capacity

Speed/Sec	Carrying capacity
15 cm	fine sand
30 cm	gravel
1.2 m	stones (1kg)
9.0 m	boulders (several tons)

The disintegration is greater near the source of river than its mouth

ii) Action of freezing: Frost is much more effective than heat in producing physical weathering

1. In cold regions, the water in the cracks and crevices freezes into ice and the volume increases to one tenth
2. As the freezing starts from the top there is no possibility of its upward expansion. Hence, the increase in volume creates enormous out ward pressure which breaks apart the rocks

iii) Alternate wetting and Drying: Some natural substances increase considerably in volume on wetting and shrink on drying. (e.g.) smectite, montmorillonite

1. During dry summer/ dry weather – these clays shrink considerably forming deep cracks or wide cracks.
2. On subsequent wetting, it swells.
3. This alternate swelling and shrinking/ wetting or drying of clay enriched rocks make them loose and eventually breaks

iv) Action of glaciers:

1. In cold regions, when snow falls, it accumulates and changes into ice sheet.
2. These big glaciers start moving owing to the change in temperature and/or gradient.
3. On moving, these exert tremendous pressure over the rock on which they pass and carry the loose materials
4. These materials get deposited on reaching the warmer regions, where its movement stops with the melting of ice

4. Action of wind:

1. Wind has an erosive and transporting effect. Often when the wind is laden with fine material viz., fine sand, silt or clay particles, it has a serious abrasive effect and the sand laden winds itch the rocks and ultimately breaks down under its force
2. The dust storm may transport tons of material from one place to another. The shifting of soil causes serious wind erosion problem and may render cultivated land as degraded (e.g.) Rajasthan deserts

5. Atmospheric electrical phenomenon: It is an important factor causing break down during rainy season and lightning breaks up rocks and or widens cracks.

Chemical Weathering of Rocks

Decomposition of rocks and minerals by various chemical processes is called chemical weathering. It is the most important process for soil formation.

Chemical weathering takes place mainly at the surface of rocks and minerals with disappearance of certain minerals and the formation of secondary products (new materials). This is called chemical transformation.

Feldspar + water \rightarrow clay mineral + soluble cations and anions

Chemical weathering becomes more effective as the surface area of the rock increases.

Since the chemical reactions occur largely on the surface of the rocks, therefore the smaller the fragments, the greater the surface area per unit volume available for reaction.

The effectiveness of chemical weathering is closely related to the mineral composition of rocks. (e.g.) quartz responds far slowly to the chemical attack than olivine or pyroxene.

Average mineralogical composition (%)

Composition	Granite	Basalt	Shale	S. Stone	L. Stone
Feldspar	52.4	46.2	30.0	11.5	-

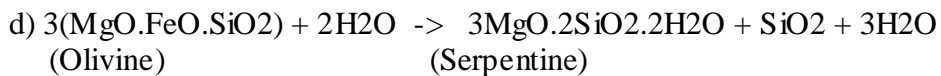
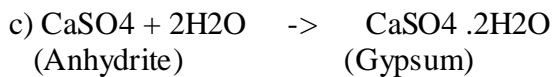
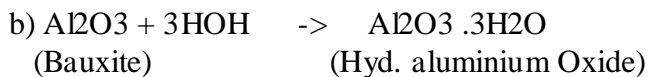
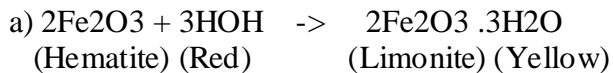
Quartz	31.3	-	2.3	66.8	-
Pyrox - amphi	-	44.5	-	-	-
FeO mineral	2.0	9.3	10.5	2.0	-
Clay mineral	14.3	-	25.0	6.6	24.0
Carbonates	-	-	5.7	11.1	76.0

Chemical Processes of weathering:

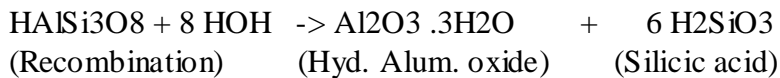
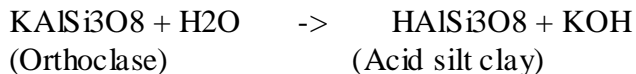
1. Hydration: Chemical combination of water molecules with a particular substance or mineral leading to a change in structure.

Soil forming minerals in rocks do not contain any water and they under go hydration when exposed to humid conditions. Up on hydration there is swelling and increase in volume of minerals. The minerals loose their luster and become soft.

It is one of the most common processes in nature and works with secondary minerals, such as aluminium oxide and iron oxide minerals and gypsum. (e.g.)



2. Hydrolysis: Most important process in chemical weathering. It is due to the dissociation of H₂O into H⁺ and OH⁻ ions which chemically combine with minerals and bring about changes, such as exchange, decomposition of crystalline structure and formation of new compounds. Water acts as a weak acid on silicate minerals.



This reaction is important because of two reasons.

1. clay, bases and Silicic acid - the substances formed in these reactions - are available to plants
2. water often containing CO₂ (absorbed from atmosphere), reacts with the minerals directly to produce insoluble clay minerals, positively charged metal ions (Ca⁺⁺,

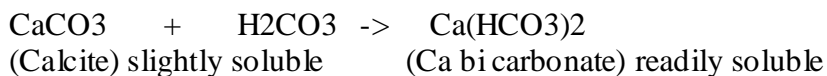
Mg⁺⁺, Na⁺, K⁺) and negatively charged ions (OH⁻, HCO₃⁻) and some soluble silica – all these ions are made available for plant growth.

3. Solution: Some substances present in the rocks are directly soluble in water. The soluble substances are removed by the continuous action of water and the rock no longer remains solid and form holes, rills or rough surface and ultimately falls into pieces or decomposes. The action is considerably increased when the water is acidified by the dissolution of organic and inorganic acids. (e.g) halites, NaCl
NaCl + H₂O → Na⁺, Cl⁻, H₂O (dissolved ions with water)

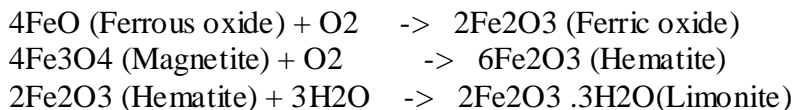
4. Carbonation: Carbon dioxide when dissolved in water it forms carbonic acid.



This carbonic acid attacks many rocks and minerals and brings them into solution. The carbonated water has an etching effect up on some rocks, especially lime stone. The removal of cement that holds sand particles together leads to their disintegration.



5. Oxidation: The process of addition and combination of oxygen to minerals. The absorption is usually from O₂ dissolved in soil water and that present in atmosphere. The oxidation is more active in the presence of moisture and results in hydrated oxides.(e.g) minerals containing Fe and Mg.



6. Reduction: The process of removal of oxygen and is the reverse of oxidation and is equally important in changing soil colour to grey, blue or green as ferric iron is converted to ferrous iron compounds. Under the conditions of excess water or water logged condition (less or no oxygen), reduction takes place.



In conclusion, during chemical weathering igneous and metamorphic rocks can be regarded as involving destruction of primary minerals and the production of secondary minerals.

In sedimentary rocks, which is made up of primary and secondary minerals, weathering acts initially to destroy any relatively weak bonding agents (FeO) and the particles are freed and can be individually subjected to weathering.

Biological Weathering of Rocks

Unlike physical and chemical weathering, the biological or living agents are responsible for both decomposition and disintegration of rocks and minerals. The biological life is

mainly controlled largely by the prevailing environment.

1. Man and Animals:

1. The action of man in disintegration of rocks is well known as he cuts rocks to build dams, channels and construct roads and buildings. All these activities result in increasing the surface area of the rocks for attack of chemical agents and accelerate the process of rock decomposition.
2. A large number of animals, birds, insects and worms, by their activities they make holes in them and thus aids for weathering.
3. In tropical and sub tropical regions, ants and termites build galleries and passages and carry materials from lower to upper surface and excrete acids. The oxygen and water with many dissolved substances, reach every part of the rock through the cracks, holes and galleries, and thus brings about speedy disintegration.
4. Rabbits, by burrowing in to the ground, destroy soft rocks. Moles, ants and bodies of the dead animals, provides substances which react with minerals and aid in decaying process
5. The earthworms pass the soil through the alimentary canal and thus bring about physical and chemical changes in soil material.

2. Higher Plants and Roots: The roots of trees and other plants penetrates into the joints and crevices of the rocks. As they grew, they exert a great disruptive force and the hard rock may break apart. (e.g.) pipal tree growing on walls/ rocks

The grass root form a sponge like mass prevents erosion and conserve moisture and thus allowing moisture and air to enter in to the rock for further action.

Some roots penetrate deep into the soil and may open some sort of drainage channel. The roots running in crevices in lime stone and marble produces acids. These acids have a solvent action on carbonates.

The dead roots and plant residues decompose and produce carbon dioxide which is of great importance in weathering.

3. Micro- organisms: In early stages of mineral decomposition and soil formation, the lower forms of plants and animals like, mosses, bacteria and fungi and actinomycetes play an important role. They extract nutrients from the rock and N from air and live with a small quantity of water. In due course of time, the soil develops under the cluster of these micro-organisms.

This organism closely associated with the decay of plant and animal remains and thus liberates nutrients for the use of next generation plants and also produces CO₂ and organic compounds which aid in mineral decomposition.

4. Soil Forming Factors

The soil formation is the process of two consecutive stages.

1. The weathering of rock (R) into Regolith
2. The formation of true soil from Regolith

The evolution of true soil from regolith takes place by the combined action of soil forming factors and processes.

1. The first step is accomplished by weathering (disintegration & decomposition)
2. The second step is associated with the action of Soil Forming Factors

Weathering Factors

Dokuchaiev (1889) established that the soils develop as a result of the action of soil forming factors

$$S = f (P, Cl, O)$$

Further, Jenny (1941) formulated the following equation

$$S = f(Cl, O, R, P, T, \dots)$$

Where,

Cl – environmental climate

o – Organisms and vegetation (biosphere)

r – Relief or topography

p – Parent material

t- Time

... - additional unspecified factors

The five soil forming factors, acting simultaneously at any point on the surface of the earth, to produce soil

Two groups – Passive i) Parent material, ii) Relief, iii) Time

Active IV) Climate, v) Vegetation & organism

Passive Soil forming factors

The passive soil forming factors are those which represent the source of soil forming mass and conditions affecting it. These provide a base on which the active soil forming factors work or act for the development of soil.

Parent Material: It is that mass (consolidated material) from which the soil has formed.

Two groups of parent material

i) **Sedentary:** Formed in original place. It is the residual parent material. The parent material differ as widely as the rocks

ii) **Transported:** The parent material transported from their place of origin. They are named according to the main force responsible for the transport and redeposition.

a) By gravity - Colluvial

b) By water - Alluvial, Marine, Locustrine

c) By ice - Glacial

d) By wind – Eolian

Colluvium: It is the poorly sorted materials near the base of strong slopes transported by the action of gravity.

Alluvium: The material transported and deposited by water is, found along major stream courses at the bottom of slopes of mountains and along small streams flowing out of drainage basins.

Locustrine: Consists of materials that have settled out of the quiet water of lakes.

Moraine: Consists of all the materials picked up, mixed, disintegrated, transported and deposited through the action of glacial ice or of water resulting primarily from melting of glaciers.

Loess or Aeolian: These are the wind blown materials.

When the texture is silty - loess; when it is sand - Eolian.

The soils developed on such transported parent materials bear the name of the parent material; viz. Alluvial soils from alluvium, Colluvial soils from Colluvium etc. In the initial stages, however, the soil properties are mainly determined by the kind of parent material.

Endodynamomorphic soils: With advanced development and excessive leaching, the influence of parent material on soil characteristics gradually diminishes. There are soils wherein the composition of parent material subdues the effects of climate and vegetation. These soils are temporary and persist only until the chemical decomposition becomes active under the influence of climate and vegetation.

Ectodynamomorphic soils: Development of normal profile under the influence of climate and vegetation.

Soil properties as influenced by parent material: Different parent materials affect profile development and produce different soils, especially in the initial stages.

1. Acid igneous rocks (like granite, rhyolite) produce light-textured soils (Alfisols).
2. Basic igneous rocks (basalt), alluvium or Colluvium derived from limestone or basalt, produce fine-textured cracking-clay soils (Vertisols).
3. Basic alluvium or Aeolian materials produce fine to coarse-textured soils (Entisols or Inceptisols).
4. The nature of the elements released during the decaying of rocks has a specific role in soil formation. (e.g.) Si and Al form the skeleton for the production of secondary clay minerals.
5. Iron and manganese are important for imparting red colour to soils and for oxidation and reduction phenomena.
6. Sodium and potassium are important dispersing agents for clay and humus colloids.
7. Calcium and magnesium have a flocculating effect and result in favorable and stable soil structure for plant growth.

2. Relief or Topography: The relief and topography sometimes are used as synonymous terms. They denote the configuration of the land surface. The topography refers to the differences in elevation of the land surface on a broad scale.

The prominent types of topography designations, as given in FAO Guidelines (1990) are:

Sr. No	Land surface	with slopes of
1	Flat to Almost flat	0 – 2 %
2	Gently undulating	2 - 5 %
3	Undulating	5 – 10 %
4	Rolling	10 – 15 %
5	Hilly	15 – 30 %
6	Steeply dissect	>30% with moderate range of elevation(<300 m)
7	Mountainous	> 30% with great range of elevation(>300 m)

Soil formation on flat to almost flat position: On level topographic positions, almost the entire water received through rain percolates through the soil. Under such conditions, the soils formed may be considered as representative of the regional climate. They have normal solum with distinct horizons. But vast and monotonous level land with little gradient often has impaired drainage conditions.

Soil formation on undulating topography: The soils on steep slopes are generally shallow, stony and have weakly- developed profiles with less distinct horizonation. It is due to accelerated erosion, which removes surface material before it has the time to develop. Reduced percolation of water through soil is because of surface runoff, and lack of water for the growth of plants, which are responsible for checking of erosion and promote soil formation.

Soil formation in depression: The depression areas in semi-arid and sub humid regions

reflect more moist conditions than actually observed on level topographic positions due to the additional water received as runoff. Such conditions (as in the Tarai region of the Uttar Pradesh) favour more vegetative growth and slower rate of decay of organic remains. This results in the formation of comparatively dark- coloured soils rich in organic matter (Mollisols).

Soil formation and Exposure/ Aspect: Topography affects soil formation by affecting temperature and vegetative growth through slope exposures (aspect}. The southern exposures (facing the sun) are warmer and subject to marked fluctuations in temperature and moisture. The northern exposures, on the other hand are cooler and more humid. The eastern and western exposures occupy intermediate position in this respect.

3. Time: Soil formation is a very slow process requiring thousands of years to develop a mature pedon. The period taken by a given soil from the stage of weathered rock (i.e. regolith) up to the stage of maturity is considered as time. The matured soils mean the soils with fully developed horizons (A, B, C). It takes hundreds of years to develop an inch of soil. The time that nature devotes to the formation of soils is termed as Pedological Time.

It has been observed that rocks and minerals disintegrate and/or decompose at different rates; the coarse particles of limestone are more resistant to disintegration than those of sandstone. However, in general, limestone decomposes more readily than sandstone (by chemical weathering).

Weathering stages in soil formation		
Sr. No	Stages	Characteristic
1	Initial	Un weathered parent material
2	Juvenile	Weathering started but much of the original material still un weathered
3	Virile	Easily weather able minerals fairly decomposed; clay content increased, slowly weather able minerals still appreciable
4	Senile	Decomposition reaches at a final stage; only most resistant minerals survive
5	Final	Soil development completed under prevailing environments

1. The soil properties also change with time, for instance nitrogen and organic matter contents increase with time provided the soil temperature is not high.
2. CaCO₃ content may decrease or even lost with time provided the climatic conditions are not arid
3. In humid regions, the H⁺ concentration increases with time because of chemical weathering.

Active Soil Forming Factors

The active soil forming factors are those which supply energy that acts on the mass for the purpose of soil formation. These factors are climate and vegetation (biosphere).

1. Climate: Climate is the most significant factor controlling the type and rate of soil formation. The dominant climates recognized are:

1. Arid climate: The precipitation here is far less than the water-need. Hence the soils remain dry for most of the time in a year.
2. Humid climate: The precipitation here is much more than the water need. The excess water results in leaching of salt and bases followed by translocation of clay colloids.
3. Oceanic climate: Moderate seasonal variation of rainfall and temperature.
4. Mediterranean climate: The moderate precipitation. Winters and summers are dry and hot.
5. Continental climate: Warm summers and extremely cool or cold winters.
6. Temperate climate: Cold humid conditions with warm summers.
7. Tropical and subtropical climate: Warm to hot humid with isothermal conditions in the tropical zone.

Climate affects the soil formation directly and indirectly.

Directly, climate affects the soil formation by supplying water and heat to react with parent material.

Indirectly, it determines the fauna and flora activities which furnish a source of energy in the form of organic matter. This energy acts on the rocks and minerals in the form of acids, and salts are released. The indirect effects of climate on soil formation are most clearly seen in the relationship of soils to vegetation.

Precipitation and temperature are the two major climatic elements which contribute most to soil formation.

Precipitation: Precipitation is the most important among the climatic factors. As it percolates and moves from one part of the parent material to another. It carries with it substances in solution as well as in suspension. The substances so carried are re deposited in another part or completely removed from the material through percolation when the soil moisture at the surface evaporates causing an upward movement of water. The soluble substances move with it and are translocated to the upper layer. Thus rainfall brings about a redistribution of substances both soluble as well as in suspension in soil body.

Temperature:

1. Temperature is another climatic agent influencing the process of soil formation.

2. High temperature hinders the process of leaching and causes an upward movement of soluble salts.
3. High temperature favors rapid decomposition of organic matter and increase microbial activities in soil while low temperatures induce leaching by reducing evaporation and there by favour the accumulation of organic matter by slowing down the process of decomposition. Temperature thus controls the rate of chemical and biological reactions taking place in the parent material.

Hanns Jenney (1941) computed that in the tropical regions the rate of weathering proceeds three times faster than in temperate regions and nine times faster than in arctic.

2. Organism & Vegetation

Organism:

1. The active components of soil ecosystem are plants, animals, microorganisms and man.
2. The role of microorganisms in soil formation is related to the humification and mineralization of vegetation
3. The action of animals especially burrowing animals to dig and mix-up the soil mass and thus disturb the parent material
4. Man influences the soil formation through his manipulation of natural vegetation, agricultural practices etc.
5. Compaction by traffic of man and animals decrease the rate of water infiltration into the soil and thereby increase the rate of runoff and erosion.

Vegetation:

1. The roots of the plants penetrate into the parent material and act both mechanically and chemically.
2. They facilitate percolation and drainage and bring about greater dissolution of minerals through the action of CO₂ and acidic substances secreted by them.
3. The decomposition and humification of the materials further adds to the solubilization of minerals
4. Forests – reduces temperature, increases humidity, reduce evaporation and increases precipitation.
5. Grasses reduce runoff and result greater penetration of water in to the parent material.

Fundamental Soil forming Processes

Humification: Humification is the process of transformation of raw organic matter into humus. It is extremely a complex process involving various organisms.

First, simple compounds such as sugars and starches are attacked followed by proteins and cellulose and finally very resistant compounds, such as tannins, are decomposed and the dark colored substance, known as humus, is formed.

Eluviation: It is the mobilization and translocation of certain constituent's viz. Clay, Fe₂O₃, Al₂O₃, SiO₂, humus, CaCO₃, other salts etc. from one point of soil body to another. Eluviation means washing out. It is the process of removal of constituents in suspension or solution by the percolating water from the upper to lower layers. The eluviation encompasses mobilization and translocation of mobile constituents resulting in textural differences. The horizon formed by the process of eluviation is termed as eluvial horizon (A₂ or E horizon).

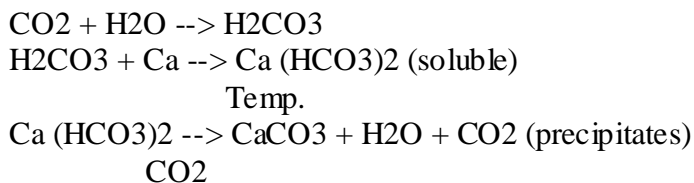
Translocation depends upon relative mobility of elements and depth of percolation.

Illuviation: The process of deposition of soil materials (removed from the eluvial horizon) in the lower layer (or horizon of gains having the property of stabilizing translocated clay materials) is termed as Illuviation. The horizons formed by this process are termed as illuvial horizons (B-horizons, especially B_t) The process leads to textural contrast between E and B_t horizons, and higher fine: total clay ratio in the B_t horizon. **Horizonation:** It is the process of differentiation of soil in different horizons along the depth of the soil body. The differentiation is due to the fundamental processes, humification, eluviation and Illuviation.

Specific Soil Forming Processes

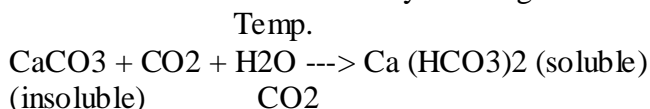
The basic pedologic processes provide a framework for later operation of more specific processes

Calcification: It is the process of precipitation and accumulation of calcium carbonate (CaCO₃) in some part of the profile. The accumulation of CaCO₃ may result in the development of a calcic horizon. Calcium is readily soluble in acid soil water and/or when CO₂ concentration is high in root zone as:



The process of precipitation after mobilization under these conditions is called calcification and the resulting illuviated horizon of carbonates is designated as B_k horizon (B_{ca}).

Decalcification: It is the reverse of calcification that is the process of removal of CaCO₃ or calcium ions from the soil by leaching



Podzolization: It is a process of soil formation resulting in the formation of Podzols and Podzolic soils. In many respects, podzolization is the negative of calcification. The calcification process tends to concentrate calcium in the lower part of the B horizon, whereas podzolization leaches the entire solum of calcium carbonates.

Apart from calcium, the other bases are also removed and the whole soil becomes distinctly acidic. In fact, the process is essentially one of acid leaching. The process operates under favorable combination of the following environments.

- i) **Climate:** A cold and humid climate is most favorable for podzolization
- ii) **Parent material:** Siliceous (Sandy) material, having poor reserves of weather able minerals, favor the operation of podzolization as it helps in easy percolation of water.
- iii) **Vegetation:** Acid producing vegetation such as coniferous pines is essential
- iv) **Leaching and Translocation of Sesquioxide:** In the process of decomposition of organic matter various organic acids are produced. The organic acids thus formed act with Sesquioxide and the remaining clay minerals, forming organic- Sesquioxide and organic clay complexes, which are soluble and move with the percolating water to the lower horizons (Bh, Bs). Aluminium ions in a water solution hydrolyze and make the soil solution very acidic.



As iron and aluminium move about, the A horizon gives a bleached grey or ashy appearance. The Russians used the term Podzols (pod means under, the Zola means ash like i.e. ash-like horizon appearing beneath the surface horizon) for such soils.

To conclude, the Podzolization is a soil forming process which prevails in a cold and humid climate where coniferous and acid forming vegetations dominate. The humus and Sesquioxide become mobile and leached out from the upper horizons and deposited in the lower horizon.

4. Laterization: The term laterite is derived from the word later meaning brick or tile and was originally applied to a group of high clay Indian soils found in Malabar hills of Kerala, Tamil Nadu, Karnataka and Maharashtra.

It refers specifically to a particular cemented horizon in certain soils which when dried, become very hard, like a brick. Such soils (in tropics) when massively impregnated with Sesquioxide (iron and aluminium oxides) to extent of 70 to 80 per cent of the total mass, are called laterite or latosols (Oxisols). The soil forming process is called Laterization or Latozation.

Laterization is the process that removes silica, instead of sesquioxides from the upper layers and thereby leaving sesquioxides to concentrate in the solum. The process operates under the following conditions.

- i) **Climate:** Unlike podzolization, the process of laterization operates most favorable in warm and humid (tropical) climate with 2000 to 2500 mm rainfall and continuous high temperature (25°C) throughout the year.
- ii) **Natural vegetation:** The rain forests of tropical areas are favorable for the process.
- iii) **Parent Material:** Basic parent materials, having sufficient iron bearing ferromagnesian minerals (Pyroxene, amphiboles, biotite and chlorite), which on weathering release iron,

are congenial for the development of laterites.

5. Gleization: The term *glei* is of Russian origin means blue, grey or green clay. The Gleization is a process of soil formation resulting in the development of a glei (or gley horizon) in the lower part of the soil profile above the parent material due to poor drainage condition (lack of oxygen) and where waterlogged conditions prevail. Such soils are called hydro or phic soils.

The process is not particularly dependent on climate (high rainfall as in humid regions) but often on drainage conditions.

The poor drainage conditions result from:

1. Lower topographic position, such as depression land, where water stands continuously at or close to the surface.
2. Impervious soil parent material, and.
3. Lack of aeration.

Under such conditions, iron compounds are reduced to soluble ferrous forms. The reduction of iron is primarily biological and requires both organic matter and microorganisms capable of respiring anaerobically. The solubility of Ca, Mg, Fe, and Mn is increased and most of the iron exists as Fe^{++} organo complexes in solution or as mixed precipitate of ferric and ferrous hydroxides.

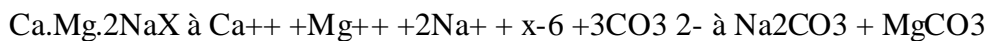
This is responsible for the production of typical bluish to grayish horizon with mottling of yellow and or reddish brown colors.

6. Salinization: It is the process of accumulation of salts, such as sulphates and chlorides of calcium, magnesium, sodium and potassium, in soils in the form of a salty (salic) horizon. It is quite common in arid and semi arid regions. It may also take place through capillary rise of saline ground water and by inundation with seawater in marine and coastal soils. Salt accumulation may also result from irrigation or seepage in areas of impeded drainage.

7. Desalinization: It is the removal by leaching of excess soluble salts from horizons or soil profile (that contained enough soluble salts to impair the plant growth) by ponding water and improving the drainage conditions by installing artificial drainage network.

8. Solonization or Alkalization: The process involves the accumulation of sodium ions on the exchange complex of the clay, resulting in the formation of sodic soils (Solonetz). All cations in solution are engaged in a reversible reaction with the exchange sites on the clay and organic matter particles.

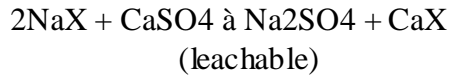
The reaction can be represented as:



+CaCO₃

(Where X represents clay or organic matter exchange sites)

9. Solodization or dealkalization: The process refers to the removal of Na⁺ from the exchange sites. This process involves dispersion of clay. Dispersion occurs when Na⁺ ions become hydrated. Much of the dispersion can be eliminated if Ca⁺⁺ and or Mg⁺⁺ ions are concentrated in the water, which is used to leach the soil. These Ca and Mg ion can replace the Na on exchange complex, and the salts of sodium are leached out as:



10. Pedoturbation: Another process that may be operative in soils is pedoturbation. It is the process of mixing of the soil. Mixing to a certain extent takes place in all soils. The most common types of pedoturbation are:

1. Faunal pedoturbation: It is the mixing of soil by animals such as ants, earthworms, moles, rodents, and man himself
2. Floral pedoturbation : It is the mixing of soil by plants as in tree tipping that forms pits and mounds
3. Argillic pedoturbation: It is the mixing of materials in the solum by the churning process caused by swell shrink clays as observed in deep Black Cotton Soils.

5. Physical Properties of Soil

Physical properties (mechanical behaviour) of a soil greatly influence its use and behaviour towards plant growth. The plant support, root penetration, drainage, aeration, retention of moisture, and plant nutrients are linked with the physical condition of the soil. Physical properties also influence the chemical and biological behaviour of soil. The physical properties of a soil depend on the amount, size, shape, arrangement and mineral composition of its particles. These properties also depend on organic matter content and pore spaces

Important physical properties of soils

1. Soil texture
2. Soil structure
3. Surface area
4. Soil density,
5. Soil porosity
6. Soil colour
7. Soil consistence

Soil Texture

Definition of Soil Texture: Soil texture refers to the relative proportion of particles or it is the relative percentage by weight of the three soil separates viz., sand, silt and clay or simply refers to the size of soil particles.

The proportion of each size group in a given soil (the texture) can not be easily altered and it is considered as a basic property of a soil.

The soil separates are defined in terms of diameter in millimeters of the particles. Soil particles less than 2 mm in diameter are excluded from soil textural determinations.

Stones and gravels may influence the use and management of land because of tillage difficulties but these larger particles make little or no contribution to soil properties such as WHC and capacity to store plant nutrients and their supply.

Gravels: 2 - 4 mm

Pebbles: 4 - 64 mm

Cobbles: 64 - 256 mm

Boulders: > 256 mm

Particles less than 2 mm are called fine earth, normally considered in chemical and mechanical analysis.

The components of fine earth: Sand, Silt and Clay (Soil separates. The size limits of these fractions have been established by various organizations. There are a number of systems of naming soil separates.

1. (a) The American system developed by USDA
2. (b) The English system or British system (BSI)
3. (c) The International system (ISSS)
4. (d) European system

i) USDA

Soil separates	Diameter in mm
Clay	< 0.002
Silt	0.002 - 0.05
Very Fine Sand	0.05 - 0.10
Fine Sand	0.10 - 0.25
Medium Sand	0.25 - 0.50
Coarse Sand	0.50 - 1
Very Coarse Sand	1 - 2 mm

ii) BSI

Soil separates	Diameter in mm
Clay	< 0.002
Fine Silt	0.002 - 0.01
Medium Silt	0.01 - 0.04
Coarse Silt	0.04 - 0.06
Fine Sand	0.06 - 0.20
Medium Sand	0.20 - 1
Coarse Sand	1 - 2 mm

iii) ISSS

Soil separates	Diameter in mm
Clay	< 0.002
Silt	0.002 - 0.02
Fine Sand	0.02 - 0.2
Coarse Sand	0.2 - 2

iv) European System

Soil separates	Diameter (mm)
Fine clay	< 0.0002 mm

Medium clay	0.0002 – 0.0006
Coarse clay	0.0006 – 0.002
Fine silt	0.002 - 0.006
Medium silt	0.006 - 0.02
Coarse silt	0.02 - 0.06
Fine sand	0.06 - 0.20
Medium sand	0.20 - 0.60
Coarse sand	0. 60 - 2.00

Sand:

1. Usually consists of quartz but may also contain fragments of feldspar, mica and occasionally heavy minerals viz., zircon, Tourmaline and hornblende.
2. Has uniform dimensions
3. Can be represented as spherical
4. Not necessarily smooth and has jagged surface

Silt:

1. Particle size intermediate between sand and clay
2. Since the size is smaller, the surface area is more
3. Coated with clay
4. Has the physico- chemical properties as that of clay to a limited extent
5. Sand and Silt forms the SKELETON

Clay:

1. Particle size less than 0.002 mm
2. Plate like or needle like in shape
3. Belong to alumino silicate group of minerals
4. Some times considerable concentration of fine particles which does not belong to alumino silicates. (e.g.) iron oxide and CaCO₃
5. These are secondary minerals derived from primary minerals in the rock
6. Flesh of the soil

Knowledge on Texture is important. It is a guide to the value of the land. Land use capability and methods of soil management depends on texture.

Particle size distribution/ determination

The determination of relative distribution of the ultimate or individual soil particles below 2 mm diameter is called as Particle size analysis or Mechanical analysis

Two steps are involved

- i) Separation of all the particles from each other ie. Complete dispersion into ultimate particles

ii) Measuring the amount of each group

Separation

Sr. No.	Aggregating agents	Dispersion method
1	Lime and Oxides of Fe & Al	Dissolving in HCl
2	Organic matter	Oxidises with H ₂ O ₂
3	High conc. of electrolytes (soluble salts)	Precipitate and decant or filter with suction
4	Surface tension	Elimination of air by stirring with water or boiling

After removing the cementing agents, disperse by adding NaOH

Measurement

Once the soil particles are dispersed into ultimate particles, measurement can be done

i) Coarser fractions - sieving - sieves used in the mechanical analysis corresponds to the desired particle size separation for 2 mm, 1 mm and 0.5 mm – sieves with circular holes, for smaller sizes, wire mesh screens are used (screening)

ii) Finer fractions - by settling in a medium the settling or the velocity of the fall of particles is influenced by viscosity of the medium. Difference in density between the medium and falling particles, size and shape of object

Stokes' Law:

Particle size analysis is based on a simple principle i.e. "when soil particles are suspended in water they tend to sink. Because there is little variation in the density of most soil particles, their velocity (V) of settling is proportional to the square of the radius 'r' of each particle.

Thus $V = kr^2$, where k is a constant. This equation is referred to as Stokes' law.

Stokes (1851) was the first to suggest the relationship between the radius of the particles and its rate of fall in a liquid. He stated that "the velocity of a falling particle is proportional to the square of the radius and not to its surface. The relation between the diameter of a particle and its settling velocity is governed by Stokes' Law:

$$V = \frac{2}{9} \frac{g r^2 (d_s - d_w)}{n}$$

Where,

V - Velocity of settling particle (cm/sec.)

g - Acceleration due to gravity cm/ sec² (981)

d_s - Density of soil particle (2.65)

d_w - Density of water (1)

n - Coefficient of viscosity of water (0.0015 at 40°C)

r - Radius of spherical particles (cm).

Assumptions and Limitations of Stokes' Law

1) Particles are rigid and spherical / smooth. This requirement is very difficult to fulfill, because the particles are not completely smooth over the surface and spherical. It is established that the particles are not spherical and irregularly shaped such as plate and other shapes.

2) The particles are large in comparison with the molecules of the liquid so that in comparison with the particle the medium can be considered as homogenous i.e. the particles must be big enough to avoid Brownian movement. The particles less than 0.0002 mm exhibit this movement so that the rate of falling is varied.

3) The fall of the particles is not hindered or affected by the proximity (very near) of the wall of the vessel or of the adjacent particles. Many fast falling particles may drag finer particles down along with them.

4) The density of the particles and water and as well as the viscosity of the medium remain constant. But this is usually not so because of their different chemical and mineralogical composition.

5) The suspension must be still. Any movement in the suspension will alter the velocity of fall and such movement is brought by the sedimentation of larger particles (> 0.08 mm). They settle so fast and create turbulence in the medium.

The temperature should be kept constant so that convection currents are not set up.

Methods of Textural determination

Numerous methods for lab and field use have been developed

i) Elutriation method – Water & Air

ii) Pipette method

iii) Decantation/ beaker method

iv) Test tube shaking method

v) Feel method - Applicable to the field - quick method - by feeling the soil between thumb and fingers

Feel Method

Evaluated by attempting to squeeze the moistened soil into a thin ribbon as it is pressed with rolling motion between thumb and pre finger or alternately to roll the soil into a thin wire

η four aspects to be seen - i) Feel by fingers, ii) Ball formation, iii) Stickiness and iv) Ribbon formation

Soil Textural Classes

To convey an idea of the textural make up of soils and to give an indication of their physical properties, soil textural class names are used. These are grouped into three main fractions viz., Sand, Silt and Clay.

According to the proportion of these three fractions a soil is given a name to indicate its textural composition. Such a name gives an idea not only of the textural composition of a soil but also of its various properties in general.

On this basis soils are classified into various textural classes like sands clays, silts, loams etc

Sands: The sand group includes all soils in which the sand separates make up at least 70% and the clay separate 15% or less of the material by weight. The properties of such soils are therefore characteristically those of sand in contrast to the stickier nature of clays. Two specific textural classes are recognized in this group sandy and loamy sand although in practice two subclasses are also used Loamy fine sand and loamy very fine sand.

Silt: The silt group includes soils with at least 80% silt and 12% or less clay. Naturally the properties of this group are dominated by those of silt. Only one textural class - Silt is included in this group.

Clays: To be designated clay a soil must contain at least 35% of the clay separate and in most cases not less than 40%. In such soils the characteristics of the clay separates are distinctly dominant, and the class names are clay, sandy clay and silty clay. Sandy clays may contain more sand than clay. Likewise, the silt content of silty clays usually exceeds clay fraction.

Loams: The loam group, which contains many subdivisions, is a more complicated soil textural class. An ideal loam may be defined as a mixture of sand, silt and clay particles that exhibits the properties of those separates in about equal proportions. Loam soils do not exhibit dominant physical properties of sand, silt or clay. Loam does not contain equal percentage of sand, silt and clay. However, exhibit approximately equal properties of sand, silt and clay.

Determination of Textural Class: In the American system as developed by the United State Department of Agriculture twelve textural classes are proposed.

The textural triangle: It is used to determine the soil textural name after the percentages of sand, silt, and clay are determined from a laboratory analysis. Since the soil's textural classification includes only mineral particles and those of less than 2mm diameter, the sand plus silt plus clay percentages equal 100 percent. (Note that organic matter is not included.) Knowing the amount of any two fractions automatically fixes the percentage of the third one.

To use the diagram, locate the percentage of clay first and project inward parallel to sand line. Do likewise for the per cent silt and project inward parallel to clay line and for sand,

project inward parallel to silt. The point at which the projections cross or intersect will identify the class name.

Some times, the intersecting point exactly falls on the line between the textural classes. Then it is customary to use the name of the finer fraction when it happens. (e.g). Soil containing 40% clay, 30% sand and 30% silt - called as clay rather than clay loam.

Importance of Soil Texture

Presence of each type of soil particles makes its contribution to the nature and properties of soil as a whole

- Texture has good effect on management and productivity of soil. Sandy soils are of open character usually loose and friable.
- Such type of the texture is easy to handle in tillage operations.
- Sand facilitates drainage and aeration. It allows rapid evaporation and percolation.
- Sandy soils have very little water holding capacity. Such soils can not stand drought and unsuitable for dry farming.
- Sandy soils are poor store house of plant nutrients
- Contain low organic matter
- Leaching of applied nutrients is very high.
- In sandy soil, few crops can be grown such as potato, groundnut and cucumbers.
- Clay particles play a very important role in soil fertility.
- Clayey soils are difficult to till and require much skill in handling. When moist clayey soils are exceedingly sticky and when dry, become very hard and difficult to break.
- They have fine pores, and are poor in drainage and aeration.
- They have a high water holding capacity and poor percolation, which usually results in water logging.
- They are generally very fertile soils, in respect of plant nutrient content. Rice, jute, sugarcane can be grown very successfully in these soils.
- Loam and Silt loam soils are highly desirable for cultivation
- Generally, the best agriculture soils are those contain 10 – 20 per cent clay, 5 – 10 per cent organic matter and the rest equally shared by silt and sand

Soil Structure

Soil conditions and characteristics such as water movement, heat transfer, aeration, and porosity are much influenced by structure. In fact, the important physical changes imposed by the farmer in ploughing, cultivating, draining, liming, and manuring his land are structural rather than textural.

Definition of Soil Structure: The arrangement and organization of primary and secondary particles in a soil mass is known as soil structure.

Soil structure controls the amount of water and air present in soil. Plant roots and germinating seeds require sufficient air and oxygen for respiration.

Bacterial activities also depend upon the supply of water and air in the soil.

Formation of soil structure: Soil particles may be present either as single individual grains or as aggregate i.e. group of particles bound together into granules or compound particles. These granules or compound particles are known as secondary particles. A majority of particles in a sandy or silty soil are present as single individual grains while in clayey soil they are present in granulated condition. The individual particles are usually solid, while the aggregates are not solid but they possess a porous or spongy character. Most soils are mixture of single grain and compound particle. Soils, which predominate with single grains are said to be structure less, while those possess majority of secondary particles are said to be aggregate, granulated or crumb structure.

Mechanism of Aggregate Formation: The bonding of the soil particles into structural unit is the genesis of soil structure. The bonding between individual particles in the structural units is generally considered to be stronger than the structural units themselves.

In aggregate formation, a number of primary particles such as sand, silt and clay are brought together by the cementing or binding effect of soil colloids. The cementing materials taking part in aggregate formation are colloidal clay, iron and aluminium hydroxides and decomposing organic matter. Whatever may be the cementing material, it is ultimately the dehydration of colloidal matter accompanied with pressure that completes the process of aggregation.

Colloidal clay: By virtue of high surface area and surface charge, clay particles play a key role in the formation of soil aggregates. Sand and silt particles can not form aggregates as they do not possess the power of adhesion and cohesion. These particles usually carry a coating of clay particles; they are enmeshed in the aggregates formed by the adhering clay particles. Colloidal particles form aggregates only when they are flocculated. There is vast difference between flocculation and aggregation. Flocculation is brought about by coalescence of colloidal particles and is the first step in aggregation.

Aggregation is some thing more than flocculation involving a combination of different factors such as hydration, pressure, dehydration etc. and required cementation of flocculated particles. The cementation may be caused by cations, oxides of Fe and Al, humus substances and products of microbial excretion and synthesis. Clay particles form aggregates only if they are wetted by a liquid like water whose molecules possess an appreciable dipole moment.

Clay - - +Water - - +Cation+ - -Clay - - +Water - - +Cation+ - -Clay -

The aggregation also depends upon the nature of clay particles, size and amount of clay particles, dehydration of clay particles, cations like calcium and anions like phosphate.

Fe and Al oxides: The colloidal Fe oxides act as cementing agent in aggregation. Al oxides bind the sand and silt particles. These act in two ways. A part of the hydroxides acts as a flocculating agent and the rest as a cementing agent.

Organic matter: It also plays an important role in forming soil aggregates.

1. During decomposition, cellulose substances produce a sticky material very much resembling mucus or mucilage. The sticky property may be due to the presence of humic or humic acid or related compounds produced.
2. Certain polysaccharides formed during decomposition.
3. Some fungi and bacteria have cementing effect probably due to the presence of slimes and gums on the surface of the living organisms produced as a result of the microbial activity

Classification of Soil Structure: The primary particles sand, silt and clay usually occur grouped together in the form of aggregates.

Natural aggregates are called peds where as clod is an artificially formed soil mass. Structure is studied in the field under natural conditions and it is described under three categories

1. Type - Shape or form and arrangement pattern of peds
2. Class - Size of Peds
3. Grade - Degree of distinctness of peds

Types of Soil Structure: There are four principal forms of soil structure

Plate-like (Platy): In this type, the aggregates are arranged in relatively thin horizontal plates or leaflets. The horizontal axis or dimensions are larger than the vertical axis. When the units/ layers are thick they are called platy. When they are thin then it is laminar.

Platy structure is most noticeable in the surface layers of virgin soils but may be present in the subsoil.

This type is inherited from the parent material, especially by the action of water or ice.

Prism-like: The vertical axis is more developed than horizontal, giving a pillar like shape. Vary in length from 1- 10 cm. commonly occur in sub soil horizons of Arid and Semi arid regions. When the tops are rounded, the structure is termed as columnar when the tops are flat / plane, level and clear cut prismatic.

Block like: All three dimensions are about the same size. The aggregates have been reduced to blocks. Irregularly six faced with their three dimensions more or less equal.

When the faces are flat and distinct and the edges are sharp angular, the structure is named as angular blocky. When the faces and edges are mainly rounded it is called sub angular blocky. These types usually are confined to the sub soil and characteristics have much to do with soil drainage, aeration and root penetration.

Spheroidal (Sphere like): All rounded aggregates (peds) may be placed in this category. Not exceeding an inch in diameter. These rounded complexes usually loosely arranged

and readily separated. When wetted, the intervening spaces generally are not closed so readily by swelling as may be the case with a blocky structural condition.

Therefore in sphere like structure, infiltration, percolation and aeration are not affected by wetting of soil.

The aggregates of this group are usually termed as granular which are relatively less porous. When the granules are very porous, it is termed as crumb. This is specific to surface soil particularly high in organic matter/ grass land soils.

Classes of Soil Structure: Each primary structural type of soil is differentiated into 5 size classes depending upon the size of the individual peds.

The terms commonly used for the size classes are:

1. 1. Very fine or very thin
2. 2. Fine or thin
3. 3. Medium
4. 4. Coarse or thick
5. 5. Very Coarse or very thick

The terms thin and thick are used for platy types, while the terms fine and coarse are used for other structural types.

Grades of Soil Structure: Grades indicate the degree of distinctness of the individual peds. It is determined by the stability of the aggregates. Grade of structure is influenced by the moisture content of the soil. Grade also depends on organic matter, texture etc. Four terms commonly used to describe the grade of soil structure are:

1. Structure less: There is no noticeable aggregation, such as conditions exhibited by loose sand.
2. Weak Structure: Poorly formed, indistinct formation of peds, which are not durable and much unaggregated material.
3. Moderate structure: Moderately well developed peds, which are fairly durable and distinct.
4. Strong structure: Very well formed peds, which are quite durable and distinct.

Soil Structure Naming: For naming a soil structure the sequence followed is grade, class and type; for example strong coarse angular blocky, moderate thin platy, weak fine prismatic.

Factors Affecting Soil Structure

The development of structure in arable soil depends on the following factors:

1) Climate: Climate has considerable influence on the degree of aggregation as well as on the type of structure. In arid regions there is very little aggregation of primary particles. In semi arid regions, the degree of aggregation is greater.

2. Organic matter: Organic matter improves the structure of a sandy soil as well as of a clay soil. In case of a sandy soil, the sticky and slimy material produced by the decomposing organic matter and the associated microorganism cement the sand particles together to form aggregates. In case of clayey soil, it modifies the properties of clay by reducing its cohesiveness. This helps making clay more crumbly.

3. Tillage: Cultivation implements break down the large clods into smaller fragments and aggregates. For obtaining good granular and crumbly structure, optimum moisture content in the soil is necessary. If the moisture content is too high it will form large clods on drying. If it is too low some of the existing aggregates will be broken down.

4. Plants, Roots and Residues: Excretion of gelatinous organic compounds and exudates from roots serve as a link. Root hairs make soil particles to cling together. Grass and cereal roots vs other roots. Pressure exerted by the roots also held the particles together.

Dehydration of soil strains the soil due to shrinkage result in cracks lead to aggregation

Plant tops and residues shade the soil prevent it from extreme and sudden temperature and moisture changes and also from rain drop impedance.

Plant residues serve as a food to microbes which are the prime aggregate builders.

5. Animals: Among the soil fauna small animals like earthworms, moles and insects etc., that burrow in the soil are the chief agents that take part in the aggregation of finer particles.

6. Microbes: Algae, fungi, actinomycetes and fungi keep the soil particles together. Fungi and actinomycetes exert mechanical binding by mycelia, Cementation by the products of decomposition and materials synthesized by bacteria.

7. Fertilizers: Fertilizer like Sodium Nitrate destroys granulation by reducing the stability of aggregates. Few fertilizers for example, CAN help in development of good structures.

8. Wetting and drying: When a dry soil is wetted, the soil colloids swell on absorbing water. On drying, shrinkage produces strains in the soil mass gives rise to cracks, which break it up into clods and granules of various sizes.

9. Exchangeable cations: Ca, Mg	-----à H, Na
Flocculating	Deflocculating
Good structure	Poor structure

10. Inorganic cements: CaCO₃ and Sesquioxides

11. Clay

12. Water

Effect of Soil Structure on other Physical Properties

Porosity: Porosity of a soil is easily changed. In plate like structure, pore spaces are less where as in crumby structure pore spaces are more.

Temperature: Crumby structure provides good aeration and percolation of water in the soil. Thus these characteristics help in keeping optimum temperature in comparison to plate like structure.

Density: Bulk density varies with the total pore space present in the soil. Structure chiefly influences pore spaces Platy structure with less total pore spaces has high bulk density where as crumby structure with more total pore spaces has low bulk density.

Consistence: Consistence of soil also depends on structure. Plate-like structure exhibits strong plasticity.

Colour: Bluish and greenish colors of soil are generally due to poor drainage of soil. Platy structure normally hinders free drainage.

Importance of Structure: Soil structure influences rather indirectly by the formation of an array of pores of various shapes and sizes. These pores are controlling factors governing water, air and temperature in soil.

Role of soil structure in relation to plant growth

1. Soil structure influences the amount and nature of porosity.
2. Structure controls the amount of water and air present in the soil. Not only the amount of water and air dependent on soil structure, but their movement and circulation are also controlled by soil structure.
3. It affects tillage practices.
4. Structure controls runoff and erosion.
5. Platy structure normally hinders free drainage whereas sphere like structure (granular and crumby) helps in drainage.
6. Crumby and granular structure provides optimum infiltration, water holding capacity, aeration and drainage. It also provides good habitat for microorganisms and supply of nutrients.

Class of Soil Structure as differentiated by size of soil peds

Class	Platy	Prism-atic	Colum-nar	Blocky	S.A. Blocky	Gran-ular	Crumb
V.Fine or V.Thin	<1	<10	<10	<5	<5	<1	<1
Fine or Thin	1-2	10-20	10-20	5-10	5-10	1-2	1-2
Medi	2-5	20-50	20-50	10-20	10-20	2-5	2-5

um	5-10	50-100	50-100	20-50	20-50	5-10	-
Coarse or Thick	>10	>100	>100	>50	>50	>10	-
V.C or V.Thic	>10	>100	>100	>50	>50	>10	-

Density of Soil: Bulk Density and Particle Density

Density represents weight (mass) per unit volume of a substance.

Density = Mass / Volume

Soil density is expressed in two well accepted concepts as particle density and bulk density. In the metric system, particle density can be expressed in terms of mega grams per cubic meter (Mg/m³). Thus if 1 m³ of soil solids weighs 2.6 Mg, the particle density is 2.6 Mg / m³ (since 1 Mg = 1 million grams and 1 m³ = 1 million cubic centimeters) thus particle density can also be expressed as 2.6 g / cm³.

Particle Density: The weight per unit volume of the solid portion of soil is called particle density. Generally particle density of normal soils is 2.65 grams per cubic centimeter. The particle density is higher if large amount of heavy minerals such as magnetite; limonite and hematite are present in the soil. With increase in organic matter of the soil the particle density decreases. Particle density is also termed as true density.

Table Particle density of different soil textural classes

Textural classes	Particle density (g / cm ³)
Coarse sand	2.655
Fine sand	2.659
Silt	2.798
Clay	2.837

Bulk Density: The oven dry weight of a unit volume of soil inclusive of pore spaces is called bulk density. The bulk density of a soil is always smaller than its particle density. The bulk density of sandy soil is about 1.6 g / cm³, whereas that of organic matter is about 0.5. Bulk density normally decreases, as mineral soils become finer in texture. The bulk density varies indirectly with the total pore space present in the soil and gives a good estimate of the porosity of the soil. Bulk density is of greater importance than particle density in understanding the physical behavior of the soil. Generally soils with low bulk densities have favorable physical conditions.

Bulk density of different textural classes

Textural class	Bulk density (g/cc)	Pore space (%)
Sandy soil	1.6	40
Loam	1.4	47
Silt loam	1.3	50
Clay	1.1	58

Factors affecting bulk density

1. Pore space: Since bulk density relates to the combined volume of the solids and pore spaces, soils with high proportion of pore space to solids have lower bulk densities than those that are more compact and have less pore space. Consequently, any factor that influences soil pore space will affect bulk density.

2. Texture: Fine textured surface soils such as silt loams, clays and clay loams generally have lower bulk densities than sandy soils. This is because the fine textured soils tend to organize in porous grains especially because of adequate organic matter content. This results in high pore space and low bulk density. However, in sandy soils, organic matter content is generally low, the solid particles lie close together and the bulk density is commonly higher than in fine textured soils.

3. Organic matter content: More the organic matter content in soil results in high pore space there by shows lower bulk density of soil and vice-versa.

6. Soil Colloids

The colloidal state refers to a two-phase system in which one material in a very finely divided state is dispersed through second phase.

The examples are:

Solid in liquid - Clay in water (dispersion of clay in water)

Liquid in gas -Fog or clouds in atmosphere

The clay fraction of the soil contains particles less than 0.002 mm in size. Particles less than 0.001 mm size possess colloidal properties and are known as soil colloids.

General Properties of Soil Colloids

1. Size: The most important common property of inorganic and organic colloids is their extremely small size. They are too small to be seen with an ordinary light microscope. Only with an electron microscope they can be seen. Most are smaller than 2 micrometers in diameter.

2. Surface area: Because of their small size, all soil colloids expose a large external surface per unit mass. The external surface area of 1 g of colloidal clay is at least 1000 times that of 1 g of coarse sand. Some colloids, especially certain silicate clays have extensive internal surfaces as well. These internal surfaces occur between plate like crystal units that make up each particle and often greatly exceed the external surface area. The total surface area of soil colloids ranges from 10 m²/g for clays with only external surfaces to more than 800 m²/g for clays with extensive internal surfaces. The colloid surface area in the upper 15 cm of a hectare of a clay soil could be as high 700,000 km²/g

3. Surface charges: Soil colloidal surfaces, both external and internal characteristically carry negative and/or positive charges. For most soil colloids, electro negative charges predominate. Soil colloids both organic and inorganic when suspended in water, carry a negative electric charge. When an electric current is passed through a suspension of soil colloidal particles they migrate to anode, the positive electrode indicating that they carry a negative charge. The magnitude of the charge is known as zeta potential. The presence and intensity of the particle charge influence the attraction and repulsion of the particles towards each other, there by influencing both physical and chemical properties.

The negative electrical charge on clays comes from

- i) Ionizable hydrogen ions and
- ii) Isomorphism substitution.

i) Ionizable hydrogen ions: Ionizable hydrogen ions are hydrogen from hydroxyl ions on clay surfaces. The -Al-OH or -Si-OH portion of the clay ionizes the H and leaves an unneutralized negative charge on the oxygen (-Al-O⁻ or - Si-O⁻). The extent of ionized hydrogen depends on solution pH; more ionization occurs in more alkaline (basic) solutions.

ii) **Isomorphous substitution:** The second source of charge on clay particles is due to the substitution of one ion for another of similar size and often with lower positive valence. In clay structures, certain ions fit into certain mineral lattice sites because of their convenient size and charge. Dominantly, clays have Si^{4+} in tetrahedral sites and Al^{3+} in octahedral sites. Other ions present in large amounts during clay crystallization can replace some of the Al^{3+} and Si^{4+} cations. Substitutions that are common are the Si^{4+} replaced by Al^{3+} , and even more extensive replacement of Al^{3+} by one or more of these: Fe^{3+} , Fe^{2+} , Mg^{2+} or Zn^{2+} . Since the total negative charge from the anions (the oxygen) remains unchanged, the lower positive charge because of substitution results in an excess negative charge at that location in the structure.

4. Adsorption of cations: As soil colloids possess negative charge they attract the ions of an opposite charge to the colloidal surfaces. They attract hundreds of positively charged ions or cation such as H^+ , Al^{3+} , Ca^{2+} , and Mg^{2+} . This gives rise to an ionic double layer.

The process, called Isomorphous substitution and the colloidal particle constitutes the inner ionic layer, being essentially huge anions; with both, external and internal layers that are negative in charge. The outer layer is made up of a swarm of rather loosely held (adsorbed) cations attracted to the negatively charged surfaces. Thus a colloidal particle is accompanied by a swarm of cations that are adsorbed or held on the particle surfaces.

5. Adsorption of water: In addition to the adsorbed cations, a large number of water molecules are associated with soil colloidal particles. Some are attracted to the adsorbed cations, each of which is hydrated; others are held in the internal surfaces of the colloidal particles. These water molecules play a critical role in determining both the physical and chemical properties of soil.

6. Cohesion: Cohesion is the phenomenon of sticking together of colloidal particles that are of similar nature. Cohesion indicates the tendency of clay particles to stick together. This tendency is primarily due to the attraction of the clay particles for the water molecules held between them. When colloidal substances are wetted, water first adheres to the particles and then brings about cohesion between two or more adjacent colloidal particles.

7. Adhesion: Adhesion refers to the phenomenon of colloidal particles sticking to other substances. It is the sticking of colloidal materials to the surface of any other body or substance with which it comes in contact.

8. Swelling and shrinkage: Some clay (soil colloids) such as smectites swell when wet and shrink when dry. After a prolonged dry spell, soils high in smectites (e.g. Vertisols) often are criss-crossed by wide, deep cracks, which at first allow rain to penetrate rapidly. Later, because of swelling, such soil is likely to close up and become much more impervious than one dominated by kaolinite, chlorite, or fine grained micas. Vermiculite is intermediate in its swelling and shrinking characteristics.

9. Dispersion and flocculation: As long as the colloidal particles remain charged, they repel each other and the suspension remains stable. If on any account they lose their charge, or if the magnitude of the charge is reduced, the particles coalesce, form flocs or loose aggregates, and settle out. This phenomenon of coalescence and formation of flocs is known as flocculation. The reverse process of the breaking up of flocs into individual particles is known as deflocculation or dispersion.

10. Brownian movement: When a suspension of colloidal particles is examined under a microscope the particles seem to oscillate. The oscillation is due to the collision of colloidal particles or molecules with those of the liquid in which they are suspended. Soil colloidal particles with those of water in which they are suspended are always in a constant state of motion. The smaller the particle, the more rapid is its movement.

11. Non permeability: Colloids, as opposed to crystalloid, are unable to pass through a semi-permeable membrane. Even though the colloidal particles are extremely small, they are bigger than molecules of crystalloid dissolved in water. The membrane allows the passage of water and of the dissolved substance through its pores, but retains the colloidal particles.

7. Types of Soil Colloids

There are four major types of colloids present in soil

1. Layer silicate clays
2. Iron and aluminum oxide clays (sesquioxide clays)
3. Allophane and associated amorphous clays
4. Humus.

Layer silicate clays, iron and aluminum oxide clays, allophane and associated amorphous clays are inorganic colloids while humus is an organic colloid.

1) Layer silicate clays: These are most important silicate clays and are known as phyllosilicates (Phyllon - leaf) because of their leaf-like or plate like structure. They are comprised of two kinds of horizontal sheets. One dominated by silicon and other by aluminum and/or magnesium.

2) Silica Tetrahedron: The basic building block for the silica-dominated sheet is a unit composed of one silicon atom surrounded by four oxygen atoms. It is called the silica tetrahedron because of its four-sided configuration. An interlocking array or a series of these silica tetrahedral tied together horizontally by shared oxygen anions gives a tetrahedral sheet.

3) Alumina Octahedron: Aluminium and/or magnesium ions are the key cations in the second type of sheet. An aluminium (or magnesium) ion surrounded by six oxygen atoms or hydroxyl group gives an eight sided building block termed octahedron. Numerous octahedra linked together horizontally comprise the octahedral sheet. An aluminum dominated sheet is known as a dioctahedral sheet, whereas one dominated by magnesium is called a trioctahedral sheet. The distinction is due to the fact that two aluminum ions in a dioctahedral sheet satisfy the same negative charge from surrounding oxygen and hydroxyls as three magnesium ions in a trioctahedral sheet.

The tetrahedral and octahedral sheets are the fundamental structural units of silicate clays. They, in turn, are bound together within the crystals by shared oxygen atoms into different layers. The specific nature and combination of sheets in these layers vary from one type of clay to another and largely control the physical and chemical properties of each clay.

Types of Silicate Clay Minerals: On the basis of the number and arrangement of tetrahedral (silica) and octahedral (alumina-magnesia) sheets contained in the crystal units or layers, silicate clays are classified into three different groups

- A. 1 :1 Type clay minerals
- B. 2:1 Type clay minerals
- C. 2: 1: 1 Type clay minerals

A) 1:1 Type Minerals: The layers of the 1:1-type minerals are made up of one tetrahedral (silica) sheet combined with one octahedral (alumina) sheet-hence the terminology. In soils, **kaolinite** is the most prominent member of this group, which includes **hallosite**, **nacrite**, and **dickite**.

The tetrahedral and octahedral sheets in a layer of a kaolinite crystal are held together tightly by oxygen atoms, which are mutually shared by the silicon and aluminum cations in their respective sheets. These layers, in turn, are held together by hydrogen bonding. Consequently, the structure is fixed and no expansion ordinarily occurs between layers when the clay is wetted.

Cations and water do not enter between the structural layers of a 1:1 type mineral particle. The effective surface of kaolinite is thus restricted to its outer faces or to its external surface area. Also, there is little isomorphous substitution in this 1:1 type mineral. Along with the relatively low surface area of kaolinite, this accounts for its low capacity to adsorb cations.

Kaolinite crystals usually are hexagonal in shape. In comparison with other clay particles, they are large in size, ranging from 0.10 to 5 μm across with the majority falling within the 0.2 to 2 μm range. Because of the strong binding forces between their structural layers, kaolinite particles are not readily broken down into extremely thin plates.

Kaolinite exhibits very little plasticity (capability of being molded), cohesion, shrinkage, and swelling.

B) 2:1-Type Minerals: The crystal units (layers) of these minerals are characterized by an octahedral sheet sandwiched between two tetrahedral sheets. Three general groups have this basic crystal structure.

- i) Expanding type: smectites and vermiculite
- ii) Non-expanding type: mica (illite)

i) Expanding Minerals: The **smectite** group is noted for interlayer expansion, which occurs by swelling when the minerals are wetted, the water entering the interlayer space and forcing the layers apart. **Montmorillonite** is the most prominent member of this group in soils, although **beidellite**, **nontronite**, and **saponite** are also found.

The flake-like crystals of smectites (e.g., Montmorillonite) are composed of an expanding lattice 2:1-type clay mineral. Each layer is made up of an octahedral sheet sandwiched between two tetrahedral (silica) sheets. There is little attraction between oxygen atoms in the bottom tetrahedral sheet of one unit and those in the top tetrahedral sheet of another. This permits a ready and variable space between layers, which is occupied by water and exchangeable cations. This internal surface far exceeds the surface around the outside of the crystal. In montmorillonite magnesium has replaced aluminum in some sites of the

octahedral sheet. Likewise, some silicon atoms in the tetrahedral sheet may be replaced by aluminum. These substitutions give rise to a negative charge.

These minerals show high cation exchange capacity, marked swelling and shrinkage properties. Wide cracks commonly form as smectite dominated soils (e.g., Vertisols) are dried. The dry aggregates or clods are very hard, making such soils difficult to till.

Vermiculites are also 2:1 type minerals in that an octahedral sheet occurs between two tetrahedral sheets. In most soils vermiculites, the octahedral sheet is aluminum dominated (dioctahedral), although magnesium dominated (trioctahedral) vermiculites are also common. In the tetrahedral sheet of most vermiculite, considerable substitution of aluminum for silicon has taken place. This accounts for most of the very high net negative charge associated with these minerals.

Water molecules, along with magnesium and other ions, are strongly adsorbed in the interlayer space of vermiculites. They act primarily as bridges holding the units together rather than as wedges driving them apart. The degree of swelling is, therefore considerable less for vermiculites than for smectites. For this reason, vermiculites are considered limited-expansion clay minerals, expanding more than kaolinite but much less than the smectites.

The cation exchange capacity of vermiculites usually exceeds that of all other silicate clays, including montmorillonite and other smectites, because of very high negative charge in the tetrahedral sheet. Vermiculite crystals are larger than those of the smectites but much smaller than those of kaolinite.

ii) Non-expanding minerals: Micas are the type minerals in this group. (e.g.) **Muscovite** and **biotite**, weathered minerals similar in structure to these micas are found in the clay fraction of soils. They are called fine-grained micas. Like sanctities, fine-grained micas have a 2:1-type crystal. However, the particles are much larger than those of the smectites. Also, the major source of charge is in-the tetrahedral sheet where aluminum atoms occupy about 20% of the silicon sites. These results in a high net negative charge in the tetrahedral sheet, even higher than that found in vermiculites, to satisfy this charge, potassium ions are strongly attracted in the interlayer space and are just the right size to fit into certain spaces in the adjoining tetrahedral sheets. The potassium thereby acts as a binding agent, preventing expansion of the crystal. Hence, fine-grained micas are quite non expansive.

The properties such as hydration, cation adsorption, swelling, shrinkage, and plasticity are much less intense in fine-grained micas than in smectites. The fine grained micas exceed kaolinite with respect to these characteristics, but this may be due in part to the presence of interstratified layers of smectite or vermiculite. In size, too, fine-grained mica crystals are intermediate between the smectities and kaolihites.

Their specific surface area varies from 70 to 100 m²/g, about one eighth that for the smectites.

C) 2:1:1 Type Minerals: This silicate group is represented by chlorites, which are common in a variety of soils. **Chlorites** are basically iron magnesium silicates with some

aluminum present. In a typical chlorite clay crystal, 2:1 layers, such as in vermiculites, alternate with a magnesium-dominated trioctahedral sheet, giving a 2:1:1 ratio. Magnesium also dominates the trioctahedral sheet in the 2:1 layer of chlorites. Thus, the crystal unit contains two silica tetrahedral sheets and two magnesium-dominated trioctahedral sheets giving rise to the term 2:1:1 or 2:2-type structure.

The negative charge of chlorites is about the same as that of fine-grained micas considerably less than that of the **smectites** or **vermiculites**. Like fine micas, chlorites may be interstratified with vermiculites or smectites in a single crystal. Particle size and surface area for chlorites are also about the same as for fine grained micas. There is no water adsorption between the chlorite crystal units, which accounts for the non expanding nature of this mineral.

Mixed and interstratified layers: Specific groups of clay minerals do not occur independently of one another. In a given soil, it is common to find several clay minerals in an intimate mixture. Furthermore, some mineral colloids have properties and composition intermediate between those of any two of the well defined minerals described. Such minerals are termed mixed layer or interstratified because the individual layers within a given crystal may be of more than one type. Terms such as "chlorite-vermiculite" and "fine-grained mica-smectite" are used to describe mixed-layer minerals. In some soils, they are more common than single-structured minerals such as montmorillonite.

2. Iron and aluminum oxide clays (sesquioxide clays): Under conditions of extensive leaching by rainfall and long time intensive weathering of minerals in humid warm climates, most of the silica and much of the alumina in primary minerals are dissolved and slowly leached away. The remnant materials, which have lower solubility, are sesquioxides. Sesquioxides (metal oxides) are mixtures of aluminum hydroxide, $\text{Al}(\text{OH})_3$, and iron oxide, Fe_2O_3 , or iron hydroxide, $\text{Fe}(\text{OH})_3$.

The Latin word sesqui means one and one-half times, meaning one and one-half times more oxygen than Al and Fe. These clays can grade from amorphous to crystalline.

Examples of iron and aluminum oxides common in soils are gibbsite ($\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$) and goethite ($\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$).

Less is known about these clays than about the layer silicates. These clays do not swell, not sticky and have high phosphorus adsorption capacity

3. Allophane and other Amorphous Minerals: These silicate clays are mixtures of silica and alumina. They are amorphous in nature. Even mixture of other weathered oxides (iron oxide) may be a part of the mixture. Typically, these clays occur where large amount of weathered products existed. These clays are common in soils forming from volcanic ash (e.g., Allophane). These clays have high anion exchange capacity or even high cation exchange capacity. Almost all of their charge is from accessible hydroxyl ions (OH^-), which can attract a positive ion or lose the H^+ attached. These clays have a variable charge that depends on H^+ in solution (the soil acidity).

4. Humus (Organic Colloid): Humus is amorphous, dark brown to black, nearly insoluble in water, but mostly soluble in dilute alkali (NaOH or KOH) solutions. It is a temporary intermediate product left after considerable decomposition of plant and animal remains. They are temporary intermediate because the organic substances remain continue to decompose slowly.

The humus is often referred to as an organic colloid and consists of various chains and loops of linked carbon atoms. The humus colloids are not crystalline. They are composed basically of carbon, hydrogen, and oxygen rather than of silicon, aluminum, iron, oxygen, and hydroxyl groups.

The organic colloidal particles vary in size, but they may be at least as small as the silicate clay particles. The negative charges of humus are associated with partially dissociated enolic (-OH), carboxyl (-COOH), and phenolic groups; these groups in turn are associated with central units of varying size and complexity.

8. Soil Moisture

Water contained in soil is called soil moisture. The water is held within the soil pores. Soil water is the major component of the soil in relation to plant growth. If the moisture content of a soil is optimum for plant growth, plants can readily absorb soil water. Not all the water, held in soil, is available to plants. Much of water remains in the soil as a thin film. Soil water dissolves salts and makes up the soil solution, which is important as medium for supply of nutrients to growing plants.

Importance of Soil Water

1. Soil water serves as a solvent and carrier of food nutrients for plant growth
2. Yield of crop is more often determined by the amount of water available rather than the deficiency of other food nutrients
3. Soil water acts as a nutrient itself
4. Soil water regulates soil temperature
5. Soil forming processes and weathering depend on water
6. Microorganisms require water for their metabolic activities
7. Soil water helps in chemical and biological activities of soil
8. It is a principal constituent of the growing plant
9. Water is essential for photosynthesis

Retention of Water by Soil: The soils hold water (moisture) due to their colloidal properties and aggregation qualities. The water is held on the surface of the colloids and other particles and in the pores. The forces responsible for retention of water in the soil after the drainage has stopped are due to surface tension and surface attraction and are called surface moisture tension. This refers to the energy concept in moisture retention relationships. The force with which water is held is also termed as suction.

Ways to Retain Water in Soil

1. Cohesion and adhesion forces: These two basic forces are responsible for water retention in the soil. One is the attraction of molecules for each other i.e., cohesion. The other is the attraction of water molecules for the solid surface of soil i.e. adhesion. By adhesion, solids (soil) hold water molecules rigidly at their soil - water interfaces. These water molecules in turn hold by cohesion. Together, these forces make it possible for the soil solids to retain water.

2. Surface tension: This phenomenon is commonly evidenced at water- air interfaces. Water behaves as if its surface is covered with a stretched elastic membrane. At the surface, the attraction of the air for the water molecules is much less than that of water molecules for each other. Consequently, there is a net downward force on the surface molecules, resulting in sort of a compressed film (membrane) at the surface. This phenomenon is called surface tension.

3. Polarity or dipole character: The retention of water molecules on the surface of clay micelle is based on the dipole character of the molecule of water. The water molecules

are held by electrostatic force that exists on the surface of colloidal particles. By virtue of their dipole character and under the influence of electrostatic forces, the molecules of water get oriented (arranged) on the surface of the clay particles in a particular manner.

Each water molecule carries both negative and positive charges. The clay particle is negatively charged. The positive end of water molecule gets attached to the negatively charged surface of clay and leaving its negative end outward. The water molecules attached to the clay surface in this way present a layer of negative charges to which another layer of oriented water molecules is attached. The number of successive molecular layers goes on increasing as long as the water molecules oriented. As the molecular layer gets thicker, orientation becomes weaker, and at a certain distance from the particle surface the water molecules cease to orientate and capillary water (liquid water) begins to appear. Due to the forces of adsorption (attraction) exerted by the surface of soil particles, water gets attached on the soil surface. The force of gravity also acts simultaneously, which tries to pull it downwards. The surface force is far greater than the force of gravity so water may remain attached to the soil particle. The water remains attached to the soil particle or move downward into the lower layers, depending on the magnitude of the resultant force.

Factors Affecting Soil Water

- 1. Texture:** Finer the texture, more is the pore space and also surface area, greater is the retention of water.
- 2. Structure:** Well-aggregated porous structure favors better porosity, which in turn enhance water retention.
- 3. Organic matter:** Higher the organic matter more is the water retention in the soil.
- 4. Density of soil:** Higher the density of soil, lower is the moisture content.
- 5. Temperature:** Cooler the temperature, higher is the moisture retention.
- 6. Salt content:** More the salt content in the soil less is the water available to the plant.
- 7. Depth of soil:** More the depth of soil more is the water available to the plant.
- 8. Type of clay:** The 2:1 type of clay increases the water retention in the soil.

Soil Water Potential

The retention and movement of water in soils, its uptake and translocation in plants and its loss to the atmosphere are all energy related phenomenon. The more strongly water is held in the soil the greater is the heat (energy) required. In other words, if water is to be removed from a moist soil, work has to be done against adsorptive forces. Conversely, when water is adsorbed by the soil, a negative amount of work is done. The movement is from a zone where the free energy of water is high (standing water table) to one where the free energy is low (a dry soil). This is called soil water energy concept.

Free energy of soil solids for water is affected by

i) Matric (solid) force i.e., the attraction of the soil solids for water (adsorption) which markedly reduces the free energy (movement) of the adsorbed water molecules.

ii) Osmotic force i.e., the attraction of ions and other solutes for water to reduce the free energy of soil solution.

Matric and Osmotic potentials are negative and reduce the free energy level of the soil water. These negative potentials are referred as suction or tension.

iii) Force of gravity: This acts on soil water, the attraction is towards the earth's center, which tends to pull the water down ward. This force is always positive. The difference between the energy states of soil water and pure free water is known as soil water potential. Total water potential (Pt) is the sum of the contributions of gravitational potential (Pg), matric potential (Pm) and the Osmotic potential or solute potential (Po).

$$Pt = Pg + Pm + Po$$

Potential represents the difference in free energy levels of pure water and of soil water. The soil water is affected by the force of gravity, presence of soil solid (matric) and of solutes.

Methods of expressing suctions: There are two units to express differences in energy levels of soil water.

i) PF Scale: The free energy is measured in terms of the height of a column of water required to produce necessary suction or pressure difference at a particular soil moisture level. The pF, therefore, represents the logarithm of the height of water column (cm) to give the necessary suction.

ii) Atmospheres or Bars: It is another common mean of expressing suction. Atmosphere is the average air pressure at sea level. If the suction is very low as occurs in the case of a wet soil containing the maximum amount of water that it can hold, the pressure difference is of the order of about 0.01 atmospheres or 1 PF equivalent to a column of water 10 cm in height. Similarly, if the pressure difference is 0.1 atmosphere the PF will be 20. Soil moisture constants can be expressed in term of PF values. A soil that is saturated with water has PF 0 while an oven dry soil has a PF 7.

Measuring Soil Moisture

Two general types of measurements relating to soil water are ordinarily used

- i) By some methods the moisture content is measured directly or indirectly
- ii) Techniques are used to determine the soil moisture potential (tension or suction)

Measuring soil moisture content in laboratory

1. Gravimetric method: This consists of obtaining a moist sample, drying it in an oven at 105°C until it loses no more weight and then determining the percentage of moisture. The gravimetric method is time consuming and involves laborious processes of sampling, weighing and drying in laboratory.

2. Electrical conductivity method: This method is based upon the changes in electrical conductivity with changes in soil moisture. Gypsum blocks inside of with two electrodes at a definite distance are apart used in this method. These blocks require previous calibration for uniformity. The blocks are buried in the soil at desired depths and the conductivity across the electrodes measured with a modified Wheatstone bridge. These electrical measurements are affected by salt concentration in the soil solution and are not very helpful in soils with high salt contents.

Measuring soil moisture potential in situ (field)

3. Suction method or equilibrium tension method: Field tensiometers measure the tension with which water is held in the soils. They are used in determining the need for irrigation. The tensiometer is a porous cup attached to a glass tube, which is connected to a mercury monometer. The tube and cup are filled with water and cup inserted in the soil. The water flows through the porous cup into the soil until equilibrium is established. These tension readings in monometer, expressed in terms of cm or atmosphere, measures the tension or suction of the soil.

If the soil is dry, water moves through the porous cup, setting up a negative tension (or greater is the suction). The **tensiometers** are more useful in sandy soils than in fine textured soils. Once the air gets entrapped in the tensiometer, the reliability of readings is questionable.

Classification of soil water

Soil water has been classified from a physical and biological point of view as Physical classification of soil water, and biological classification of soil water.

Physical classification of soil water

1. Gravitational water: Gravitational water occupies the larger soil pores (macro pores) and moves down readily under the force of gravity. Water in excess of the field capacity is termed gravitational water. Gravitational water is of no use to plants because it occupies the larger pores. It reduces aeration in the soil. Thus, its removal from soil is a requisite for optimum plant growth. Soil moisture tension at gravitational state is zero or less than 1/3 atmosphere.

Factors affecting gravitational water

i. Texture: Plays a great role in controlling the rate of movement of gravitational water. The flow of water is proportional to the size of particles. The bigger the particle, the more rapid is the flow or movement. Because of the larger size of pore, water percolates more easily and rapidly in sandy soils than in clay soils.

ii. Structure: It also affects gravitational water. In platy structure movement of gravitational water is slow and water stagnates in the soil. Granular and crumbly structure helps to improve gravitational water movement. In clay soils having single grain structure, the gravitational water, percolates more slowly. If clay soils form aggregates (granular structure), the movement of gravitational water improves.

2. Capillary water: Capillary water is held in the capillary pores (micro pores). Capillary water is retained on the soil particles by surface forces. It is held so strongly that gravity cannot remove it from the soil particles. The molecules of capillary water are free and mobile and are present in a liquid state. Due to this reason, it evaporates easily at ordinary temperature though it is held firmly by the soil particle; plant roots are able to absorb it. Capillary water is, therefore, known as available water. **The capillary water is held between 1/3 and 31 atmosphere pressure.**

Factors affecting capillary water: The amount of capillary water that a soil is able to hold varies considerably. The following factors are responsible for variation in the amount of capillary water.

i. Surface tension: An increase in surface tension increases the amount of capillary water.

ii. Soil texture: The finer the texture of a soil, greater is the amount of capillary water holds. This is mainly due to the greater surface area and a greater number of micro pores.

iii. Soil structure: Platy structure contains more water than granular structure.

iv. Organic matter: The presence of organic matter helps to increase the capillary capacity of a soil. Organic matter itself has a great capillary capacity. Undecomposed organic matter is generally porous having a large surface area, which helps to hold more capillary water. The humus that is formed on decomposition has a great capacity for absorbing and holding water. Hence the presence of organic matter in soil increases the amount of capillary water in soil.

3. Hygroscopic water: The water that held tightly on the surface of soil colloidal particle is known as hygroscopic water. It is essentially non-liquid and moves primarily in the vapour form.

Hygroscopic water held so tenaciously (31 to 10000 atmospheres) by soil particles that plants can not absorb it. Some microorganism may utilize hygroscopic water. As hygroscopic water is held tenaciously by surface forces its removal from the soil requires a certain amount of energy. Unlike capillary water which evaporates easily at atmospheric temperature, hygroscopic water cannot be separated from the soil unless it is heated.

Factors affecting hygroscopic water: Hygroscopic water is held on the surface of colloidal particles by the dipole orientation of water molecules. The amount of hygroscopic water varies inversely with the size of soil particles. The smaller the particle,

the greater is the amount of hygroscopic water it adsorbs. Fine textured soils like clay contain more hygroscopic water than coarse textured soils.

The amount of clay and also its nature influences the amount of hygroscopic water. Clay minerals of the montmorillonite type with their large surface area adsorb more water than those of the kaolinite type, while illite minerals are intermediate.

Biological Classification of Soil Water

There is a definite relationship between moisture retention and its utilization by plants. This classification based on the availability of water to the plant. Soil moisture can be divided into three parts.

i. Available water: The water which lies between wilting coefficient and field capacity. It is obtained by subtracting wilting coefficient from moisture equivalent.

ii. Unavailable water: This includes the whole of the hygroscopic water plus a part of the capillary water below the wilting point.

iii. Super available or superfluous water: The water beyond the field capacity stage is said to be super available. It includes gravitational water plus a part of the capillary water removed from larger interstices. This water is unavailable for the use of plants. The presence of super-available water in a soil for any extended period is harmful to plant growth because of the lack of air.

Soil Moisture Constants

Earlier classification divided soil water into gravitational, capillary and hygroscopic water. The hygroscopic and capillary waters are in equilibrium with the soil under given condition. The hygroscopic coefficient and the maximum capillary capacity are the two equilibrium points when the soil contains the maximum amount of hygroscopic and capillary waters, respectively. The amount of water that a soil contains at each of these equilibrium points is known as soil moisture constant.

The soil moisture constant, therefore, represents definite soil moisture relationship and retention of soil moisture in the field. The three classes of water (gravitational, capillary and hygroscopic) are however very broad and do not represent accurately the soil - water relationships that exists under field conditions.

1. Though the maximum capillary capacity represents the maximum amount of capillary water that a soil holds, the whole of capillary water is not available for the use of the plants.
2. A part of it, at its lower limit approaching the hygroscopic coefficient is not utilized by the plants.
3. Similarly a part of the capillary water at its upper limit is also not available for the use of plants.

Hence two more soil constants, viz., field capacity and wilting coefficient have been introduced to express the soil-plant-water relationships as it is found to exist under field conditions.

1. Field capacity: Assume that water is applied to the surface of a soil. With the downward movement of water all macro and micro pores are filled up. The soil is said to be saturated with respect to water and is at maximum water holding capacity or maximum retentive capacity. It is the amount of water held in the soil when all pores are filled.

Sometimes, after application of water in the soil all the gravitational water is drained away, and then the wet soil is almost uniformly moist. The amount of water held by the soil at this stage is known as the field capacity or normal moisture capacity of that soil. It is the capacity of the soil to retain water against the downward pull of the force of gravity. At this stage only micropores or capillary pores are filled with water and plants absorb water for their use. At field capacity water is held with a force of $1/3$ atmosphere. Water at field capacity is readily available to plants and microorganism

2. Wilting coefficient: As the moisture content falls, a point is reached when the water is so firmly held by the soil particles that plant roots are unable to draw it. The plant begins to wilt. At this stage even if the plant is kept in a saturated atmosphere it does not regain its turgidity and wilts unless water is applied to the soil. The stage at which this occurs is termed the Wilting point and the percentage amount of water held by the soil at this stage is known as the Wilting Coefficient. It represents the point at which the soil is unable to supply water to the plant. Water at wilting coefficient is held with a force of 15 atmospheres.

3. Hygroscopic coefficient: The hygroscopic coefficient is the maximum amount of hygroscopic water absorbed by 100 g of dry soil under standard conditions of humidity (50% relative humidity) and temperature (15°C). This tension is equal to a force of 31 atmospheres. Water at this tension is not available to plant but may be available to certain bacteria.

4. Available water capacity: The amount of water required to apply to a soil at the wilting point to reach the field capacity is called the "available" water. The water supplying power of soils is related to the amount of available water a soil can hold. The available water is the difference in the amount of water at field capacity ($- 0.3$ bar) and the amount of water at the permanent wilting point ($- 15$ bars).

5. Maximum water holding capacity: It is also known as maximum retentive capacity. It is the amount of moisture in a soil when its pore spaces both micro and macro capillary are completely filled with water. It is a rough measure of total pore space of soil. Soil moisture tension is very low between $1/100$ th to $1/1000$ th of an atmosphere or pF 1 to 0.

6. Sticky point moisture: It represents the moisture content of soil at which it no longer sticks to a foreign object. The sticky point represents the maximum moisture content at which a soil remains friable. Sticky point moisture values vary nearly approximate to the

moisture equivalent of soils. Summary of the soil moisture constants, type of water and force with which it held is given in following table.

Soil moisture constants and range of tension and PF

Sr. No.	Moisture class	Range of tension in atmosphere	Equivalent PF range
1	Chemically combined	Very High	-
2	Water vapour	Held at saturation point in the soil air	-
3	Hygroscopic	31 to 10,000	4.50 to 7.00
4	Hygroscopic coefficient	31	4.50
5	Wilting point	15	4.20
6	Capillary	1/3 to 31	2.54 to 4.50
7	Moisture equivalent	1/3 to 1	2.70 to 3.00
8	Field capacity	1/3	2.54

Entry of Water into Soil

1. Infiltration: Infiltration refers to the downward entry or movement of water into the soil surface. It is a surface characteristic and hence primarily influenced by the condition of the surface soil. Soil surface with vegetative cover has more infiltration rate than bare soil. Warm soils absorb more water than colder ones. Coarse surface texture, granular structure and high organic matter content in surface soil, all help to increase infiltration. Infiltration rate is comparatively lower in wet soils than dry soils.

Factors affecting infiltration

1. Clay minerals
2. Soil Texture
3. Soil structure
4. Moisture content
5. Vegetative cover
6. Topography

2. Percolation: The movement of water through a column of soil is called percolation. It is important for two reasons.

- i) This is the only source of recharge of ground water which can be used through wells for irrigation
- ii) Percolating waters carry plant nutrients down and often out of reach of plant roots (leaching)

Percolation is dependent of rainfall. In dry region it is negligible and under high rainfall it is high. Sandy soils have greater percolation than clayey soil. Vegetation and high water table reduce the percolation loss

3. Permeability: It indicates the relative ease of movement of water with in the soil. The characteristics that determine how fast air and water move through the soil are known as permeability. The term hydraulic conductivity is also used which refers to the readiness with which a soil transmits fluids through it.

Soil Water Movement

- i) Saturated Flow
- ii) Unsaturated Flow
- iii) Water Vapour Movement

Saturated flow: This occurs when the soil pores are completely filled with water. This water moves at water potentials larger than -33 k Pa . Saturated flow is water flow caused by gravity's pull. It begins with *infiltration*, which is water movement into soil when rain or irrigation water is on the soil surface. When the soil profile is wetted, the movement of more water flowing through the wetted soil is termed *percolation*.

Hydraulic conductivity can be expressed mathematically as

$$V = kf$$

Where,

V = Total volume of water moved per unit time

f = Water moving force

k = Hydraulic conductivity of soil

Factors affecting movement of water

1. Texture, 2. Structure, 3. Amount of organic matter, 4. Depth of soil to hard pan,
5. Amount of water in the soil, 6. temperature and 7. Pressure

Vertical water flow:

The vertical water flow rate through soil is given by *Darcy's law*. The law states that the rate of flow of liquid or flux through a porous medium is proportional to the hydraulic gradient in the direction of flow of the liquid.

$$QW = -k \frac{(dw) At}{Ds}$$

Where,

QW = Quantity of water in cm^3

k = rate constant (cm/s)

dw = Water height (head), cm

A = Soil area (cm^2)

t = Time

ds = Soil depth (cm)

9. Soil Air

Soil air is a continuation of the atmospheric air. Unlike the other components, it is constant state of motion from the soil pores into the atmosphere and from the atmosphere into the pore space. This constant movement or circulation of air in the soil mass resulting in the renewal of its component gases is known as soil aeration.

Composition of Soil Air: The soil air contains a number of gases of which nitrogen, oxygen, carbon dioxide and water vapour are the most important. Soil air constantly moves from the soil pores into the atmosphere and from the atmosphere into the pore space. Soil air and atmospheric air differ in the compositions. Soil air contains a much greater proportion of carbon dioxide and a lesser amount of oxygen than atmospheric air. At the same time, soil air contains a far great amount of water vapour than atmospheric air. The amount of nitrogen in soil air is almost the same as in the atmosphere.

Composition of soil and atmospheric air

Percentage by volume			
	Nitrogen	Oxygen	Carbon dioxide
Soil Air	79.2	20.6	0.3
Atmospheric Air	79.9	20.97	0.03

Factors Affecting the Composition of Soil Air:

1. Nature and condition of soil: The quantity of oxygen in soil air is less than that in atmospheric air. The amount of oxygen also depends upon the soil depth. The oxygen content of the air in lower layer is usually less than that of the surface soil. This is possibly due to more readily diffusion of the oxygen from the atmosphere into the surface soil than in the subsoil. Light texture soil or sandy soil contains much higher percentage than heavy soil. The concentration of CO₂ is usually greater in subsoil probably due to more sluggish aeration in lower layer than in the surface soil.

2. Type of crop: Plant roots require oxygen, which they take from the soil air and deplete the concentration of oxygen in the soil air. Soils on which crops are grown contain more CO₂ than fallow lands. The amount of CO₂ is usually much greater near the roots of plants than further away. It may be due to respiration by roots.

3. Microbial activity: The microorganisms in soil require oxygen for respiration and they take it from the soil air and thus deplete its concentration in the soil air. Decomposition of organic matter produces CO₂ because of increased microbial activity. Hence, soils rich in organic matter contain higher percentage of CO₂.

4. Seasonal variation: The quantity of oxygen is usually higher in dry season than during the monsoon. Because soils are normally drier during the summer months, opportunity for gaseous exchange is greater during this period. This results in relatively high O₂ and low CO₂ levels. Temperature also influences the CO₂ content in the soil air.

High temperature during summer season encourages microorganism activity which results in higher production of CO₂.

Exchange of Gases between Soil and Atmosphere

The exchange of gases between the soil and the atmosphere is facilitated by two mechanisms

1. Mass flow: With every rain or irrigation, a part of the soil air moves out into the atmosphere as it is displaced by the incoming water. As and when moisture is lost by evaporation and transpiration, the atmospheric air enters the soil pores. The variations in soil temperature cause changes in the temperature of soil air. As the soil air gets heated during the day, it expands and the expanded air moves out into the atmosphere. On the other hand, when the soil begins to cool, the soil air contracts and the atmospheric air is drawn in.

2. Diffusion: Most of the gaseous interchange in soils occurs by diffusion. Atmospheric and soil air contains a number of gases such as nitrogen, oxygen, carbon dioxide etc., each of which exerts its own partial pressure in proportion to its concentration.

The movement of each gas is regulated by the partial pressure under which it exists. If the partial pressure on one of the gases (i.e. carbon dioxide) is greater in the soil air than in the atmospheric air, it (CO₂) moves out into the atmosphere. Hence, the concentration of CO₂ is more in soil air.

On the other hand, partial pressure of oxygen is low in the soil air, as oxygen present in soil air is consumed as a result of biological activities. The oxygen present in the atmospheric air (partial pressure of O₂ is greater) therefore, diffuses into the soil air till equilibrium is established. Thus, diffusion allows extensive movement and continual change of gases between the soil air and the atmospheric air. Oxygen and carbon dioxide are the two important gases that take in diffusion

Importance of Soil Aeration

1. Plant and root growth: Soil aeration is an important factor in the normal growth of plants. The supply of oxygen to roots in adequate quantities and the removal of CO₂ from the soil atmosphere are very essential for healthy plant growth.

When the supply of oxygen is inadequate, the plant growth either retards or ceases completely as the accumulated CO₂ hampers the growth of plant roots. The abnormal effect of insufficient aeration on root development is most noticeable on the root crops. Abnormally shaped roots of these plants are common on the compact and poorly aerated soils. The penetration and development of root are poor. Such undeveloped root system cannot absorb sufficient moisture and nutrients from the soil

2. Microorganism population and activity: The microorganisms living in the soil also require oxygen for respiration and metabolism. Some of the important microbial activities such as the decomposition of organic matter, nitrification, Sulphur oxidation etc depend

upon oxygen present in the soil air. The deficiency of air (oxygen) in soil slows down the rate of microbial activity.

For example, the decomposition of organic matter is retarded and nitrification arrested. The microorganism population is also drastically affected by poor aeration.

3. Formation of toxic material: Poor aeration results in the development of toxin and other injurious substances such as ferrous oxide, H₂S gas, CO₂ gas etc in the soil.

4. Water and nutrient absorption: A deficiency of oxygen has been found to check the nutrient and water absorption by plants. The energy of respiration is utilized in absorption of water and nutrients. Under poor aeration condition (this condition may arise when soil is water logged), plants exhibit water and nutrient deficiency

5. Development of plant diseases: Insufficient aeration of the soil also leads to the development of diseases. For example, wilt of gram and dieback of citrus and peach.

10. Organic Colloids – Soil Organic Matter

Soil organic matter (SOM) comprises an accumulation of

- i) Partially disintegrated and decomposed plant and animal residues
- ii) Other organic compounds synthesized by the soil microbes upon decay.

OM content of a well drained mineral soil is LOW: 1 – 6 % by weight in the top soil and even less in the subsoil.

Sources of Soil Organic Matter

The primary sources of SOM are plant tissues

1. The tops and roots of trees
2. Shrubs, grasses, remains of harvested crops and Soil organisms

Animals are secondary sources of Organic Matter.

1. Waste products of animals
2. Remains of animals after completion of life cycle.

Factors Affecting Soil Organic Matter

1. Climate: Temperature and rainfall exert a dominant influence on the amounts of N and OM found in soils.
 - a) Temperature: The OM and N content of comparable soils tend to increase if one moves from warmer to cooler areas. The decomposition of OM is accelerated in warm climates as compared to cooler climates. For each 10 C decline in mean annual temperature, the total OM and N increases by two to three times.
 - b) Rainfall: There is an increase in OM with an increase in rainfall. Under comparable conditions, the N and OM increase as the effective moisture becomes greater.
2. Natural Vegetation: The total OM is higher in soils developed under grasslands than those under forests.
3. Texture: Fine textured soils are generally higher in OM than coarse textured soils.
4. Drainage: Poorly drained soils because of their high moisture content and relatively poor aeration are much higher in OM and N than well drained soils.
5. Cropping and Tillage: The cropped lands have much low N and OM than comparable virgin soils. Modern conservation tillage practices helps to maintain high OM levels as compared to conventional tillage.
6. Rotations, residues and plant nutrients: Crop rotations of cereals with legumes results in higher soil OM. Higher OM levels, preferably where a crop rotation is followed.

Role of organic matter

1. Organic Matter creates a granular condition of soil which maintains favorable condition of aeration and permeability.
2. Water holding capacity of soil is increased and surface runoff, erosion etc., are reduced as there is good infiltration due to the addition of OM.
3. Surface mulching with coarse OM lowers wind erosion and lowers soil temperatures in the summer and keeps the soil warmer in winter.
4. OM serves as a source of energy for the microbes and as a reservoir of nutrients that are essential for plant growth and also hormones, antibiotics.
5. Fresh OM supplies food for earthworms, ants and rodents and makes soil P readily available in acid soils.
6. Organic acids released from decomposing OM help to reduce alkalinity in soils; organic acids along with released CO₂ dissolve minerals and make them more available.
7. Humus (a highly decomposed OM) provides a storehouse for the exchangeable and available cations.
8. It acts as a buffering agent which checks rapid chemical changes in pH and soil reaction.

Humus

Humus is a complex and rather resistant mixture of brown or dark brown amorphous and colloidal organic substance that results from microbial decomposition and synthesis and has chemical and physical properties of great significance to soils and plants.

Humus Formation

The humus compounds have resulted from two general types of biochemical reactions: Decomposition and Synthesis.

1. Decomposition:

- a) Chemicals in the plant residues are broken down by soil microbes including lignin.
- b) Other simpler organic compounds that result from the breakdown take part immediately in the second of the humus-forming processes, biochemical synthesis.
- c) These simpler chemicals are metabolized into new compounds in the body tissue of soil microbes.
- d) The new compounds are subject to further modification and synthesis as the microbial tissue is subsequently attacked by other soil microbes.

2. Synthesis: Involve such breakdown products of lignin as the phenols and quinones.

- a) These monomers undergo polymerization by which polyphenols and polyquinones are formed.
- b) These high molecular weight compounds interact with N-containing amino compounds and forms a significant component of resistant humus.
- c) Colloidal clays encourage formation of these polymers.
- d) Generally two groups of compounds that collectively make up humus, the humic group and the nonhumic group.

Properties of Humus

- 1. The tiny colloidal particles are composed of C, H, and O₂. The colloidal particles are negatively charged (-OH, -COOH or phenolic groups), has very high surface area, higher CEC (150 – 300 cmol/kg), 4 - 5 times higher WHC than that of silicate clays.
- 2. Humus has a very favorable effect on aggregate formation and stability.
- 3. Impart black colour to soils.
- 4. Cation exchange reactions are similar to those occurring with silicate clays.

Clay – Humus Complex

Humus, the organic amorphous colloid supplies both basic and acidic ions which is transitory and ultimately disappears from soil. Clay, the inorganic crystalline colloid supplies chiefly the basic nutrient ions is more or less stable. Both these colloids form the soil colloidal complex and are extremely active and form important sources of plant nutrients.

It is believed that humus and clay exist in the soil as clay – humus complex, the two being held together by cations like Ca, Fe, etc. Depending upon the nature of binding cation, two types of Clay – humus complex have been recognized. The colloidal complex bound by Ca ions is more stable and is responsible for the favorable physical condition of the soil, particularly its structure. The other type where Fe acts as the binding agent creates a poor physical condition of the soils.

11. Ion Exchange

As soils are formed during the weathering processes, some minerals and organic matter are broken down to extremely small particles. Chemical changes further reduce these particles until they cannot be seen with the naked eye. The very smallest particles are called colloids.

The mineral clay colloids are plate like in structure and crystalline in nature. In most soils; clay colloids exceed organic colloids in amount.

Colloids are primarily responsible for the chemical reactivity in soils. The kind of parent material and the degree of weathering determine the kinds of clays present in the soil. Since soil colloids are derived from these clays, their reactivity is also influenced by parent material and weathering. Each colloid (inorganic and organic) has a net negative (-) charge developed during the formation process. This means it can attract and hold positively (+) charged particles. An element with an electrical charge is called an ion.

Potassium, sodium (Na), hydrogen (H), Ca and Mg all has positive charges. They are called cations and ions with negative charges, such as nitrate and sulfate, are called anions.

Negatively charged colloids attract cations and hold them like a magnet holds small pieces of metal. This characteristic explains; why nitrate-N is more easily leached from the soil than ammonium-N. Nitrate has a negative charge like soil colloids. So, NO_3^- is not held by the soils, but remains as a free ion in soil water to be leached through the soil profile in some soils and under some rainfall conditions. The charges associated with soil particles attract simple and complex ions of opposite charge. Thus, a given colloidal mixture may exhibit not only a maze of positive and negative surface charges but also an equal complex complement of simple cations and anions such as Ca^{2+} and SO_4^- - that are attracted by the particle charges.

The adsorbed anions are commonly present in smaller quantities than the cations because the negative charges generally predominate on the soil colloid.

Mechanism of Cation Exchange: The exchange of cations has been explained on the basis of the electro-kinetic theory of ion exchange. According to this theory, the adsorbed cations forming the outer shell of the ionic double layer are supposed to be in a state of oscillation when suspended in water, forming a diffuse double layer. Due to these oscillations, some of the cations move away from the surface of the clay micelle. In the presence of the solution of an electrolyte a cation of the added electrolyte slips in between the inner negative layer and the outer oscillating positive ion. The electrolyte cation is now adsorbed on the micelle and the surface cation remains in solution as an exchanged ion. Thus the exchange of cations takes place. Cations are positively charged nutrient ions and molecules. While, clay particles are negatively charged constituents of soils. These negatively charged particles (clay) attract, hold and release positively charged nutrient ions (cations). Organic matter particles also have a negative charge to attract cations. Sand particles carry little or no charge and do not react. Cations held by

soils can be replaced by other cations. This means they are exchangeable. For example, Ca^{++} can be exchanged for H^+ and/or K^+ and vice versa.

Cation Exchange Capacity (CEC): The CEC is the capacity of soil to hold and exchange cations. The cation exchange capacity is defined simply as the sum total of the exchangeable cations that a soil can adsorb. The higher the CEC of soil the more cations it can retain. Soils differ in their capacities to hold exchangeable K^+ and other cations.

The CEC depends on amount and kinds of clay and organic matter present. A high-clay soil can hold more exchangeable cations than a low-clay soil. CEC also increases as organic matter increases. Clay minerals usually range from 10 to 150 meq/100 g in CEC values. Organic matter ranges from 200 to 400 meq/100 g. So, the kind and amount of clay and organic matter content greatly influence the CEC of soils. Clay soils with high CEC can retain large amounts of cations against potential loss by leaching. Sandy soils, with low CEC, retain smaller quantities.

This makes timing and application rates important in planning a fertilizer programme. For example, it may not be wise to apply K on very sandy soils in the middle of a monsoon, where rainfall can be high and intense. Fertilizer application should be split to prevent leaching and losses through erosion. Also, splitting N applications to meet peak crop demand are important to reduce the potential for nitrate leaching on sands as well as finer-textured soils.

Means of Expression: The cation exchange capacity is expressed in terms of equivalents or more specifically, as milli equivalents per 100 gram and is written as meq/100g. The term equivalent is defined as one gram of atomic weight of hydrogen (or the amount of any other ion) that will combine with or displace this amount of hydrogen for monovalent ions such as Na^+ , K^+ , NH_4^+ and Cl^- , the equivalent weight and atomic weight are same, since they can replace one H ion. Divalent cations such as Ca^{++} and Mg^{++} can take the place of two H^+ ions. The milliequivalent weight of a substance is one thousandth of its atomic weight. Since the atomic weight of hydrogen is about 1 gram. The term milliequivalent (meq.) may be defined as 1 milligram of hydrogen.

This unit of exchangeable cations i.e. milliequivalent per 100 g of soil (meq/100g) was used prior to 1982. In the newer metric system the term equivalent is not used, however, now moles are the accepted chemical unit. All the calculation and concepts of "equivalents" are still mentally used but the notation must be written differently. The old "equivalent" is represented by moles (+) or mole, which indicates a monovalent ion portion. For example, to write 12.5 meq/100 g in the newer metric system, it can be written as: 12.5 c mol (+) kg^{-1} of soil (centimoles) or 125 m mol (+) kg^{-1} of soil (millimoles).

Replacing power of cations: The replacing power of cations varies with the type of ion, its size and degree of hydration, valence and concentration and the kind of clay mineral involved, as it is controlled by number of factors no single order of replacement can be given. All other factors being equal the replacing power of monovalent cations increases in the following order: $\text{Li} < \text{Na} < \text{K} < \text{Rb} < \text{Cs} < \text{H}$ and for divalent cations: $\text{Mg} < \text{Ca} < \text{Sr} < \text{Ba}$. In case of mixture of monovalent and divalent cations as they exist in normal soils the replacing power increases in the following order: $\text{Na} < \text{K} < \text{NH}_4 < \text{Mg} < \text{Ca} < \text{H}$. This means Na is more easily replaced than K and K more easily than NH_4 and so on.

Percent base saturation: The percent of total CEC occupied by the major cations has been used in the past to develop fertilizer programs. The idea is that certain nutrient ratios or 'balances' are needed to ensure proper uptake by the crop for optimum yields. Research has shown, however, that cation saturation ranges and ratios have little or no utility in a vast majority of agricultural soils. Under field conditions, ranges of nutrients can vary widely with no detrimental effects, so long as individual nutrients are present in sufficient levels in the soil to support optimum plant growth.

Importance of Cation Exchange: Cation exchange is an important reaction in soil fertility, in causing and correcting soil acidity and basicity, in changes altering soil physical properties, and as a mechanism in purifying or altering percolating waters. The plant nutrients like calcium, magnesium, and potassium are supplied to plants in large measure from exchangeable forms.

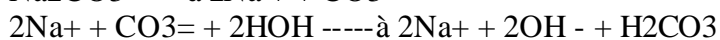
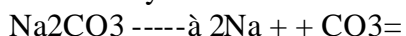
Soil reaction

Soil reaction is one of the most important physiological characteristics of the soil solution. The presence and development of micro-organisms and higher plants depend upon the chemical environment of soil. Therefore study of soil reaction is important in soil science.

There are three types of soil reactions: 1. Acidic 2. Alkaline and 3. Neutral

1. Acidic: It is common in region where precipitation is high. The high precipitation leaches appreciable amounts of exchangeable bases from the surface layers of the soils so that the exchange complex is dominated by H ions. Acid soils, therefore, occur widely in humid regions and affect the growth of plants markedly.

2. Alkaline: Alkali soils occur when there is comparatively high degree of base saturation. Salts like carbonates of calcium, magnesium and sodium also give a preponderance of OH ions over H ions in the soil solution. When salts of strong base such as sodium carbonate go into soil solution and hydrolyze, consequently they give rise to alkalinity. The reaction is as follows:



since sodium hydroxide dissociates to a greater degree than the carbonic acid, OH ions dominate and give rise to alkalinity. This may be as high as 9 or 10. These soils most commonly occur in arid and semi-arid regions.

3. Neutral: Neutral soils occur in regions where H ions just balance OH ions.

Soil pH: The reaction of a solution represents the degree of acidity or basicity caused by the relative concentration of H ions (acidity) or OH ions present in it. Acidity is due to the excess of H ions over OH ions, and alkalinity is due to the excess of OH ions over H ions. A neutral reaction is produced by an equal activity of H and OH ions. According to the theory of dissociation, the activity is due to the dissociation or ionization of compounds into ions.

Method of expressing acidity or alkalinity

Equivalent quantities of all acids or alkalis contain the same number of total H or OH ions, respectively. But, when they are dissolved in water they do not ionize to the same extent. The amount of acid or alkali ionized depends upon the content of free H and OH ions. When the dissociation is high e.g., hydrochloric acid (a strong acid), it dissociates to a larger extent than the weak acetic acid. Acetic acid dissociates only to about 10 % as compared to hydrochloric acid. In a 1N solution of hydrochloric acid there will be 1 gram of H ions per liter, while in a normal solution of acetic acid there will be 1/10 gram of H ions per liter. But in titration 1 ml of 1 N hydrochloric acid and 1 ml of 1 N acetic acid will require 1 ml of 1 N sodium hydroxide for neutralization separately, because the total acidity is the same and titration determines both the ionized and unionized H or OH ions.

Soil acidity: There are three kinds of acidity.

- (i) Active acidity is due to the H⁺ ion in the soil solution.
- (ii) Salt replaceable acidity represented by the hydrogen and aluminum that are easily exchangeable by other cations in a simple unbuffered salt solution such as KCl and
- (iii) Residual acidity is that acidity which can be neutralized by limestone or other alkaline materials but cannot be detected by the salt-replaceable technique. Obviously, these types of acidity all add up to the total acidity of a soil.

i. Active acidity: The active acidity is a measure of the H⁺ ion activity in the soil solution at any given time. However, the quantity of H⁺ ions owing to active acidity is very small compared to the quantity in the exchange and residual acidity forms. For example, only about 2 kg of calcium carbonate would be required to neutralize the active acidity in a hectare-furrow slice of an average mineral soil at pH 4 and 200/0 moisture. Even though the concentration of hydrogen ions owing to active acidity is extremely small, it is important because this is the environment to which plants and microbes are exposed.

ii. Salt replaceable (exchangeable) acidity: This type of acidity is primarily associated with the exchangeable aluminum and hydrogen ions that are present in largest quantities in very acid soils. These ions can be released into the soil solution by an unbuffered salt such as KCl.

$Al^{3+} + 4KCl \rightarrow AlCl_3 + HCl$ L⁻⁻⁻ H + L[~] (Soil Solid) (Soil Solution) (Soil Solid) (Soil solution), in moderately acid soils, the quantity of easily exchangeable aluminum and hydrogen is quite limited. Even in these soils, however, the limestone needed to neutralize this type of acidity is commonly more than 100 times that needed for the soil solution (active acidity). At a given pH value, exchangeable acidity is generally highest for smectites, intermediate for vermiculites, and lowest for kaolinite. In any case, however, it accounts for only a small portion of the total soil acidity as the next section will verify.

iii. Residual acidity: Residual acidity is that which remains in the soil after active and exchange acidity has been neutralized. Residual acidity is generally associated with aluminum hydroxy ions and with hydrogen and aluminum atoms that are bound in non exchangeable forms by organic matter and silicate clays. If lime is added to a soil, the pH

increases and the aluminum hydroxy ions are changed to uncharged gibbsite as follows.
 OH^- OH^- $\text{Al}(\text{OH})_2^+$ $7 \text{ Al}(\text{OH})_2^+$ $7 \text{ Al}(\text{OH})_3$ In addition, as the pH increases bound hydrogen and aluminum can be released by calcium and magnesium In the lime materials [$\text{Ca}(\text{OH})_2$ is used as an example of the reactive calcium liming material] The residual acidity is commonly far greater than either the active or salt replaceable acidity. Conservative estimates suggest that the residual acidity may be 1000 times greater than the soil solution or active acidity in a sandy soil and 50000 or even 100000 times greater in a clayey soil high in organic matter. The amount of ground limestone recommended to at least partly neutralize residual acidity is commonly 4-8 metric tons (Mg) per hectare furrow slice (1.8-3.6 tons/AFS).

It is obvious that the pH of the soil solution is only "the tip of the iceberg" in determining how much lime is needed. Buffering and Soil Reaction Buffer action: Buffering refers to resistance to a change in pH. If 1 ml HCl (of 0.1 N) is added to one liter of pure distilled water of pH 7.0, the resulting solution would have a pH of about 5.0. If on the other hand, the same amount of acid is added to a liter of soil suspension the resulting change in pH would be very small. There is, a distinct resistance to change in pH. This power to resist a change in pH is called buffer action. A buffer solution is one which contains reserve acidity and alkalinity and does not change pH with small additions of acids or alkalis. Buffer capacity: The colloidal complex acts as a powerful buffer in the soil and does not allow rapid and sudden changes in soil reaction. Buffering depends upon the amount of colloidal material present in soil. Clay soils rich in organic matter are more highly buffered than sandy soils. Buffer capacity of the soil varies with its cation exchange capacity (C.E.C.). The greater the C.E.C. the greater will be its buffer capacity. Thus heavier the texture and the greater the organic matter content of a soil, the greater is the amount of acid or alkaline material required to change its pH. Importance of Buffering to Agriculture: Changes In soil reaction (pH) have a direct influence on the plants and it also affects the availability of plant nutrients. Deficiency of certain plant nutrients and excess availability of others in toxic amounts would seriously upset the nutritional balance in the soil. Buffering prevents sudden changes and fluctuation in soil pH, so it regulates the availability of nutrients and also checks direct toxic effect to plants.

Factors Controlling Soil Reactions

Soil reaction varies due to following factors

1. Nature of soil colloids: The colloidal particles of the soil influence soil reaction to a very greatest extent. When hydrogen (H^+) ion forms the predominant adsorbed cations on clay colloids, the soil reaction becomes acid.
2. Soil solution: The soil solution carries a number of salts dissolved in capillary water. The cations of the salts intermingle with those of the diffuse double layer of the clay particle and increase the concentration. The concentration of cations in bulk of the solution is more or less (or nearly) the same as that near the particle surfaces. For an unsaturated soil (Clay), the more compact the layer the greater is the number of hydrogen ions dissociating into the solution. This increases the acidity of the soil solution or lowers its pH. Under field conditions, the concentration of salts varies with the moisture content of the soil. The more dilute the solution, the higher the pH value. Hence the pH tends to

drop as the soil gets progressively dry. Soil reaction is also influenced by the presence of CO₂ in soil air. As the CO₂ concentration increases, the soil pH falls and increases the availability of the nutrients. Under field conditions, plant roots and micro-organisms liberate enough CO₂, which results in lowering the pH appreciably. This principle of increasing the concentration of CO₂ in soil air is also used in the reclamation of alkali soils.

3. **Climate:** Rainfall plays an important role in determining the reaction of soil. In general, soils formed in regions of high rainfall are acidic (low pH value), while those formed in regions of low rainfall are alkaline (high pH value).

4. **Soil management:** Cultural operations in general tend to increase soil acidity. They make an acid soil more acidic, and an alkaline soil less alkaline. As a result of constant cultivation, basic elements are lost from the soil through leaching and crop removal. This leads to change the soil reaction to the acid side.

5. **Parent materials:** Soils developed from parent material of basic rocks generally have higher pH than those formed from acid rocks (e.g. granite). The influence of parent material is not very important as it is completely masked by the climatic conditions under which the soil is developed.

6. Precipitation: As water from rainfall passes through the soil, basic nutrients such as calcium (Ca) and magnesium (Mg) are leached. They are replaced by acidic elements including Al, H and manganese (Mn). Therefore, soils formed under high rainfall conditions are more acid than those formed under arid conditions.

7. Decomposition of organic matter: Soil organic matter is continuously being decomposed by micro-organisms into organic acids, carbon dioxide (CO₂) and water, forming carbonic acid. Carbonic acid, in turn, reacts with the Ca and Mg carbonates in the soil to form more soluble bicarbonates, which are leached away, leaving the soil more acid.

8. Native vegetation: Soils often become more acid when crops are harvested because of removal of bases. Type of crop determines the relative amounts of removal. For example, legumes generally contain higher levels of bases than do grasses. Calcium and Mg contents also vary according to the portion(s) of the plant harvested. Many legumes release H⁺ ions into their rhizosphere when actively fixing atmospheric N₂. The acidity generated can vary from 0.2 to 0.7 pH units per mole of fixed N.

9. Soil depth: Except in low rainfall areas, acidity generally increases with depth, so the loss of topsoil by erosion can lead to a more acid pH in the plough layer. The reason is that more subsoil is included in the plow layer as topsoil is lost. There are areas, however, where subsoil pH is higher than that of the topsoil.

10. Nitrogen fertilization: Nitrogen from fertilizer, organic matter, and manure and legume N fixation produces acidity. Nitrogen fertilization speeds up the rate at which acidity develops. At lower N rates, acidification rate is slow, but is accelerated as N fertilizer rates increase.

11. Flooding: The overall effect of submergence is an increase of pH in acid soils and a decrease in basic soils. Regardless of their original pH values, most soils reach pH of 6.5 to 7.2 within one month after flooding and remain at the level until dried. Consequently, liming is of little value in flooded rice production. Further, it can induce deficiencies of micronutrients such as zinc (Zn).

Influence of Soil Reaction on Availability of Nutrients

The unproductiveness of acid and alkali soils is very often due to the lack of available plant nutrients. In highly acid soils (low pH), the availability of some of the nutrients such as aluminum, iron, manganese etc., is increased to a point to become toxic to the plant. At the same time the supplies of available calcium, nitrogen, phosphorus etc., are reduced to starvation level (become unavailable). The same is the case at high pH (alkaline conditions), plant growth suffers due to the unavailability of nutrients like nitrogen, phosphorus and some minor elements (e.g., iron, manganese, boron etc). Another indirect effect occurs through the activity of microorganisms. Most microorganisms function at their best within a pH range 6.0 to 7.5. If soil reaction is changed beyond this range, the microorganisms become functionless. Consequently the supply of some of the essential plant nutrients like nitrogen is considerably reduced.

1. Nitrogen: Plant absorbs most of their nitrogen in the form of nitrate of which availability depends on the activity of nitrifying bacteria. The micro- organisms responsible for nitrification are most active when the pH is between 6.5 and 7.5. They are adversely affected if the pH falls below 5.5 and rises above 9.0. Nitrogen fixing bacteria (like Azotobactor) also fail to function below pH 6.0. The decomposition of organic matter which is the primary source of nitrogen is also slowed down under acidic condition.

2. Phosphorus: Its availability is at its highest when the reaction is between 6.5 and 7.5. When the reaction is above or below this range, availability is reduced. In the strongly acidic soil (pH 5.0 or less), iron, aluminum, manganese and other bases are present in a soluble state and in more quantity. The phosphates of these elements are formed and become unavailable.

3. Potassium: The availability of potassium does not influence by soil reaction to any great extent. In acid soil potassium is lost through leaching. The unavailability of K is due to the conversion of exchangeable to non-exchangeable potassium in alkaline soil. Particularly if the alkalinity is due to CaCO₃ (brought about by over liming in acid soil), the solubility of soil potassium is depressed.

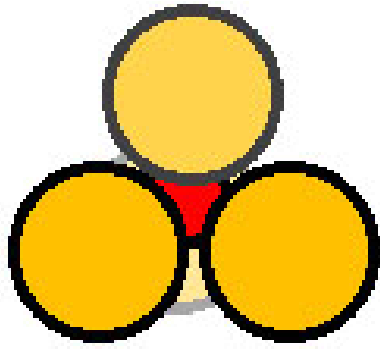
4. Calcium and magnesium: Acid soils (base unsaturated) are poor in / available calcium and magnesium. In alkaline soil (pH not exceeding 8.5) the availability of Ca and Mg nutrients are always high. When the pH is above 8.5, the availability of these nutrients again decreases.

5 Iron, aluminum and manganese: When the pH is low the solubility of iron, aluminum and manganese compounds are increased. and hence they are readily available in acid soils. At the pH range 5.5 to 7.0, iron and manganese are present in the soluble ferrous (Fe⁺⁺) and manganous (Mn⁺⁺) forms. At pH below 5.5 the solubility of these compounds

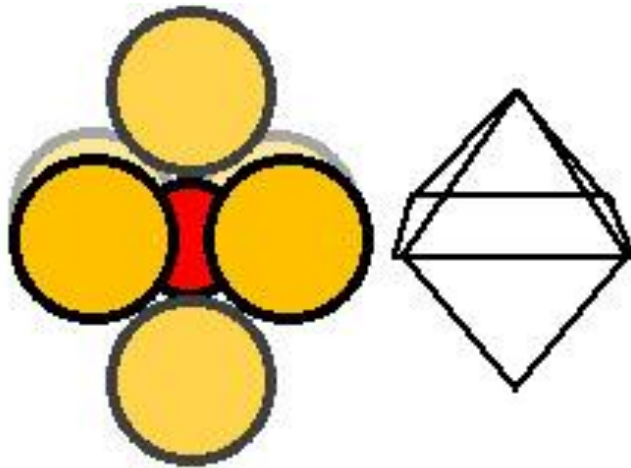
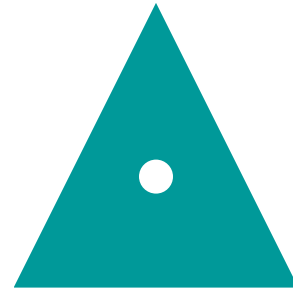
considerably increased with the result that they have a toxic influence on plant growth. Under neutral and alkaline conditions, iron and manganese are usually present in ferric (Fe^{3+}) and manganese (Mn^{+++}) states. Hence in soils with pH 7.5 and above, they become unavailable and sometimes produce deficiency diseases like chlorosis in plants.

6. Sulphur: The availability of sulphur is not affected by soil reaction as sulphur compounds are soluble in low pH range. However, it is more soluble in acid soil and lost in leaching. Acid conditions, which retard the decomposition of organic matter, therefore, retard the release of available sulphur. The availability of sulphur present in organic matter depends upon the decomposition of organic matter.

7. Micronutrients: In general, the availability of boron, copper and zinc is reduced in alkaline soils and that of molybdenum in acid soils. The availability of these nutrients progressively decreases as the soil pH increases. Their availability also decreases under highly acid condition when the pH is below 5.0. Zinc availability in alkaline soils from insoluble zinc salts (calcium zincates) is reduced. Zinc and copper are adsorbed on the clay colloids and not easily displaced and hence not available for plant growth. The availability of molybdenum is reduced under acid soils. It is more available in neutral and alkaline soils.



Tetrahedron



Octahedron

