SAC 101 Fundamentals of Soil Science (2+1)

Unit I

Soil as a natural body, Pedological and edaphological concepts of soil. Components of soil. Soil genesis: Composition of Earth's crust- soil forming rocks and minerals – Primary and secondary minerals. Weathering of rocks and minerals. Factors of soil formation. Soil forming processes. Soil Profile.

Unit II

Soil physical properties: Soil texture, structure, density and porosity, soil colour, consistence and plasticity. Soil water retention, movement and availability. Soil air, composition, gaseous exchange-problem and its effect on crop growth. Source, amount and flow of heat in soil, Soil temperature and crop growth.

Unit III

Soil physico chemical and chemical properties: Soil reaction-pH, soil acidity and alkalinity, buffering, effect of pH on nutrient availability. Electrical conductivity. Soil colloids - inorganic and organic. Silicate clays: constitution and properties, sources of charge, ion exchange, cation and anion exchange capacity and base saturation.

Unit IV

Soil organic matter: composition, properties and its influence on soil properties. Humic substances - nature and properties. Soil Biology : Soil organisms : macro and micro organisms, their beneficial and harmful effects. Soil enzymes. Soil pollution – Types and behaviour of pesticides. Inorganic contaminants. Prevention and mitigation of soil pollution.

PRACTICAL SCHEDULE

Study of soil profile in field. Study of soil sampling tools, collection of representative soil sample, its processing and storage. Study of soil forming rocks and minerals. Determination of soil density, moisture content and porosity. Determination of soil colour. Determination of soil texture by feel and Bouyoucos Methods. Studies of capillary rise phenomenon of water in soil column and water movement in soil. Demonstration of heat transfer in soil. Preparation and standardization of laboratory reagents, indicators and buffers. Determination of soil pH and electrical conductivity. Determination of cation exchange capacity of soil. Estimation of organic matter content of soil. Study of soil map.

Lecture Schedule:

1. Soil definition - Soil as a three dimensional natural body, Pedological and edaphological concepts of soil

2. Components of soil – soil a three phase system- Composition of Earth's crust.
3. Soil genesis: soil forming rocks-definition, formation, Classification of rocks- igneous, sedimentary and metamorphic rocks
4. Brief description of important rocks - mineralogical composition
5. Minerals- definition, occurrence, classification of important soil forming primary minerals - silicate and non silicate minerals, ferro and non-ferro magnesium minerals
6. Formation of secondary minerals - clay minerals and amorphous minerals
7. Weathering - Rocks and minerals - Physical, chemical and biological weathering
8. Factors of soil formation- Passive and active soil forming factors soil forming factors
9. Soil forming process- Fundamental - Simenson's four fold soil forming process - eluviation, illuviation, translocation and humification
10. Specific Soil forming processes - podzolization, laterization, salinization, alkalization, calcification, decalcification and pedoturbation
11. Soil Profile – Horizons, Master horizons and subordinate horizons, subdivisions, Lithological discontinuity.
12. Soil physical properties: Soil texture - particle size distribution - textural classes - textural triangular diagram - significance of soil texture
13. Soil structure - classification - genesis - factors influencing structural stability - significance of soil structure
14. Soil bulk density, particle density and porosity - factors influencing – significance.
15. Soil colour - causes and measurement - Munsell colour chart - factors influencing soil colour – Significance of soil colour.
16. Soil consistence - cohesion, adhesion, plasticity, Atterberg's constants - upper and lower plastic limits, plasticity number- significance of soil consistence
17. **Mid semester Examination**
18. Soil water- forms of water, units of expression and pF scale
19. Soil water potentials - gravitational, matric, osmotic- Soil moisture constants and Soil moisture measurements.
20. Movement of soil water - Saturated and unsaturated flow - infiltration, hydraulic conductivity, percolation, permeability and drainage
21. Soil air, composition, gaseous exchange – Problem and its effect on crop growth.
22. Source, amount and flow of heat in soil, soil temperature and crop growth. and crop growth.

23. Soil reaction (pH) - definition, pH scale, soil acidity and alkalinity, buffering, effect of pH on nutrient availability and factors affecting soil pH
24. Soil Electrical Conductivity - Factors affecting EC – its significance
25. Soil colloids - inorganic and organic
26. Silicate clays: constitution and classification - 1:1, 2:1 expanding and non expanding - 2:2 clay minerals, amorphous minerals and their properties
27. Sources of charge, ion exchange – positive and negative charge – isomorphous substitution, pH dependant charge.
28. Ion exchange - Cation and anion exchange capacity and base saturation
29. Soil organic matter: composition, properties and its influence on soil properties
30. Humic substances – fractionation, nature and properties, Theories of humus formation.
31. Soil Biology- Soil organisms: macro and micro organisms, their beneficial and harmful effects, Soil enzymes
32. Soil carbon sequestration and carbon trading
33. Soil pollution - behaviour of pesticides and inorganic contaminants
34. Prevention and mitigation of soil pollution

Practical schedule:

1. Study of soil sampling tools, collection of representative soil sample, its processing and storage.
2. Study of soil profile in field.
3. Study of soil forming rocks and minerals.
4. Determination of soil density and porosity.
5. Determination of soil colour and moisture content and porosity.
6. Determination of soil texture by feel and Bouyoucos Methods
7. Determination of soil texture by International pipette method.
8. Studies of capillary rise phenomenon of water in soil column and water movement in soil (Infiltration Rate)
9. Studies of capillary rise phenomenon of water in soil column and water movement in soil (Hydraulic conductivity)
10. Determination of soil temperature and demonstration of heat transfer.
11. Preparation and standardization of laboratory reagents, indicators and buffers
12. Determination of soil pH and electrical conductivity.
13. Determination of cation exchange capacity of soil - I.
14. Determination of cation exchange capacity of soil - II

15. Estimation of soil organic carbon.
16. Study of soil map (India and Tamil Nadu)
17. **Final Practical Examination**

References

1. Brady, N.C. and Raymond, C.Weil. 2013. The Nature and Properties of Soils (14th Edition). Pearson Education, Inc. Publishing as Prentice Hall.
2. Fundamentals of Soil Science. 2009. ISSS Publication, New Delhi.
3. Sehgal, J. 2005. Pedology concepts and applications, Kalyani Publishers, New Delhi.
4. Das, D.K. 2013. Introductory Soil science, Kalyani Publishers, New Delhi.

Lecture 1 : Soil definition – Soil as three dimensional natural body – Pedological and Edaphological concepts

Soil science is the study of soil as a natural resource on the surface of the earth including soil formation, classification and mapping; physical, chemical, biological, and fertility properties of soils; and these properties in relation to the use and management of soils.

Sometimes terms which refer to branches of soil science, such as pedology (formation, chemistry, morphology and classification of soil) and edaphology (influence of soil on organisms, especially plants), are used as if synonymous with soil science. The diversity of names associated with this discipline is related to the various associations concerned. Indeed, engineers, agronomists, chemists, geologists, physical geographers, ecologists, biologists, microbiologists, sylviculturists, sanitarians, archaeologists, and specialists in regional planning, all contribute to further knowledge of soils and the advancement of the soil sciences.

Soil scientists have raised concerns about how to preserve soil and arable land in a world with a growing population, possible future water crisis, increasing per capita food consumption, and land degradation.

Soil occupies the pedosphere, one of Earth's spheres that the geosciences use to organize the Earth conceptually. This is the conceptual perspective of pedology and edaphology, the two main branches of soil science. Pedology is the study of soil in its natural setting. Edaphology is the study of soil in relation to soil-dependent uses. Both branches apply a combination of soil physics, soil chemistry, and soil biology. Due to the numerous interactions between the biosphere, atmosphere and hydrosphere that are hosted within the pedosphere, more integrated, less soil-centric concepts are also valuable. Many concepts essential to understanding soil come from individuals not identifiable strictly as soil scientists. This highlights the interdisciplinary nature of soil concepts.

Soil Science

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

Soil Science has six well defined and developed disciplines

Soil fertility: Nutrient supplying properties of soil.

Soil chemistry: Chemical constituents, chemical properties and the chemical reactions.

Soil physics: Involves the study of physical properties

Soil microbiology: Deals with micro organisms, its population, classification, its role in transformations.

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.

Soil Pedology : Dealing with the genesis, survey and classification

Views on Soil (Science)

The term SOIL was derived from the Latin Word "SOLUM" Means FLOOR

- For a Layman soil is dirt or debris
- For an Agriculturist soil is a habitat for plant growth (to grow crops)
- For a Mining Engineer soil is a debris covering the rocks
- For a Civil engineer soil is a material on which road bed or house bed is formed
- For a home owner soil is a mellow or loamy or hard material

Definitions

Generally soil refers to the loose surface of the earth as identified from the original rocks and minerals from which it is derived through weathering process.

Whitney (1892): Soil is a nutrient bin which supplies all the nutrients required for plant growth.

Hilgard (1892): Soil is more or less a loose and friable material in which plants, by means of their roots, find a foothold for nourishment as well as for other conditions of growth"

Dokuchaiev (1900): Russian scientist - Father of soil science - Soil is a natural body composed of mineral and organic constituents, having a definite genesis and a distinct nature of its own.

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".

Jenny (1941): Soil is a naturally occurring body that has been formed due to combined influence of climate and living organisms acting on parent material as conditioned by relief over a period of time.

Ruffin and Simonson (1968): Soil is a mixture of Earth's uppermost mantle of weathered rock and organic matter.

Buckman and Brady (1969): Soil is a dynamic natural body on the surface of the earth in which plants grow, composed of mineral and organic materials and living forms.

Soil Science Society of America (1970)

(i) Soil is 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.

Dr. W.E.H. Blum

Soils not only serve for agriculture and forestry, but also for filtering, buffering and transformation activities between the atmosphere and the ground water, protecting the food chain and drinking water against pollution and biodiversity.

As soil provides nutrients, water, air and anchorage and supports life on Earth, it can be called as Soul of Infinite Life (SOIL).

List of International Soil Scientists

1. Van Helmont (1577 – 1644)
2. Theoder De Saussure
3. John Woodward
4. Boussingault (1802 – 1882)
5. J.V. Liebig (1803 – 1873)
6. J.B.Laws & J.H. Gilbert (1855)
7. J.T.Way (1856)
8. R.Warrington (1876)
9. E.W. Hilgard (1860)
10. V.V. Dokuchaiev (1846-1903)
11. K.D.Glinga (1914)
12. C.F.Marbut (1927)
13. Hens Jenny (1941)

List of Indian Scientists

1. J.W.Leather (1906)
2. Madam Scholasky (1932)

3. Wadia et al. (1935)
4. Viswanath & Ukil (1943)

Soil as a three dimensional body

Soil is a three dimensional body having length, breadth and depth. They form a continuation over the land surface and differ in properties from place to place. Its upper boundary is air or water and lower boundary is the rock lithosphere.

Composition of soil on volume basis

Soil consists of four major components *viz.,*

Mineral matter, (ii) Organic matter, (iii) Water (iv) Air.

Soil contains about 50 per cent solid space and 50 per cent pore space. Total solid space of the soil is occupied by mineral matter and organic matter by about 45 per cent and 5 per cent, respectively. The total pore space of the soil is occupied by air and water on 50:50 basis. The air present is 25 per cent, and water is 25 per cent. The proportion of air and water will vary under natural conditions depending upon the weather and environmental factors.

Mineral matter in soil

The size and composition of mineral matter in soils are variable due to nature of **parent rock** from which it has been derived.

It is generally composed of very fine broken rock fragments and minerals either dominated by distinct minerals like **quartz** and **feldspar**. The rock fragments are disintegrated and broken portions of the massive rocks from which the regolithhave been found through weathering.

In general, the primary minerals *viz.,***Quarts, biotite, muscovite** etc. dominate the coarser fractions of soil. On the other hand, the secondary minerals *viz.,***Silicate clays and hydrous oxide clays of iron and aluminum** etc. are present as very finer fraction of clay in the soil.

Organic matter in soil

Soil organic matter exists as partly decayed and partially synthesized plant and animal residues. Such organic residues are continually being broken down as a result of microbial activity.

It acts as a **chelate**. A chelate is any organic compound that can bind to a metal by more than one bond and form a ring or cyclic structure through that bonding.

Due to this mechanism, the availability of the metallic elements will be increased. Organic matter reduces the soil erosion. By improving the soil physical property, water holding capacity and nutrient supplying power of soil.

Soil water

Water is held within the soil pores with varying degrees of forces depending on the amount of water present. Soil water plays a major role in plant growth relationship. The movement and retention of water is the soil in primarily influenced by the characteristics of the soil *viz.,*Texture, structure, nature of inorganic and organic colloids, type and amount of exchangeable cations and size and total amount of pore spaces etc.

Soil water is also involved in making dissolved salts and that makes up the soil solution. The soil solution acts as an important medium for supplying different essential nutrient elements to growing plants.

Soil air

Air space or pore spaces in a soil consists of that portion of the soil volume not occupied by soil soilds, either mineral or organic.

Soil air contains various gases like carbon dioxide, very small amount of oxygen and nitrogen. Soil air contains much more carbon dioxide and small amounts of oxygen due to microbial respiration present in the soil.

Soil can be compared to various systems of animals

Digestive system : Organic matter decomposition
Respiratory system : Air circulation & exchange of gases
Circulatory system (blood) : Water movement within the soil
Excretory system : Leaching out of excess salts
Brain : Soil clay
Colour : Soil colour
Height : Soil depth

Approaches of Soil Study

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

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.

Edaphological Approach: Edophology (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 fibre as their ultimate goal. They must determine the reasons for variation in the productivity of soils and find means for improvement.

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Lecture 2: Origin of earth – theories – planetesimal and nebular hypothesis – Composition of earth' crust

Origin of earth

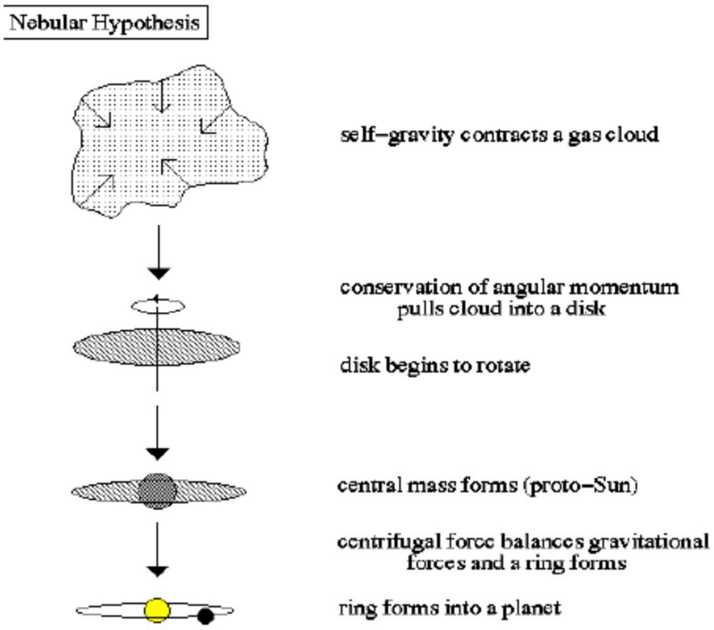
Earth is one of the 9 planets (8 excluding Pluto) orbiting the SUN in the Solar System. The universe is composed of several Galaxies. Our solar system is part of milky way galaxy which is disk shaped with about 1, 00,000 million stars of varying sizes. Our solar system consists of 9 planets and 31 satellites, a belt of asteroids.

Theories about origin of Earth

Nebular hypothesis (Kant and Laplace, 1755)

This hypothesis suggests that the solar system formed through the condensation of a nebula which once encircled the Sun. The outer planets formed first, followed by Mars, the Earth, Venus, and Mercury. This hypothesis suggests a sequential origin from outermost planet to innermost. As per this hypothesis, Mars evolved earlier than the Earth.

This hypothesis is widely accepted to explain origin of different planets. During the past, the entire solar systems existed as a hot gaseous mass called nebula rotating in space. With time, the gaseous mass (nebula) cooled and contracted. Due to contraction, there developed a bulge at the equatorial region. This bulge subsequently separated into several rings.



The ring coalesces in the form of a globe and continues to revolve around the nebula. In similar manner, ten rings were formed of which nine of them gave rise to nine planets.

One broke down into smaller fragments to form the group of planetoids, while the remnant of the pre-existing nebula formed the central incandescent mass of the solar system and is known as the Sun. The planets were originally gaseous but were subsequently cooled down into liquid and ultimately to the solid state.

Planetismal hypothesis (Thomas Chamberlain and Forest Moulton, 1905)

Planets were formed as a result of mutual interaction between the sun and another star of suitable size. This is the theory of biparental origin of the solar system. This theory postulates that due to the near approach of a larger star, tidal distortions were raised upon the surface of the sun and these in conjunction with the eruptive force prevalent in the sun (known as the solar prominence) brought about a description of the mass of the sun and a number of gaseous bulbs were shot forth, in space, to great distances. These gaseous solar materials thus ejected in space were subjected to immediate chilling, resulting in the formation of a number of minute solid particles known as planetesimals.

These planetesimals continued to rotate round the sun in highly elliptical orbits. The orbits must have happened to intersect one another and at points of intersection, they must have collided whereby the small planetesimals continued to coalesce gradually giving rise to the planets. During collision and coalescence of the planete·simals, large quantities of heat must have been generated and were dissipated in space before the next collision could occur and accordingly the planets must have been solid all the time during the growth. This theory accepts that the collision must have taken place in quick succession and accumulation of heat might have caused a melting of the masses of the planets.

Age of the earth

So far scientists have not found a way to determine the exact age of the Earth directly from Earth rocks because Earth's oldest rocks have been recycled and destroyed. Nevertheless, scientists have been able to determine the probable age of the Solar System and to calculate an age for the Earth by assuming that the Earth and the rest of the solid bodies in the Solar System formed at the same time and are, therefore, of the same age.

The ages of Earth and Moon rocks and of meteorites are measured by the decay of long-lived radioactive isotopes of elements that occur naturally in rocks and minerals and that decay with half lives of 700 million to more than 100 billion years to stable isotopes of other elements. These dating techniques, which are firmly grounded in physics and are known collectively as radiometric dating, are used to measure the last time that the rock being dated was either melted or disturbed sufficiently to rehomogenize its radioactive elements.

These ancient rocks have been dated by a number of radiometric dating methods and the consistency of the results give scientists confidence that the ages are correct to within a few percent. An interesting feature of these ancient rocks is that they are not from any sort of ‘primordial crust’ but are lava flows and sediments deposited in shallow water, an indication that Earth history began well before these rocks were deposited. In Western Australia, single zircon crystals found in younger sedimentary rocks have radiometric ages of as much as 4.3 billion years, making these tiny crystals the oldest materials to be found on Earth so far. The source rocks for these zircon crystals have not yet been found. The ages measured for Earth's oldest rocks and oldest crystals show that the Earth is at least 4.3 billion years in age but do not reveal the exact age of Earth's formation.

The best age for the Earth (4.54 Ga)

Present condition of the earth

Our earth is supposed to have been formed in similar manner. It must have been hot and plastic when it was first formed. At present the outside of the earth is quite cool and solid but the interior is very hot and in a fluid condition. The outer cool layer, which we know as the earth’s crust rests upon a denser molten substratum in which various gases are dissolved at high pressure.

Interior of Earth

The Earth Ball consists of 3 concentric rings namely Crust, Mantle and Core.

Crust: 5 to 56 km on the surface of Earth. Density of rocks is 2.6 to 3.0 g cc^{-1}. 5 to 11 km in oceans and 35 to 56 km in the continents. The crust has been divided into two sub-zones called Sial and Sima. The Sial is a heterogeneous mixture of rocks. The Sima is a homogenous plastic or semi-plastic concentric layer that behaves like a solid. The Sial floats on the Sima, which in turn floats on the lower concentric layer called the Mantle. Sial contains about 65-75 % silica. Aluminium is the next important element in the Sial, represented by the most common rocks like granite, and rhyolite. Silica decreases to about 50-60 percent in the Sima where aluminium has largely been replaced by magnesium with minor quantities of iron. Sima is represented by the most common rocks like basalt and gabbro at the upper level and by olivine rich rocks at the lower level. ·

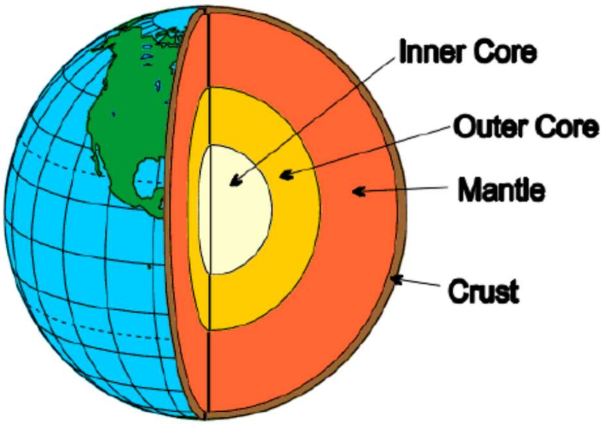
Mantle

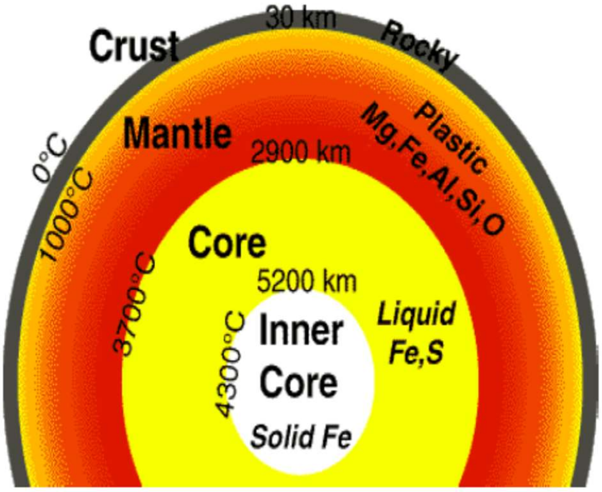
The Mantle is a thick layer of solid rock below the crust. It is 2900 km in thickness. The rock is composed of silicon, oxygen, aluminium, iron and magnesium. The density of rock is 3.0 to 4.5 g cc^{-1}. The upper part of the mantle has a temperature of about 870°C. The temperature gradually increases with depth, through the mantle, to about 4400°C where the mantle meets the next lower section, the outer core.

Core

Core is the innermost portion of Earth, 3500 km in thickness. The rock contain molten metals like nickel and iron. Average density of rock is 9.0 to 12.0 g cc^{-1}. The outer core is about 2250 km thick and is composed of melted iron and nickel. Its temperature varies from

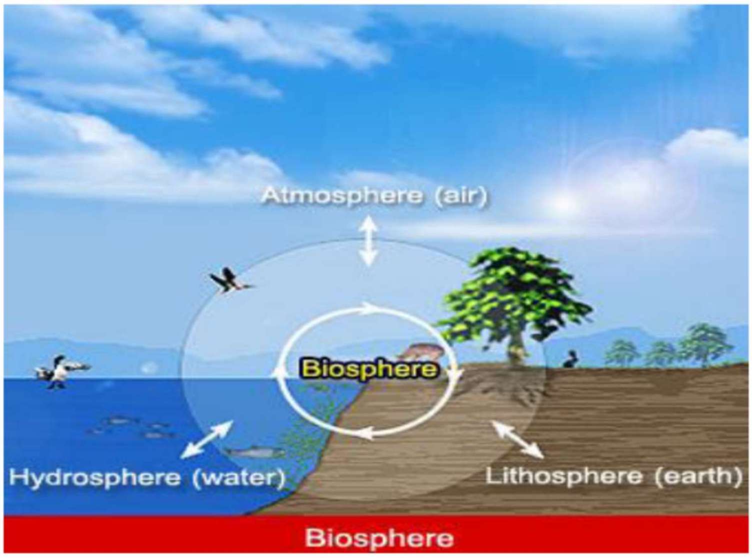
2200°C to about 5000°C. The ball shaped inner core makes up the center of the earth. It is about 1300 km thick and consists of solid iron and nickel.





Exterior of Earth

Solid lithosphere, Liquid hydrosphere and gaseous atmosphere. The atmosphere is of 320 km above the lithosphere / hydrosphere. (70% of earth surface is covered by water (Hydrosphere).



Composition of atmospheric air

	By volume (%)	By weight (%)
N_2	78.08	76.5
O_2	20.9	23.1
CO_2	0.033	0.04
Other gases	0.93	1.36

(H, NH_3, H_2S, SO_2, O_3, He, Argon, Neon, Krypton, Xenon)

Hydrosphere

It is the layer of water surrounding the lithosphere. It is present in the form of seas and oceans. It covers 70% of the earth leaving only about 30% above sea level. The surface of the waters of the various seas is in one level in contrast with the surface of the land. This surface is known as the sea level.

The seawater has a higher specific gravity than terrestrial water due to the salts it contains in solution. The average density is 1.026, but it varies slightly from place to place. Sea water contains 3.5% salts (minerals) It is least dense at the places where river enters the sea and very heavy at places where evaporation is high.

Lithosphere

It is the inner most body within the gaseous and watery envelops. That portion of the lithosphere, which rises above the seawater, is visible to us and is known as land. The land is only about ¼th of the total surface of the earth. Most of this land is situated in the northern hemisphere. The lithosphere consists of two portions, viz.,

1. The upper or outer cool solid surface.
2. The inner hot and molten mass.

It is the heaviest of the three spheres. Its mean density is 5.5 compared to that of water as one. The outer crust has a density of about 2.5 to 3.0, while the inside core, consists of much heavier materials. The outer solid layer, called as the earth's crust is estimated to be about 10 to 20 miles thick. It consists of the various rocks together with a more or less thin mantle of soil enveloping them. It is on this crust that life, both animal and plant sustains. The inner mass, which forms the interior of the earth, is in molten condition. According to one belief, the whole of the inner core is a molten mass of materials, upto the centre. According to another view, the interior of the earth consists of a molten magma, about 50 to 100 miles thick, surrounding a gaseous centre. A gradation exists from the central gaseous nucleus, through the intermediate molten mass, to the outer solid crust.

The interior of the earth is intensely hot and is at the same time under enormous pressure. There is progressive increase in temperature with depth. Though, the increase is not uniform at all depths and all places, there is a rise of 1oF for every 64 feet on an average. Assuming the temperature to increase at this rate, at a depth of 25 to 50 miles, it should be sufficiently high to melt all substances known to us. This indicates that the interior of the earth is in molten condition.

Composition of Earth crust

The Earth's crust is principally compassed of mineral matter. This mineral matter is made up of various elements combined together to form compounds. Some elements exist as such without forming compounds. Almost all the elements known to us at present, except the inert gases, are present in the earth's crust. The elements do not exist in the earth's crust as such. Each element is in combination with one or more other elements to form definite chemical compounds known as **minerals**. Many of these minerals in turn combine together to form aggregates, which we know as **rocks**. Almost all the mineral mater is present in the form of rocks in the earth's crust. Rock is composed of elements, which in turn are made up of atoms. Out of 106 elements knows, 8 are sufficiently abundant as to constitute about 99 percent by weight of the Earth's Crust (upto 16 Km). The elements are geochemically distributed into five main groups based on their **bonding characters**.

Lithophile elements- which ionize readily or form stable oxyanions, viz. O,Si, Ti, Fe, Mn, A1, H, Li, Na, K, Rb, Cs, Be, Mg, Ca, Sr, Ba, B, Ga, Ge, Sn, Sc, Y, F,C1, Br, I, C, HF,Th, P,V, Nb, Ta, Cr,W, U, Zr, (Mo), (Cu), (Zn), (Pb), (T1), (As), (Sb), (Bi), (S), (Se), (Te), (Ni), (Co) and rare earths.

Chalcophile element – which tend to form covalent bonds with sulphide, viz, S,Se, Te, (Fe), Ni, Co, Cu,Zn, Pb, Mo, Ag,Sb,(Sn), Cd, In,T1,Pb, As,Bi, Re, (Mn), (Ga) and Ge).

Siderophile elements – which readily form metallic bonds, viz, Fe, Ni, Co,Ru, Rh,Pd, Ir, Os and Au.

Atmosphile elements- which tend to remain in atmospheric gases, viz. N, (O) He, Ne, Ar, Kr, Xe.

Biophile elements- which tend to be associated with living organisms, viz. C,H,O,N, P,S,C1, I, B, Ca, Mg, K, Na, Mn, Fe, Zn, Cu, Ag, Mo, Co, Ni, Au, Be, Cd, Se, T1,Sn, Pb, As and V.

Average composition of the Earth's Crust (% by weight)

Non- metallic	Oxygen (O_2)	46.60%	74.32% (3/4th of total)
	Silica (Si^{4+})	27.72%	
Metallic	Aluminium (Al^{3+})	8.13%	25.68% ¼ th of total)
	Iron (Fe^{2+})	5.00%	
	Calcium (Ca^{2+})	3.63%	
	Sodium (Na^{+})	2.83%	
	Potassium (K^{+})	2.59%	
	Magnesium (Mg^{2+})	2.09%	
	Others	1.41%	

Lecture 3 & 4 : Soil genesis: soil forming rocks-definition, formation, Classification of rocks- igneous, sedimentary and metamorphic rocks, Brief description of important rocks - mineralogical composition

What is Rocks?

Rocks are the materials that form the essential part of the Earth's solid crust. "Rocks are hard mass of mineral matter comprising one or more rock forming minerals". Rocks are formed from the molten material known as magma. The study of rocks is called Petrology (in Greek, petra means rock, logos means science). Petrology deals with the description of rocks; petrogenesis is the study of the origin of rocks.

Formation of rocks

1. Cooling and consolidation of molten magma within or on the surface of earth = Igneous or Primary rocks
2. Transportation and cementation of primary rocks = Sedimentary or Secondary rocks
3. Alteration of the existing primary and secondary rocks = Metamorphic rocks.

Petrology	-Study of rocks (in Greek, *petra* means rock, *logos* means science)
Petrography	- Description of rocks
Petrogenesis	-Study of the origin of rocks.

1. Igneous rocks (primary or massive rocks)

These are first formed in the earth crust due to the solidification of molten magma. Based on the mode of formation, they are further classified as extrusive and intrusive rocks.

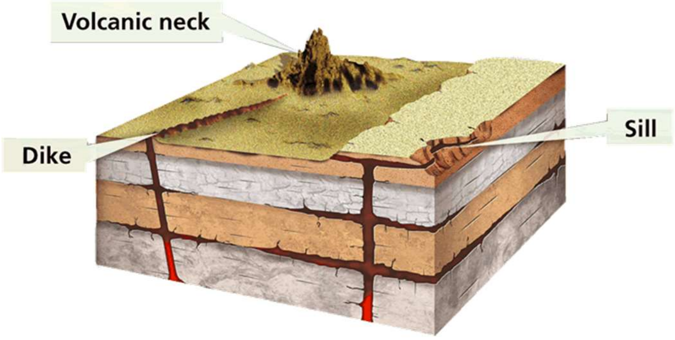
Extrusive rocks or volcanic rocks

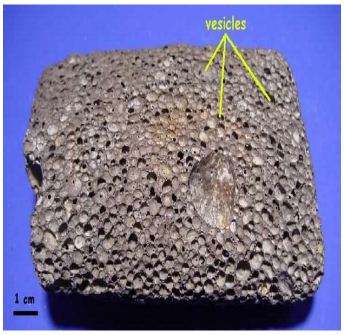
These rocks are formed due to the consolidation of magma on the surface of the earth. The magma, when it flows on the Earth surface is called LAVA. Eg. Basalt.

Intrusive rocks or plutonic rocks

These rocks are produced due to solidification of magma below the surface of the earth.

Plutonic – intrusive rocks solidifies at greater depth and Hypabassal rocks solidifies at shallow depth from the surface. E.g. Granite, syenite, diorite, Gabbro etc. Rocks formed in vertical cracks are called dykes and in horizontal cracks are called sills.





Vesicular rocks:

Molten magma cools on the surface. Steam of water is entrapped into rocks and forms vesicles.

Based on the silica content, rocks are also classified as

1. Acid rocks : >65% SiO_2 **(Granite, Rhyolite)**
2. Intermediate: 56 to 65% SiO_2
   (Sub acid rocks 60 to 65% SiO_2 **(Syenite &Trachyte)**
   (Sub basic rocks 56 to 60 % SiO_2 **(Diorite & Andesite)**
3. Basic rocks : 40 to 55% **(Gabbro & Basalt)**

General characters of Igneous rocks

S.No	Rocks	Origin	Essential Mineral	Common minerals	Average specific gravity	Remarks
1	Granite	Plutonic holocrystalline	Quartz (20 to 30%)	Hornblende, Magnetite, Mica	2.64	Light coloured white or reddish
2	Syenite	Plutonic holocrystalline	Quartz, orthoclase	Hornblende, Magnetite, Mica	2.8	Light coloured white or reddish
3	Diorite	Plutonic holocrystalline	Quartz	Hornblende, Magnetite, Mica	2.85	Darker
4	Gabbro	Plutonic holocrystalline	Labradorite, augite, olivine	Hornblende, Ilmenite	3	Blackish
5	Dolerite	Hypabasal	Labradorite, augite, olivine	Hornblende, Ilmenite	3	Blackish
6	Basalt	Volcanic crystalline with glassy mass	Labradorite, augite, olivine	Hornblende, Ilmenite	3	

2. Sedimentary rocks

These rocks are formed from the consolidation of sediments accumulated through wind or water action at the surface of the earth. Many are deposited in layer or formed through chemical reactions as precipitates from aqueous solutions. Sediments may contain various size particles cemented together by substances like SiO_2, Fe_2O_3 or lime. These rocks are also called as clastic rocks.

Formation of Sedimentary Rocks

Sedimentary rocks are formed by four stages. These are 1.Weathering, 2.Transportation, 3.Deposition and 4. Diagenesis.

1. **Weathering:** Igneous rocks and other rocks disintegrate owing to physical, chemical and biological weathering. The products of these weathering agencies provide the basic materials (gravels, sand, silt and mud) for the formation of secondary rocks.
2. **Transportation:** The disintegrated materials are transported by water, wind, glaciers, runoff and gravity. The dust clouds in dry months carry finer fractions by wind it is called **saltation**. The movement of loose material by water is called **soil creep**.
3. **Deposition or sedimentation:** The transported materials are deposited when the carrying agent has no longer energy enough to move it further. It is called **graded bedding**.
4. **Diagenesis:** The transformation of unconsolidated sediments to hard rock. It involves compaction and cementation. In compaction process, the weight of the upper layer give very high compression to lower layers. It forms sedimentary rocks. The fine grained deposits under such conditions are transformed to clay, shales, etc. In cementation process, the deposit materials contain cementing agents like lime, silica and iron oxide. The water helps to bind the materials and form the sedimentary rocks.

Classification of sedimentary rocks

Based on their origin, the sedimentary rocks are grouped into two main classes

(1) Fragmental, Detrital or Mechanically formed

(2) Chemically formed

- (a) Inorganically formed
- (b) Organically or biochemically formed

1. Fragmental, Detrital or Mechanically formed

These sedimentary rocks are formed by the deposition and cementation of erosion products of pre-existing rocks. These weathering materials are transported by air, water or ice. These fragmental material derived and deposited in beds of varying thickness and cemented. E.g. **Breccia**, **Shale**, **Conglomerate** and **Sandstone**.

2. Chemically formed

(a) Inorganically formed

These rocks are formed by evaporation or precipitation of material dissolved in sea water or lack water. **Halite** (rock salt), **Gypsum** and **Anhydrite** rocks are formed by evaporation process. **Limestone** and **Dolomite** are formed by precipitation and flocculation. The less soluble minerals like gypsum and anhydrite are first to precipitate. In evaporation process, the more soluble halite is precipitated.

$$Ca(HCO_3)_2 \xleftrightarrow[\text{Temp.}\downarrow,\ CO_2\ \uparrow]{\text{Temp.}\uparrow,\ CO_2\ \downarrow} CaCO_3 + CO_2 + H_2O$$

(b) Organically or Biochemically formed

These rocks are formed by accumulation and potential decomposition of organic remains under anaerobic conditions. Carbonated rocks such as peat, lignite, bituminous, semi-bituminous coal and anthracite (hard coal) are formed by this process.

Based on the origin, the sedimentary rocks are classified as

1. Residual : Laterite
2. Transported
   a. Deposited as solids in suspension : Sandstone, shale
   b. Deposited by chemical precipitation : Limestone, ironstone
   c. Deposited through agency of organic matter: Peat, Phosphatic deposits

Based on the grain size, sedimentary rocks are classified as

1. Rocks with boulder pebbles sized minerals (Rudaceous) : Conglomerate
2. Rocks with sand size particles (Arenaceous) : Sandstone
3. Rocks with silt size particles (silt rocks): Siltstone
4. Rocks with clay size particles (Argillaceous): Shale

Some important sedimentary rocks

S.No	Rock	Mineral composition	Colour and structure
1	Sandstone	Mainly quartz with some $CaCO_3$, iron oxides and clay	Light to red, granular
2	Shale	Clay minerals, quartz and some organic matter	Light to dark thinly laminated
3	Limestone	Mainly calcite with some dolomite, iron oxides, clay, phosphate and organic matter	Light grey to yellow, fine grained and compact
4	Conglomerate	Composed of more or less rounded fragments. Fragments are more angular than rounded, the rock is called breccia. Calcite or quartz cement that binds the rock together.	Black, brown, buff, light to dark grey, orange, rust, white, yellow colors

Sedimentary rocks are also formed through the agency of water. Also called as **aqueous rocks**. Formed from sediments brought by water. The sediment may contain various types of substances and varying sizes of particles. The particles are cemented by silica, iron oxide and lime to give a consolidated form. The rocks are mostly deposited in layers or strata, so called as **stratified rocks**. Sometimes they are formed by cooling, evaporation or by direct chemical precipitation.

Sedimentary rocks divided into six groups as follows

1. **Arenaceous rocks**

Formed of the deposits of coarse grained particles. They are composed of siliceous material derived from the disintegration of older rocks. The fragmental material so derived is deposited in beds of varying thickness through the agency of water. Depending upon the nature of cementing material present, some arenaceous rocks are hard and refractory, but most are loose and fall away very easily. E.g. **Sandstone, Grit, Conglomerate** and **Breccia**.

2. **Argillaceous rocks**

Consist of small sized particles such as clay. They are composed of hydrated silica of alumina in admixture with sand, various other silicates and calcareous matter. When clay is deposited mainly of silicate of alumina, it is known as kaolin or China clay. E.g. **Clay, Mudstone, Shale** and **Fuller’s earth.**

3. **Calcareous rocks**

Consist of carbonate of lime or lime and magnesia. They may be of sedimentary origin or formed by chemical precipitation or by organic agency. When they are of organic agency, and they are composed mainly of debris from plant and animal life. They are formed either by growth and decay of organisms in situ or by the transport and subsequent accumulation of their remains. The rocks so formed are found in layers, which vary considerably in depth of thickness.

When formed by chemical precipitation, the calcareous material is deposited in the form of layers/sheets from waters containing calcium carbonate in solution. The precipitate when first formed is usually soft and chalky, but soon acquires a hard, compact structure and crystalline texture. E.g. **Limestone, Chalk, Magnesian, Ferruginous limestones, Dolomite, marks of various varieties** and **coral.**

4. Carbonaceous rocks

These are formed from decomposing vegetation under anaerobic conditions. When plants undergo decomposition under restricted air supply, with greater portion of the carbonaceous matter is retained and the material is slowly converted into coal. E.g. **Peat, Lignite, Coal** and **Anthracite.**

5. Siliceous rocks

Siliceous rocks of organic origin formed from parts of minute plants and animals like diatoms, radiolaria etc. Some are soft and friable and crumble to powder very easily. Others like flint and chert are hard and compact. E.g. **Flint** and **Chert**

6. Precipitated salts

Consist mainly of deposits formed as rock masses either by cooling, evaporation or by chemical precipitation. Water charged with acid or alkaline material, acting under pressure as it does under subterranean regions, dissolves various mineral substances from rocks with which it comes in contact. The salts thus formed deposit as rocks and such rocks vary in composition. They are

i. **Oxides** : Hematite, limonite, bauxite and quartz.
ii. **Carbonates** : Stalactite, stalagmite, magnetite and limestone.
iii. **Sulphates** : Gypsum and anhydrite
iv. **Phosphates** : Phosphorite
v. **Chlorides** : Rock salt.

3. Metamorphic rocks

These are formed from igneous and sedimentary rocks under the influence of heat, pressure, chemically active liquids and gases. Change may occur in mineral composition or texture or both. The change due to water is called **hydro metamorphism**, due to heat is called **thermal metamorphism**, due to pressure is called **dynamo metamorphosis** and due to pressure and heat is called **dynamo thermal metamorphosis**.

The changes that are brought about are both physical and chemical in character. In some cases the metamorphism is so pronounced that the new rock looks quite different from the original. The action of water tends to remove some material or introduce new materials.

By the introduction of a cementing material like silica, lime or iron oxide, loose sand may be turned into sandstone or sandstone into a quartzite. By the removal of certain constituents by percolating waters, basalt or granite may be converted into a laterite.

The action of heat hardens the rock and develops new crystals in it. Crystalline marble is produced this way from amorphous limestone by the action of heat and pressure. Due to pressure, the crystals of the original rock get pressed or flattened and the new rock is foliated.

When foliation is slight, the layers are inseparable and it is called as **gneiss**. If foliation is complete with distinct and separable layers it is called as **schist**.

Original rock		**Metamorphosed rock**
Sand stone	-	Quartzite
Shale	-	Slate/mica, schist
Lime stone	-	Marble
Granite	-	Granite gneiss
Dolerite	-	Hornblende gneiss

Classification of metamorphic rocks

They are classified in to three groups, and they are 1. Foliated (Parallel structure), 2. Non foliated (Massive structure) and 3. Granular

1. Foliated

It means leaved or leafy. The rocks contain micas and ferromagnesian minerals and show foliation, as the minerals are flattened and arranged in parallel layers. E.g. **Gneiss, Schist, Phyllite and Slate**

2. Non foliated

The rocks, that contain quartz and feldspars, do not show foliation even under pressure because of their coarse-sized crystals. E.g. **Anthracite, Talc, schist and Amphibolite**

3. Granular

The rocks comprising mostly of equidimensional grains. E.g. **Quartz and Marble**

Classification of metamorphic rocks by texture

Grain size	Grain size		
	Fine	**Medium**	**Coarse**
Poorly foliated	Hornfels	Marble, quartzite	Marble, quartzite
Well foliated	Slate	Schist	Gneiss
Well foliated and sheared	Mylonite	Mylonite, schist	Augen gneiss

Some important metamorphic rocks

S.No	Rock	Mineral composition	Colour and structure
1	Gneiss	Formed from granite	Alternating light and dark colours, banded and foliated
2	Schist	Formed from basalt or shale	As original rock, foliated
3	Quartzite	Formed from sandstone	Light to brown, compact and uniform texture, foliated structure
4	Slate	Formed from shale	Grey to black, compact and uniform texture, foliated structure
5	Marble	Formed from lime stone	Light red, green, black, compact fine to coarse texture, foliated structure

Rocks composition of the upper 5 km of the Earth's crust

Sedimentary Rocks		
Shale	52%	74%
Sandstones	15%	
Limestone and dolomite	7%	
Igneous Rocks		
Granite	15%	18%
Basalt	3%	
Others	8%	8%
Total (%)	100	100

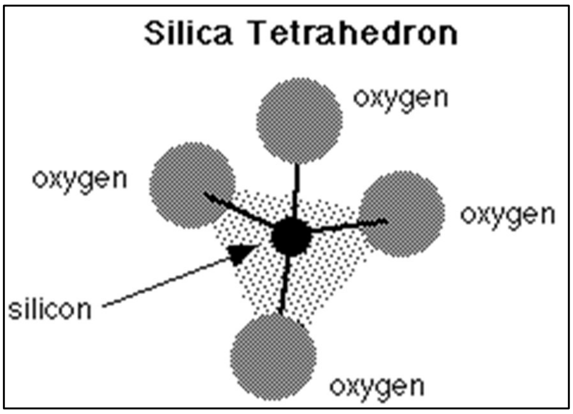
Rocks composition as a whole of the Earth's crust

Igneous rocks	**95%**
Sedimentary rocks	**5%**

Shales (4.0%)
Sandstones (0.75%)
Limestones (0.025%)

Lecture 5: Minerals- definition, occurrence, classification of important soil forming primary minerals - silicate and non silicate minerals, ferro and non-ferro magnesium minerals

Minerals



Minerals are naturally occurring solids with a definite chemical composition and crystal structure. “Solid substances composed of atoms having an orderly and regular arrangement”. When molten magma solidifies, different elements present in them freely arrange in accordance with the attractive forces and geometric form. Silica tetrahedron is the fundamental building blocks for the formation of different minerals. (SiO_2). Different silicate minerals are ortho silicates, ino-silicates, phyllosilicates and tectosilicates. There are nonsilicate minerals also. These are different oxides, carbonates, sulphates, phosphates *etc*.

Primary Minerals

Minerals that are original components of rocks called as **primary minerals.** E.g. **Feldspar, Mica**, etc.

Secondary Minerals

Minerals are formed from changes in primary minerals, and rocks are called as **secondary minerals.** E.g. **Clay minerals**.

Essential Minerals

Minerals those are chief constituents of rocks are called as **essential minerals.** E.g. **Feldspars, Pyroxenes, Micas** etc.

Accessory Minerals

Minerals those which are present in small quantities, whose presence or absence will not alter the properties of rocks are called **accessory minerals** (E.g. **Tourmaline**, **Magnetite** etc).

Types of minerals

Minerals	Important constituents	Percent distribution
A.Primary minerals		
I. Ferromagnesium minerals		
i. Ortho-ino-silicates		
Olivines	Fe, Mg	**16.8 %**
Pyroxenes	Ca, Na, Fe, Mg	
Amphiboles	Ca, Na, Fe, Mg, Al, OH	

ii. Phyllosilicates		
Biotite	K, Fe, Mg, Al, OH	**3.6%**
Muscovite	K, Al, OH	
II. Non-ferromagnesians		
• Tectosilicates		
Feldspars, such as:		
Anorthite	Ca, Al	**61.0%**
Albite	Na, Al	
Orthoclase	K, Al	
Quartz		**11.6%**
B. Secondary clay minerals	Na, K, Ca, Mg, Fe, Al, OH	
C. Others		**6%**
Total		**100%**

A. Primary Minerals

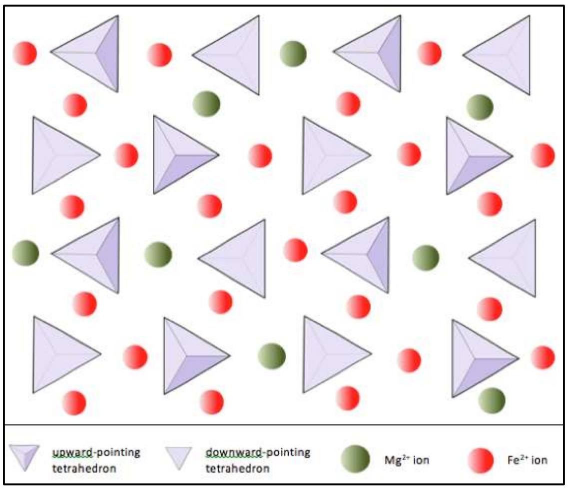
Primary minerals are formed from crystallization of molten magma, and it is composed of oxygen (46.6%) and silicon (27.72%). Primary minerals are silicate minerals, and it contains the basic component of silicon tetrahedron linkage ($(SiO_4)^{4-}$).

I. Ferro magnesium minerals (Silicate minerals)

i. Ortho/ Neosilicates

The minerals in this group are composed of single tetrahedral linked together by Mg or Fe. To effect a break down, it is considered sufficient to sever the weaker Mg-O or Fe-O bonds.

Non-withstanding the bond energy considerations susceptibility of the minerals in this group to breakdown by weathering appears to vary considerably from one mineral to another, E.g. **Zircon** makes the mineral comparatively hard.



ii. Inosilicates

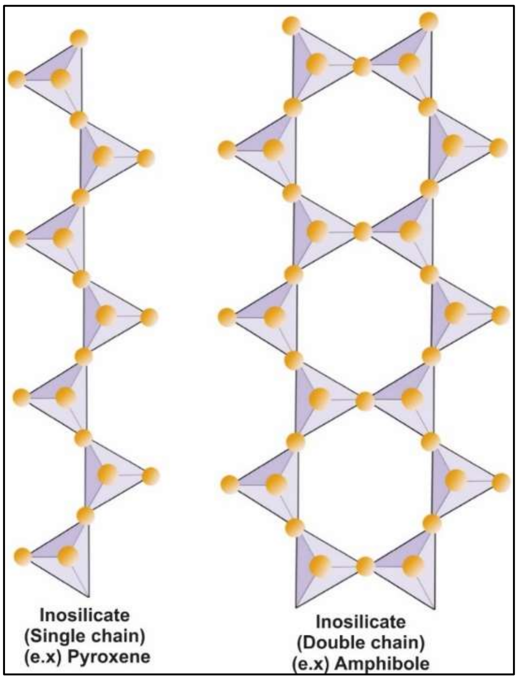
The inosilicate group has single-chain (pyroxenes) and double chain (amphiboles) silica tetrahedral linked together by Ca, Mg, or Fe. Because of the presence of many weak spots provided by the Ca-O, Mg-O, or Fe-O bonds, these minerals tend to weather rapidly.

Pyroxenes and Amphiboles

The pyroxenes and amphiboles are two groups of ferromagnesian minerals (heavy group) the structure of which consists of long chains of linked silica tetrahedral.

The pyroxenes consist of a **single chain** (2 oxygen shared in each tetrahedron) whereas amphiboles consist of a **double chain** (alternately 2 and 3 oxygen atoms shared with successive tetrahedral).

These chain silicates are sometimes referred to inosilicates.



i. Phyllosilicates

The phyllosilicates, or sheet silicates, are an important group of minerals that includes the micas, chlorite, serpentine, talc, and clay minerals. The clay minerals as one of the primary products of chemical weathering and one of the more abundant constituents of sedimentary rocks. The basic structure of the phyllosilicates is based on interconnected six member rings of SiO_4^{-4} tetrahedra that extend outward in infinite sheets. Three out of the 4 oxygens from each tetrahedra are shared with other tetrahedra. This leads to a basic structural unit of $Si_2O_5^{-2}$.

Most phyllosilicates contain hydroxyl ion, OH^-, with the OH located at the center of the 6 membered rings. Thus, the group becomes $Si_2O_5(OH)^{-3}$. When other cations are bonded to the SiO_4 sheets, they share the apical oxygens and the (OH) ions which bond to the other cations in octahedral coordination. This forms a layer of cations, usually Fe^{+2}, Mg^{+2}, or Al^{+3}, that occur in octahedral coordination with the O and OH ions of the tetrahedral layer. The triangles become the faces of the octahedral groups that can bind to the tetrahedral layers.

The octahedral layers take on the structure of either Brucite [$Mg(OH)_3$], if the cations are +2 ions like Mg^{+2} or Fe^{+2}, or Gibbsite [$Al(OH)_3$], if the cations are +3 like Al^{+3}. In the brucite structure, all octahedral sites are occupied and all anions are OH^{-1}. In the Gibbsite structure every 3rd cation site is unoccupied and all anions are OH^{-1}.

This gives rise to 2 groups of sheet silicates

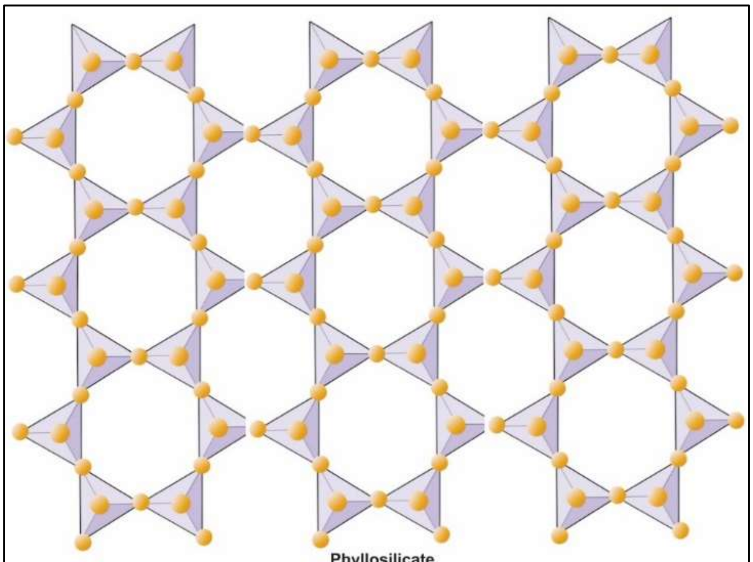
1. The **trioctahedral** sheet silicates where each O or OH ion is surrounded by 3 divalent cations, like Mg^{+2} or Fe^{+2}.
2. The **dioctahedral** sheet silicates where each O or OH ion is surrounded by 2 trivalent cations, usually Al^{+3}.

Some of the minerals, E.g. **Biotite** (black mica) and **Muscovite** (white mica), are relatively susceptible to weathering, whereas others, like clay minerals, are resistant weathering products and further breakdown of clays is difficult. Disruption of interlayer ions, or through cleavage of A1-O bonds in tetrahedral and octahedral positions.

Micas

Micas occur extensively in soils. They primarily originated from the parent rock from which the soil is derived. The **sheet structures** are phyllosilicates, of which mica are a typical example.

Generally soils are inherited from well-ordered and imperfectly ordered micas. Well-ordered micas are derived from sedimentary rocks.



Phyllosilicate
Sheet structure
(e.x) Mica

The most common well-ordered micas are **muscovite, paragonite, biotite and phlogopite** (trioctahedral).

The imperfectly ordered micas contain less potassium and more water as compared to well-ordered micas and this type of micas are most abundant in the clay fraction of soils.

Among the ordered micas, biotite weathers more rapidly than muscovite. In imperfectly ordered micas, many of the illite type specimens as well as the disordered micas of soils exhibits some mixed-layering with phases of vermiculite, smectite group of minerals, chlorite and intergrades of several of these species.

Non ferromagnesium minerals

i. Tectosilicate

The tectosilicates or framework silicates have a structure wherein all of the 4 oxygens of SiO_4^{-4} tetrahedra are shared with other tetrahedra. The ratios of Si to O is thus 1:2. Since the Si - O bonds are strong covalent bonds and since the structure is interlocking, the tectosilicate minerals tend to have a high hardness.

Feldspars

Feldspars are anhydrous alumino silicates of K, Na and Ca and occasionally of other large cations such as Ba. The feldspar structure consists of tetrahedral which are attracted by sharing each oxygen atom between neighbouring tetrahedran. The tetrahedral contain mainly Silicons with sufficient Al substitution. It belongs to the group of minerals that are light in weight.

(i) Potassium feldspars ($KAlSi_3O_8$)

It is include **orthoclase, microcline, adularia and sanidine**. Orthoclase and microcline are more common in the plutonic and metamorphic rocks. The potassium feldspars occur commonly in the silts and sands of soils and also abundant in clay-size.

Quartz (SiO_2)

Quartz is a mineral composed of silicon and oxygen atoms in a continuous framework of SiO_4 silicon–oxygen tetrahedra, with each oxygen being shared between two tetrahedra, giving an overall chemical formula of SiO_2. Quartz is the second most abundant mineral in Earth's continental crust, behind feldspar.

It is very densely packed and occurs in a high degree of purity.The hardness of the mineral is 7. It is strongly resistant to weathering as the structure is densely packed, electrically neutral and free from any substitution. It is the most abundant mineral next to feldspars. The specific gravity the mineral is 2.65. **Serpentine** is a hydrous magnesium silicate occurs more commonly as a secondary product. **Garnets** are characteristic of metamorphic rocks and are very hard and most resistant to weathering.

Non-silicate minerals

i. Oxides and Hydroxides

The oxides are usually harder than any other mineral, except the silicates. The most important soil-forming oxide minerals are

(E.g.) Hematite (Fe_2O_3)
Limonite (Fe_2O_3 $3H_2O$)
Goethite (FeO (OH) H_2O)
Gibbsite ($Al_2O_3H_2O$)

The red, yellow or brown colours in soils are due to the presence of goethite and hematite, which occur as coatings on the surface of soil particles.

ii. Carbonates

The basic compounds, like $Mg(OH)_2$ and $Ca(OH)_2$, combine with CO_2 or carbonic acid (H_2CO_3) to form carbonates. E.g. **Calcite** ($CaCO_3$), **Dolomite** ($CaMgCO_3$), **Siderite** ($FeCO_3$)

iii. Sulphates

Sulphur is a complex group formed by the combination of 1 sulphur and 4 oxygen ions, which further reacts with Ca to form calcium sulphate (anhydrite, $CaSO_4$). On hydration it forms gypsum ($CaSO_4 . 2H_2O$).

Physical properties of minerals

1. Color
2. Streak
3. Fracture/ cleavage
4. Hardness
5. Luster
6. Crystal form
7. Taste
8. Specific gravity
9. Magnetism
10. Effervescence (fizz)
11. Birefringence
12. Fluorescence

Additional reading

Color

- Denotes the natural colour of the mineral
- The most obvious, but least reliable.
- Calcite has more colours
- Sulfur and Pyrite have same colour

Streak

Refers to the colour of the powder form of the mineral When an unknown mineral is rubbed against a piece of unglazed porcelain (streak plate) it produces a colored line.

- Hematite - red
- Magnetite - Black
- Talc – white

Fracture and Cleavage

These terms describe the way a mineral breaks Fracture is the nature of the surface produced as a result of its breakage.

Conchoidal - curved surface

Uneven - Uneven surface

Hackly - Jagged surface

Earthy - Like chalk

Even - Smooth

Cleavage

Some minerals break along certain well defined planes called cleavage planes.

Gypsum - 1 set

Calcite - 2 sets

Flourite - 3 sets

Hardness

This is how resistant a mineral is to being scratched. We use the Mohs scale to classify a given minerals hardness. Try to scratch the unknown mineral with various items, such as a fingernail (hardness of about 2.5), a coin (3), a steel nail (5.5) and a steel file (7).

MOHS SCALE OF HARDNESS

Mineral	**Hardness**	**Mineral**	**Hardness**
Talc	1	Feldspar	6
Gypsum	2	Quartz	7
Calcite	3	Topaz	8
Fluorite	4	Corundum	9
Apatite	5	Diamond	10

Luster

The way a mineral reflects light Metallic (Magnetite); sub-metallic, Vitreous (Opal), Resinous (Pyrite), Pearly, Adamentine (Diamond), silky (Asbestos) and greasy.

Crystal form

Crystal structure is the result of regular grouping of atoms that are homogeneous. A crystal is a polyhedral form, which means it is a geometric solid. It has a specific set of faces, corners and edges, which is consistent with the geometric packing of the atoms.

There are 6 basic crystal forms

1. Isometric 2. Tetragonal 3. Hexagonal
4. Orthorhombic 5. Monoclinic 6. Triclinic

Taste

This property is used to identify the mineral halite (salt).

Specific Gravity

This characteristic relates to the minerals density. If the mineral is heavy for its size, then it has a high specific gravity.

Magnetism

Is the mineral magnetic (try using a compass), or is it attracted by a magnet? This property is characteristic of Magnetite.

Effervescence

When some minerals are exposed to acids, they begin to fizz (calcite).

Birefringence

This is also known as double refraction. Birefringent minerals split the light into two different rays which gives the illusion of double vision in this Iceland Spar Calcite.

Fluorescence

Some minerals display the phenomenon of photoluminescence. They "glow" when exposed to UV light. Eg. Opal and Fluorite.

Lecture 6: Formation of secondary minerals – Clay minerals and amorphous minerals

Formation of secondary minerals, Clay minerals & Amorphous minerals

The secondary minerals are formed at the Earth's surface by weathering of the preexisting primary minerals under variable conditions of temperature and pressure. During weathering, water accompanied by CO_2, from the atmosphere plays an important role in processes, such as hydrolysis, hydration and solution. As a result the primary minerals are altered or decomposed.

Feldspar + water — clay mineral + cations + anions + soluble silica

Because of weathering, many elements are released into solution; a part of which may be used as a source of plant nutrients, a part may be leased out into the groundwater; still another part together with other constituents of the environment (like CO_2, H_2O) may recombine to form secondary minerals. The most commonly formed secondary minerals are clay minerals (e.g. illite, montmorillonite, kaolinite, etc.) and iron and aluminium oxides. Other secondary minerals observed in soils, especially in arid and semi-arid (dry) regions are gypsum, calcite, attapulgite and apatite.

SILICATES

- Clay minerals: hydrous aluminosilicates, with layer structure similar to micas, e.g. illite, montomorillonite, kaolinite, etc.

NON-SILICATES

--Oxides, hydroxides or oxyhydrates of Si, Al and Fe

Haematite - Fe_2O_3

Goethite; Limonite $FeO(OH)n\ H_2O$

Gibbsite $Al_2O_3.\ H_2O$

Clay Minerals

Clay minerals in soils are formed from primary minerals due to weathering processes. These clay minerals are of size <0.002 mm and are considered to be the most reactive part of soil. Important soil properties like nutrient and water holding capacity are controlled by clay minerals. These minerals are layered silicates consisting of silica tetrahedron and aluminium octahedron.

1) 1 silicon tetrahedron + 1 aluminium octahedron = 1:1 clay mineral (Kaolinite)

2) 2: 1 non-expanding clay mineral i. Black mica (Biotite)

ii. White mica (Muscovite)

iii. Weathered mica (illite)

3) 2: 1 expanding clay mineral i.Partially expanding (Vermiculite).

ii.Fullyexpanding (Montmorillonite)

4) 2: 2 clay mineral (chlorite)

Of the naturally-occuring inorganic crystalline minerals found in the clay fraction of soils, the most commonly observed are layer silicates (illite, montmorillonite, chlorite, vermiculite, kaolinite). Besides, O, OH, A1 and Si, they contain Mg, Fe and K in large amounts. They are variable in colour (white, grey, light yellow) depending on their chemical composition. In soils, the clays and oxihydrates of iron which form coatings on mineral grains impart shades of yellow, brown or red colour to soils.

The clay minerals carry a significant negative electrical charge on their surfaces and have a structure like that of mica .In some cases, the groups of sheets are not firmly bounded together and water molecules can enter in their crystal lattice. This can cause considerable swelling due to change in soil moisture content. This is case in Vertisols (black cotton Soils) of India and NE Iraq, where deep and wide cracks on the surfaces are suggestive of the shrink-swell characteristics of soil clays.

Owing to the negative electrical charge on the clay surfaces, the cations are attracted to regions of electrical charge around the clay minerals these cations do not get bounded permanently and can be exchanged for other cations. The amount of charge varies depending upon the type of clay mineral and it is referred to as the cation exchange capacity. Because of this exchange, there is always a balance between the concentration of cations in soil water and those adsorbed on the surfaces of the particles. Rain water percolating through the soil leaches out many meta cations (K, Na, Ca, Mg) together with the existing soil-water and replaces it with new water containing H+ ions and may render the soils acidic in reaction. The H+ concentration is expressed in terms of pH. A soil with high pH (>7) is alkaline while a soil with low pH (7) is acidic in nature.

 pH is expressed in terms of negative log to the base 10 of H+ ion concentration.

Distribution

While primary minerals are observed in all rocks and in sand and silt fractions of soils, the secondary minerals dominantly occur in the clay fractions of almost all soils and in sedimentary rocks, especially shales. The kind and proportion of mineral(s) observed in a soil depend on the kind of parent material and weathering intensity (to which it has been exposede.) The most common clay mineral observed is illite. Apart from illite, smectite predominates in the cracking-clay soils (of Australia, northern Iraq and central India northeast Africa), kaolinite in the highly-weathered soils of the inter tropical zones (of southern India, South America, S.E. Asia) and southern Iraq, western India). In view of their high surface area and negative charge on them, they are considered as a source of cation adsorption and cation release which are so important in acidic soil fertility.

Non-Silicates

Oxides, Hydroxides or Hydrous-oxides group

We have already seen that oxygen is present in great abundance (46.7%) in the Earth's crust. The oxide minerals are found by the direct combination of elements (present in the Earth's crust) with oxygen.

The oxides are usually harder than any other mineral, except the silicates. The most important soils-forming oxide minerals are:

Haematite : Fe_2O_3
Limonite : $Fe_2O_3 .3H_2O$
Goethite : FeO (OH).nH_2O
Gibbsite : $A1_2O_3.H_2O$

Haematite, Fe_2O_3

It varies in colour from red to blackish and has reddish streak. It has a metallic luster and hardness (H) of about 5. Its presence in rocks is indicative of quick chemical change. Haematite alters to limonite, magnetite, pyrite and siderite. It occurs as coatings on sand grains and acts as a cementing agent. It swells on absorbing water to form hydrated iron oxide, i.e. limonite, $2Fe_2O_3$ $3H_2O$ and goethite, FeO (OH).nH_2O.

Limonite or Bog Iron, $2Fe_2O_3,3H_2O$

It is hydrated ferric oxide, yellow to brown in colour and is of wide occurrence. It is the final product of most iron minerals and hence is resistant any further change, except for absorption of water. It is an important colouring and cementing agent in soils, iron. Limonite is a common alteration product of pyrite, magnetite, hornblende and pyroxene. It may be present in the form of iron concentration.

Goethite, FeO(OH)nH_2O

Most materials, called limonite, are goethite with some adsorbed water. It is usually white but may pink or grey in colour. Its hardness is 5.3

Gibbsite (Hydragillite), $A1_2O_3H_2O$

It is the most common aluminium compound in soils. Its natural colour is white. It is abundantly observed in highly-weathered soils of the tropical environment, supporting Laterites (Oxisols). It's present in soils suggests extreme degree of weathering and leaching under well drained conditions. The red, yellow or brown colour in soils is due to the presence of goethite and hematite which occur as coating on the surfaces of soil particles, especially clay.

Carbonate Group

The basic compounds, like Mg $(OH)_2$, and Ca(OH) combine with CO_2 or carbonic acid to form carbonates as under:

Calcite, $CaCO_3$

A white mineral, with hardness of 3, is widely distributed in sedimentary rocks, like limestone and decomposes easily to calcium bicarbonate as:

$$CaCO_3 + CO_2 + H_2O \text{ --- } Ca(HCO_3)_2 \text{ (soluble in water)}$$

Dolomite, Ca Mg $(CO_3)_2$

Dolomite is less-readily decomposed than calcite; it is the chief source of Mg in soils.

Siderite, $FeCO_3$

It is an alteration product of other iron-bearing minerals, having hardness of 4 and may itself alter to hematite or limonite. It is an important mineral in waterlogged soils.

Sulphate Group

Sulphate is a complex group formed by the combination of 1 sulphur and 4 oxygen ions, which further reacts with Ca to form calcium sulphate (anhydrite, $CaSO_4$) On hydration it forms gypsum ($CaSO_4.2H_2O$).

Gypsum, $CaSO_4$ $2H_2O$

It is a common mineral in desert soils and in sedimentary rocks having a hardness of 2. It is slightly soluble in water and gets most-easily leached. It precipitates as very fine, powdery mycelium from ground waters rich in Ca and SO_4 ions (as observed in the Mesopotamian Plain of Iraq where hyper aridic prevail). In India, it is used as an amendment to reclaim sodic soils and also acts as a source of Ca and S for plants. Under the hot, aridic climatic environments of Iraq, the presence of gypsum in high amounts is a problem, as it causes civil structures to collapse and makes sink-holes in soils, resulting in loss of irrigation water.

Phosphate Group

Apatite, Rock Phosphate

It is a primary source of phosphorus in soils. Its hardness is 5 in mho's scale. It decomposes readily under the influence of carbonic acid. It becomes immobile in calcareous soils as it readily combines with clays, with clays, Fe-A1 hydrous oxides, calcium carbonate to form rock phosphate. It also precipitates under acidic environment, as Fe and/or A1-phosphate.

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Lecture 7: Weathering of rocks and minerals - Physical, chemical and biological weathering

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 it's upper portion. May be defined as the unconsolidated and more or less chemically weathered mineral material from which soil are developed.

Weathering

Two basic processes

Physical /mechanical **Chemical**
(disintegration) **(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.

Physical weathering

The rocks are disintegrated and are broken down to comparatively smaller pieces, without producing any new substances.

1. Physical condition of rocks

The permeability of rocks is the most important single factor. Coarse textured (porous) sand stone weather more readily than a fine textured (almost solid) basalt. 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.

- 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.

- This alternate expansion and contraction weakens the surface of the rock and crumbles it because the rocks do not conduct heat easily.
- The minerals within the rock also vary in their rate of expansion and contraction

1. The cubical expansion of quartz is twice as feldspar
2. Dark coloured rocks are subjected to fast changes in temperature as compared to light coloured rocks.

- The differential expansion of minerals in a rock surface generates stress between the heated surface and cooled un expanded parts resulting in fragmentation of rocks.
- 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

- Moving water has the great cutting and carrying force.
- It forms gullies and ravines and carries with the suspended soil material of variable sizes.
- 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	**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

- In cold regions, the water in the cracks and crevices freezes into ice and the volume increases to one tenth.
- 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, montmorilonite.

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

iv). Action of glaciers

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

4. Action of wind

- 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.
- 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

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 → 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.

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.

$2Fe_2O_3 + 3H_2O \longrightarrow 2Fe_2O_3.3H_2O$

(Hematite) (Limonite)

$Al_2O_3 + 3H_2O \longrightarrow Al_2O_3.3H_2O$

(Bauxite)

$CaSO_4 + 2H_2O \longrightarrow CaSO_4.2H_2O$

(Anhydrite) (Gypsum)

2. Hydrolysis

Most important process in chemical weathering. It is due to the dissociation of H2 O 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.

Feldspar + Water → Clay mineral + Soluble cations and anions

$KAlSi_3O_8 + HOH \longrightarrow HAl\ Si_3O_8 + KOH$

(Orthoclase) (Acid silicate clay)

$2HAl\ Si_3O_8 + 8HOH \longrightarrow Al_2O_3.3H_2O + 6H_2SiO_3$

(Recombination) (Silicic acid)

This reaction is important because of two reasons

- clay, bases and silicic acid - the substances formed in these reactions - are available to plants
- water often containing CO_2 (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^{-}, HCO_3^{-}) 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.

$$NaCl + H_2O \longrightarrow Na^+ Cl^-, H_2O \text{ (Dissolved ions with water)}$$

4. Carbonation: Carbon di oxide 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.

$$CO_2 + H_2O \longrightarrow H_2CO_3$$

$$\underset{\text{(Insoluble)}}{CaCO_3} + H_2CO_3 \longrightarrow \underset{\text{(Soluble)}}{Ca(HCO_3)_2}$$

5. Oxidation

The process of addition and combination of oxygen to minerals. The absorption is usually from O_2 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.

$$2FeO + O_2 \longrightarrow 2Fe_2O_3$$

$$\underset{\text{(Magnetite)}}{4Fe_3O_4} + O_2 \longrightarrow \underset{\text{(Haematite)}}{6Fe_2O_3}$$

$$\text{Further, } \underset{\text{(Haematite)}}{2Fe_2O_3} + 3H_2O \longrightarrow \underset{\text{(Limonite)}}{2Fe_2O_3.3H_2O}$$

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.

$$2Fe_2O_3 - O_3 \longrightarrow 2FeO$$

(Haematite) Ferrous oxide (reduced form)

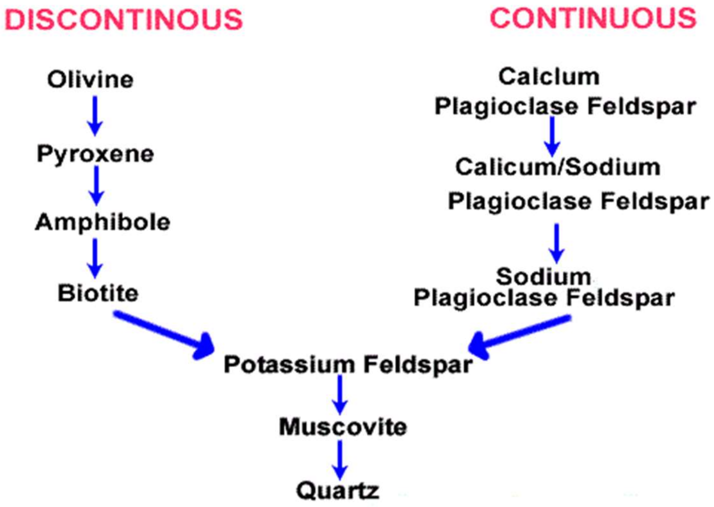
In conclusion, during chemical weathering igneous and metamorphic rocks can be degarded as involving destruction of primary minerals and the production of secondary minerals.

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.

Weathering Sequence of Minerals

Climatic and biotic conditions determine the relative stability of the various soil forming minerals. Under all climatic conditions a definite listing of minerals in relation to their resistance to weathering cannot be made. So considering the various environmental conditions and diversification the following order of weathering resistance of the sand and silt-size minerals are:

Quartz (most resistant) > Muscovite, K-feldspars > Na and Ca-feldspars >Biotite, Hornblende and Augite>Olivine > Dolomite and Calcite > Gypsum and it is subjected to change according to environmental conditions. Goldich (1938) described the weathering sequence.



Products of Weathering Minerals

The weathering products of most common rock-forming silicate minerals (Ferromagnesian and Non-ferromagnesian) are presented below:

Group	Mineral	Composition	Weathering Products	
			Minerals	**Others**
Ferromagnesian	Olivine	$(Fe,Mg)_2 SiO_4$	Serpentine, limonite, hematite, quartz	Some silica in solution. Carbonates of Fe and Mg (soluble)
	Pyroxenes, amphiboles and biotite	Fe. Mg, Ca-silicates of Al	Clay, calcite, limonite, hematie, quartz	As above
Non-ferromagnesian	Plagioclase (feldspar)	Calcic Ca ($Al_2Si_2O_3$) sodic $NaAlSi_3O_8$	Clay, quartz, calcite	Some silica in solution, carbonates of Na and Ca (soluble)
	Orthoclase	$KAlSi_3O_8$	Clay quartz	Some silica in solution and carbonate of K (soluble)
	Quartz	SiO_2	Quartz grains	Some silica in solution

Biological Weathering

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

- 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.
- A large number of animals, birds, insects and worms, by their activities they make holes in them and thus aid for weathering.
- 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.

- Rabbits, by burrowing in to the ground, destroy soft rocks. Moles, ants and bodies of the dead animals, provide substances which react with minerals and aid in decaying process.
- The earthworms pass the soil through the alimentary canal and thus brings about physical and chemical changes in soil material.

2. Higher Plants and Roots

The roots of trees and other plants penetrate into the joints and crevices of the rocks. As they grew, they exert a great disruptive force and the hard rock may broken apart. (e.g) pipal tree growing on walls/ rocks. The grass roots form a sponge like mass, prevent 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. These organisms closely associated with the decay of plant and animal remains and thus liberate nutrients for the use of next generation plants and also produce CO_2 and organic compounds which aid in mineral decomposition.

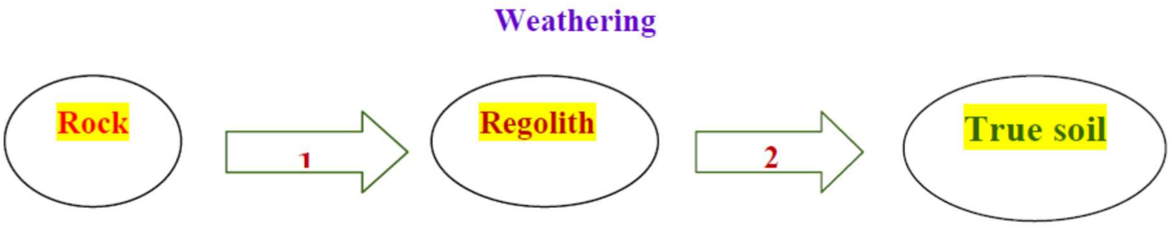
Lecture 8. Factors of soil formation- Passive and active soil forming factors soil forming factors

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.

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



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

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

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

- **Sedentary**

Formed in original place. It is the residual parent material. The parent material differ as widely as the rocks.

- **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 (river), Marine (sea / ocean), Locustrine (lake)
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.

Lacustrine

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 - loss; 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.

- Acid igneous rocks (like granite, rhyolite) produce light-textured soils (Alfisols).
- Basic igneous rocks (basalt), alluvium or colluvium derived from limestone or basalt, produce fine-textured cracking-clay soils (Vertisols).
- Basic alluvium or aeolian materials produce fine to coarse-textured soils (Entisols or Inceptisols).
- The nature of the elements released during the decaying of rocks has a specific role in soil formation. (e.g.) Si and Al forms the skeleton for the production of secondary clay minerals.
- Iron and manganese are important for imparting red colour to soils and for oxidation and reduction phenomena.
- Sodium and potassium are important dispersing agents for day and humus colloids.
- 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:

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 –3 0 %
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).

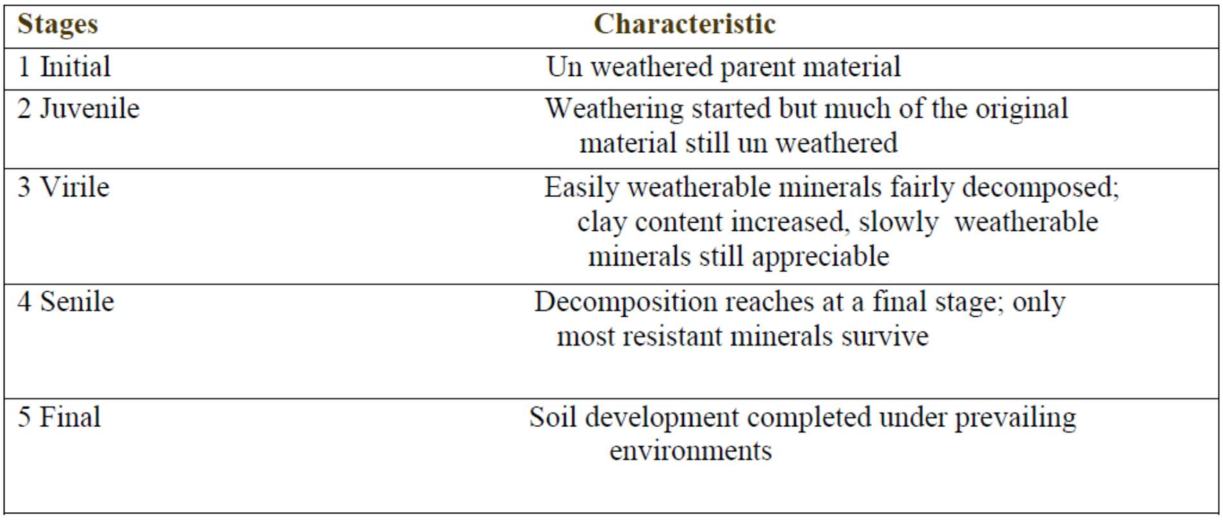
Soil formation and Exposure/ Aspect

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 Pedologic 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



Stages	Characteristic
1 Initial	Un weathered parent material
2 Juvenile	Weathering started but much of the original material still un weathered
3 Virile	Easily weatherable minerals fairly decomposed; clay content increased, slowly weatherable minerals still appreciable
4 Senile	Decomposition reaches at a final stage; only most resistant minerals survive
5 Final	Soil development completed under prevailing environments

- The soil properties also change with time, for instance nitrogen and organic matter contents increase with time provided the soil temperature is not high.
- $CaCO_3$ content may decrease or even lost with time provided the climatic conditions are not arid
- In humid regions, the H^+ concentration increases with time because of chemical weathering.

B. 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:

- **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.
- **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.
- **Oceanic climate:** Moderate seasonal variation of rainfall and temperature.
- **Mediterranean climate:** The moderate precipitation. Winters and summers are dry and hot.
- **Continental climate:** Warm summers and extremely cool or cold winters.

- **Temperate climate:** Cold humid conditions with warm summers.
- **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

- Temperature is another climatic agent influencing the process of soil formation.
- High temperature hinders the process of leaching and causes an upward movement of soluble salts.
- 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.

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

- The active components of soil ecosystem are plants, animals, microorganisms and man.

- The role of microorganisms in soil formation is related to the humification and mineralization of vegetation.
- The action of animals especially burrowing animals to dig and mix-up the soil mass and thus disturb the parent material.
- Man influences the soil formation through his manipulation of natural vegetation, agricultural practices etc.
- 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

- The roots of the plants penetrate into the parent material and act both mechanically and chemically.
- They facilitate percolation and drainage and bring about greater dissolution of minerals through the action of CO_2 and acidic substances secreted by them.
- The decomposition and humification of the materials further adds to the solubilization of minerals
- Forests – reduces temperature, increases humidity, reduce evaporation and increases precipitation.
- Grasses reduce runoff and result greater penetration of water in to the parent material.

Lecture 9: Soil forming process- Fundamental - Simenson's four fold soil forming process -eluviation, illuviation, translocation and humification

The pedogenic processes, although slow in terms of human life, yet work faster than the geological processes in changing lifeless parent material into true soil full of life. The pedogenic processes are extremely complex and dynamic involving many chemical and biological reactions, and usually operate simultaneously in a given area. One process may counteract another, or two different processes may work simultaneously to achieve the same result. Different processes or combination of processes operate under varying natural environment.

The collective interaction of various soil forming factors under different environmental conditions set a course to certain recognized soil forming processes.

The basic process involved in soil formation (Simonson, 1959) includes the following.

- Gains or Additions of water, mostly as rainfall, organic and mineral matter to the soil.
- Losses of the above materials from the soil.
- Transformation of mineral and organic substances within the soil.
- Translocation or the movement of soil materials from one point to another within the soil. It is usually divided into
- Movement of solution (leaching) and
- Movement in suspension (eluviation) of clay, organic matter and hydrous oxides.

A. 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 coloured substance, known as humus, is formed.

Eluviation

It is the mobilization and translocation of certain constituent's viz. Clay, Fe_2O_3, Al_2O_3,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 (A2or 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 Bt) The process leads to textural contrast between E and Bt horizons, and higher fine: total clay ratio in the Bt 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.

Lecture 10: Specific Soil forming processes - podzolization, laterization, salinization, alkalization, calcification, decalcification and pedoturbation

Specific Soil Forming Processes

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

1. Calcification

It is the process of precipitation and accumulation of calcium carbonate ($CaCO_3$) in some part of the profile. The accumulation of $CaCO_3$ may result in the development of a calcic horizon. Calcium is readily soluble in acid soil water and/or when CO_2 concentration is high in root zone as:

$$CO_2 + H_2O \longrightarrow H_2CO_3$$

$$H_2CO_3 + Ca \longrightarrow Ca(HCO_3)_2 \text{ (soluble)}$$

Temp ↑

$$Ca(HCO_3)_2 \longrightarrow CaCO_3 + H_2O + CO_2$$

$CO_2 \downarrow$ (precipitates)

The process of precipitation after mobilization under these conditions is called calcification and the resulting illuviated horizon of carbonates is designated as Bk horizon (Bca).

2. Decalcification

It is the reverse of calcification that is the process of removal of $CaCO_3$ or calcium ions from the soil by leaching.

$$\underset{\text{(Insoluble)}}{CaCO_3 + CO_2 + H_2O} \xrightarrow[CO_2 \uparrow]{\text{Temp}} \underset{\text{(soluble)}}{Ca(HCO_3)_2}$$

3. 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 weatherable 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.

$$2Al + 6H_2O \longrightarrow 2Al\,(OH)_3 + 6H^+$$

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 horizon s 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 sesquioxides (iron and aluminium oxides) to extent of 70 to 80 per cent of the total mass, are called laterites 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 orphic 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:

- Lower topographic position, such as depression land, where water stands continuously at or close to the surface.
- Impervious soil parent material, and.
- 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

$$Ca.Mg.2NaX \longrightarrow Ca^{++} + Mg^{++} + 2Na^{+} + X^{-6} + 3CO_3^{2-} \longrightarrow Na_2CO_3 + MgCO_3 + CaCO_3$$

(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 an be eliminated if $Ca^{++}$and or Mg^{++} ions are concentrated in the water, which is used to leach the soonest. These Ca and Mg ion can replace the Na on exchange complex, and the salts of sodium are leached out as:

$$2NaX + CaSO_4 \longrightarrow \underset{\text{(Leachable)}}{Na_2\,SO_4} + CaX$$

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:

- **Faunal pedoturbation:** It is the mixing of soil by animals such as ants, earthworms, moles, rodents, and man himself.
- **Floral pedoturbation :** It is the mixing of soil by plants as in tree tipping that forms pits and mounds.
- **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.

Lecture 11: Soil Profile – Horizons, Master horizons and subordinate horizons, subdivisions, Lithological discontinuity.

Definition of soil profile

The vertical section of the soil showing the various layers from the surface to the unaffected parent material is known as a soil profile. The various layers are known as horizons. A soil profile contains three main horizons.

They are named as horizon A, horizon B and horizon C.

- The surface soil or that layer of soil at the top which is liable to leaching and from which some soil constituents have been removed is known as horizon A or the horizon of eluviation.
- The intermediate layer in which the materials leached from horizon A have been redeposited is known as horizon B or the horizon of illuviation.
- The parent material from which the soil is formed is known as horizon C.

A Study of soil profile is important as it is historic record of all the soil forming processes and it forms the basis for the study in pedagogical investigations. Soil profile is the key for the soil classification and also forms the basis for the practical utility of soils.

A hypothetical mineral soil profile will include O, A, B, C and R master horizons and all the possible sub-horizons.

Master horizons and sub horizons

O horizon - It is called as organic horizon. It is formed in the upper part of the mineral soil, dominated by fresh or partly decomposed organic materials.

This horizon contains more than 30% organic matter if mineral fraction has more than 50 % clay (or) more than 20 % organic matter if mineral fraction has less clay.

The organic horizons are commonly seen in forest areas and generally absent in grassland, cultivated soils.

O_1 - Organic horizon in which the original forms of the plant and animal residues can be recognized through naked eye.

O_2 - Organic horizon in which the original plant or animal matter can not be recognized through naked eye.

A horizon - Horizon of organic matter accumulation adjacent to surface and that has lost clay, iron and aluminium.

A1 - Top most mineral horizon formed adjacent to the surface. There will be accumulation of humified organic matter associated with mineral fraction and darker in Colour than that of lower horizons due to organic matter.

A2 - Horizon of maximum eluviation of clay, iron and aluminium oxides and organic matter. Loss of these constituents generally results in accumulation of quartz and other sand and silt size resistant minerals. Generally lighter in Colour than horizons above and below.

A3 - A transitional layer between A and B horizons with more dominated properties of A1 or A2 above than the underlying B horizon. This horizon is sometimes absent in solum.

B horizon - Horizon in which the dominant features are accumulation of clay, iron, aluminium or humus alone or in combination. Coating of sesquioxides will impart darker, stronger of red Colour than overlying or underlying horizons.

B1 - A transitional layer between A and B. More like A than B.

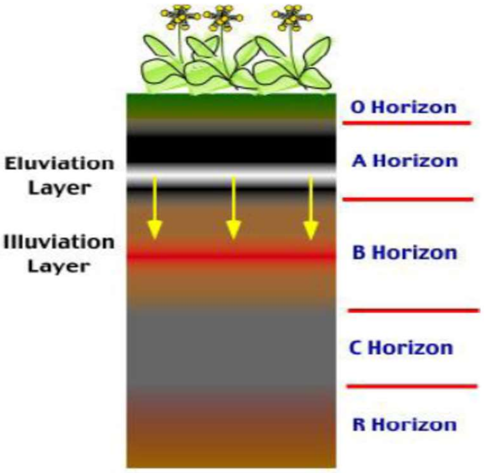
B2 - Zone of maximum accumulation of clay, iron and aluminium oxide that may have moved down from upper horizons or may have formed in situ. The organic matter content is generally higher and Colour darker than that of A2 horizon above.

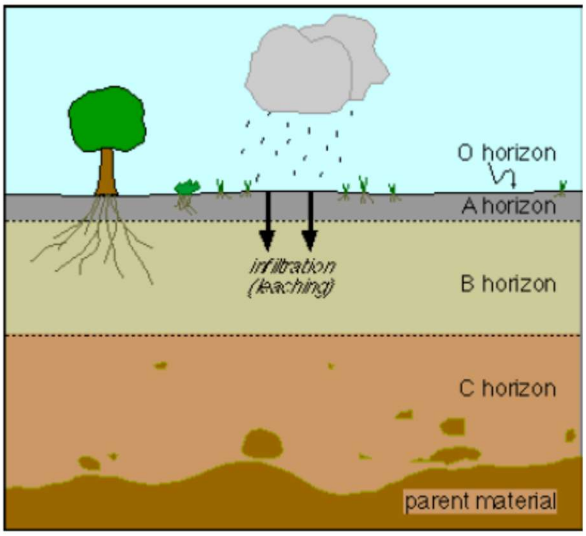
B3 - Transitional horizon between B and C and with properties more similar to that of overlying B2 than underlying C.

C horizon - It is the horizon below the **solum** (A + B), relatively less affected by soil forming processes. It is outside the zone of major biological activity. It may contain accumulation of carbonates or sulphates, calcium and magnesium.

R - Underlying consolidated bed rock and it may or may not be like the parent rock from which the solum is formed.

Besides, lower case letters are used to indicate the special features of master horizons. This case letters follow the subdivisions of master horizons. eg. Ap - ploughed layer eg. B2t - illuvial clay When two or more genetically unrelated (contrasting) materials are present in a profile as in the case of alluvial or colluvial soils then the phenomenon is known as lithological discontinuity. This is indicated by the use of Roman letters as prefixes to the master horizons. eg. Ap, B2, II B22, IIIC.





Special Features

Soil Individual or Polypedon: The Soil Survey Staff (1960) defined the soil individual or polypedon (Pedon, Ground) as a natural unit of soil that differs from its adjoining unit on the landscape in one or more properties. The term pedon has been proposed for small basic soil entities that are part of the continuum mantling the land.

Lithological Discontinuity

When two or more genetically unrelated (contrasting) materials are present in a profile as in the case of alluvial or colluvial soils, then the phenomenon is known as lithological discontinuity. This is indicated by the use of Roman letters as prefixes to the master horizons.

Soil monolith - A vertical section of a soil profile removed from the field and mounted for display or study.

Transitional Horizons

Two kinds of transitional horizons are recognized. In one, the horizon is dominated by properties of one master horizon but has subordinate properties of another. Two capital latter symbols are used, such as AB, EB, BE, or BC. The master horizon symbol that is given first designates the kind of master horizon whose properties dominate the transitional horizon.

In the other, distinct parts of the horizon have recognizable properties of the two kinds of master horizons indicated by the capital letters. The two capital letters are separated by a virgule (/), as E/B, B/E, or B/C. The first symbol is that of the horizon that makes up the greater volume.

E.g**. AB** - A horizon with characteristics of both an overlying A horizon and an underlying B horizon, but which is more like the A than the B.

A **pedon** is the smallest volume that can be called "a soil". The set of pedons must fit within the

range of one series and occur in a contiguous group to form a polypedon.

A **polypedon** is therefore, defined as a contiguous similar pedons bounded on all sides by "notsoil or by pedons of unlike characters. It is a real physical soils body which has a minimum area of more than 1 sq. km and an unspecified maximum area.

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Lecture 12: Phases of soils – solid, liquid and gaseous phases – properties of soil – defining the physical, chemical and biological properties,

Soil is a three- phase, porous media, composed of solids, liquids and voids (empty spaces between the solids). The typical amount of total porosity (ratio of void volume to total volume) in a mineral soil ranges from about 40% to 60%. This means approximately 40 to 60% of the volume of a mineral soil is actually empty space between the solid particles (voids).

Mineral matter in soils (Solid phase)

The size and composition of mineral in soils are variable due to nature of parent rock from which it has been derived. In general, the primary viz., Quartz, biotite, muscovite etc., dominate the coarser fractions of soil, on the other hand, the secondary minerals viz., silicate clays and hydrous oxide clays of iron and aluminium etc., are present as very finer fractions, clay in the soils.

Organic matter in soils

Soil organic matter exists as partly decayed and partially synthesized plant and animal residues. Such organic residues are continually being broken down as a result of microbial activity in soil. Organic matter is a store house of nutrients soil. Besides these, organic matter is responsible for most desirable soil physical properties.

Soil water (Liquid phase)

Soil water plays a very significant role in soil plant growth relationships. Water is held within the soil pores with varying degrees of forces depending on the amount of water present. Soil water is also present along with dissolved salts and makes up the soil solution. This soil solution (water plus dissolved salts) acts as an important medium for supplying different essential nutrients for growing plants.

Soil air (Gaseous phase)

Air spaces or pore spaces (voids) in a soil consists of that portion of the soil volume not occupied by soil solids, either mineral or organic. During rainy season, water replaces air from the soil pore spaces, but as soon as water disappeared by downward movement, surface evaporation and transpiration etc., air gradually the water as it is lost from the pore spaces. Soil air influence the growth of the plants as well as the activity of different beneficial micro organisms present in soil.

Properties of soils:

A soil is a product of nature and is derived from a set of combination of weathering agents and varying intensities. Accordingly, a soil may possess a well defined set of properties which act upon their surrounding, be it the crop, microbes or for that matter the neighbouring soil particles. The various soil properties can be grouped into four categories. They are

1. Physical properties
2. Chemical properties
3. Physico – chemical properties
4. Biological properties

Soil physical properties

Physical properties (mechanical behaviour) of a soil greatly influence its use and behavior 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 8. Soil water 9. Soil air 10. Soil temperature

Physico-chemical properties

Certain soil properties are responsible both for physical processes and chemical reactions. They operate in close association so that the resultant change is noticed.

5. Soil pH
6. Electrical conductivity
7. Cation Exchange Capacity
8. Anion Exchange Capacity
9. Soil colloids

Chemical properties

The soil properties which result in chemical changes are called chemical properties

4. Soil fertility
5. Nutrient cycles

6. Nutrient contents and their transformations
7. Soil organic matter
8. Mineralization and immobilization

Soil biological properties

9. Changes that occur due to the living organisms
10. Soil biochemical changes
11. Soil biota
12. Soil fauna flora

Lecture 12.1: Soil physical properties: Soil texture - particle size distribution - textural classes - textural triangular diagram - significance of soil texture

Definition

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.

Particle size distribution

The proportion of each size group in a given soil (the texture) cannot 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 water holding capacity (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

A particle (sand, silt and clay) less than 2 mm is size is called fine earth, normally considered in chemical and mechanical analysis. The size limits of these fractions have been established by various organizations.

There are a number of systems of naming soil separates.

(i) The American system developed by USDA
(ii) The English system or British system (BSI)
(iii) The International system (ISSS)
(iv) European system

- **USDA system**

Soil separates	**Diameter (mm)**
Clay	<0.002 mm
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	0.50-1.00
Very Coarse Sand	1.00-2.00

- **BSI system**

Soil separates	**Diameter (mm)**
Clay	<0.002 mm
Fine Silt	0.002-0.01
Medium Silt	0.01-0.04
Coarse Sand	0.04-0.06
Fine Sand	0.06-0.20
Medium Sand	0.20-1.00
Coarse Sand	1.00-2.00

- **ISSS system**

Soil separates	**Diameter (mm)**
Clay	<0.002 mm
Silt	0.002-0.02
Fine Sand	0.02-0.2
Coarse Sand	0.2-2.0

- **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

Major characteristics of soil separates

Sand fraction

Particles diameter is in the range of 2 mm to 0.02 mm (ISSS) or 2 mm to 0.05 mm (USDA) are known as sand fraction. They are the largest soil separates. They are spherical or irregular in shape, non cohesive, non sticky and non plastic materials. They have no swelling and shrinkage property. They are estimated by sieves. Soils dominated by sand fraction have low water holding capacity and posses good drainage and aeration.

Sand particles have low specific surface area (0.0011 square meter per gram (m^2/gm) for coarse sand and 0.1 m^2/gm for fine sand), they have relatively very low chemical activity for which they are resistant to weathering. They have no cation exchange capacity. Thus the nutrient supplying ability of this fraction is very low.

Usually consists of quartz but may also contain fragments of feldspar, mica and occasionally heavy minerals *viz.,* zircon, tourmaline and hornblende.

Silt fraction

Particle diameter is in the range of 0.02 to 0.002 mm (ISSS) or 0.05 to 0.002 (USDA) are known as silt fraction. They are the intermediate soil separate between sand and clay fractios. They are irregular in shape. These particles have negligible plasticity, cohesion (stickiness), swelling, shrinkage and adsorptive capacity. They are determined by sedimentation method. Soils dominated by silt fraction are more compact and may form surface curst under the impact of rain resulting poor drainage condition and poor aeration status of soil.

Silt fraction have medium specific surface (about 1.1 m^2/gm), relatively low chemical activity, negligible cation exchange capacity and low nutrient supplying ability. They contain primary minerals like quartz (SiO_2), feldspars (alumina silicates) and micas (iron and aluminum silicates) which are resistant to weathering. Sand and Silt forms the SKELETON structure.

Clay fraction

Clay particles are the smallest soil separates (diameter is less than 0.002 mm). They are platy or spherical in shape, highly sticky and highly plastic materials. Some clays swell upon wetting and shrink upon drying. The swelling and shrinking properties are evidenced from the cracks found in dry soil after loss of rain or irrigation water. They are determined by sedimentation method. Soils having high amount of clay particles possess high retention of water and poor aeration. Unless these soils are properly handled or tilted at friable moisture content they become very hard and cloddy when dry.

The specific surface area of clay particle is very high (11 to 1100 m^2/gm). A grain of fine colloidal clay has about 10,000 times as much specific surface as the same weight of medium sized sand. They have relatively high chemical activity. Clay particles are composed of simple aluminosilicates with varying quantities of iron, calcium, magnesium, potassium, sodium, ammonium etc. in fixed or exchangeable form. They exchangeable form is related for use of plant. Coarse clay particles are composed or quartz and hydrous oxides of iron and aluminium.

Influence of soil separates on some properties and behavior of soils

S.No	Properties/Behaviour	Rating associated with soil separates		
		Sand	Silt	Clay
1	Specific surface	Low	Medium	High
2	Total porosity	Low	Medium	High
3	Macro pores	High	Medium	Low
4	Micro pores	Low	Medium	High
5	Internal drainage and drainage rate	High	Slow to medium	Very slow
6	Aeration status	Good	Medium	Poor
7	Plasticity	Non plastic	Slightly plastic	Very plastic
8	Stickiness	Non sticky	Slightly sticky	Very sticky
9	Swelling-shrinking potential	Nill	Very low	Low to very high depending on clay mineralogy
10	Water holding capacity	Low	Medium to high	High
11	Infiltration and permeability	High	Slow to medium	Very slow
12	Unsaturated hydraulic conductivity	Low	Medium	High
13	Buffering capacity	Low	Medium	High
14	Ability to store and supply plant	Poor	Medium	High
15	Ability to hold organic matter	Low	Medium to high	High to medium
16	Decomposition of added organic matter	Rapid	Moderate	Slow
17	Compatibility	Very low	Medium	High
18	Susceptibility to wind erosion	Moderate	High	Low
19	Susceptibility to water erosion	Low	High	Low if aggregated; high if not
20	Mineralogical characteristics			
	(a) Dominant minerals	Quartz	Quartz	Aluminosilicates
	(b) Other minerals	Feldspars & Mica (hematite and goetite as coating on sand grains)	Feldspars and mica	Quartz, gibbsite and goetite

Soil Textural Classes

The groups based on the relative proportions of the various soil separates are known as soil textural classes. According to the proportion of sand, silt and clay, a soil is given a name to indicate its textural composition. Soil textural classes may broadly be grouped into three, such as sandy soils, loamy soils and clayey soils.

On the basis of these, additional class names have been devised. There are two main systems of naming soils according to their texture. In U.S Department of Agricultural Classification System, textural classes are Sand, loamy sand, sandy loam, loam, silt loam, silt clay loam, sandy clay, silty clay and clay. The classification is from coarse to fine texture are given below

Common Name	Texture	Basic soil textural class names
Sandy soils	Coarse	Sandy
		Loamy sands
	Moderately coarse	Sandy load
		Fine sandy loam
Loam soils	Medium	Very fine sandy loam
		Loam
		Silt loam
		Silt
	Moderately fine	Clay loam
		Sandy Clay loam
		Silty Clay loam
Clayey soils	Fine	Sandy Clay
		Silty Clay
		Clay

Sandy soils

This group includes soils having at least 70% sand separates and 15% or less clay separates by weight. The basic soil textural class names of this group are sands and loamy sands which are coarse in texture. These soils have relatively low chemical activity for which they are resistance to weathering. As the soils of this group have very low content of clay separates, they have low specific surface area and they are non sticky and non plastic in nature.

The water holding capacity and the amount of plant available water of these soils are low but drainage condition and aeration status of them are high. They have low cation exchange capacity for which their nutrient supplying ability and nutrient retention power are very low. The productivity of this soil in normally low.

Loamy soils

An ideal loam may be defined as a mixture of sand, silt and clay separates in about equal proportions. This group has six subdivisions due to varying quantities of sand, silt and clay separates in the soil. A loam in which sand is dominant is named as silt loam. Loam and silt loam are medium in texture. In the same way a loam in which both sand clay are dominant is termed as sandy clay loam. Silt and clay are dominant is called as silty clay loam and clay is dominant is known as clay loam. Sandy clay loam, silty clay loam and clay loam are moderately fine in texture.

An ideal loam soil feels gritty and is cohesive. It has good drainage, medium water retention capacity, medium plant available water content, low plasticity, low stickiness, medium cation exchange capacity. The nutrient retention power and nutrient supplying ability of this soil is medium. This is an ideal soil for crop production.

Clayey soils

Clayey soils mist contain at least 35% of clay separate and in most cases not less than 40% of clay separate. These soils are fine in texture. The class names of this group are sandy clay, silty clay and clay. Sandy clay may contain more sand than clay. Similarly silt content of silty clay is usually more silt than clay content.

As these soils contain high amount of clay, their specific surface and cation exchange capacity are very high and because of these, their capacity for adsorption of water and nutrients is high. They are highly sticky and highly plastic. As they have high total porosity and most of these pores are micro pores, water retention capacity and plant available water content of these soils are very high but internal drainage condition or soil aeration is very poor. It is difficult to till these soils. If tillage operation is not done at proper soil water content, clods are formed during tillage operation. As the dry clods are very hard, it is difficult to break those clods by laddering.

Silty soils

This group includes soils which have at least 80% silt and 12% or less clay. The only textural class name of this group is silt which is medium in texture. This soil feels silky and buttery and possesses some plasticity, cohesion (stickiness) and adsorptive capacity. The surface of this soil is normally compact and crusty. This soil has medium plant available water fair internal drainage condition or soil aeration.

Heavy soils and light soils

Sometimes fined textured soils (E.g. **Clayey soils**) are termed as heavy soils and coarse textured soils (E.g. **Sandy soils**) are termed as light soils. Heavy and light soils do not reflect the weight of the soils. As relatively more power is needed to plough fine textured soil. E.g. **Clayey soils** because they become plastic and sticky when wet, hard and coherent when dry and difficult to cultivate, they are called heavy soils.

As relatively less power is needed to plough coarse textured soils i.e., sandy soils because they are loose and easy to cultivate, they are called light soils. Actually, clayey soils have more total pore space and less bulk density than those of sandy soils. So clayey soils are actually lighter in weight than sandy soils i.e. the weight of clayey soils is less than that of equal volume of sandy soils.

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 i.e. Complete dispersion into ultimate particles

ii) Measuring the amount of each group.

S. No	Aggregating agents	Dispersion method
1	Lime and Oxides of Fe & Al	Dissolving in HCl
2	Organic matter	Oxidises with H_2O_2
3	High concentration .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 particles. 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:

$$V = \frac{2gr^2 (ds-dw)}{9n}$$

Where ,

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

g - acceleration due to gravity cm/ sec^2 (981)

ds - density of soil particle (2.65)

dw - density of water (1)

η - coefficient of viscosity of water (0.0015 at 4°C)

r - radius of spherical particles (cm).

Assumptions and Limitations of Stokes' Law

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. 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. Ie the particles must be big enough to avoid Brownian movement.

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. Bouyoucos hydrometer method (Bouyoucos 1927)

vi. Feel method - Applicable to the field, it is a quick method and by feeling the soil between thumb and fingers.

i. Elutriation method - Water & Air

Elutriation is a process for separating particles based on their size, shape and density, using a stream of gas or liquid flowing in a direction usually opposite to the direction of sedimentation. This method is mainly used for particles smaller than 1 µm. The smaller or lighter particles rise to the top (overflow) because their terminal sedimentation velocities are lower than the velocity of the rising fluid. The terminal velocity of any particle in any medium can be calculated using Stokes' law if the particle's Reynolds number is below 0.2. Counter flow centrifugation elutriation is a related technique to separate cells.

ii. Pipette method (Robinsion's pipette method/Mechanical analysis)

Mechanical analysis is the procedure for determining the particle size distribution of a soil. Particles can be separated into size groups by sieving through graded sieves up to a particle of approximately 0.05 mm.

To separate finer particles, the method of sedimentation is used. The binding or cementing agents which binds the soil particles together namely organic matter and carbonates are removed first. The soil is dispersed as a colloid and the size separation is done following the Stoke's law.

iii. Decantation/ beaker method

This method for beaker decantation is a technique used to separate a mineral sample into two size fractions according to the differences in settling velocities of the particles. It is a technique which can be used to accurately split a sample at a pre-determined cut size but has some disadvantages in that is a time consuming method especially when splitting at fine sizes.

Also a dilute solids content is required to stop natural coagulation from occurring. A spherical shape is assumed for the calculations.

iv. Test tube shaking method

A simple texture analysis can be conducted by collecting a soil sample in a glass jar or test tube and mixing it with an equal volume of water. The dry sample should be measured (linear from the bottom of the jar to the top of the sample or volumetrically) before water is added. After adding water, shake the container vigorously and then place in a location that will allow it to sit undisturbed for twenty-four hours.

The sand particles will settle to the bottom and the silt above that. Clay will settle on top but can take up to two weeks to precipitate out of suspension, which is why it is important to measure the dry sample before adding the water. If the volume of the original sample is known, it is easy to determine the percentage of sand and silt by measuring each layer and calculating what portion it is of the whole sample. Once the percent sand and silt is determined, use the texture analysis chart to classify the sample.

v. Bouyoucos hydrometer method (Bouyoucos 1927)

Bouyoucus (1927) introduced this method for determining the size distribution of particles having diameters less than 0.074 mm. Hence silt and clay separates can be determined by this method. This method is very popular for rapid determination of soil texture especially in soil taxonomic studies.

The main principle of hydrometer method is based on continuous reduction of density of the soil suspension with time at the rate the particles drops below the level of hydrometer. At any time density of the suspension is lowest near the top and increases towards the bottom. Bouyoucos hydrometer calibrated at 20°C and may be used over the range 15°C with suitable corrections, measures the density of the suspension at a given depth and time. A hydrometer is placed in a well dispersed soil suspension at desired time interval and the readings are taken for determining silt plus clay fraction and clay fraction. Particles bigger than silt separates are determined by direct sieving or by subtraction.

vi. Feel Method

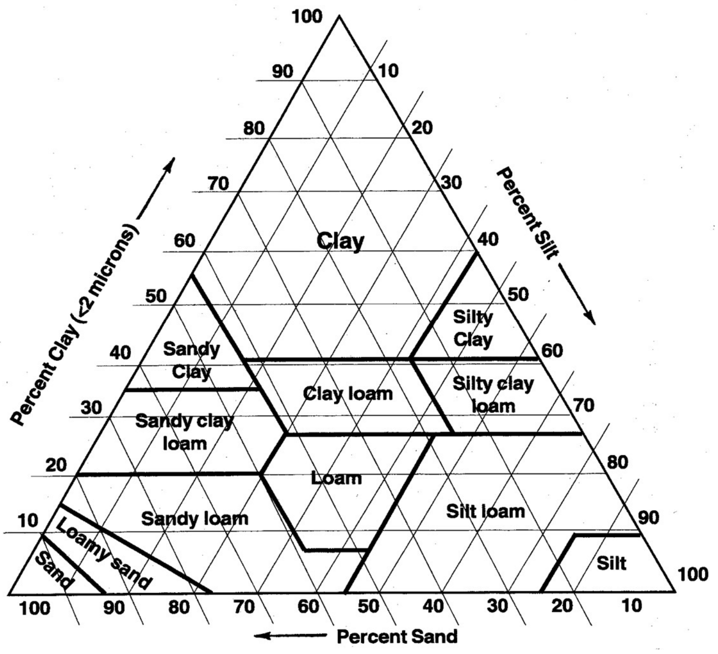
Evaluated by attempting to squeeze the moistened soil into a thin ribbon as it is pressed with rolling motion between thumb and pre finger or alternately to roll the soil into a thin wire. Four aspects to be seen they are i. Feel by fingers, ii. Ball formation , iii. Stickiness and iv. Ribbon formation.

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. Sometimes, the intersecting point exactly fall 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**.



Percentage of sand, silt and clay in the principal textural classes based upon USDA and ISSS fractions system.

Textural name	Based upon USDA fraction system			Based upon ISSS fraction system		
	Sand (%)	**Silt (%)**	**Clay (%)**	**Sand (%)**	**Silt (%)**	**Clay (%)**
Sand*	85-100	0-15	0-10	88-100	0-8	0-8
Loamy sand*	70- 90	0-30	0-15	65-92	4-24	0-12
Sandy loam*	43-80	0-50	0-20	70-90	0-12	8-20
Loam	23-52	28-50	7-27	50-76	10-25	11-26
Silty loam	0-50	50-88	0-27	25-75	25-50	0-26
Silt	0-20	88-100	0-12	0-50	50-100	0-26
Sandy clay loam	45-80	0-28	20-35	64-82	0-10	18-31
Clay loam	20-45	15-33	27-40	0-48	25-73	26-40
Silty clay loam	0-20	40-73	27-40	0-48	25-73	26-40
Sandy clay	45-65	0-20	35-45	50-73	0-6	27-50
Silty clay	0-20	40-60	40-60	0-34	25-60	40-74
Clay	0-45	0-40	40-100	0-62	0-25	32-100

* When soil contain 50 percent or more fine sand or less than 25 percent course sand the word fine is used before textural name. When soil contains 50 percent or more of very fine sand the word very fine is used before textural class names.

Importance of Soil Texture

- Knowledge of soil texture is extremely important in determining the suitability of soils for the production of different crop plants, for foundation of buildings and for construction of roads.

- 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’t stand drought and unsuitable for dry farming. Sandy soils are poor store house of plant nutrients. It 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 and 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 and 30% silt called as clay rather than clay loam.

Lecture 13: Soil structure - classification - genesis - factors influencing structural stability - significance of soil structure

Definition

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.

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.

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 structureless, 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 cannot form aggregates as they do not possess the power of adhesion and cohesion. These particles usually carry a coating of 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 something 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. 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.

$$\text{Clay}^- - {}^+\text{Water}^- - {}^+\text{Cation}^+ - {}^-\text{Clay}^- - {}^+\text{Water}^- - {}^+\text{Cation}^+ - {}^-\text{Clay}^-$$

Anions

Among the anions phosphate plays a significant role in aggregation of soil. The polyvalent phosphate ion is bonded with colloidal soil particles through iron and aluminium possibly as

$$\text{Clay} - \text{Al (OH)}.\ \text{HPO}_4.\ \text{(OH) Al} - \text{Clay}$$

The cementation effect of precipitated phosphates of iron and aluminium may form aggregates after their sorption on colloidal soil particles. Added phosphate sometimes react with active fraction of soil through hydroxyl group and binds the phosphates with colloidal soil particles.

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. During decomposition, cellulosic substances produce a sticky material very much resembling mucus or mucilage.

The sticky properly may be due to the presence of humic or humic acid or related compounds produced.

Certain polysaccharides formed during decomposition. 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

The primary particles of sand, silt and clay - usually occur grouped together in the form of aggregates. Natural aggregates are called peds whereas 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 Structure

There are four principal forms of soil structure. 1. Platy, 2. Prism, 3. Block, 4. Spheroidal

1. 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.

2. 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.** The tops are flat / plane, level and clear cut structure is called **prismatic**.

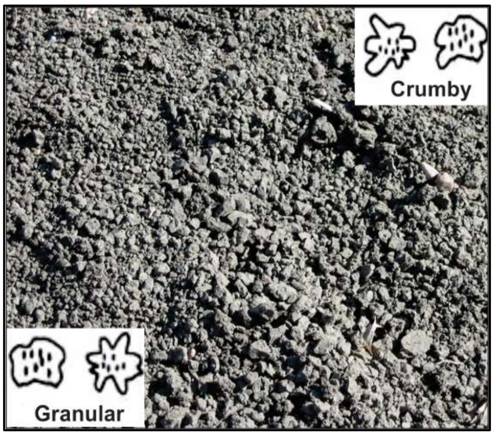
3. Block like

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.

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

S.No	Class	Size of peds (mm)				
		Platy	Prismatic or Columnar	Blocky or Sub angular blocky	Granular	Crumb
1	Very fine or very thin	<1	<10	<5	<1	<1
2	Fine or thin	1-2	10-20	5-10	1-2	1-2
3	Medium	2-5	20-50	10-20	2-5	2-5
4	Coarse or thick	5-10	50-100	20-50	5-10	-
5	Very coarse or very thick	>10	>100	>50	>10	-

Grades of 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.

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 un aggregated 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.

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 crumby.

3. Tillage

Cultivation implements break down the large clods into smaller fragments and aggregates. For obtaining good granular and crumby 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 bind the soil particles together. Root hairs make soil particles to cling together. Eg. Grass and cereal roots. 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, Calcium Ammonium Nitrate (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

The Ca and Mg cations are good aggregating agents and produce good structure. The H and Na are deflocculating agents produce poor structure.

10. Inorganic cements

$CaCO_3$ and Sesquioxides are good cementing agents and produce good structure.

11. Clay

Both structural form and stability are influenced by the mineralogical composition of clay because some expanding type clays such as smectities swell when wet and shrink when dry; but some non expanding type clays such as kaolinite, halloysite etc. do not show the property of swelling and shrinking during wetting and drying. Vermiculite clay is intermediate in its swelling and shrinking characteristics. Soils dominated by montmorillonite clays are good swelling and shrinking capacity and produce good structure

12. Water

Water is important factor for soil structure. Alternate wetting and drying causes disruption of soil mass into finer fragment and develop cracks or surface of weakness by causing strains and pressure in the soil body due to shrinking to swelling.

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.

Evaluation of soil structure

The methods of evaluation of soil structure may be grouped into two, such as **direct method** and **indirect method**. The direct methods are based on the measurement of shape, size and arrangement of soil aggregates while the indirect methods are based on the determination of extent and stability of aggregates and properties related to soil structure.

A. Direct methods

Direct methods of evaluation of soil structure are 1. Microscopic method and 2. Macroscopic method or field method.

1. Microscopic method

In this method thin sections of soils are examined in the laboratory in a petrographic microscope under polarized light which clearly show the nature, shape and size of secondary particles and voids. Macroscopic photograph may be taken for permanent record.

2. Macroscopic method (or field method)

A chunk of dry soil is excavated from the field and is allowed to fall gently on a wooden plank or smooth pucca floor from about 3 feet height. The chunk breaks into small pieces of peds of different sizes and shapes form the cleavages. The geometric shapes of the aggregates determine the type of soil structure. The size of the aggregates determine the class of soil structure. The distinctness with which peds are observed such as whether the broken pieces of chunk contain completely distinct peds or mixture of peds and unaggregated particles or entirely unaggregated materials.

B. Indirect method

In these methods soil structure is evaluated by determining i) Size distribution and stability of aggregates commonly known as **aggregate analysis** and ii) Properties related to soil structure.

i) Aggregate analysis

An aggregate analysis aims to measure the percentage of water stable secondary particles in the soil and the extent to which finer mechanical separates are aggregated into coarser fraction. The techniques employed for aggregate analysis are wet sieving, dry sieving, elutration and sedimentation.

a) Wet sieving

In this method the soil is slowly wetted by capillarity for 30 minutes. The wetted soil samples are then transferred into the sieve having largest opening in the nest and nest of sieves are immersed in water. The sieves are slowly raised and lowered vertically in the water by mechanical method through a distance of 1.25 inch at a rate of 30 oscillations per minutes for 30 minutes. The sieves in the nest have an opening from 8.0 mm to 0.1 mm (sizes 8.0, 5.0, 2.0, 1.0, 0.5, 0.25 and 0.1mm). Oven dry weight of the fractions of the aggregates retained an each sieve is determined for knowing the proportionate distribution of the aggregates. The fraction retained on each sieve represents the size range *viz.,* the size of the opening of the sieve on which the fraction of soil is retained and the size of the sieve just above through which it has passed. A wet sieving apparatus (yoder type).

b) Dry sieving

Dry sieving measures the size and stability of dry aggregates or clods which are weakly held together as in arid soils. The different size fractions of the soil passed through a serves of sieves gives the aggregate size distribution. The principle of this method lies in determining the proportion of aggregates which are stable against vibrating action stimulating the scouring effect of wind. Problem related to clogging of sieves and the mechanical action required in the sieve operation for breaking up the week aggregates may be overcome by use of improved rotary nest of sieves as described by **Chepel (1962)**. Dry sieving of aggregates gives important index for determining the susceptibility of soils to wind erosion. It is also used to evaluate tillage practices for seed bed preparation.

c) Elutration Method

Elutration method separates different soil fractions (different fraction of dispersed soil particles during mechanical analysis and different fraction of soil aggregates during aggregate analysis) by ‘washing out’ soil particles or soil aggregates in a rising current of water. So, the principle of elutration method is based upon the separation of soil particles or soil aggregates by washing out them in a raising stream of water. The apparatus is such that the velocity of flow can be controlled and separation is accomplished by treating the velocity that balances the rate of fall of particles as calculated from the Stockes law. Water is allowed to flow through vertical vessels with a given velocity in each successive vessel. The particles having settling velocity less than the upward velocity of water in a particular vessel or tube connecting the vessels will be carried out of the tube.

The method may be used for separating aggregates having diameters between 1 and 0.02 mm. This is mainly used for separation of aggregates where wet sieving cannot be adopted.

d) Sedimentation method

This method is based upon the settling velocity of a particle suspended in water. This method is used for finer sizes of aggregates which cannot be separated by sieving. The methods adopted in sedimentation method are pipette method and the hydrometer method. Here soil suspension is made in distilled water without dispersion. Primary particle size determination of the same soil is also done to know the actual weight of aggregated fractions.

Importance of Structure

- Soil structure influences the amount and nature of porosity.
- **It is the plumbing system for the soil which controls** water flow and air flow.
- 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.
- **It provides space, and a protected home, for**roots, germinating seeds and soil fauna (macro and micro).
- It affects the impact of land use on the soil factors like the amount of run-off and erosion, amount of nutrients lost in drainage and run-off and/or erosion from the soil and amount of pollutants lost from farms in erosion, run-off and/or drainage.
- It affects tillage practices.
- Platy structure normally hinders free drainage whereas sphere like structure (granular and crumby) helps in drainage.
- Crumby and granular structure provides optimum infiltration, water holding capacity, aeration and drainage.
- It also provides good habitat for microorganisms and supply of nutrients

Lecture 14: Soil bulk density, particle density and porosity - factors influencing – significance

Density is the weight pen unit volume of a substance. It is expressed as gram per cubic centimeter or pound per cubic feet or mega gram per cubic meter (Mg m^{-3}).

Two density measurements like particle density and bulk density are common for soils.

$$\text{Density (D)} = \frac{\text{Mass (M)}}{\text{Volume (V)}} \quad \text{gm / cc}$$

Soil density is expressed in two well accepted concepts

1. Bulk density or Apparent specific gravity
2. Particle density or Absolute specific gravity

Bulk density(or) Apparent specific gravity

The mass (weight) per unit volume of a dry soil (volume of solid and pore spaces). It is expressed in gm/cc.

$$\text{Bulk density} = \frac{\text{Weight of soil}}{\text{Volume of soil}} \text{ gm / cc}$$

Soil bulk density, like all density measurement, is an expression of the mass to volume relationship for a given material. Soil bulk density measures total soil volume. Thus, bulk density takes into account solid space as well as pore space.

The bulk density of a soil is always smaller than its particle density. The bulk density of normal soils range between 1.00 to 1.6 g cc^{-1} with an average of 1.35 g cc^{-1}. Generally in normal soils bulk density ranges from 1 to 1.60 g cc^{-}1. For good plant growth, bulk densities should be below 1.4 g $cc^{-1}$for clay, 1.6 g $cc^{-1}$for sands and 0.5 g $cc^{-1}$for pure organic matter.

The density of water is 1.0 Mg m^{-3} and mineral soils are usually heavier than water. However, organic soils generally have a bulk density less than water. Soils that are loose, porous or well-aggregated will have lower bulk densities than soils that are compacted or non aggregated.

This is because pore space weighs less than solid space (soil particles). Sandy soils have less total pore space than clayey soils. So generally they have higher bulk densities.

Bulk density normally decreases, as mineral soils become finer in texture. The bulk density decreases with increase in 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^{-1})**	**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. Bulk density in an indirect measure of spacae and is affected primarily by soil texture and structure.
2. In sandy soils, organic contents are low and the bulk densities are commonly higher than in the finer textured soils. As aggregation and clay content increases, bulk density decreases.

Clayey	**Loamy**	**Sandy**
Well aggregated	Moderately aggregated	No aggregatged
High organic matter content	Moderate organic matter content	Low organic matter content

3. Tillage operations do not affect texture, but they alter structure (soil particle aggregation). Primary tillage operation, such as ploughing, generally decreases bulk density and increases pore space, which is beneficial. Secondary tillage (cultivation) generally increases bulk density and decreases pore space.
4. Cropped soils generally have higher bulk density than uncropped soils.
5. The movement of machinery over the field forces solid particles into spaces once occupied by water or air, resulting in less pore space and increased bulk density.
6. Organic matter and decomposed products bind the primary soil particles to form the soil aggregates resulting an increase in total pore space and decrease in the mas of unit volume of soil as because here pore spaces occur within and between the soil aggregates. Manure additions in large amounts tend to lower the surface bulk density of mineral soils because of the addition of low bulk density material promotes soil aggregation.
7. Soil bulk density increases with soil depth primarily because of less organic matter and decreased aggregation.
8. Compactness of the soil resulting decrease in total space and increase in bulk density.

Methods of estimation of bulk density

a) Measuring cylinder method

A known weight of soil (w) is taken in a graduated measuring cylinder and the volume is noted (V_1). A known volume of water (V_2) is added and the total volume of soil with water is noted (V_3). From the data, the following parameters are calculated.

$$\text{Bulk density} = \frac{W}{V_1} \text{ g cc}^{-1}$$

$$\text{Pore space} = V_1 + V_2 - V_3$$

$$\text{Particle density} = \frac{W}{V_3 - V_2} \text{ g cc}^{-1}$$

b) Wax coating / clod method

A small undisturbed soil clod is weighed and then coated with molten wax. The volume of the clod is determined by inserting the clod into a measuring cylinder filled water (water displacement method).

c) Keen-Reczkowski cup method

A brass cylinder with perforated bottom is fitted with a filter paper and filled with air dry soil and placed in a tray containing water. The wet soil weight is noted. The expanded soil over the cup is removed and the weight of expanded soil and soil in the cup are weighed with moisture and after oven drying. From the data observed, the bulk density, particle density, pore space, volume ofexpansion and water holding capacity are determined.

d) Core sampler method

Core sampling cylinder with guard rings are inserted using a core sampler. The undisturbed soil in the cylinder is oven dried and weighed. From the weight of soil and volume of cylinder, the bulk density of soil is calculated. This is the true field bulk density determined in an un-disturbed soil core.

e) Penetrability test

The resistance offered by the soil to the force delivered by hand using a pin through the profile is assessed and bulk density is rated (rough estimate only).

Pin enters easily and neatly	1 g cc^{-1}	Soil is very loose
Pin enters easily with slight force	1.2 g cc^{-1}	Soil is loose
Half of the pin enters	1.4 g cc^{-1}	Moderately compact
Little part of pin enters	1.6 g cc^{-1}	Compact

Pin will not enter	1.8 g cc^{-1}	Very compact

Particle density (or) Absolute specific gravity

The weight per unit volume of the solid portion of soil is called as particle density. It is also called as true density and it is expressed in gram per cubic centimeter (gm/cm^3) or Megagram per meter cube ($Mg\ m^{-3}$).

$$\text{Particle density} = \frac{\text{Weight of soil}}{\text{Volume of soil solids}} \text{ gm / cc}$$

Soil particle density is a measure of the mass per unit volumes of the soil solids only. Generally particle density of normal soils is 2.65 g/cc or $Mg\ m^{-3}$. The particle density of most mineral soils is in the range of 2.60 to 2.75 g/cc or $Mg\ m^{-3}$. Soil particle density is determined using Archimede's principle.

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.

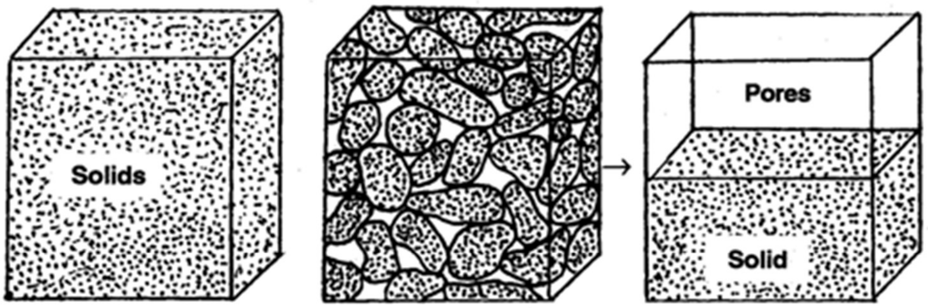
Particle density of different soil textural classes

Textural classes	**Particle density (g cc^{-1})**
Coarse sand	2.655
Fine sand	2.659
Silt	2.798
Clay	2.837

Factors affecting particle density

1. Texture and structure do not affect particle density.
2. However, organic matter, which is a soil solid, readily influences particle density. Organic matter much less per unit than soil minerals. Soils high organic matter has lower particle densities than soils that are low in organic matter.
3. Soil particle density generally increases with soil depth because of the decrease in organic matter.
4. Particle density varies with the type of soil minerals present as well as the amount of organic matter.
5. Generally quartz, feldspars, and colloidal silicates dominate the mineral fraction of soils.When large density of most minerals, such as hornblende or magnetite, are present, the soil particle density is greater than 2.65 g/cc or $Mg\ m^{-3}$.

6. Soils formed in volcanic parent materials, such as pumice or ash, generally have particle densities less than 2.65 g/cc or Mg m^{-3}.



Particle Density

100% solid
Weight = 2.66 g
Volume = 1 cm^3

Bulk Density

50% solid, 50% pore space
Weight = 1.33 g
Volume = 1 cm^3

Porosity of Soil

Pore-spaces (also called voids) in a soil consist of that portion of the soil volume not occupied by solids, either mineral or organic. The pore-space under field conditions, are occupied at all times by air and water.

The amount or ratio of pore space in a soil is determined by the arrangement of soilparticles like sand, silt and clay. In sandy soils, the particles are arranged closely and the pore space is low. In clay soils, the particles are arranged in popous aggregates and the pore space is high. Presence of organic matter increases the pore space.

Pore-spaces directly control the amount of water and air in the soil and indirectly influence the plant growth and crop production.

In general there are broadly two of pores in soils

(i) Macro pores

(ii) Micro or capillary pores.

13. **Macro pores**

Dimensions is larger than about 0.08 mm. Large-sized pores are referred as macro-pores which allow air and water movement readily. Sands and sandy soils have a large number of maco-pores. It is found in between the granules.

14. **Micro-pores or capillary pores**

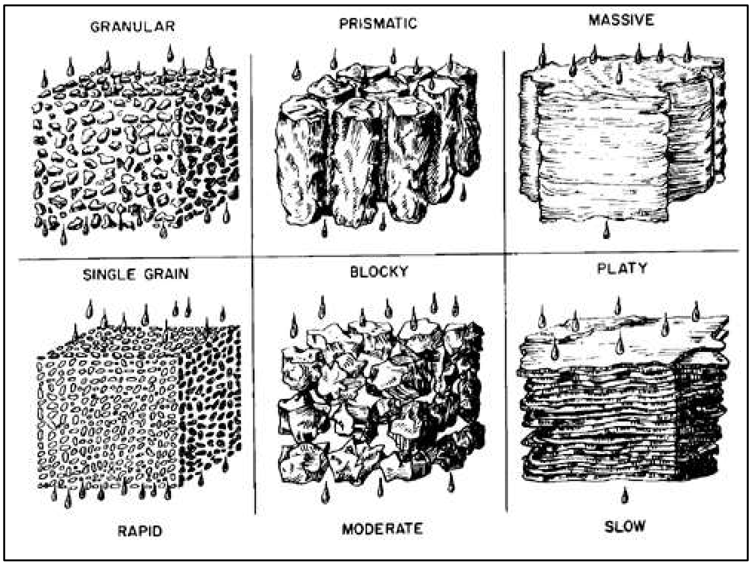
Dimensions is smaller than about 0.08 mm. Smaller sized pores are generally as micro or capillary pores in which movement of air and water is restricted to some extent. Clays and clayey soils have a greater number of micro or capillary pores. It has got more importance in the plant growth relationship. It is found within the granules.Loamy soils will have 50% porosity and have equal portion of macro and micro pores.

Biopores

Macropores created by roots, earthworms and other organisms constitute a very important type of pores termed biopores. For optimum plant growth, approximately equal amount of macro and micro-pores which favourably influences aeration, permeability, drainage and water retention is essential.

Factors Affecting Porosity

1) **Soil Structure:** A soil having granular or crumb structure contains more pore spaces than that of prismatic and platy soil structure. So the well aggregated soil structure has more porosity than that of structure less or single grained soils.



2) Soil Texture: In sandy soils the total pore space is small, where as in finetextured clay and clay loams total pore space and micro pores are higher.

3) Organic Matter Content: Soils containing high organic matter possesses high porosity because of compactness in the sub-soil.

4) Depth of Soil: With increase in depth of soil the porosity will decrease becauseof compactness in sub -soil.

5) Organisms: Macro organisms like earth worms, rodents, insects etc., increasemacro pores.

6) Cropping: Intensive cultivation tends to lower the porosity of soil as comparedto fallow soils. The decrease in porosity is due to reduction in organic mattercontent.

7) Puddling: Due to puddling under sufficient soil moisture, the soil surface layer is made dense and compact. Eventually, the porosity of this surface soil is reduced by the infiltration of muddy surface materials.

8) Arrangement of soil particles: When the spheres like particles are arranged in columnar form (one after another on the surface forming column like shape) it gives the most open packing system resulting very low amount of pore spaces. When such particles are arranged in the pyramidal form it gives the most close packing system resulting in high amount of pore spaces.

Significance of bulk density on crop growth

Generally soils with low bulk density have better physical condition than those with higher bulk densities. Texture and structure of a soil, its total pore space and organic content are all related to bulk densities. Soil density can be modified with aeration.In agricultural terms the bulk density of the soil can be used to give an indication of the porosity and structure of the soil which will, govern O_2 and H_2O movement in the soil.

It is also a measurement of the degree of compaction of the soil, which gives a comparative basis to indicative the strength of similar materials.One of the most important factors agriculturally in terms of bulk density is plant growth, if the soil has a high bulk density (compaction) the seed will be restricted in emergence and root growth which will affect the total plant growth and yield.

The use of tractors will directly affect the soils bulk density causing extreme compaction especially if the soil is wet, however lately the use of controlled traffic has decreased these problems.

Soil volume - weight relationship

1) Void ratio (e)	Ratio between pores (void) and solids
2) Porosity (N)	Ratio between pores (voids) and total volume of soil
3) Water content (w)	Ratio between weight soil water and soil solids expressed in percentage
4) Degree of saturation(S)	Ratio between volume of water and volume of pores (voids) expressed in percentage
5) Specific gravity (G)	Ratio between unit weight of unit volume of soil and the weight of unit volume of water at 4°C
6) Bulk density (B.D)	Ratio between dry weight of soil and the total volume of soil mass
7) Particle density (P.D)	Ratio between dry weight of soil and the volume of soil solids
8) Moisture content	Ratio between weight of soil water and bulk weightof wet soil expressed in percentage (on wet weight basis) (or) ratio between weight of soil water and oven dry weight of soil expressed in percentage (on oven dry weight basis)

Porosity = 1 - bulk density/particle density)

Void ratio = porosity/(1-porosity)

Lecture 15: Soil colour - causes and measurement - Munsell colour chart - factors influencing soil colour – Significance of soil colour

Soil Colour

The colour is the property of an object that depends on the wavelength of light it reflects or emits. Soils like other objects, reflect light of a great variety of wavelengths causing variety of colours. For example, the wavelength of green colour is 0.49 - 0.57 microns, yellow colour is 0.57 - 0.60 microns and red colour is 0.62 - 0.75 microns. Soil colour may vary from place to place and from horizon to horizon within the same profile. It also varies with soil moisture content and organic matter content. Moist soil are usually darker than dry soil. Within small geographic regions black colour of a soil is normally due to presence of high amount of organic matter. The soil colour may be inherited from its parent material for example, red soil inherited from red sandstone are the result of soil forming processes, such as red soils developed from granite, gneiss or schist.

Some other factors like ferric iron oxides produce yellow or red soil. Dark brown or black color in soil indicates that the soil has a high organic matter content. The colours of soils containing carbonate of calcium and magnesium are white, gray or sometimes olive green. Color development and distribution of color within a soil profile are part of weathering.

Soil colour indicates many soil features. A change in soil colour from the adjacent soils indicates a difference in the soil`s mineral origin (parent material) or in the soil development. Soil colour varies among different kinds as well as within the soil profile of the same kind of soil. It is an important soil properties through which its description and classification can be made.

Kinds of soil colour

Lithochromic

Soil colour is inherited from its parent material and that is referred to as lithochromic, E.g. red soils developed from red sandstone.

Pedochromic

Soil colour also develops during soil formation through different soil forming processes and that is referred to as acquired or pedochromic colour. E.g. red soils developed from granite or schist.

Factors affecting soil colour

There are various factors or soil constituents that influence the soil colour which are as follows.

1. Organic matter

Soils containing high amount of organic matter show the colour variation from black to dark brown.

2. Iron compounds

Soil containing higher amount of iron compounds generally impart red, brown and yellow tinge colour.

3. Silica, lime and other salts

Sometimes soils contain either large amounts of silica and lime or both. Due to presence of such materials in the soil the colour of the soil appears like white or light coloured.

4. Mixture of organic matter and iron oxides

Very often soils contain a certain amount of organic matter and iron oxides. As a result of their existence in soil, the most common soil colour is found and known as brown.

5. Alternate wetting and drying condition

During monsoon period due to heavy rain the reduction of soil occurs and during dry period the oxidation of soil also takes place due to development of such alternating oxidation and reduction condition, the colour of soil in different horizons of the soil profile is variegated or mottled. This mottled colour is due to residual products of this process especially iron and manganese compounds.

6. Oxidation-reduction conditions

When soils are waterlogged for a longer period, the permanent reduced condition will develop. The presence of ferrous compounds resulting from the reducing condition in waterlogged soils impart bluish and greenish colour.

Therefore, it may be concluded that soil colour indirectly indicative of many other important soil properties. Besides soil colour directly modify the soil temperature. E.g. Dark coloured soils absorb more heat than light coloured soils.

Determination of soil colour

Soil colours are determined by comparison with standard colour charts. Generally **Munsell soil colour charts** are used for determination of soil colour. The most commonly used charts consist of 199 different coloured chips symmetrically arranged on hue cards according to their Munsell notation in a loose note book. The Munsell colour system has three simple variable *viz.,* **hue, value and chroma**, which combine to describe all colours.

Red, brown, yellow, yellowish-red, grayish-brown, and pale red are all good descriptive colors of soil, but not very exact. Just as paint stores have pages of color chips, soil scientists use a book of color chips that follow the Munsell System of Color Notation**.**

This colour chart is commonly used for this purpose. The colour of the soil is a result of the light reflected from the soil.

Soil colour notation is divided into three parts

Hue - It denotes the dominant spectral colour (red, yellow, blue and green).

Value - It denotes the lightness or darkness of a colour (the amount of reflected light).

Chroma - It represents the purity of the colour (strength of the colour).

The Munsell colour notations are systematic numerical and letter designations of each of these three variables (Hue, Value and Chroma). For example,the numerical notation 2.5 YR6/6 suggests a hue of 2.5 YR, value of 5 and chroma of 6. The equivalent or parallel soil colour name for this Munsell notation is `red`.

Interpreting Soil Color

Color can be used as a clue to mineral content of a soil. Iron minerals, by far, provide the most and the greatest variety of pigments in earth and soil (see the following table).

Properties of Minerals

Mineral	Formula	Size	Munsell	Color
Goethite	FeOOH	(1-2 mm)	10YR 8/6	yellow
Goethite	FeOOH	(~0.2 mm)	7.5YR 5/6	strong brown
Hematite	Fe_2O_3	(~0.4 mm)	5R 3/6	red
Hematite	Fe_2O_3	(~0.1 mm)	10R 4/8	red
Lepidocrocite	FeOOH	(~0.5 mm)	5YR 6/8	reddish-yellow
Lepidocrocite	FeOOH	(~0.1 mm)	2.5YR 4/6	red
Ferrihydrite	$Fe(OH)_3$		2.5YR 3/6	dark red
Glauconite	$K(Si_xAl_{4-x})(Al,Fe,Mg)O_{10}(OH)_2$		5Y 5/1	dark gray
Iron sulfide	FeS		10YR 2/1	black
Pyrite	FeS_2		10YR 2/1	black (metallic)
Jarosite	$K Fe_3 (OH)_6 (SO_4)_2$		5Y 6/4	pale yellow

Todorokite	MnO_4		10YR 2/1	black
Humus			10YR 2/1	black
Calcite	$CaCO_3$		10YR 8/2	white
Dolomite	$CaMg(CO_3)_2$		10YR 8/2	white
Gypsum	$CaSO_4 \times 2H_2O$		10YR 8/3	very pale brown
Quartz	SiO_2		10YR 6/1	light gray

Soil color is a study of various chemical processes acting on soil. These processes include the weathering of geologic material, the chemistry of oxidation-reduction actions upon the various minerals of soil, especially iron and manganese, and the biochemistry of the decomposition of organic matter. Other aspects of Earth science such as climate, physical geography, and geology all influence the rates and conditions under which these chemical reactions occur.

Importance of soil colour

i. Colour is one of the most useful and important characteristics for identification and classification of soils.
ii. Colours help to distinguish the different horizons of a soil profile.
iii. Soil colour is an indicator of the soil moisture regimes under which a soil was developed. When a dry soil becomes moist, soil colours become darker.
iv. The oxidation and reduction changes the colour of iron containing minerals.
v. The productivity of a soil can be judged from its colour. Generally the darker a soil, the higher is it productivity. Light colour usually contain quarts which has no nutritional value.
vi. Dark coloured soils absorb heat more readily and warm up quickly than light coloured soil. But dark coloured soil contain high amount of organic matter and water holding capacity.

Lecture 16: Soil consistence - cohesion, adhesion, plasticity, Atterberg's constants - upper and lower plastic limits, plasticity number- significance of soil consistence

Definition

Soil consistence is defined as “the resistance of a soil at various moisture contents to mechanical stresses or manipulations”. It combines both the ‘cohesive’ and ‘adhesive’ forces, which determine the ease with which a soil can be reshaped or ruptures.

Soil consistency is also an important physical and dynamic property which varies with the variation of soil moisture and applied stress. Consistency is the behaviour of soil under stress. This stress is commonly evidenced by feeling the soil, manipulating it by hand, or by tillage operation. Soil consistency is considered a combination of soil properties dependent upon the forces of attraction between soil particles as influenced by soil moisture.

Forces causing soil consistency

Two main forces are responsible for soil consistency. They are i) Cohesion and ii) Adhesion.

i) Cohesion

Force holding a solid or liquid together, owing to attraction between like molecules. Decreases with rise in temperature.

ii) Adhesion

Molecular attraction that holds the surfaces of two substances of different nature (E.g. Water and soil particles) in contact.

Forms of soil consistency

Atterberg (1912) suggested four forms of soil consistency excluding the viscous state and five forms of consistency including the viscous state as a result of differences in soil water contents. They are

1. Hard or harsh consistency

This occurs in dry soil (pF > 4.5). At this consistency soil are very hard and coherent because of cementation effect between the dried particles and soil particles once apart do not stick together again.

2. Soft or friable consistency

This occurs in moist soil (pF 4.5 - 2.8). Soil particles stick together only lightly. The adsorbed water molecules decrease the coherence and make the soil mass soft and friable. The friable consistency represents the optimum moisture condition for tillage.

3. **Plastic consistency**

This occurs in fairly wet soil (pF 2.8 - 0.5). At this consistency soils are tough, exhibit considerable cohesion can be moulded without loosing its coherence and have the capacity to change its shape without cracking.

4. **Sticky consistency**

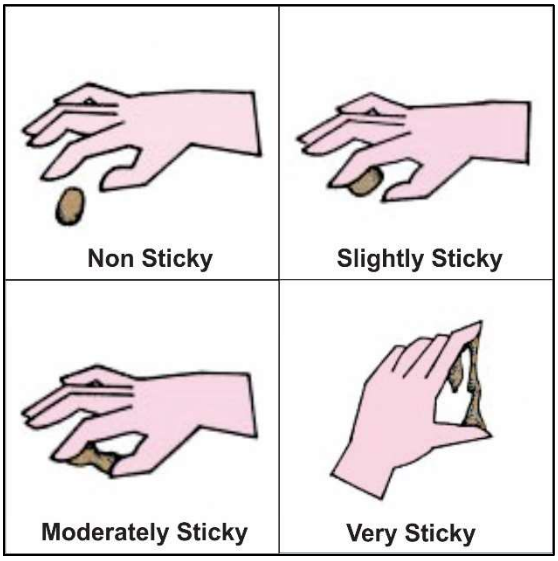
This occurs in very wet soil (pF 0.5 - 0). At this consistency soils adhere to various objects and exhibit the properties of stickiness. This is because the force of adhesion of soil to foreign object like plough is greater than cohesion within the soil itself.

5. **Viscous consistency**

Viscous consistency is for saturate soil (pF 0). In viscous consistency soil flows either under pressure or merely by gravity. At this consistency tillage by ordinary procedure is impossible because here soils run together with water.

Soil consistency description

Soil consistence is described at three moisture levels namely ‘wet’, ‘moist’ and ‘dry’.



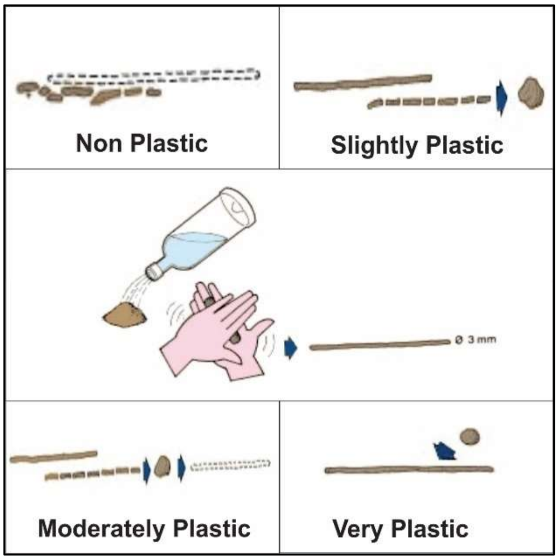
i. **Wet soils:** Consistency is denoted by terms stickiness and plasticity. Stickiness is grouped into four categories namely
   - i) **Non sticky**- Soils do not adhere to the fingers after release of pressure
   - ii) **Slightly sticky**- Soils adhere to one of the two fingers. The soils do not stretch appreciably after the release of pressure
   - iii) **Moderately sticky**- Soil adhere to both the fingers and stretch slightly after the release of pressure
   - iv) **Very sticky**– Soils strongly adhere to both the fingers and stretch when the fingers are separated.

Plasticity

The capacity soil to be moulded (to change its shape depending on stress) and to retain the shape even when the stress is removed. Soils containing more than about 15% clay exhibit plasticity.

There are four degrees in plasticity namely

i) **Non plastic** - A wire of 4 cm long and 6 mm thick that supports its own weight, if held on one end, cannot be formed



ii) **Slightly plastic** - A wire of 4 cm long and 6 mm thick, that supports its own weight if held on one end, is formed. But a wire of 4 cm long and 4 mm thick that supports its own weight can not be formed.

iii) **Moderately plastic** - A wire 4 cm long and 4 mm thick that support its own weight if held on one end, can be formed but a wire of 4 cm long and 2 mm thick that supports its own weight if held on one end, cannot be formed. Modrate pressure is needed to deform that molded soil mass.

iv) **Very plastic** - A wire of 4 cm long and 2 mm thick that supports is own weight if held on one end can be formed. Very high pressure is required for deformation of the moulded soil mass.

ii. Moist soil: Moist soil with least coherence adheres very strongly and resists crushing between the thumb and forefinger. The different categories are

i. Loose - Non coherent and soil particles exist as individual particles

ii. Very friable - Coherent, but very easily crushed (equivalent stress is < 8 Newtons)

iii. Friable - Easily crushed (equivalent stress is 8-20 Newtons)

iv. Firm - Crushable with moderate pressure (equivalent stress is 20-40 Newtons)

v. Very firm - Crushable only under strong pressure (equivalent stress is 40 – 80 Newtons)

vi. Extremely firm - Completely resistant to crushing (equivalent stress is 80 – 160 Newtons). It is based on type and amount of clay and humus present is soil.

iii. Dry soil: In the absence of moisture, the degree of resistance is related to the attraction of particles for each other. The different categories are

i) Loose - Non coherentand soil particles exist as individual particles

ii) Soft - Breaks with slight pressure and becomes powder (equivalent stress is < 8 Newtons)

iii) Slightly hard - Break under moderate pressure (equivalent stress is 8-20 Newtons)

iv) Moderately Hard- Breaks with difficulty with pressure (equivalent stress is 20-40 Newtons)

v) Hard - Very resistant to pressure (equivalent stress is 40 – 80 Newtons)

vi) Veryhard - Extreme resistance and cannot be broken. (equivalent stress is 80 – 160 Newtons)

Significance of soil consistency

Under field situations, soil consistency gives an indication of root impedance and workability.

It has thus a major bearing on

- The productive capacity of the soil
- The suitability of the soil for various purposes
- The flow path through which the soil water moves within the soil and between the landforms.
- How the soil will respond to various management strategies

In general, sandy soil will have loose consistency. On the other hand, the loam and clay loams have greater diversity of consistency properties and can range from being soft to strong.

Factors affecting soil consistency

i) **Soil texture:** As sand and silt particles have insignificant molecular attraction, consistency of coarse textured soils is negligible. Soils with sufficient clay exhibit the phenomena of consistency. Consistency of soil increases with increase in clay content. If consistency of a dry soil is destroyed by separating the soil particles, it can be re-established by rewetting the soil through which the particles are again oriented.

ii) **Type of clay:** Expanding type clay such as montmorillonite has more consistency than non expanding type clay such as kaolinite as because montmorillonite clay has more surface charge and more specific surface than kaolinite.

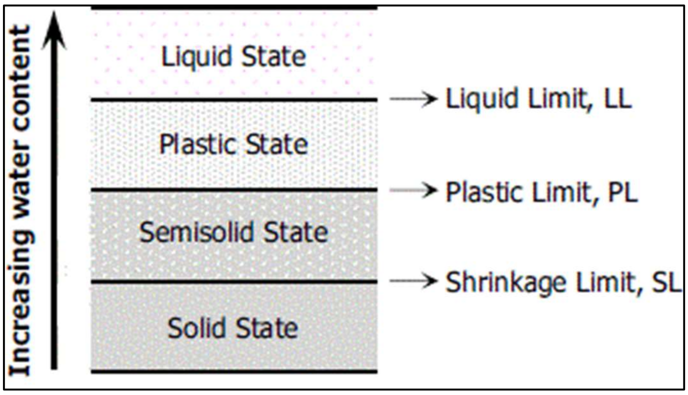
iii) **Organic matter:** It has more consistency than sand and silt but less than clay because cohesion of organic matter is more than sand and silt but less than clay.

iv) **Structure:** Well aggregated soil has less consistency than a puddled soil because well aggregated soil has less areas of contact between the individual particles than a puddled soil. Puddling orients the clay particles in such a way that number of micro pores as well as total pores space is reduced sharply. When a puddled soil becomes dry, its consistence is sharply increased because of the increase of surface contacts.

Atterberg's constants

Atterberg (1912) studied plasticity from the point of view of moisture range over which plasticity range is maintained.

The constants are also known as **Atteberg's limits**.



There are three limits namely

1. Shrinkage limit or lower plastic limit,
2. Plastic limit and
3. Liquid limit or upper plastic limit.

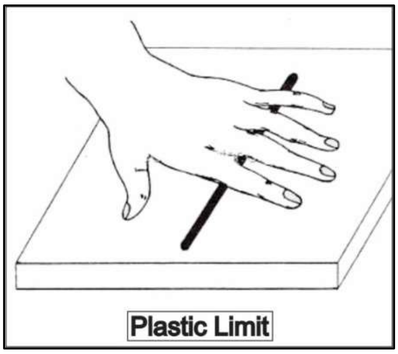
From the upper and lower limits, plasticity number or index is calculated.

Plasticity number or index is an indirect measure of the forces required to mould the soil.


Lower Plastic Limit

Shrinkage limit or Lower plastic limit (SL)

It is the moisture content below which the soil becomes friable. The maximum water content at which a reduction in moisture will not cause a decrease in the volume of the soils. This defines the arbitrary limit between solid and semi-solid states (The boundary between the semi-solid and solid states).


Plastic Limit

Plastic limit (PL)

It is the moisture content at which the soil begin to exhibit plasticity. Soil cannot be deformed without cracking. Soils should not be ploughed at moisture content above the plastic limit (The boundary between the plastic and semi-solid states).


Casagrade

Liquid limit or Upper plastic limit (LL)

The moisture content, at which soil stops to be plastic, becomes semi-fluid and tends to flow like a liquid (The boundary between the liquid and plastic states). The determination of liquid limit by **Casa grade** device.

Plasticity number

The difference between the moisture content of upper and lower plastic limits. Different soils are characterized by a specific plastic number or index of plasticity.

Factors affecting Atterberg's constants

1. Clay content

Plasticity is a function of finer soil fractions and moisture content. Because of the plate like shape of clay particles and the lubricating effect of water, the fine soil fractions tend to slide over each other. High clay content increases the moisture content of different plastic limits and increases the plasticity index / number.

2. Nature of clay minerals

Quartz and feldspars are non plastic. Kaolinite, illite, talc, muscovite, biotite, vermiculite and montmorillonite clays are plastic nature.

3. Nature of exchangeable cations

Sodium saturated soils have lower plastic limits than potassium, calcium and magnesium saturated soils.

4. Organic matter content

Organic matter favours plasticity.

Plasticity index

The plasticity index (PI) is a measure of the plasticity of a soil. The plasticity index is the size of the range of water contents where the soil exhibits plastic properties. The PI is the difference between the liquid limit and the plastic limit (PI = LL-PL). Soils with a high PI tend to be clay, those with a lower PI tend to be silt, and those with a PI of 0 (non-plastic) tend to have little or no silt or clay.

Soil descriptions based on PI

PI	Description
0	Non plastic
1-5	Slightly plastic
5-10	Low plasticity
10-20	Medium plasticity
20-40	High plasticity
>40	Very high plasticity

Liquidity index

The **liquidity index** (LI) is used for scaling the natural water content of a soil sample to the limits. It can be calculated as a ratio of difference between natural water content, plastic limit, and liquid limit: LI=(W-PL)/(LL-PL) where W is the natural water content.

Consistency index

The **consistency index** (CI) indicates the **consistency** (firmness) of a soil. It is calculated as CI = (LL-W)/(LL-PL) , where W is the existing water content. Soil at the liquid limit will have a **consistency index** of 0, while soil at the plastic limit will have a **consistency index** of 1.

Flow index

The flow index (FI), which is the slope of the water content versus $\log_{10}$ (number of blows) plot in the percussion cup method or the slope of the water content versus $\log_{10}$ (depth of penetration) plot of the cone penetration method of determining the liquid limit of soils is a measure of soil plasticity.

This flow index is nothing but it indicates the rate at which the soil looses shearing resistance with an increase in the water content.

Sticky point

The moisture content at which the attractive power of the soil for water is satisfied. The sticky point moisture percentage is near the liquid limit (slightly higher for less plastic soils and slightly lower for more plastic soils.

Plasticity chart for fine-grained soils

Casagrande (1932) studied the relationship of the plasticity index to the liquid limit of a wide variety of natural soils.

Many properties of clays and silts (the cohesive soils) such as their compressibilitycan be correlated with the liquid limit and the plasticity index. This correlation has been expressed in Casagrande's plasticity chart for fine-grained soils. It is based on the following observations:

- As the liquid limit of soils increases, the plasticity and compressibility of soils also increase.
- The values LL = 30 percent and LL = 50 percent differentiate between the various degrees of plasticity of inorganic soils.
- At equal LL values, the dry strength of inorganic soils increases generally with an increasing plasticity index.

The plasticity chart for fine-grained soils is divided into six sections by the oblique line A drawn so that the PI = 0.73 (LL - 20) and two vertical lines drawn at LL = 30 percent and LL = 50 percent.

Each section of the chart characterizes a group of soils with well-defined mechanical characteristics. The three sections above line A are inorganic clays of low, medium or high plasticity. The three sections below line A are inorganic silts of varying compressibility, organic silts and organic clays. These form the basis of a useful soil classification system.

Significance of Atterberg limits

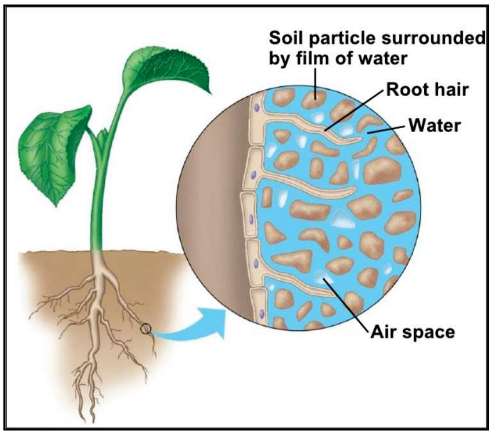
- Atterberg limits and constants may serve as an index for physical classification of soils.
- It is an index of the accumulation of clay within a profile and express the difference in consistency of various soils.
- It is a best index to soil tilth.
- The shrinkage limit signifies that water content at which soils are friable.
- The sticky point represents the maximum water content at which the soil will scour during tillage operations.

Relationships among soil moisture condition, soil consistency and some other related properties of medium to fine textured soils

Soil moisture condition	Dry	Moist	Wet		Saturated
			Fairly wet	Very wet	
1. Soil moisture tension					
i) pF	>4.5	4.5-2.8	2.8-0.5	0.5-0	0
ii) Bar	>31	31-0.62	0.62-0.003	0.003-0.001	0.001
2. Consistency form	Hard or harsh	Friable or soft	Plastic	Sticky	Viscous
3. Relative degree of consistency	Very high	Low	High	Low	Very low
4.Forces mainly responsible for consistency	Cohesion	Adhesion	Surface tension and Cohesion	Negligible tension	Tension free water
5.Soil condition	Hard and coherent	Soft	Stick together	Stick to foreign particles	Soil is as water suspension
6.Strength	High	Fairly high	Low	Very low	None
7.Draft requirement for tillage	Heavy	Light	Heavy	Light with slow traction	None
8.Result of tillage	Clod formation	Optimum for tillage	Puddles	Soil adhere to plough	Soil run together

Lecture 17: Soil water- forms of water, units of expression and pF scale

Soil water



Water contained in soil is called soil moisture. The water is held within the soil pores. Water enters soil through precipitation. It falls to the ground and accumulates in a very thin layer right at the surface. From this thin film at the surface, the water infiltrates into the soil, where it accumulates in empty pore spaces. Some of the water in the soil percolates down through the intermediate zone into the deeper groundwater, where it flows away from places where the water table is higher. Much of the soil water is drawn back up to the surface by plants and by normal evaporation.

Soil water is an important component of the soil which influences soil organisms and plant growth. Unlike open source water like ponds and lakes, soil water exists in different forms. Some forms are available to plants while others are not. Soil water with dissolved salts is called soil solution. Soil solution supplies nutrients to growing plants.

Importance of Soil Water

- Water is essential for life including soil organisms and plants.
- Soil water serves as a solvent and carrier of food nutrients for plant growth.
- Yield of crop is first determined by the amount of water available.
- Availability of other food nutrients.
- Soil water itself is a nutrient.
- Soil water regulates soil temperature.
- Soil water is essential for soil forming processes and weathering.
- Microorganisms require water for their metabolic activities.
- Soil water helps in chemical and biological activities of soil.
- It is a principal constituent of the growing plant.
- Water is essential for photosynthesis.

Water serves four functions in plants

It is the major constituents of plant protoplasm (85-95%). It is essential for photosynthesis and conversion of strarches to sugars. It is the solvent in which nutrients move into and through plant parts to capture sunlight.In fact, the soil water is a great regulator of physical, chemical and biological activities in the soil.

Plants absorb some water through leaf stomata (openings), but most of the water used by plants is absorbed by the roots from the soil. For optimum water used, it is vital to know how water moves into and through the soil, how the soil stores water, how the plant absorbs it, how nutrients are lost from the soil by percolation, and how to measure soil water content and losses.

Soil also serves as a regulated reservoir for water because it receives precipitation and irrigation water. A representative cultivated loam soil contains approximately 50% solid particles (sand, silt, clay and organic matter), 25% air and the rest 25% mater. Only half of this water is available to plants because of the mechanics of water storage in the soil.

Structure of water

Water occurs from troposphere (15 km above mean sea level) to 3 to 11 km deep from soil surface (ocean). Water can participate in a series of reactions occurring in soils and plants, only because of its structural behavior. Water is simple compound, its individual molecules containing one oxygen atom (with two negative charges) and two much smaller hydrogen atoms (with positive charges each). The elements are bonded together covalently, each hydrogen or proton sharing its single electron with the oxygen. Instead of the atoms being arranged linearly (H-OH) the hydrogen atoms are attached to the oxygen as a V shaped.

The volume (space) occupied by water molecule is determined by the oxygen ion as the diameter of oxygen is 2.64 Å which is several times larger than hydrogen ions. The two hydrogen ions take practically no space.

The hydrogen atoms are attached to oxygen at an angle of 103° to 106° with the oxygen nucleus as the apex. The positive charges of the hydrogen are partially neutralized by the negative charges of oxygen and hence there will be always a net negative charge on the oxygen side and positive charge on the hydrogen side. This makes water molecule a **dipole**.

Water is a giant polymer of hydrogen bonded water molecules. Water molecules are bondedwith each other very tightly. The melting point, boiling point, dielectric constant, specific heat and viscosity are higher than similar liquids.

Heat of vapourization

Evaporation is an endothermic reaction. About 580 calories per gran water are absorbed in evaporation. To evaporate 1 cm of water over one hectare area requires 50 million K calories of energy.

Heat of fusion

Freezing is an exothermic reaction.

Water is a solvent

Due to its dipole nature and high dielectric constant, the dissolved salts cannot reorient itself again as salt. The energy required to displace the hydrogen bonds of water is called dielectric constant.

Density of water

Highest density is at 4°C. Above and below 4°C, the density decreases. At 25°C, the densityof water is around 0.97 to 0.98 g cc^{-1}.

Retention of Water by Soil

The soil holds moisture due to their colloidal properties andaggregation 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 due togravity has stopped) are surface tension and surface attraction (surface moisture tension) and are described in terms of different energy concepts. The force with which water is held is also termed as suction.

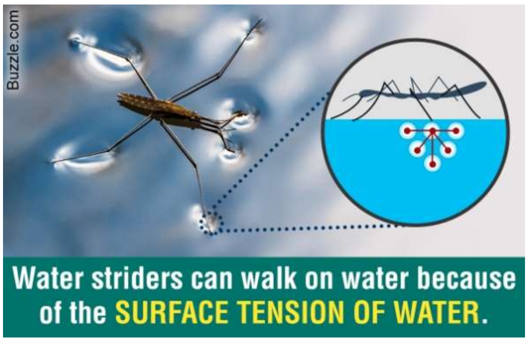
The water retained in the soil by following ways

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 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 evidencedat 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,



Water striders can walk on water because of the SURFACE TENSION OF WATER.

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 claymicelle is based on the dipole character of the molecule of water. The water molecules areheld 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 itdownwards. 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:** Fine texture soil have more pore space more surface area and greater 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 have more water retention in the soil.
4. **Density of soil:** Higher the density of soil have lower moisture content.
5. **Temperature:** Cooler temperature retain higher moisture.
6. **Salt content:** More salt content of soil reduce the water availability to plant.
7. **Depth of soil:** More depth of soil increase the water availability to plant.
8. **Type of clay:** The 2:1 type of day increases the water retention in the soil.

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.

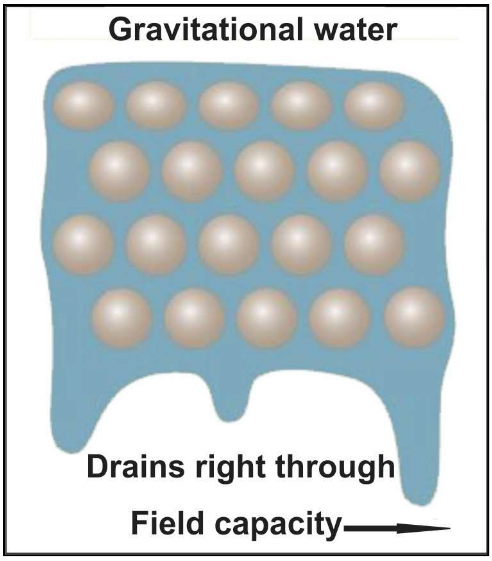
A. Physical classification of soil water

1. Gravitational water
2. Capillary water
3. Hygroscopic 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.



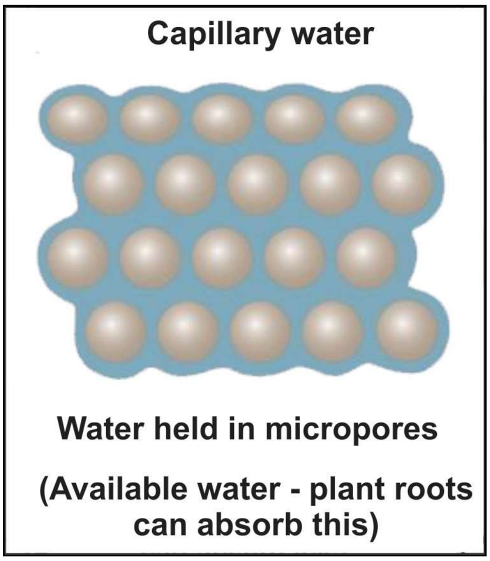
Factors affecting gravitational water

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 crumby 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

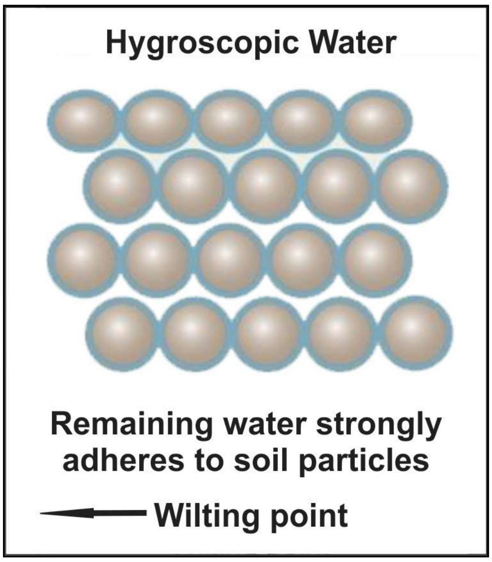
The finer the texture of a soil, greater is the amount of capillary water holds. This is mainly due to the greater surface area and a greater number of micro pores.

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

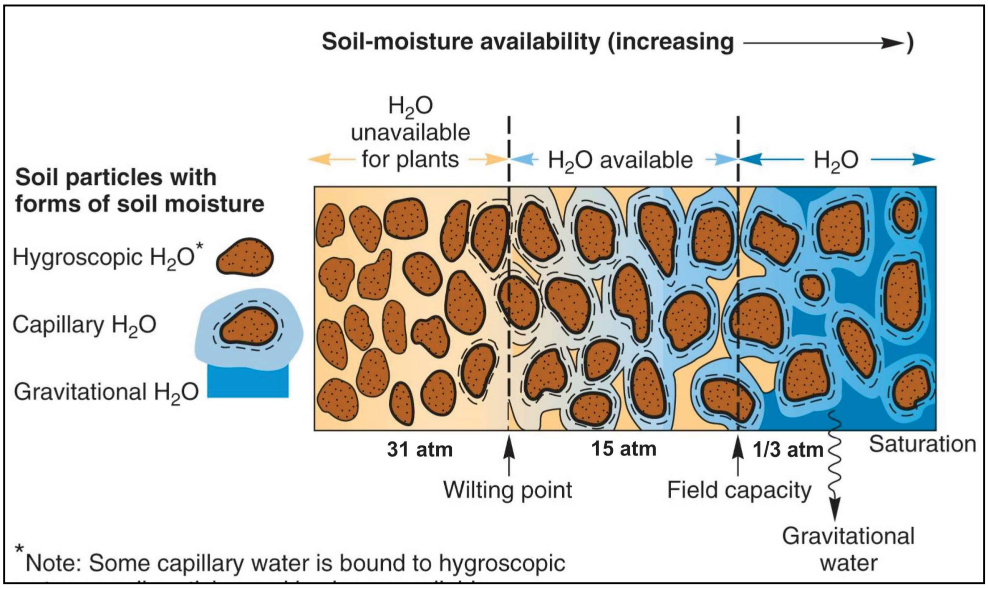
The water that held tightly on the surface of soil colloidal particle is known as hygroscopic water. It is essentially non-liquid and moves primarily in the vapour form.

Hygroscopic water held so tenaciously (31 to

10000 atmosphere) by soil particles that plants cannot 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 montmoril1onite type with their large surface area adsorb more water than those of the kaolinite type, while illite minerals are intermediate.

B. 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

ii. Unavailable water

iii. Super available or superfluous water

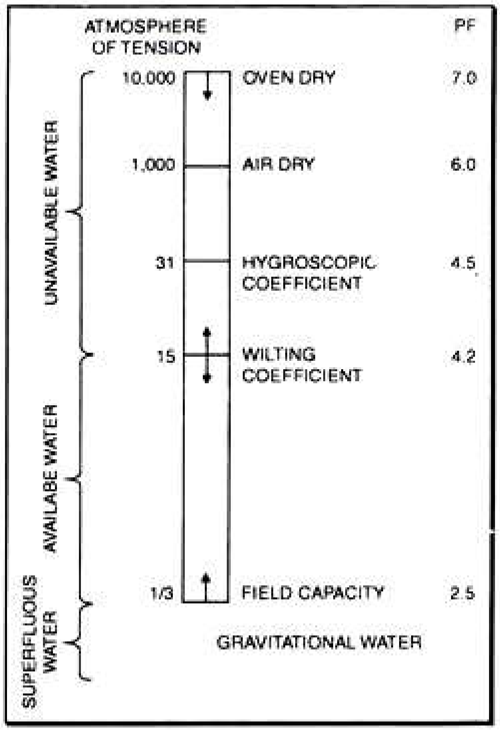
i. Available water

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.

Relationship between soil moisture and tension

Determination of soil moisture

The soil moisture is determined *viz.,* the amount of water in the given soil and the stress or tension under which the water is held by the soil. The relationship between these two properties throughout the entire moisture range gives a good deal of insight in to the physical characteristics of a soil.

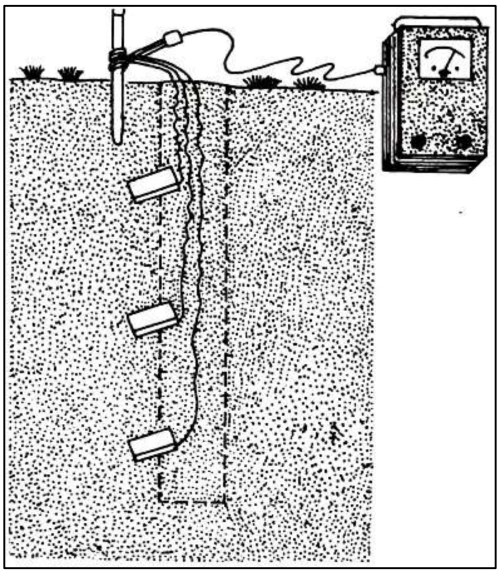
i. Gravimetric method

This is a standard method in which the soil sample is collected from the field (moist condition) and weighed. The weighed sample is oven dried at about 105-110°C for overnight and weighed again.The weight difference is considered as water content. An error may result from oxidation of organic matter and also while cooling the sample after taking out the same from oven before weighing. Several methods have been developed to simplify the determination of the amounts of soil moisture. More recently infrared radiation is utilized for drying soil samples quickly.

$$\%\ \text{moisture} = \frac{\text{Weight of fresh soil} - \text{Weight of oven dry soil}}{\text{Weight of oven dry soil}} \times 100$$

ii. Electric resistance method or electrical conductivity method

In this method two platinum electrodes are embedded in gypsum block at a definite distance apart. These gypsum blocks require calibration for uniformity before use. These blocks are buried in the soil at a required depth and conductivity is measured with a modified Wheatstone bridge.

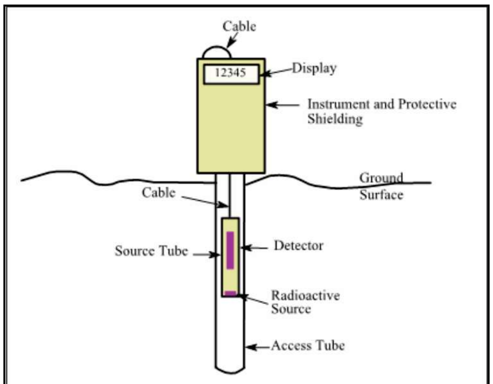


The electrical resistance of the soil decreases with increase in water content. The soil water content is obtained with the help of a calibration curve, for the same block of electrical resistance against known soil water content.

This method is useful to measure the percentage of soil moisture from field capacity to wilting percentage. This is not an appropriate method in soils containing high salt concentration.

iii. Neutron scattering method

The neutron moisture meter consists of americium and beryllium or radium and beryllium as the neutron source and boron tri fluoride (BF3) gas as a detector.



Neutron Moisture Meter

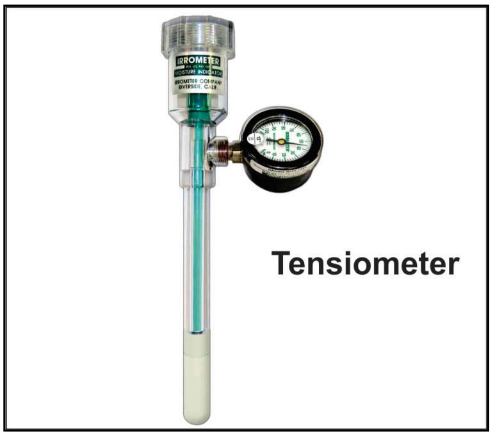
The fast moving neutrons emitted in to the soil collide with hydrogen atoms of water molecules, lose their energy and their movement will be slowed down.

The slow neutrons are caught by BF3 gas and counted with the help of rate meter or electronic microprocessor. The more slowed down neutrons return to the probe the greater is the water content of soil.

This instrument has to be calibrated for each soil. This method is rapid, less laborious, non-destructive and repeated measurements can be made at the same depth. However, this method is not suitable for moisture determination of surface soil.

Soil moisture tension determination

i. Tensiometer method



Tensiometers measure the matric potential of the soil moisture. The tensiometer consists of a porous cup made of ceramic material and a mercury manometer attached to the water filled cup through a water reservoir tube.

When the porous cup is placed in soil and equilibrated, water tends to move out of the cup under the suction exerted by soil. These suction readings are thus calibrated on the gauge to a specific soil to interpret the percent of moisture.

The use of tensiometer is confined up to the minimum matric potential of -0.8 bars, but much of the available water is retained by soil between -0.8 bars and -15 bars, particularly in fine textured soils. Beyond the tensiometric range, the matric potential is usually interpolated from the water retention characteristics curve using the measured soil moisture content.

ii. Psychrometer method

The total soil water potential is determined with the help of soil psychrometer. As the liquid water in soil is always in equilibrium with its vapour in the ambient air, the energy status of water vapour (vapor potential) must be equivalent to the energy status of soil water.

At constant temperature and gravitational potential, the vapour potential is largely contributed by matric and osmotic potentials. The vapour potential of soil can be precisely estimated with the help of a thermocouple psychrometer.

iii. Pressure plate Apparatus

It is a form of tensiometer used under laboratory conditions. A core of soil is placed firmly on a porous plate to which a suction is applied.

The soil core eventually reaches an equilibrium with porous plate. The soil is weighed and relationship between tension and moisture content is determined. Suitable in 0-1 bars tension.

iv. Pressure membrane apparatus

Pressure membrane apparatus is widely used for the investigation of soil moisture suction. The results obtained are usually expressed in the form of moisture content versus suction (pF) curves. They have wide application in engineering, agriculture and other fields, where moisture movement, pore water pressures and related phenomena are being considered. The pressure membrane apparatus can be used to much higher soil suction values than is possible with the simpler suction plate apparatus, where the suction is limited by the instability of a water column to a pressure of about -13 psi (0.92 kg/cm^2) and it is used to measure the tension values as high as 100 bars. Helpful to measure energy moisture relation of a number of samples over a wide energy ranges in a short time.

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 (to the base 10) of the height of water column (cm) to give the necessary suction.

pF Concept

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. To express the soil moisture suction or tension **Schofield 1935** has suggested to use the logarithm of this tension with a symbol pF, an exponential function of free energy difference. It is defined as the logarithm of the height of water column (cm) to give the necessary suction.

(or)

Logarithm to the base 10 of the numerical value of the negative pressure of the soil moisture expressed in centimeters.

$$pF = \log(-h)$$

pF is the logarithmic expression of the force with which water is held in soils expressed in terms of height of a water column (cm) required to produce equal tension.

P = logarithmic value F= free energy

Height of water column (cm)	pF
1	0
100	2
346	2.53 (field capacity)
15849	4.1 (wilting point)
31623	4.5 (Hygroscopic coefficient)

Classification of soil moisture on the basis of pF

Water of constitution and inter layer water	pF >7.0
Hygroscopic water	pF 7.0 – 4.5
Capillary water	pF 4.5 – 2.5
Gravitational water	pF 2.5 – 0.0
Ground water	Tension free

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 atmosphere or 1pF 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.

Lecture 19: Soil water potentials - gravitational, matric, osmotic- Soil moisture constants and Soil moisture measurements

SOIL WATER POTENTIAL

The retention and movement of water in soils, its uptake and translocation inplants and its loss to the atmosphere are all energy related phenomenon. The morestrongly water is held in the soil the greater is the heat (energy) required. In otherwords, 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.

Different kinds of energy are involved including potential, kinetic and electrical. By using the term ‘free energy’ (ability to do work) energy status of water can be characterized to indicate the strength with which water is held. Several concepts have been used.

The pressure required to force the water off soil and was measured in atmospheres of pressure needed to remove water. The opposite of pressure moisture suction or tension. The soil water potential is used and it may be defined as the work the water can do when it movies from its present state to a pool of water in the reference state. The movement of water in soil takes place from a higher free energy to a lower free energy level. It is expressed that there is great variability in the free energy levels of water in soils. So the tendency for soil water to move from one soil zone to another due to variation in free energy levels.

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.

Concept of water potential

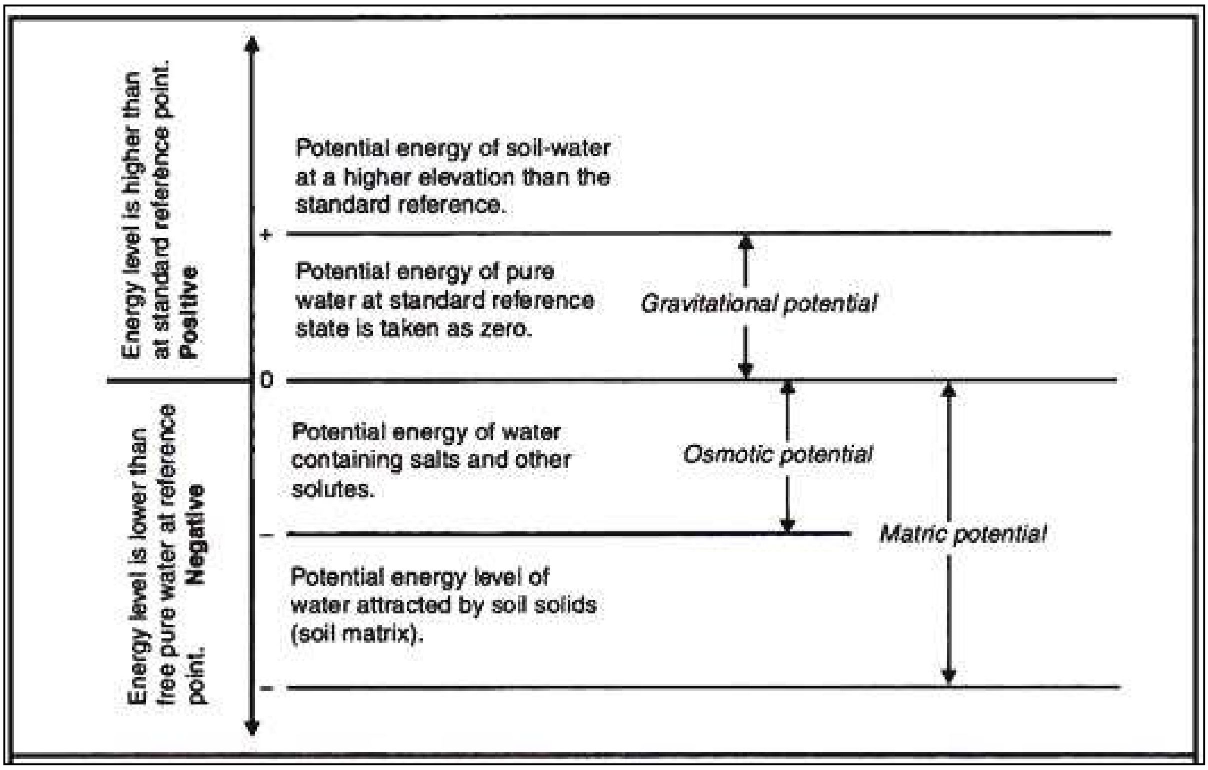
Water in soil has potential energy as well as kinetic energy. Kinetic energy is very small. Potential energy may be defined as the capacity to do work. Pure water has the maximum capacity to do work. Water in soil is held by adsorptive, osmotic and pressure gradient forces and also has relatively lower capacity to do work. However, work is necessary for the movement of water from one position to another against the force filed to which it is subjected. Extraction of water by plant roots is an example of work done on soil water. Since the term potential refers to the work done per unit quantity, it can be used quantitatively to the work done by water or work done on water as a function of its energy status. Work is positive when water loses energy and is negative when it gain energy due to movement.

The energy with which the water is held by the soil is as important as the amount of water in a soil. This energy at any given temperature usually is measured with reference to a flat surface of pure water at some specified elevation and at a particular pressure. Pure water in a saturated soil sample at the same elevation, pressure and temperature as the reference has a total water potential of zero.

As the water content of the soil decreases, the force with which the remaining water is held by the soil particles (adhesion) increases. Since energy must be added to this water to restore to the reference state, its potential energy is said to be negative. Similarly, water potential of a soil at a lower elevation than the reference is negative. If it is higher than the reference level, its water potential can be positive. The same holds good for samples at different pressures than the reference. Solutes in the soil-water also reduce its potential energy.

Soil Water potential is the potential energy of water per unit volume relative to pure water in reference conditions. Water potential quantifies the tendency of water to move from one area to another due to osmosis, gravity, mechanical pressure, or matrix effects such as capillary action (which is caused by surface tension). The concept of water potential has proved useful in understanding and computing water movement within plants, animals, and soil. Water potential is typically expressed in potential energy per unit volume and very often is represented by the Greek letter ψ.

Water potential integrates a variety of different potential drivers of water movement, which may operate in the same or different directions. Within complex biological systems, many potential factors may be operating simultaneously. For example, the addition of solutes lowers the potential (negative vector), while an increase in pressure increases the potential (positive vector). If flow is not restricted, water will move from an area of higher water potential to an area that is lower potential. A common example is water with a dissolved salt, such as sea water or the fluid in a living cell. These solutions have negative water potential, relative to the pure water reference. With no restriction on flow, water will move from the locus of greater potential (pure water) to the locus of lesser (the solution). The flow proceeds until the difference in potential is equalized or balanced by another water potential factor, such as pressure or elevation.



Components of water potential

Many different factors may affect the total water potential, and the sum of these potentials determines the overall water potential and the direction of water flow

$$\psi = \psi_0 + \psi_\pi + \psi_p + \psi_g + \psi_v + \psi_m$$

where

ψ_0 - is the reference correction

ψ_π - is the solute or osmotic potential

ψ_p - is the pressure component

ψ_g -is the gravimetric component

ψ_v - is the potential due to humidity

ψ_m - is the potential due to matrix effects (E.g. **Fluid cohesion and surface tension**.)

All of these factors are quantified as potential energies per unit volume, and different subsets of these terms may be used for particular applications (E.g. **Plants or soils**). Different conditions are also defined as reference depending on the application: for example, in soils, the reference condition is typically defined as pure water at the soil surface.

As defined by the International Society of Soil Science (1963), the total potential of soil-water is "the amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at a specified elevation at atmospheric pressure to the soil-water (at the point under consideration).

This total water potential (ψ_t) can be divided into parts to distinguish between the actions of different force fields. The algebraic sum of these parts or component potentials must always equal to the total water potential.

The four component potentials distinguished are

1. The matric or capillary potential (ψ_m) which results from the interaction of soil particles surfaces with water
2. The osmotic potential (ψ_0) which results from the solutes dissolved in the soil- water
3. The gravitational (ψ_g) which results from elevation with respect to reference level
4. The pressure potential (ψ_p) which results from external pressure on the soil-water.

In unsaturated soils, the pressure potential (ψ_p) is usually considered zero and in saturated soils the matric potential (ψ_m) is usually considered zero. Osmotic potential (ψ_0) is of considerable importance with respect to plant growth but of less consequence where water movement is considered.

If the quantity of water involved in the above definition of water potential is a mass, water potential has units as joules kg^{-1} or ergs g^{-1}. The quantity of water can be expressed in terms of a volume, in which case the water potential has units such as ergs cm^{-3}, which is dimensionally the same as pressure.

Types of Soil-Water Potential

Soil-water potential (soil-water tension) is a measure of the tenacity with which water is retained in the soil and shows the force per unit area that must be exerted to remove water from the soil. It is measured in terms of potential energy of water in the soil measured, usually, with respect to free water. It is expressed in atmospheres, the average air pressure at sea level. Other pressure units such as cm or mm of water or mercury are also used.

1 atmosphere = 1036 cm of water = 76.39 mm of mercury.

1 bar = 10^6 dynes cm^{-2}. = 1036 cm of water.

1milli bar = 1/1000 bar.

i. Total Water Potential (ψ_t)

The total water potential is the sum of the gravitational potential, the matric potential, the pressure potential and the osmotic/solute potential and any other external potential which may be working on the system.

$$\psi_t = \psi_g + \psi_m + \psi_p + \psi_o$$

where,

ψ_g - Gravitational potential - Only matters when the soil is saturated

ψ_m - Matric potential - Water potential of soils

ψ_p - Pressure potential - Negligible in soils

ψ_o - Osmotic potential/solute potential (ψ_S). It matters when the soil is salty.

ii. Gravitational Potential (Ψ_g)

The gravitational potential, (ψ_g), is that portion of the total water potential that is due to the gravitational force field of the earth and is dependent on the vertical location of the water relative to the reference level. When the water is above the reference level, its gravitational potential is positive, because it will tend to flow toward the reference level due to the force of gravity.

iii. Pressure Potential (ψ_p)

In a soil-water system, the pressure is usually the result of overlying water or submergence depth (h) and atmospheric pressure is the reference. Thus, in a soil-water system, the pressure potential will be positive in a saturated soil and zero in an unsaturated soil. In a plant-water system, the pressure potential is the result of the resistance to expansion of the cell walls.

The pressure potential in a plant-water system normally will be positive, but under dry conditions or when the soil-water has a solute potential lower than the solute potential of the plant sap, the plant-water potential may become negative and cause plasmolysis, a separation of the cell membranes from the cell walls. In plants, the pressure potential is sometimes called the turgor pressure (TP).

iv. Matric Potential (ψ_m)

The matric potential (ψ_m) is that portion of the total water potential associated with the more or less solid colloidal matrix of the system. It has been defined in the literature as both a negative pressure and a positive suction head. The matric potential includes the forces of adsorption at the soil-water interfaces and the forces caused by surface tension at the air-water interfaces.

Free water has zero matric potential and will move into a dry soil because of these forces, so the matric potential is negative for an unsaturated soil and zero for a saturated soil. Thus, the removal of water from a soil-water system decreases the matric potential of the water remaining in the system.

v. Osmotic Potential (Ψ_o)/ Solute Potential (Ψ_s)

The solute potential (ψ_S) is that portion of the total water potential associated with the combined effects of all solute species present in the system. It would exist in a soil system between water at the surface and water deeper in the soil when evaporation is occurring at the surface because the concentration of solutes would gradually increase at the surface as only pure water would leave by evaporation.

Generally, however, the solute potential is important only when plants are included in the system because the plant roots act as a semi-permeable membrane separating systems of different solute concentrations. Pure water can move through a semi-permeable membrane much easier than the solutes can, and pure free water tends to move through the membrane into the solution, thus, the solute potential is negative in the solution, and is reduced by the addition of more solutes.

Methods for Measuring Soil-Water Potential

Soil-water potential can be measured in the field with tensiometers, electrical resistance blocks, heat dissipation sensors and psychrometers.

1. Tensiometers: 0 to -80 kPa (-0.8 to ~ -800 cm)
2. Electrical resistance blocks: 0 to -890 or -1,500 kPa (0 to -10 or -15 bars)
3. Thermocouple psychrometers: -98 to -3,000 kPa (-1 to -30 bars)
4. Heat dissipation sensors: -9.8 to -100,000 kPa (-0.1 to -1,000 bars).

The above methods are discussed elaborately under measurement of soil moisture. These instruments measure the energy potential of soil-water either in negative units of pressure or with positive units of tension, which is the opposite of pressure. More energy is required to extract water from soil at lower or more negative water potentials. Bars, atmospheres (atm), pounds per square inch (psi) and kilo Pascals (kPa) are several examples of common pressure units.

Pressure potential in different units (Marshall *et al* 1996 and Hillel 1982)

Soil, Plant and atmospheric conditions	**Bars and atmospheres (bars) and (atm)**	**Pounds per square inch (psi)**	**Kilo Pascals (kpa)**	**Relative energy potential**
Saturated soil	0	0	0	High
Field capacity	-0.33	-5	-33	Medium
Available water	-0.33 to -15	-5 to -225	-33 to -1500	Medium
PWP	At or below -15	At or below -225	At or below -1500	Low
Air dried soil	-31	-465	-3100	Low
Oven dried soil	Below -31	Below -465	Below -3100	Low
Root tissue	-3 to -20	-45 to -300	-300 to -2000	Low
Leaf tissue	-15 to -30	-225 to -450	-1500 to -3000	Low
Atmosphere	-100 to -500	-1500 to -7500	-10,000 to -50,000	Very low

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. 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.

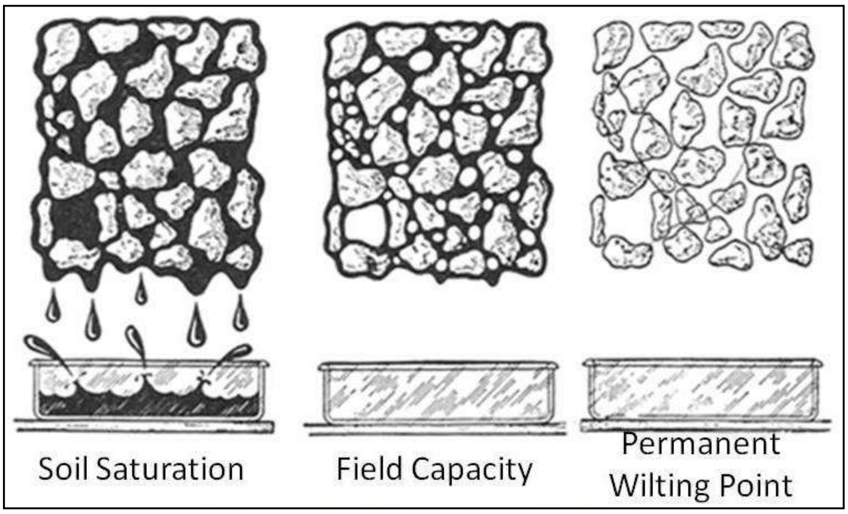
A part of it, at its lower limit approaching the hygroscopic coefficient is not utilized by the plants. Similarly a part of the capillary water at its upper limit is also not available for the use of plants. Hence two more soil constants, *viz.,* field capacity and wilting coefficient have been introduced to express the soil-plant-water relationships as it is found to exist under field conditions.

1. Field capacity

Field Capacity is the amount of soil moisture or water content held in the soil after excess water has drained away and the rate of downward movement has decreased. This usually takes place 2 to 3 days after rain or irrigation in pervious soils of uniform structure and texture.

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 or Permanent wilting point

The level of soil moisture at which water becomes unavailable to plants and permanent wilting ensues.

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 atmosphere.

3. Hygroscopic coefficient

The percentage of water that is absorbed and held in equilibrium by a soil in a saturated atmosphere. 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. Available water capacity is the water held in soil between its field capacity (- 0.3 bar) and 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 1/100th to 1/1000th 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.

Moisture equivalent

It is defined as the percentage of water held by one centimeter thick moist layer of soil subjected to a centrifugal force of 1000 times of gravity for half an hour.

Soil moisture constants and range of tension and pF

S.No	Moisture class	Tension (atm)	pF
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

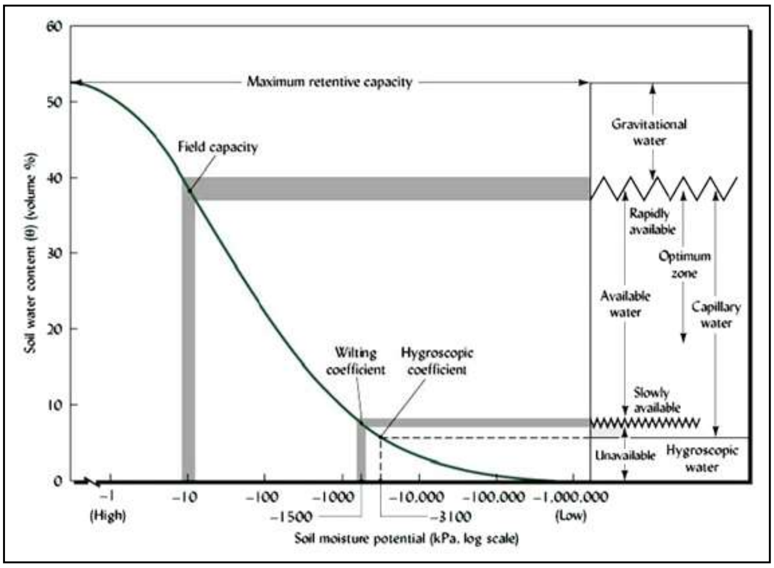
7	Moisture equivalent	1/3 to 1	2.70 to 3.00
8	Field capacity	1/3	2.54
9	Sticky point	1/3 (more or less)	2.54
10	Gravitational	Zero or less than 1/3	<2.54
11	Maximum water holding capacity	Almost zero	----

Soil moisture retention curve

The soil moisture retention curve (pF curve) gives the relation between soil moisture suction and soil moisture content.

A soil is at F.C. (field capacity) or has a pF-value of 2, some 2 to 3 days the soil has been saturated by rainfall or irrigation.When the soil becomes dry and plants cannot take up water anymore the soil is at W.P (wilting point) or has a pF=4.2.

The amount of water held by a soil in the root zone between F.C. and W.P. and which can be used by plants is described as available water. (F.C.- W.P.= available water) For sand, loam and clay the values are 6, 20 and 17 volume percent respectively.

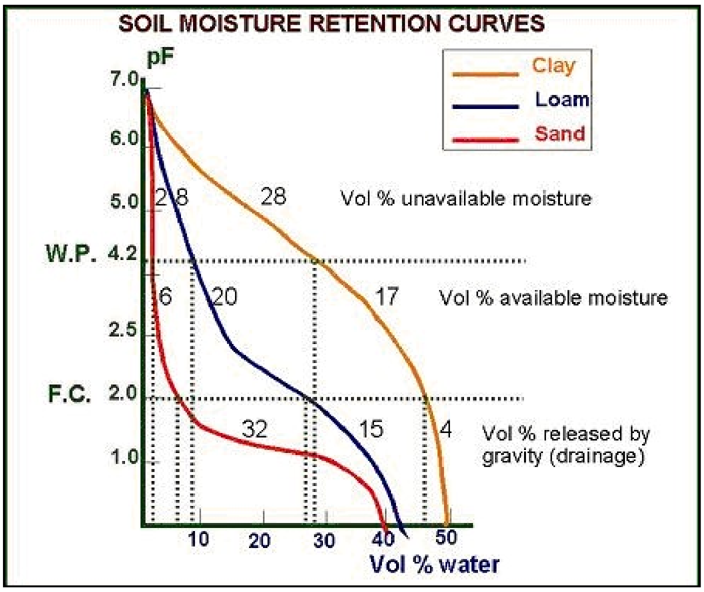


Plant Available Water Concept

Field capacity is the term used to describe the upper limit of plant available water and permanent wilting point is considered the lower limit.

Field capacity is the amount of water held in soil after excess water has drained away and the rate of downward movement has materially decreased, which usually takes place within 2-3 days after a rain or irrigation in pervious soils of uniform structure and texture.

In practice field capacity is often estimated by taking the soil water content at1/3 or 1/10 bar. The permanent wilting point is the root zone soil wetness at which the wilted plant can no longer recover turgidity when placed in a saturated atmosphere for

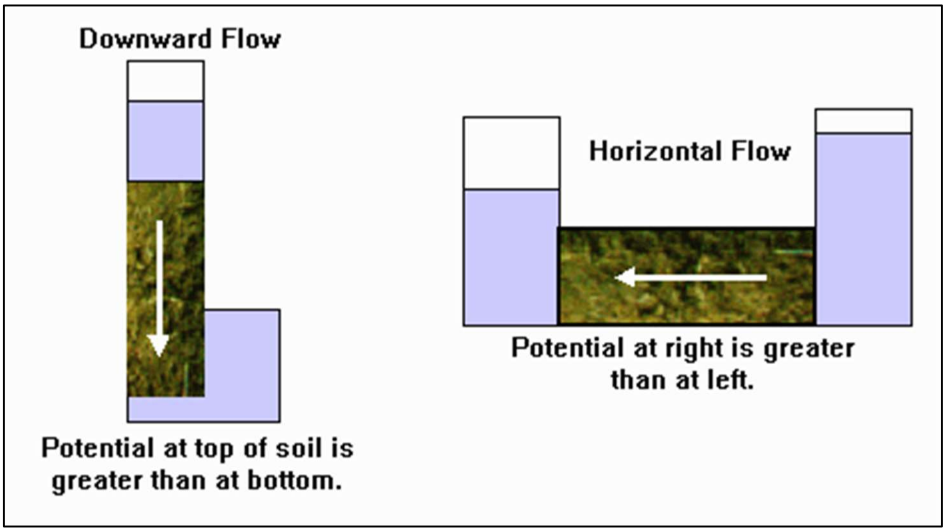


The permanent wilting point is commonly approximated as the soil water content at 15 bar. Plant available water is considered the amount of water held between field capacity and the permanent wilting point.

Lecture 20: Movement of soil water - Saturated and unsaturated flow - infiltration, hydraulic conductivity, percolation, permeability and drainage

Soil acts as a sponge to take up and retain water. Movement of water into soil is called *infiltration,* and the downward movement of water within the soil is called *percolation, permeability*or *hydraulic conductivity*. Pore space in soil is the conduit that allows water to infiltrate and percolate. It also serves as the storage compartment for water.

A unit volume or mass of water tends to move from an area of higher potential energy to one of lower potential energy.



The force of gravity is one factor. Just as water at a higher elevation on a street tends to run down to a lower elevation due to gravity, so water in a soil tends to move downward due gravity. A second factor is the attraction of the soil surfaces for water. When water is added to the bottom of a dry pot of soil, the water moves up into the soil due to this attraction of the soil surfaces for water. The energy level of the water in contact with the soil particles is less than that of the pool of water in the pan so it moves up into the soil. As the soil in the pot becomes wet, this attraction is reduced so that by the time the pores are completely filled, the soil no longer attracts additional water. If a soil is saturated, a third source of potential energy can exist in the form of external pressure such as that provided by a pump or a layer of water in a flooded area. These are the main sources of potential energy in soil water.

Direction of Water Movement

The total potential energy of water is the sum of the potentials from all sources. Potential energy per unit mass or per unit volume or per unit weight is known as the potential of the water. So water free to move will move from a region where it has higher total potential to one of lower total potential. The potential due to gravity is known as the gravitational potential and that due to the soil particles is the matric potential. Soils whose pores are not filled have matric potentials less than zero. Saturated soils under the influence of external hydrostatic pressures have matric potentials which are greater than zero. The total potential at any point is just the sum of the gravitational and matric (or pressure) potentials at that point. The distribution of total potential within a soil allows us to determine if water will move and the direction of movement for any soil system. If the total potentials are equal, no movement will occur.

Rate of Water Movement

The water movement as the product of a driving force causing water to move and a factor representing the ease with which water moves in the soil. This was formalized by Henry Darcy in 1856 as

$$q = -K \frac{dTH}{dx}$$

where q is the volume of water flowing through a unit cross-sectional area of soil per unit time, K is the saturated hydraulic conductivity of the soil, TH is the total hydraulic head and x is the position coordinate in the direction of flow. This equation is known as Darcy's Law. For uniform saturated soils, it is useful to write this equation as

$$q = K \frac{TH_A - TH_B}{L_{AB}}$$

where TH_A is the total head at the inlet end of the soil, TH_B is the total head at the outlet end of the soil column, and L_{AB} is the distance between the inlet and outlet. The hydraulic conductivity, K, represents the ease with which water flows through a soil. Its value depends upon the soil properties and the properties of the soil water. The driving force, df, is represented by

$$df = -\frac{dTH}{dx} \quad \text{or} \quad \frac{TH_A - TH_B}{L_{AB}}$$

Hydraulic Conductivity

Since some of the volume of a soil is occupied by soil particles, water does not flow through a soil as easily as it does through an open pipe. The ease of water movement is known as the hydraulic conductivity of the soil. Soils with large pores such as sandy soils tend to conduct water more easily than soils with smaller pores such as clay soils. Thus sands have higher saturated conductivity values than clays. As soils dry out, water in the largest pores is lost first. Water movement must then occur through smaller pores and in films near soil particles. These pores and films present more resistance so the conductivity of a soil decreases as it dries out. In nature the wetness of the soil often changes with position in the soil. Thus, the conductivity changes with position as well.

Driving Force

The rate of water movement in soils increases with increasing driving force. In saturated soils the driving force is a result in differences in elevation and positive external pressures in the soil. In unsaturated soils, the attraction of the soil surfaces for water is often a major component to the driving force. Driving forces are commonly much larger in unsaturated soils than in saturated soils. The driving force commonly changes in magnitude with changing soil properties and wetness.

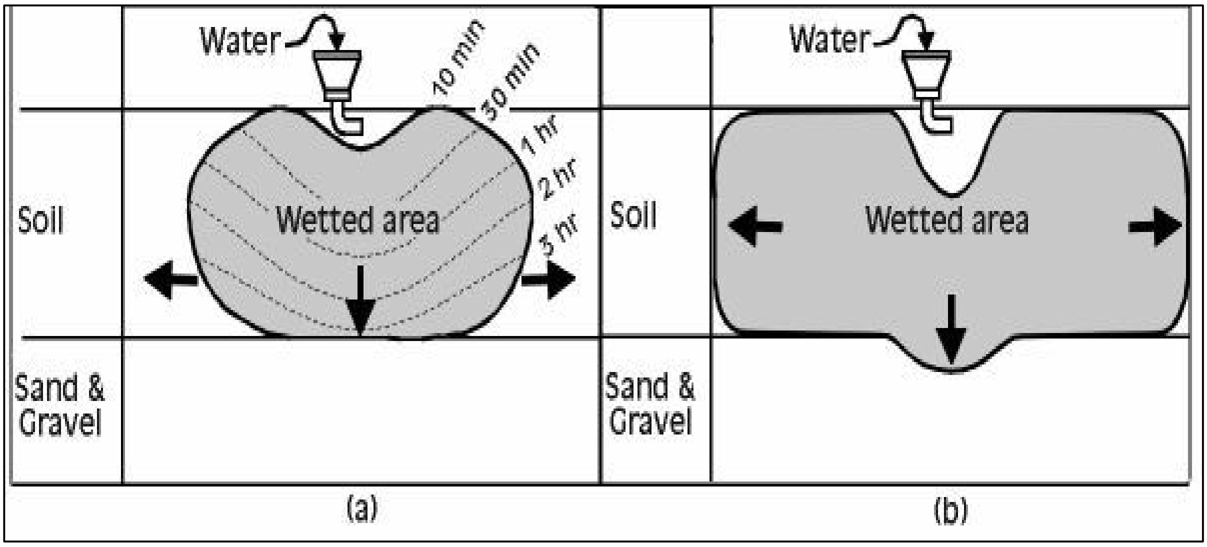
E.g. The driving force causing water to move between two points in the soil is the difference between the total potential at those points divided by the distance between the points. For the soil illustrated below, this becomes

$$df = \frac{TH_A - TH_B}{L_{AB}}$$

where df is the driving force and L_{AB} is the distance between points A and B in the soil column. Note that the driving force is dimensionless when the potential is taken on a unit weight basis or as hydraulic head.

Water movement in coarse sand and gravel

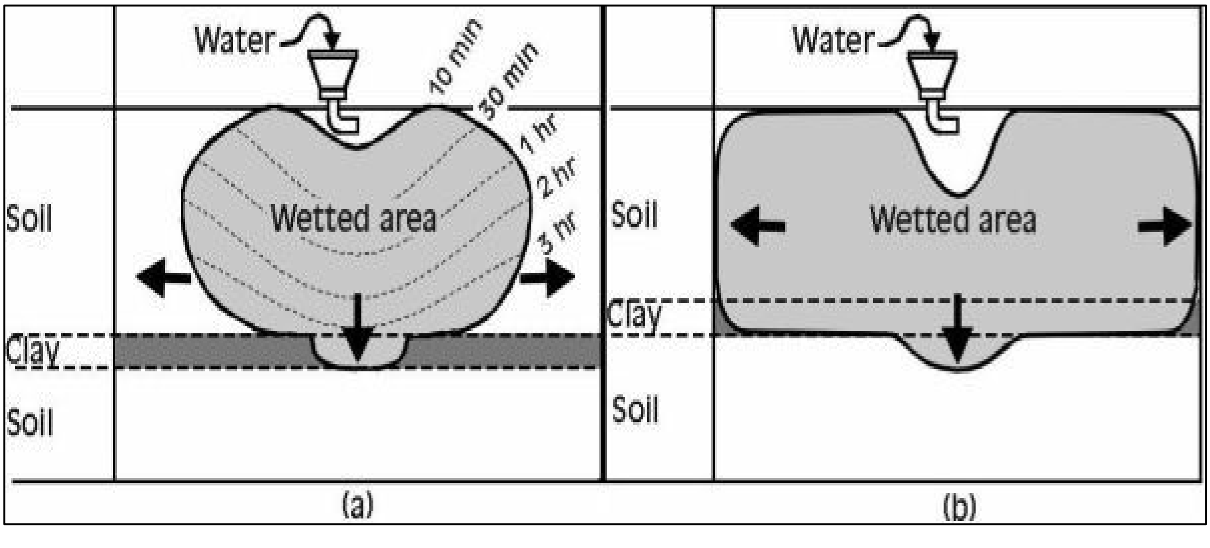
How water moves once in the soil is an important factor in determining the suitability of land for irrigation. Movement or redistribution of water in the soil is dependent upon the size, number and continuity of the soil pores.



Water movement through fine-textured soil into underlying sand and gravel does not occur until the finer material above the gravel is fully saturated. Because the smaller pores, in the finer material in the upper layer, have a greater attraction for the water than the relatively larger pores on the underlying layer, the water moves laterally and fills the upper layer before moving into the coarse material below. Remember, the fine textured soil was able to move water higher in the soil column. After the upper layer becomes saturated, water enters the underlying layer. The practical implication of this demonstration is that in shallow soils underlain by sands, like those found in the Platte River Valley, water movement is actually slowed by the underlying coarse sand and gravel layer.

Water movement in clay

The water moves rapidly into the clay. Compared to the overlying layer, the smaller pores associated with clay layer have a greater attraction on the water. This causes the clay to wet immediately when the wetting front reaches the layer. Although the clay layer wets rapidly, the small pores hold the water tightly and effectively retard the advance of the wetting front. The slowing of the wetting front causes lateral water movement in the overlying coarse soil. Finally, after the clay layer is saturated, the wetting front will move below the clay. The claypan restricts the downward (or upward) movement of water. If a claypan is at or near the soil surface, excessive runoff may become a problem during rainfall or irrigation events even though the soil below the clay pan is dry. A subsurface clay layer also can cause the soil above it to become fully saturated forming a **perched water table**. Perched water tables often cause drainage and aeration problems in the upper soil layers.



Properly managing the soil water is the goal of both dryland and irrigated producers. Familiarity with soil water terms and the processes that control soil water management are critical for proper management.

Entry of Water into Soil

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.

Intakeor **infiltration rate**is the speed at which water can be taken into soil during an irrigation or rainfall event.

Factors affecting infiltration

i. Clay minerals
ii. Soil Texture
iii. Soil structure
iv. Moisture content
v. Vegetative cover
vi. Topography

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).

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.

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 is known as permeability. The term hydraulic conductivity is also used which refers to the readiness with which a soil transmits fluids through it.

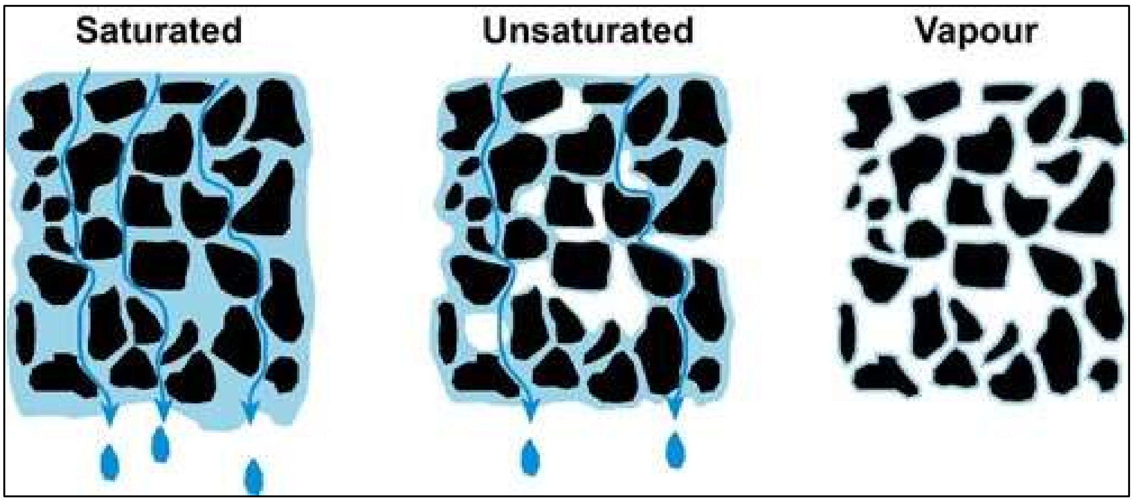
Soil Water Movement

Soil water moved in three ways they are

i) Saturated Flow

ii) Unsaturated Flow

iii) Water Vapour Movement



i. Saturated flow

This occurs when the soil pores are completely filled with water. This water moves at water potentials larger than - 33 kPa. 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 floe of the liquid.

$$QW = - k (dw) At / Ds$$

Where,

QW - Quantity of water in cm^{-1}
K - Rate constant (cm/s)
dw - Water height (head), cm
A - Soil area (cm^2)
t - Time
ds - Soil depth (cm)

(ii) Unsaturated Flow

It is flow of water held with water potentials lower than -1/3 bar. Water will move toward the region of lower potential (towards the greater "pulling" force). In a uniform soil this means that water moves from wetter to drier areas. The water movement may be in any direction.

The rate of flow is greater as the water potential gradient (the difference in potential between wet and dry) increases and as the size of water filled pores also increases. The two forces responsible for this movement are the attraction of soil solids for water (adhesion) and capillarity. Under field conditions this movement occurs when the soil macropores (noncapillary) pores with filled with air and the micropores (capillary) pores with water and partly with air.

Factors Affecting the Unsaturated Flow

Unsaturated flow is also affected in a similar way to that of saturated flow. Amount of moisture in the soil affects the unsaturated flow. The higher the percentage of water in the moist soil, the greater is the suction gradient and the more rapid is the delivery.

(iii) Water Vapour Movement

The movement of water vapour from soils takes place in two ways 1. Internal movement, 2. External movement.

1. Internal movement: The change from the liquid to the vapour state takes place within the soil, that is, in the soil pores.

2. External movement: The phenomenon occurs at the land surface and the resulting vapour is lost to the atmosphere by diffusion and convection.

The movement of water vapour through the diffusion mechanism taken place from one area to other soil area depending on the vapour pressure gradient (moving force). This gradient is simply the difference in vapour pressure of two points a unit distance apart. The greater this difference, the more rapid the diffusion and the greater is the transfer of water vapour during a unit period.

Soil conditions affecting water vapour movement:

There are mainly two soil conditions that affect the water vapour movement namely moisture regimes and thermal regimes. In addition to these, the various other factors which influence the moisture and thermal regimes of the soil like organic matter, vegetative cover, soil colour etc. also affect the movement of water vapour.

The movement takes place from moist soil having high vapour pressure to a dry soil (low vapour pressure). Similarly the movement takes place from warmer soil regions to cooler soil region. In dry soils some water movement takes place in the vapour form and such vapour movement has some practical implications in supplying water to drought resistant plants.

Drainage

The frequency and duration of periods when the soil is free from saturation with water. It controls the soil cum water relationship and the supply of nutrients to the plants.

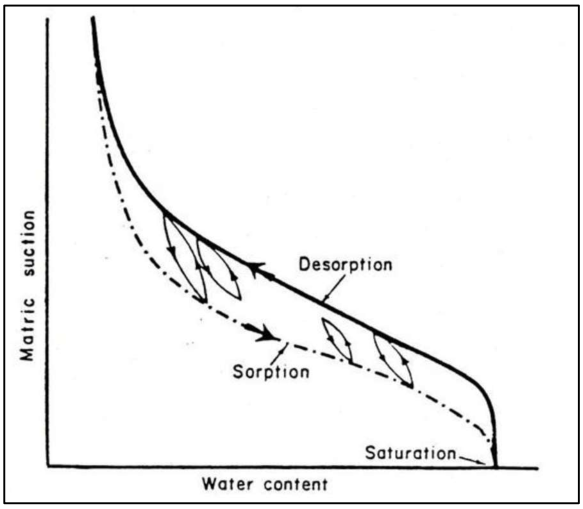
Drainage class

1. Very poorly drained
2. Poorly drained
3. Imperfect
4. Moderately well
5. Well
6. Somewhat excessive
7. Excessive

Hysterisis

The moisture content at different tensions during wetting of soil varies from the moisture content at same tensions during drying. This effect is called as hysterisis. This is due to the presence of capillary and non capillary pores. The moisture content is always low during sorption and high during desorption.

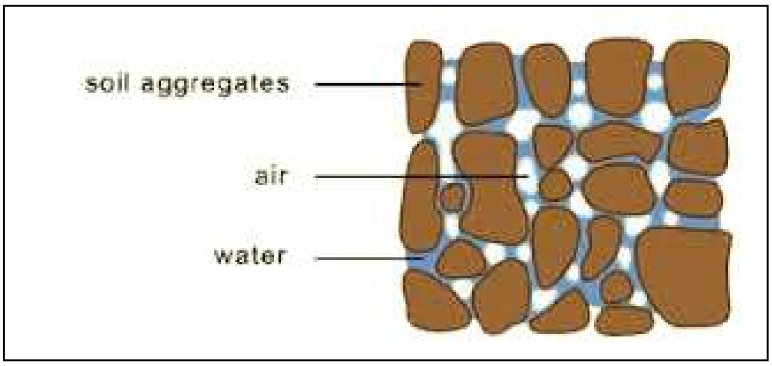
Hysterisis phenomenon exists in soil minerals as a consequence of shrinking and swelling. Shrinking and swelling affect pore size on a micro basis as well as on the basis of overall bulk density. So, hysterisis phenomenon occurs due to factors like shape and size of soil pores and their interconnection with each other pore configuration, nature of soil colloids bulk density of soil and entrapped air. The most important factor affecting hysterisis is the entrapment of air in the soil under rewetting condition. This clogs some pores and prevent effective contact between others.



Lecture 21: Soil air, composition, gaseous exchange – Problem and its effect on crop growth

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.

Soil aeration is the exchange of carbondioxide and oxygen gases between the soil por space and aerial atmosphere. There are two types of pores involved in the soil aeration-pores between the crumbs called inter crumb pores and pores within the crumbs called crumb pores.



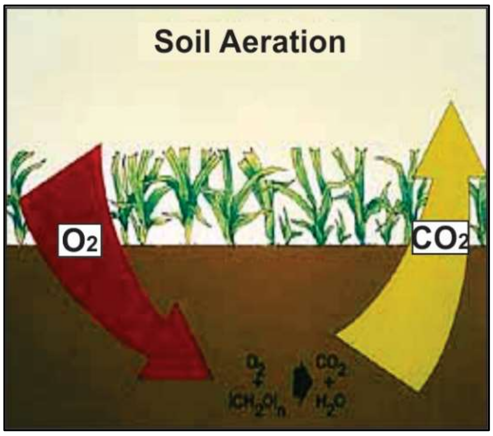
Reason for poor aeration

There are generally two reasons by which poor aeration results.

(1) Excess moisture

When a soil is subjected to excess moisture, waterlogged condition is developed. This situation is generally foundpoorly drained, fine-textured soils having a minimum of macro pores through which water can move very rapidly.

(II) Gaseous interchange



The inadequate interchanges of gases between the soil and free atmosphere depends on two factors

(a) The rate of biochemical reactions influencing the soil gases.

(b) The actual rate at which each gas is moving into or out of the soil.

The exchange of gases between the soil and the atmosphere is facilitated by two mechanisms: (a) Mass flow and (b) Diffusion

(a) Mass flow

Mass flow of air is apparently due to pressure differences between the atmosphere and the soil air. Temperature may influence the renewal of soil air by two ways.

(1) There may be temperature variations within the soil between the different horizons.

(2) The soil and the atmosphere usually have different temperatures. This differential temperature also permits an exchange between the atmosphere and soil air in the immediate surface.

(b) Diffusion

Diffusion is the molecular transfer of gases. The molecules of gases are in a state of movement in all directions. Through this process each gas tends to move in a direction determined by its own partial pressure.

Composition of soil air

The gaseous constituents of soil atmosphere are as important for crop growth as nutrients and water. Aerobic respiration in roots, micro organisms and soil fauna involve the continuous consumption of O_2 and evaluation of CO_2.

The soil air contains a variety of gases like O_2, N_2, Ar, CO_2 and water vapour etc. The composition of soil air, though quite variable is similar to that of the atmospheric air, but differs in one respect.

Composition of soil and atmospheric air

Name of gas	**Percentage by volume**	
	Soil air	**Atmospheric air**
Oxygen	20.00	21.00
Nitrogen	78.60	78.03
Carbondioxide	0.50	0.03
Argon	0.90	0.94

Factors Affect Composition of soil air

The Composition of soil air is affected by a variety of factors like physical properties of soil, Soil condition, type of vegetation, seasons, amount of organic matter, microbial activity, depth of the soil and temperature etc.

Oxygen

The amount of O_2 in soil air is much less than that of the atmospheric air. Plant roots and different micro-organisms require O_2 for their metabolism which they take from the soil air and thereby decreases the concentration of O_2 in the soil environment.

The depth of soil also modify the amount of O_2 in the soil and with the increase in depth the amount O_2 is less. This may be due to more slow diffusion of O_2 from the sub-soil horizon to the atmosphere through overlying soil layers.

The seasonal variation also affects the composition of soil air. In the dry season or summer season the quantity of O_2 is usually higher than that of the rainy season. Because in the summer season the opportunity for the gaseous exchange is greater as compared to monsoon season resulting higher O_2 and low CO_2 concentration in the former season.

Cultural and other soil management practices affect the composition of soil air by modifying various physical properties of soil like bulk density, porosity, soil structure etc. Light texture soil contains much more O_2 than the heavy texture soil. As a rule, the higher the bulk density the more compact the soil, the more poorly defined structure and the smaller, the amount of pore spaces and restrict plant growth.

Carbon dioxide

During decomposition of organic matter CO_2 is evolved and therefore the concentration of CO_2 will be very high in soils high in organic matter. The production of CO_2 generally will be very high through increased activity of soil micro organisms during summer season. The concentration of CO_2 is generally higher in the sub-soil due to slow rate of exchange of gases in the lower horizon as compared to the surface soil.

Water vapour

The content of water vapour is much more in the soil air as compared to the atmospheric air. The capillary water in the soil saturate the soil air with the water vapour. If the soil moisture falls below hygroscopic coefficient, the water vapour also decreases.

The aeration requirements of crop plants vary considerably. The most crop plants can grow well in a soil atmosphere with less than 21% of O_2 present in the free air, whereas in the soli containing O_2 content below 10%, the growth of most of the crops is restricted.

Characterization of soil aeration

There are various parameters that can be used for characterizing soil aeration.

1. The volume percentage of soil air or air capacity is that part of the pore space which is filled with air. This is generally determined by applying the tension equivalent to a water column of 50 cm to a saturated soil on a tension table.
2. Gaseous composition.
3. **The oxygen diffusion rate (ODR):** It is determined by using the platinum microelectrode technique where the diffusing oxygen is allowed to reduce at the platinum electrode at a given electric potential. The rate of diffusion of oxygen to the

platinum electrode is used as an index of the rate of diffusion of oxygen through the water film to the roots.

4. Oxidation- reduction potential, indicating the oxidized or reduced condition of the soil.
5. Composition of the soil for its reduced components.

Importance of soil air in plant growth and biological activities in soil

The importance of soil aeration in relation to soil fertility and plant responses are generally divided as follows

(a) Growth of plants and the development of roots

The growth of plants are adversely affected by poor aeration namely (i) The development of plant roots are restricted or inhibited, (ii) The absorption of water and nutrients is decreased, (iii) The formation of toxic substances are encouraged by poor aeration, especially under puddled waterlogged condition in case of rice cultivation.

(b) Adaptations of plant

The ability of different plant species to grow in soils with low soil air porosity varies greatly. Certain plants such as rice roots are adapted to growth in aerated soils have been found to carry out some anaerobic respiration if oxygen is deficient, the effects are detrimental. The relatively low rate and energy output of anaerobic respiration and the accumulation of toxic products are perhaps the most important reasons.

(c) Microbial population and activity

Soil aeration influence the activity of soil micro-organisms and also affects the rate of decomposition of organic matter. The poor soil aeration decreases the microbial activity as well as oxidation of organic matter. This decrease may be due to lack of O_2 than with an excess of CO_2.

(d) Production of toxic substances

Heavy rainfall, excessive irrigation or puddling together with flooding brings about a reduced condition of the soil and subsequently reduction of elements like, Fe and Mn and slow rate or incomplete of decomposition of organic matter will take place in the soil. Thus production of various organic toxic acids like, lactic, butyric and citric acids etc. Due to production of such toxic substances plant roots get injury and thereby lost their ability to absorb nutrients and water for their growth.

(e) Absorption of water and nutrients

Effects of aeration on the uptake of nutrients by plants may result from changes in nutrient availability that occur in the soil in response to the aeration or they may result from changes in metabolic statues of the plants. The micronutrients like Cu and Zn show deficiency under poor aeration (waterlogged) conditions. Poor aeration decreases the uptake of water.

(f) Incidence of diseases

Changes in aeration may modify the susceptibility of the plant, the virulence of the organism or both, so that disease incidence may vary with conditions of soil aeration. The effect of aeration has to do mainly with the sensitivity of the disease organism and not the plant.

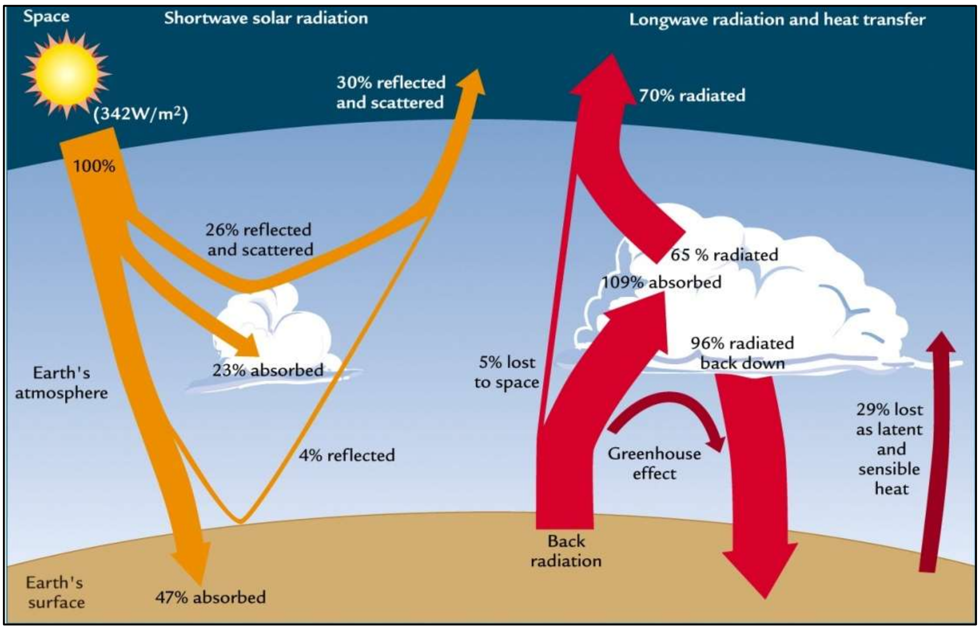
Lecture 22: Source, amount and flow of heat in soil, soil temperature and crop growth

Soil Temperature

Soil temperature is simply the measurement of the warmth in the soil. Ideal soil temperatures for planting most plants are 65 to 75 F. (18 to 24^{O}C).

Heat is a form of energy and temperature is a measure of the heat energy. The heat energy refers to kinetic energy or random motion (vibration) of molecules of a substance. The speed of vibration is directly proportional to temperature. Soil temperature is an important plant growth factor like air, water and nutrients. Soil temperature affects plant growth directly and also indirectly by influencing moisture, aeration, structure, microbial and enzyme activities, rate of organic matter decomposition, nutrient availability and other soil chemical reactions.

Specific crops are adapted to specific soil temperatures. Apple grows well when the soil temperature is about 18^{O}C, maize 25^{O}C, potato 16 to 21^{O}C. Major part of this energy is absorbed in the atmosphere, absorbed by plants and also scattered. Only a small part of it reaches soil. Thermal energy is transmitted in the form of thermal infrared radiation from the sun across the space and through the atmosphere.



Sources of soil heat

There are various sources of soil heat namely, solar radiation, biochemical reactions, conduction, precipitation, exposure and vegetation etc.

i. Solar radiation

Radiant energy from the sun is the power source that determines the thermal regime of the soil and the growth of plants. Agriculture is the exploitation of solar energy in the presence of an adequate water supply and sufficient plant nutrients to maintain plant growth.

Much of the radiation from the sun is dissipated and distributed by components of the earth's atmosphere as it passes downward toward the earth. The heat absorbed by the surface of the earth from the solar radiation is affected by several parameters like latitude, distribution of land and water, slope of the land etc.

ii. Bio-chemical reactions

In the soil atmosphere a variety of chemical reactions are going on and during such reaction liberation of large amount of heat in the soil environment results. Besides decomposition of organic matter and other crop residues in the soil and other microbial processes liberate large amounts of heat in the soil and thus contributes soil heat.

iii. Conduction

Conduction is a process in which heat is transmitted from the hotter to the colder part of a body or from a hotter body to a colder body in contact with each other, by internal molecular motion without any transference of material particles.

The inner atmosphere of the earth is very hot, the conduction of the heat to the soil environment is very slow. Generally, during night, the upper most surface soil becomes cooler than sub-surface soil. Thus heat flows from the regions of sub-soil to the region of surface soil (cooler soil layer).

(iv) Convection

It is a process in which heat is transferred from one point to another by actual movement of the heated material particles from a place of higher temperature to another place of lower temperature. Liquids and gasses are generally heated by the process of convection.

(v) Radiation

Radiation is the process in which heat can transmit from a hotter body to a colder body without intervention of any material medium. The radiant heat travels in straight lines with the speed of light.

(vi) Precipitation

During the winter season precipitation increases soil heat because of its (precipitation) higher specific heat.

(vii) Exposure

Exposure is of little importance in the tropics because of the high elevation of the sun. It is of significance in the middle latitude where the elevation is lower. The greater the percentage of diffuse sky radiation in the global radiation, the smaller is the difference in the incoming solar energy per unit area for slopes of different exposures.

(viii) Vegetation

Vegetation plays a significant role of soil heat because of the insulating properties of plant cover. Bare soil is unprotected from the direct rays of the sun and becomes % try warm during the hottest part of the day.

Method of measuring temperature

Soil thermometers are the most common tool for measuring soil temperature.

Factors affecting soil temperature

The average annual soil temperature is about 1°C higher than mean annual air temperature. Soil temperature is influenced by climatic conditions. The factors that affect the transfer of heat through the atmosphere from sun affect the soil temperature also.

Environmental factors

Solar radiation

The amount of heat received from sun on Earth's surface is cal cm^{-2} min^{-1}. But the amount of heat transmitted into soil is much lower. The heat transmission into soil depends on the angle on incident radiation, latitude, season, time of the day, steepness and direction of slope and altitude. The insulation by air, water vapour, clouds, dust, smog, snow, plant cover, mulch etc., reduces the amount of heat transferred into soil.

Soil factors

Thermal (Heat) capacity of soil

The amount of energy required to raise the temperature by $1^{O}C$ is called **heat capacity.** When it is expressed per unit mass (Calories per gram), then it is called as **specific heat.** The specific heat of water is 1.00 cal g^{-1} where the specific heat of a dry soil is 0.2 cal g^{-1}.Increasing water content in soil increases the specific heat of the soil and hence a dry soil heats up quickly than a moist soil.

Heat of vaporization

The evaporation of water from soil requires a large amount of energy, 540 kilo calories kg^{-1} soil. Soil water utilizes the energy from solar radiation to evaporate and thereby rendering it unavailable for heating up of soil. Also the thermal energy from soil is utilized for the evaporation of water, thereby reducing the soil temperature.

This is the reason that surface soil temperatures will be sometimes 1 to 6°C lower than the sub-surface soil temperature. That is why the specific heat of a wet soil is higher than dry soil.

Thermal conductivity and diffusivity

This refers to the movement of heat in soils. In soil, heat is transmitted through conduction. Heat passes from soil to water about 150 times faster than soil to air. So the movement of heat will be more in wet soil than in dry soil where the pores will be occupied with air. Thermal conductivity of soil forming materials is 0.005 thermal conductivity units, and that of air is 0.00005 units, water 0.001 units. A dry and loosely packed soil will conduct heat slower than a compact soil and wet soil.

Biological activity

Respiration by soil animals, microbes and plant roots evolve heat. More the biological activity more will be the soil temperature.

Radiation from soil

Radiation from high temperature bodies (Sun) is in short waves (0.3 to 2.2 μ) and that from low temperature bodies (soil) is in long waves (6.8 to 100 μ). Longer wavelengths have little ability to penetrate water vapour, air and glass and hence soil remains warm during night hours, cloudy days and in glass houses.

Soil colour

Colour is produced due to reflection of radiation of specific wavelengths. Dark coloured soils radiate less heat than bright coloured soils. The ratio between the incoming (incident energy) and outgoing (reflected energy) radiation is called **albedo**. The larger the albido, the cooler is the soil. Rough surfaced soil absorbs more solar radiation than smooth surface soils.

$$\text{Albido} = \frac{\text{Reflected energy}}{\text{Incident energy}}$$

Soil structure, texture and moisture

Compact soils have higher thermal conductivity than loose soils. Natural structures have high conductivity than disturbed soil structures. Mineral soils have higher conductivity than organic soils. Moist soil will have uniform temperature over depth because of its good conductivity than dry soils.

Soluble salts

Indirectly affects soil temperature by influencing the biological activities, evaporation etc.

Surface roughness and stoniness

A soil with a rough surface will absorb more solar radiation and usually be warmer than one with a smooth surface. Similarly stony soils and rock gardens will absorb more heat per unit area and normally be warmer than non stony soils or non rocky surfaces.

Controlling Soil Temperature

Plants grow best in a certain range of temperatures. Most plants stop growing at soil temperatures under 50 degrees. Temperature also affects the speed of chemical reactions. Higher temperatures provoke more rapid soil reactions, up to a point. For example, some organic fertilizers release nutrients only at relatively warm temperatures. On the other hand, the availability of chemical fertilizers is only slightly reduced at low soil temperatures. Also, the beneficial effects of soil organisms occur only in warm soil.

Because soil temperature depends on sunlight, it fluctuates according to season and climate. Other factors that influence soil temperature are described below, with those easiest for the gardener to manipulate being listed first.

Mulch

Mulch acts as an insulation. It keeps heat from being absorbed during the day and released at night. It keeps the soil cool in the summer and warm in the winter. The surface of the soil can reach 120 degrees under hot summer sun, much too hot for plant roots to grow. A mulch cools the soil, allowing roots to grow right to the surface.

In the winter, mulch delays the soil's freezing and thawing. This is important in preventing heaving, the loosening of soil that results from alternate freezing and thawing. To allow soil to warm early in spring, remove the mulch when the weather warms.

Compaction

Loose soil contains air spaces that insulate the soil, acting a little like a much. Compact soil heats faster in the sun and cools faster at night than cultivated soil.

Moisture

Water in the soil conducts heat rapidly. Water the soil in the spring to help the day's heat be conducted to the depths.

Shade

Since the sun is the source of most soil heat, shading the soil keeps it cooler during the day. The soil can shade with ground covers or other plants, or with structures.

Exposure

The direction the soil faces influences the soil temperature. North facing slopes are coolest. East, west, and south-facing slopes follow in order of coolness.

The soil can also be used as a source of heat to keep plants from freezing in light frosts. Damp, compacted soil with a clean surface radiates the most heat to plants above it. When protecting plants from early frosts, cover them with fabric or cardboard and clean the soil under them of mulch and plants. This allows heat to be stored in the soil during the day and radiated to the plants and cover at night.

Effect of soil temperature on plant growth

1. In affects plant growth directly, that is all crops practically slow down their growth below the soil temperature of about 9 °C and above the soil temperature of above 50 °C.
2. For germination of different seeds requires different ranges of soil temperature E.g. Maize begins to germinate at soil temp of 7 to 100C.
3. Most of the soil organisms function best at an optimum soil temperature of 25 to 35 °C
4. The optimum soil temperature for nitrification is about 32 °C.
5. It also influences soil moisture content, aeration and availability of plant nutrients.

Lecture 23: Soil reaction (pH) - definition, pH scale, soil acidity and alkalinity, buffering, effect of pH on nutrient availability and factors affecting soil pH

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.

Kinds of soil acidity

i) Active acidity

The acidity develops due to hydrogen (H^+) and aluminum $(Al)^{3+}$ ions concentration of the soil solution.

ii) Exchange acidity

The acidity develops due to adsorbed hydrogen (H^+) and aluminum $(Al)^{3+}$ irons on the soil colloids.

iii) Total acidity

The summation of active, exchange and residual acidity.

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:

$$Na_2CO_3 \longrightarrow 2Na^+ + CO_3^-$$

$$2Na^+ + CO_3^- + 2HOH \longrightarrow 2Na^+ + 2OH^- + H_2CO_3$$

Since sodium hydroxide dissociates to a greater degree than the carbonic acid, OH ions dominate and give rise to alkalinity. This may be pH 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.

The soil reaction describes the degree of acidity or alkalinity of a soil. Water molecule is very stable and dissociates as

$$H_2O \longrightarrow H^+ + OH^- .$$

The ionization constant of water $Kw = [H^+]\,[OH^-]$ is called ionic product of water. The value of which is 10-14 at 22^0C.

$$\text{Log } 1/ Kw = \log 1/H^+ + \log 1/OH^{-1} = 14$$

Thus when the hydrogen ion concentration of an aqueous solution increases, its hydroxyl ion concentration decreased to maintain the ionic product of water.

pH and its method of Expression

pH

Sorenson (1909) suggested the term pH (pH puissance dehydrogen or pourvoir hydrogen), which means the power of hydrogen. pH is the negative logarithm of hydrogen ion activity.

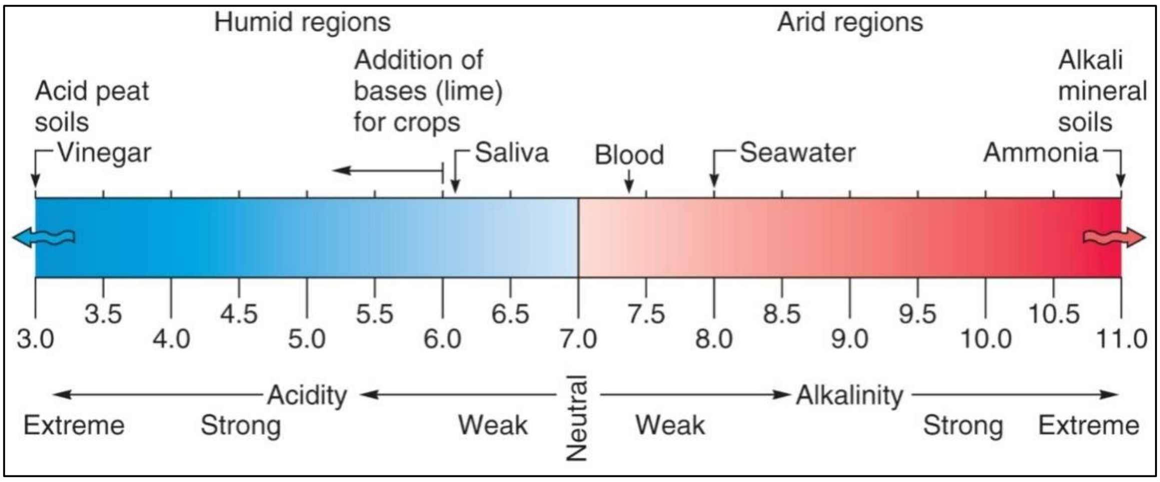
$$pH = - \log 10\ (H^+)$$

pH 6.0 is ten times more acidic than pH 7.0

pH scale

The pH value represents the amount of free or active acidity and not the total acidity. That means it represents the intensity of acidity of a soil solution.

In this scale, the pH value ranges from 0 to 14 where pH value of 0 represents the highest limit of active acidity; pH ‘7’ represents neutrality and pH ‘14’ represents the highest degree of alkalinity or basicity.



Relationship between H^+ and OH^-

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

Classification of soils based on pH

Based on the pH value of soil solution, the soils have been classified into the following categories.

pH Range	Category (Rating)
<4.5	Extremely acidic
4.5 – 5.0	Very strongly acidic
5.1 – 5.5	Strongly acidic
5.6 – 6.0	Medium acidic
6.1 – 6.5	Slightly acidic
6.6 – 7.5	Neutral
7.5 – 7.8	Mildly alkaline
7.9 – 8.4	Moderately alkaline
8.5 – 9.0	Strongly alkaline
>9.1	Very strongly alkaline

The factors affecting soil pH are

1. Nature of soil colloids

When the soil colloids have H^+ as adsorbed ion on the exchange complex, the soil reactions becomes acidic.

2. Percentage base saturation (PBS) and kind of adsorbed base

Soils having low PBS show acidic reaction.Sodium saturated soils have much higher pH values than Ca and Mg saturated soils.

3. Soil solution

Under field conditions, the concentration of salts varies with the moisture content of the soil. More dilute the solution of a base unsaturated soil; more is the number of hydrogen ions that dissociate into the soil solution. The pH lowers with increase in concentration of CO_2 of soil solution due to the formation of carbonic acid.

4. Climate

In general, soils formed in the areas of high rainfall are acidic, while those formed in areas of low rainfall are alkaline.

5. Soil management

As a result of constant cultivation basic cations are lost from the soil through crop removal and leaching making the soil acidic. Acidic fertilizers like ammonium sulphate make the soil acidic, while basic fertilizers like $NaNO_3$ make the soil alkaline.

6. Oxidation

Reduction state of soilunder anaerobic or reduced conditions as in water logged soils the ferric and manganic ions are reduced to ferrous (Fe^{2+}) and manganous (Mn^{2+}) states, thus increasing pH, under such conditions sulphides are converted to SO_4^- thus lowering the pH upon formation of acid.

Significance pH

Influence of soil reaction on the availability of plant nutrients

Soil reaction is the important factor which governs the availability of various nutrients by influencing the soil properties like physical, chemical and biological etc.

Soil reaction and microbial activity

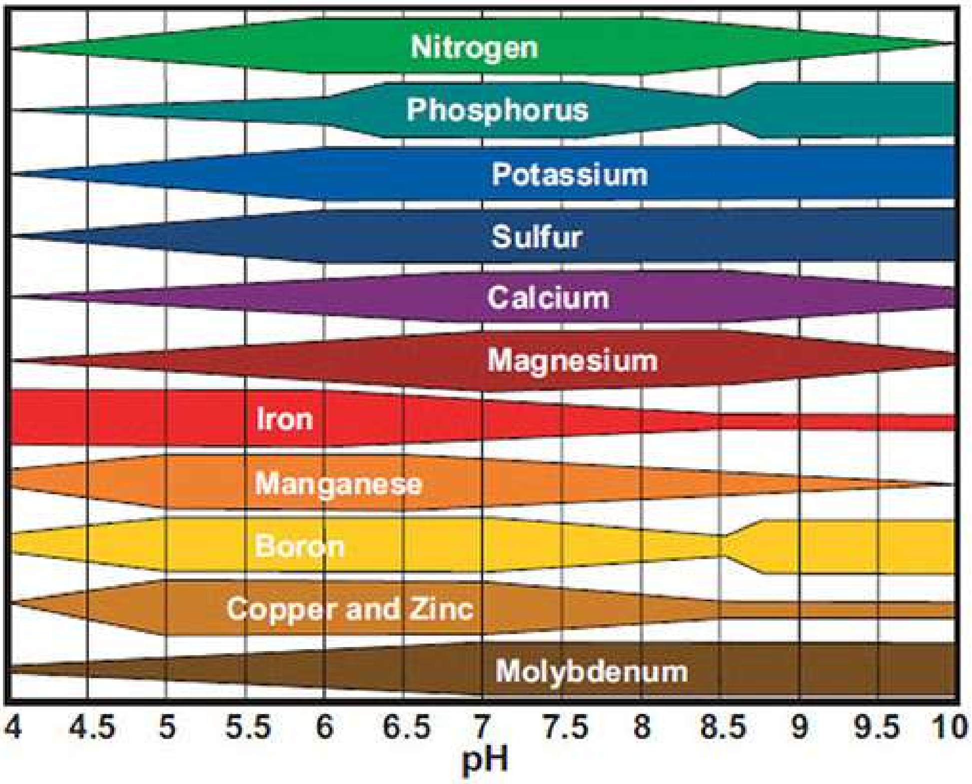
The activity of microbes is influenced by the variations in soil pH. Bacteria and actinomycetes prefer near neutral to slightly alkaline reaction (pH 6.5 – 8.0). Fungi work satisfactorily at all pH ranges. They face a large competition at higher pH values with bacteria and actinomycetes. Hence they grow better in acidic reaction of pH between 4.5 to 6.5.

Nitrogen

Plants absorb nitrogen in the form of NO_3^- whose formation depends on the ability of nitrifying bacteria. The microbes responsible for nitrification are active when the soil pH is between 6.5 to 7.5. Nitrogen fixing bacteria also fail to function below a soil pH of 6.0.

Phosphorus

Phosphorus availability is high when soil pH is between 6.0 to 7.5. At pH values less than 5.0, soluble phosphates are fixed as complex and insoluble compounds of Fe, Al and Mn. At pH values of more than 7.5, complex insoluble calcium phosphates like apatites. Excess calcium also hinders the phosphorus absorption and utilisation by the plants.



Calcium and magnesium

Calcium and magnesium and other basic cationslike potassium become deficient due to their leaching.

Sulphur

The bacteria responsible for the oxidation of sulphides to sulphates can function satisfactorily at all pH values.

Micronutrients

The availability of micronutrients like zinc, iron, copper and manganese are more in the acidic range. Under acidic conditions as in the humid regions because of high rainfall due to leaching of bases, aluminium, iron and micro nutrients become toxic. They are more soluble at low pH. At a soil pH of less than 5.5, aluminium starts to leave the structure of silicate clays. High levels of soluble aluminium are injurious to crops. Aluminium toxicity increases water stress during dry period.

Measuring Soil pH

Soil pH provides various clues about soil properties and is easily determined. The most accurate method of determining soil pH is by a pH meter. A second method which is simple and easy but less accurate then using a pH meter, consists of using certain indicators or dyes.

Many dyes change color with an increase or decrease of pH making it possible to estimate soil pH. In making a pH determination on soil, the sample is saturated with the dye for a few minutes and the color observed. This method is accurate enough for most purposes. Kits (pH) containing the necessary chemicals and color charts are available from garden stores.

Buffering capacity of soils

Buffering refers to the resistance to a change in pH. Buffering in soils is their resistance against any change in the concentration of hydrogen or other cations in the soil solution. This power to resist a change in pH is due to the Buffering action.

In soils, the clay and humus act as buffer systems. Soil exchange complex has reserve and active acidity and there exists equilibrium between these two. If the active acidity is neutralized by the addition of lime, the reserve acidity will release exchangeable H^+ into the soil solution to maintain the equilibrium and resists any change in soil reaction.

The buffering capacity is greater in clay soils than in sandy soils. The colloidal complex behaves in a way similar to a mixture of weak acid and their salts thus acting as a buffer. In soils, the buffering action may be due to one or several of the following buffering agents carbonates, bicarbonates and phosphates present in the soil. Colloidal complex with associated cations. Organic acids being released continuously in the soil act as weak acids.

H^+ released from the soil exchange complex provides the reserve acidity.Soils are a mixture of acidoids and basoids and hence act as amphoteric. The basis of buffering capacity lies in the adsorbed cations (both acidic and basic). There is equilibrium between the adsorbed hydrogen and aluminium ions and H^+ in the soil solution.

This can be represented as follows.

$$\underset{\text{(Reserve acidity)}}{\text{Adsorbed } H^+ \text{ and } Al^{3+}} \longleftrightarrow \underset{\text{(Active acidity)}}{\text{Soil solution containing } H^+ \text{ and } Al^{3+} \text{ ions}}$$

Factors affecting buffering capacity of soils

Buffering capacity of the soil depends on the factors like amount and kind of clay, organic matter content, cation exchange capacity, carbonates, bicarbonates, phosphate content and organic acids.

1. Amount of soil colloids: Soils containing large amounts of clay and organicmatter are said to be highly buffered and require large amounts of lime for affecting a certain change in pH than an acid soil containing smalleramounts of clay and organic matter.
2. Cation exchange capacity: Buffering capacity of the soil increased with increase in cation exchange capacity.
3. Soils having higher amounts of calcium carbonate, organic acids andphosphates show higher buffering capacity.
4. Soils containing large amounts of 1:1 type clays are generally less stronglybuffered than soils in which the predominant clay minerals are 2:1 type.
5. The degree of buffering is highest between the soil pH of 4.5 and 6.0 anddrops off below and above this range.

Importance of buffering in agriculture

The importance buffering in soils is mainly two folds

i) The stabilization of soil pH

ii) Calculations of amount of amendment necessary to affect a certainchange in soil reaction.

An abrupt change in pH causes a radical modification in soil environment and if it fluctuates too widely, the plants and microorganisms would suffer seriously. A change in soil reaction not only has a direct influence on the plants but also exerts an indirect influence on soil environment by creating sudden changes in the availability of nutrients.

Deficiencies of certain plant nutrients and excess of others in toxic amounts would seriously upset the nutritional balance in the soil. The amounts of amendments necessary to affect a given alteration in soil reaction also relate to the effectiveness of pH stabilization. The greater the buffering capacity of the soil, the larger must be the amounts of amendments to affect a given change in pH.

Lecture 24: Soil Electrical Conductivity - Factors affecting EC – its significance

Soil Electrical Conductivity

Soil electrical conductivity, which is known as EC, is the ability of soil to conduct electrical current. EC is expressed in milli Siemens per meter (mSm^{-1}) or deci Seimen per meter (dSm^{-1}). EC is used to measure the concentration of soluble salts present in the soil solution. Soil salinity is a measure of the total amount of soluble salts in soil. As salinity levels increase, plants extract water with great difficulty from soil, aggravating water stress condition and sometimes and sometimes plant starts drying up due to reverse osmosis.

The electrical conductivity of soils varies depending on the amount of moisture held by soil particles. Sands have a low conductivity, silts have a medium conductivity, and clays have a high conductivity. Consequently, EC correlates strongly to soil particle size and texture.

EC related other soil properties

Water-holding capacity/drainage

Droughty areas typically have distinct textural differences from those with excess water; these can be identified using EC. Soils in the middle range of conductivity, which are both medium-textured and have medium water-holding capacity, may be the most productive.

Cation exchange capacity (CEC)

CEC is related to percent of clay and organic matter. As the percent of clay and organic matter increase, the CEC also increases.

Depth to claypan or rock outcropping

The response of conductivity to the presence of clay has been used to accurately predict the depth of topsoil over a clay layer or rock outcropping.

Porosity

The greater the total soil porosity, the more easily it conducts electricity. Soil with a high clay content has more total pore space than sandier soils when other soil parameters remain constant.

Salinity

An excess of dissolved salts in the soil is readily detected by electrical conductivity.

Temperature

As temperature decreases to the freezing point of water, soil EC decreases slightly. Below freezing, soil pores become increasingly insulated from each other, and overall soil EC declines rapidly.

Factors affecting EC

The condition of electricity in soil takes place through the moisture filled pores that occur between individual soil particles.

Therefore, the EC of soil is determined by the following soil properties.

4. Porosity: When the soil porosity is greater, electricity is more easily conducted. Soil with high clay content has higher porosity than sandier soil.
5. Water content: Dry soil is much lower in conductivity than moist soil.
6. Cation exchange capacity (CEC), mineral soil containing high levels of organic matter (humus) and/or 2:1 clay minerals such as montmorillonite, illite, or vermiculite have a much higher ability to retain positively charged ions (such as Ca, Mg, K, Na, NH_4, or H) than soil lacking these constituents. The presence of these ions in the moisture-filled soil pores will enhance soil EC in the same way that salinity does.
7. Mineral soil containing high levels of organic matter and or 2:1 minerals such as montmorillonite, illite or vermiculite have a much higher ability to retain positively charged ions (such as Ca, Mg, Na, NH_4 or H) than soil lacking these constituents. The presence of these ions in the moisture filled soil pores will increase soil EC.
8. Temperature: As temperature decreases toward the freezing point of water, soil EC decreases slightly. Below freezing, soil pores become increasingly insulated from each other and overall soil EC declines rapidly. Therefore, salt problems are more severe under hot, dry conditions than cool, humid conditions.
9. Other conditions: Salinity tolerance is influenced by many plant, soil and environment factors and their interrelationships. Generally fruits, vegetables and ornamentals are more salt sensitive than forage or field crops. In addition, certain varieties, cultivars or rootstalks may tolerate higher salt levels than others.

7) Plants are more sensitive to high salinity during seeding stages, immediately after transplanting and when subject to other stresses like pests and diseases.

8) Climate and irrigation also influence salinity tolerance.

Determination of EC

The determination of soil Electrical Conductivity by EC meter. The basic principle is the conductometry measurement of electrical conductivity of ionic solutions by applying electric current between two electrodes.

Solutions of strong acids, strong bases and most salts are good conductors of electric current and obeys Ohm’s law which states that the current flowing through a conductor between two points is directly proportional to the potential difference across the two points.

For the same type of material, the resistance is directly proportional to the length of the conductor and inversely to the length of the conductor and inversely to the cross section or diameter.

Interpretation based on EC measurement

Sl.No.	**EC (dSm-1)**	**Salinity effect**	**Interpretation**
1.	0.0-1.0	harmless	Salinity effect negligible
2.	1.1 – 3.0	Critical	Limit yield of sensitive crops
3.	3.1 and above	injuries	Limit yield of many crops

The relationship between EC and total soluble salt (TSS) in soil is

$$\mathbf{TSS = EC \times 640}$$

Lecture 25: Soil colloids - inorganic and organic

Soil Colloids

Soil colloids are defined as particles less than 0.001 mm in size not all clay particles are strictly colloids. Inorganic colloids usually make up the bulk of soil colloids, but the humus particles or organic colloids are the most reactive chemically and generally have a greater influence on soil properties per unit weight than the clay particles.

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 (Dispersion of clay in water) and Liquid in gas (Fog or clouds in atmosphere).

One of the most important properties of colloids is their ability to adsorb (attach to), hold, and release the mineral ions which form plant nutrients and dissolved in soil water. The nutrients are in the form of ions. Among the many ions in soil water, one important group consists of bases, which are ions of four elements: calcium (Ca^{++}), magnesium (Mg^{++}), potassium (K^{+}), and sodium (Na^{+}).

Colloids have a negative charge as a result of their physical and chemical composition. This negative charge attracts cations (nutrient atoms which all have a positive charge and water molecules), and holds them on the surface of the colloid.

Significance of soil colloids

The organic and inorganic contaminates are often transported via colloidal particles. Majority of surface area and electrostatic charge in a soil resides in the less than 1μm size fraction, with particles with radii between 20 and 1000 mm constituting the major part of soil surface area.

Since major part of the surface area is in the colloidal fraction of the soil, almost all surface controlled processes including adsorption reactions, nucleation and precipitation involve colloids. In addition to these chemical processes, colloids are mobile in soils, and thus affect not only the chemical transport of otherwise immobile chemicals, but also exert a strong influence on soil hydraulic properties.

Genesis of Clay colloids

The silicate clays are developed from the weathering of a wide variety of minerals by the two distinct process.

1. Alteration- A slight physical and chemical alteration of certain primary minerals

Changes in particle size. Alteration of muscovite mica to fine grained mica is the good example. As weathering occurs muscovite mineral is broken down in size to the colloidal range, part of the potassium is lost and some silicon is added from weathering solutions. Net result is the less rigid crystal structure and an electronegative charge. The fine mica colloid has a 2:1 structure, only have been altered in this process.

2. Recrystallization- Decomposition of primary minerals with subsequent recrystallization of certain of their products in to the silicate clays

Complete breakdown of clay structures and re-crystallization of clay minerals from product of this breakdown. It is the result of much more intense weathering than that required for alteration. Formation of Kaolinite (1:1) from solutions containing soluble aluminum and silicon that came from the breakdown of the primary minerals having 2:1 type structure.

Relative stages of weathering

The contact of rocks and water produces clays, either at or near the surface of the earth" (from Velde, 1995).

$$\text{Rock} + \text{Water} \rightarrow \text{Clay}$$

The CO_2 gas can dissolve in water and form carbonic acid, which will become hydrogen ions H^+ and bicarbonate ions, and make water slightly acidic.

$$CO_2 + H_2O \rightarrow H_2CO_3 \rightarrow H^+ + HCO_3$$

The acidic water will react with the rock surfaces and tend to dissolve the K ion and silica from the feldspar. Finally, the feldspar is transformed into kaolinite.

Feldspar + hydrogen ions +water → clay (kaolinite) + cations, dissolved silica

$$2KAlSi_3O_8 + 2H^+ + H_2O \rightarrow Al_2Si_2O_5(OH)_4 + 2K^+ + 4SiO_2$$

Fine grained micas and magnesium rich chlorites represent earlier weathering stages of the silicates and kaolinite and ultimately iron and aluminum oxides the most advanced stages.

General Properties of Soil Colloids

1. Size

The inorganic and organic colloids are smaller than 2 micrometers in diameter. These particles cannot be seen using an ordinary light microscope but can be seen only with an electron microscope.

2. Surface area

Because of their small size, all soil colloids have a larger external surface area per unit mass. The external surface area of 1 g of colloidal clay is 1000 times that of 1 g of coarse sand. Certain silicate clays have extensive internal surfaces occurring 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 m^2/g for clays with only external surfaces to more than 800 m^2/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 as 700,000 km^2g^{-1}.

3. Surface charges

Both external and internal surfaces of soil colloids carry negative and/or positive charges.Most of the organic and inorganic soil colloids carry a negative 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 sources of negative charge on clays comes from

i) Ionizable hydrogen ions

ii) Isomorphous substitution.

i) Ionizable hydrogen ions

These are hydrogen from hydroxyl (OH) 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 (-AlO- or - SiO-). The extent of ionized hydrogen depends on solution pH and hence these negative charges are **pH dependent charges**. More ionization occurs in alkaline (basic) solutions.

ii) Isomorphous substitution

This is due to the substitution of a cation of higher valence with another cation of lower valence but similar size in the clay crystal structure. In clay crystals some ions fit exactly into 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 $A1^{3+}$ and Si^{4+} cations. Common substitutions are the Si^{4+} replaced by $A1^{3+}$, and replacement of $A1^{3+}$ by Fe^{3+}, Fe^{2+}, Mg^{2+} or Zn^{2+}. As the total negative charge from the anions (oxygen) remains unchanged, the lower positive charge of the substituted cations result in excess negative charges on clay crystals.

4. Adsorption of cations

As soil colloids possess negative charge they attract and attach the ions of positive charge on the colloidal surfaces. They attract cations like H^+, $A1^{3+}$, Ca^{2+} and Mg^{2+}. This gives rise to an ionic double layer.

The Isomorphous substitution in the colloidal particle makes the external and internal layers of clay minerals negatively charged and these surfaces act as huge anions, which form the inner layer of the double layer. The outer layer is made up of a swarm of loosely held (adsorbed) cations attracted to the negativelycharged surfaces.

5. Adsorption of water

A large number of water molecules are associated with soil colloidal particles. Some water molecules are attracted to the adsorbed cations and the cation is said to be in hydrated state.Others water molecules are held in the internal surfaces of the colloidal clay particles. These water molecules play a critical role in determining both the physical and chemical properties of soil.

6. Cohesion (Attractive force between similar molecules or materials)

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

7. Adhesion (Attractive force between different molecules or materials).

Adhesion refers to the attraction of colloida1 materials to the surface of any other body or substance with which it comes in contact.

8. Swelling and shrinkage

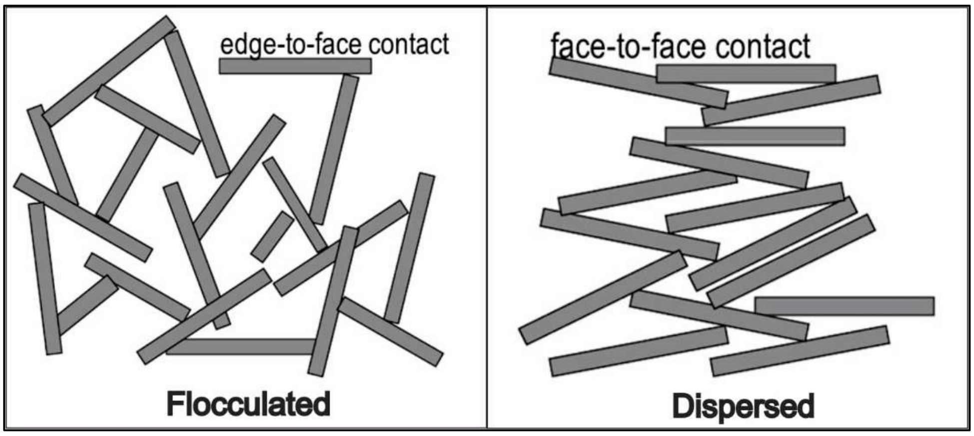
Some soil clay colloids belonging to smectite group like Montmorillonite swell when wet and shrink when dry. After a prolonged dry spell, soils high in smectite clay (E.g. **Black soil or Vertisols**) often show crises cross wide and deep cracks.

These cracks first allow rain to penetrate rapidly. Later, because of swelling, the cracks will close and become impervious. But soils dominated by kaolinite, chlorite, or fine grained micas do not swell or shrink. Vermiculite is intermediate in its swelling and shrinking characteristics.

9. Dispersion and flocculation

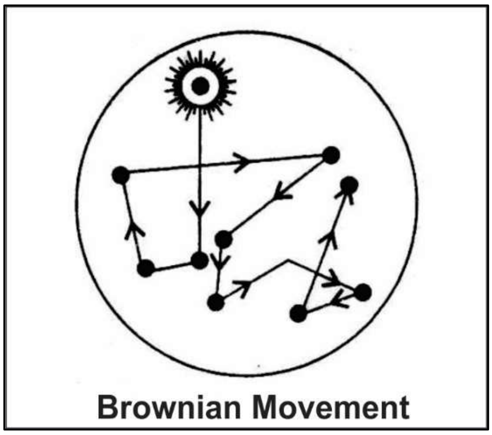
As long as the colloidal particles remain negatively charged, they repel each other and the suspension remains stable. If on any account they loose their charge, or if the magnitude of the charge is reduced, the particles coalesce, form flock or loose aggregates, and settle down.

This phenomenon of coalescence and formation of flocks is known as flocculation. The reverse process of the breaking up of flocks into individual particles is known as de-flocculation 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 crystalloids, 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.

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

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 primary source of charge on organic colloids is considered to be the gain or loss of H^+ from functional groups on the surface of colioids. The functional groups include hydroxyl (-OH), carboxyl (-COOH), phenolic ($-C_6H_4OH$) and amine ($-NH_2$). The charge that develops from such groups depends largely on the pH of the ambient solution, which regulates the degree of protonation or deprotonation of the group. The soil colloids that are capable of developing pH dependant charge include layer silicates, oxides and hydroxides of Fe and Al and organic matter.

Difference between organic and inorganic colloids

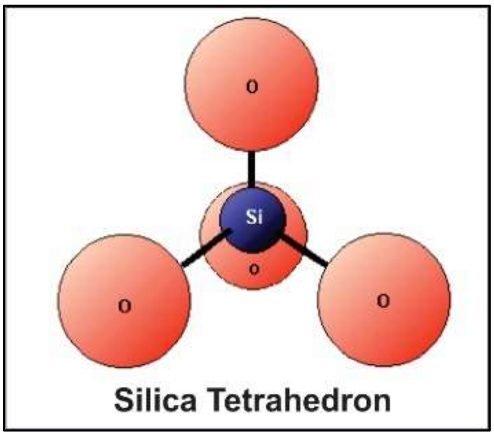
Humus	**Clay**
Made up of C, H, O	Made up of Si, Ai, O
Complex amorphous organic colloid	Inorganic and crystalline
More dynamic formed and destroyed more rapidly	Clays are stable relatively
Complex structure not well known	Clays have definite and well known structure

Lecture 26: Silicate clays: constitution and classification - 1:1, 2:1 expanding and non expanding - 2:2 clay minerals, amorphous minerals and their properties

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

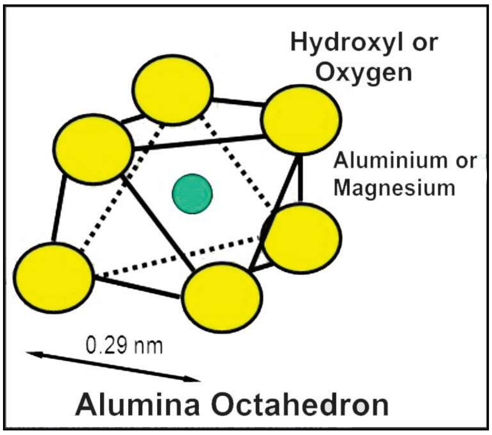
These important silicate clays are also known as phyllosilicates (Phyllon - leaf) because of their leaf-like or plate like structure. These are made up of two kinds of horizontal sheets. One dominated by silicon and other by aluminum and/or magnesium.



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 tetrahedra tied together horizontally by shared oxygen anions gives a tetrahedral sheet.



Alumina octahedron

Aluminium and/or magnesium ions are the key cations surrounded by six oxygen atoms or hydroxyl group giving an eight sided building block termed octahedron. Numerous octahedra linked together horizontally comprise the octahedral sheet.

An aluminum-dominated sheet is known as a di-octahedral sheet, whereas one dominated by magnesium is called a tri-octahedral sheet. The distinction is due to the fact that two aluminum ions in a di-octahedral sheet satisfy the same negative charge from surrounding oxygen and hydroxyls as three magnesium ions in a tri-octahedral sheet.

The tetrahedral and octahedral sheets are the fundamental structural units of silicate clays.These sheets 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 control the physical and chemical properties of each clay.

Classification of layer silicate clays

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

1) 1 :1 Type clay minerals
2) 2:1 Type clay minerals
3) 2: 1: 1 (or) 2:2 Type clay minerals

The basic building blocks of tetetrahedral and octahedral sheets are the silica tetrahedron and the aluminum octahedran. The Si^{4+} cation occurs in fourfold and tetrahedral coordination with oxygen while the Al^{3+} is generally found in six fold or octahedral coordination. Layer silicate minerals are sometimes defined on the basis of the number of certain positions occupied by cations.

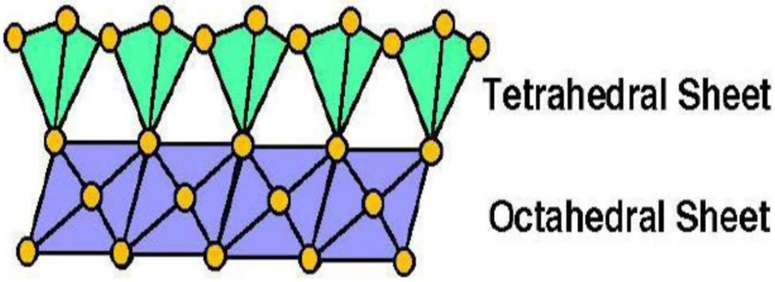
When two-thirds of the octahedral positions are occupied, the mineral is called **dioctahedra**l; when all 3 positions are occupied it is called **trioctahedral**.

1) 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. In soils, **kaolinite group** is the most prominent 1:1 clay mineral, which includes kaolinite, 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 restricted to its outer faces or to its external surface area. Also, there is little isomorphous substitution in this 1:1 type mineral. Because of low surface area and little isomorphous substitution the capacity to adsorb cations is also low.



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 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.

2) 2:1 Type Minerals

The crystal units (layers) of these minerals are characterized by an octahedral sheet sandwiched between two tetrahedral sheets.

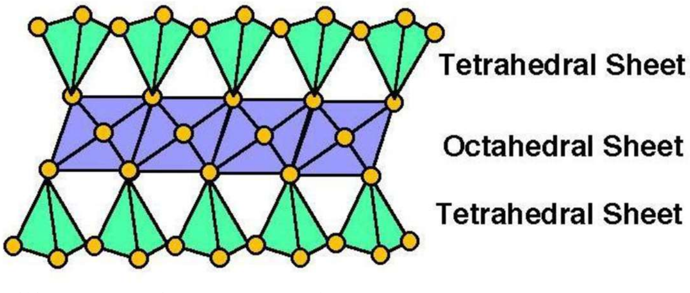
Three general groups have this basic crystal structure.

i. Expanding type: Smectite group and vermiculite

ii. Non-expanding type: mica group (illite)

i. Expanding Minerals (2:1)

The smectite group of minerals is noted for their interlayer expansion and swelling when wetted. The water enters the interlayer space and forces the layers apart. Montmorillonite is the most prominent member of this group in soils. Beidellite, nontronite, and saponiteare also found in soils.



The flake-like crystals of smectite (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 exceeds the external surface of clay crystal. In montmorillonite, magnesiumreplaces replaced aluminum in some sites of 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, swelling and shrinkage properties. Wide cracks commonly form in smectite dominated soils (E.g. **Vertisols**) when 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 (di-octahedral), although magnesium dominated (tri-octahedral) vermiculites are also present.

In the tetrahedral sheet of most vermiculite, aluminum is substituted by silicon in most of the sites. 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 smectite. For this reason, vermiculites are considered limited expansion clay minerals, expanding more than kaolinite but much less than the smectite.

The **cation exchange capacity** (CEC) of vermiculite is higher than all other silicate clays, including montmorillonite and other smectite because of very high negative charge in the tetrahedral sheet. Vermiculite crystals are larger than those of the smectite but much smaller than those of kaolinite.

ii) Non-expanding minerals (2:1)

Micas are the type of minerals in this group 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 or illite. The basic structure of illite or micaceous mineral is similar to that of montmorillonite. However, the particles are much larger than those of the smectite.

Some of the silicon ions are replaced by aluminium ions in the tetrahedral sheet (20%). This results in a net negative charge in the tetrahedral sheet which is compensated by potassium ions. The potassium as a binding agent, preventing expansion of the crystal. Hence, fine grained micas are quite non-expanding.

The properties such as hydration, cation adsorption, swelling, shrinkage and plasticity are less intense in fine grained micas.The specific surface area varies from 70 to 100 m^2g^{-1}, about one eighth that for the smectite.

S.No	Property	Montmorillonite	Illite	Kaolinite
1	Structure	2:1 lattice (expanding type)	2:1 lattice (non- expanding type	1:1 lattice (non- expanding type)
2	Size (micron)	0.01-1.0 (fine particle)	0.1-2.0 (medium particle)	0.1-5.0 (Coarse article)
3	Total surface area (m2g-1)	700-800	100-120	5-20
4	Shape	Irregular flakes	Irregular flakes	Hexagonal crystals
5	Substitution	Substitution in alumina sheet by Mg or Fe	Substitution in silica layer by aluminium	Substitution
6	Non-exchangeable cations	Megnesium	Potassium	None
7	Base(cation exchange capacity)(me/100 gm)	80-100	15-40	3-15
8	Cohesion	High	Medium	Low
9	Plasticity	High	Medium	Low
10	Swelling capacity	High	Medium	Low
11	Porosity	Low	Medium	High
12	Permeability	Low	Medium	High
13	CAtion Exchange capacity	80-100	5-40	3-15
14	Anion Exchange capacity	Low	Medium	High

3) 2:1:1 Type Minerals

This silicate group is represented by **chlorites**. 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 tri-octahedral sheet, giving a 2:1:1 ratio. Magnesium also dominates the tri-octahedral sheet in the 2:1 layer of chlorites. Thus, the crystal unit contains two silica tetrahedral sheets and two magnesium-dominated tri-octahedral 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 mica and less than smectite 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 minerals.

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 "mica - smectite" are used to describe mixed layer minerals.

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 alumina in primary minerals are dissolved and slowly leached away. The remnant materials, which have lower solubility are called sesquioxides.

Sesquioxides (metal oxides) are mixtures of aluminum hydroxide, $Al(OH)_3$, and iron oxide, Fe_2O_3, or iron hydroxide, $Fe(OH)_3$. 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 ($Al_2O_3.3H_2O$) and geothite ($Fe_2O_3.H_2O$). 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).

Lecture 27 & 28 : Sources of charge, ion exchange – positive and negative charge – isomorphous substitution, pH dependant charge, Ion exchange - Cation and anion exchange capacity and base saturation

Iron exchange

Ion adsorption and subsequent exchange are important processes that take place between soil colloidal particles (clays, organic matter, sesquioxides, and amorphous minerals) and various ions. Soil colloids serve very much as a modern bank. They are the sites within the soil where ions of essential plant nutrients are held and protected from excessive loss by leaching.

Subsequently, the nutrients can be "withdrawn" from the colloidal "bank" sites and taken up by plant roots. In turn, these elements can be "deposited" or returned to the colloids through the addition of commercial fertilizers, lime, manures, and plant residues.

The charges associated with soil particles attract ions (simple and complex) of opposite charge. In temperate region soils, negative charges generally predominate on the soil particles (colloids), hence adsorbed cations are present in larger quantities than anions. In more highly weathered soils (E.g. **Tropics**) where 1:1 type clays and Fe/Al oxides are the most dominant type of colloids, anion adsorption and exchange is relatively more prominent.

Ion exchange reactions - Cation exchange, anion exchange and base saturation - significance

The Ion Exchange phenomenon was first identified by **Harry Stephen Thompson** in England during 1850. When soil was leached (washed) with ammonium sulphate, calcium sulphate was detected in the leachate.

The ammonium ion in the solution replaced calcium in the soil.

$$(NH_4)2SO_4 + \text{Soil Ca} \longrightarrow \text{Soil } (NH_4)_2 + CaSO_4$$

The process by which ions are exchanged between solid and liquid phases and /or between solid phases if in close contact with each other is called ion exchange. The common exchangeable cations are Ca^{2+} mg^{2+}, H^+, K^+, NH_4^+ and Na. The common anions are SO_4^{2-}, Cl^-, NO_3^-.

The ion exchange in soil is due to the presence of residual positive and negative charges on the soil colloids. These residual charges result due to the process of isomorphous substitution and ionization of hydrogen and hydroxide functional groups. The negative charges attract positively charged ions and the positive charges attract negatively charged ions from soil solutions coming in contact with colloids.

The ions thus attracted are reversible and are on equivalent proportions. Exchange of cation is called cation exchange and exchange of anion is called anion exchange. The cation exchange phenomenon was first discovered by **Thomasway (1850).**

Ion exchange is the second most important reaction is nature. The first one is photosynthesis by green plants. The capacity of the soil to hold cation is called cation exchange capacity. The unit is C mol (P^+)/kg. The capacity to hold anion is called Anion exchange capacity (AEC). The unit of expression is C mol (e^-) / kg.

Mechanism of cation exchange

Clay colloids have negative charges. Cations are attracted to the clay particles. These cations are held on the held on the clay surfaces electro statistically. They are held by small particles of clay and organic matter. These small particles are called Micelle (Micro cell).

The cations that can be replaced on exchange site by other cations are called exchangeable cations. They are weekly held and they are in direct contact with the soil solution. They can be exchanged fairly easily.

Ions that are held very lightly with the colloid may be traped between layers of clay micelle. They do not pass to the soil solution very easily. They are called non exchangeable cations.

When any cation is added to the soil such as Ca^{++}, K^+ or NH_4 through fertilizers and soil amendments they are exchanged with those ions held on the colloid. When calcium is added to an acid soil the following reaction takes place.

$$Ca^{++}\ \boxed{\textbf{Clay Micelle}} + 2H^+ \longleftrightarrow \begin{matrix} H^+ \\ H^+ \end{matrix}\ \boxed{\textbf{Clay Micelle}} + Ca^{++}$$

Similarly when H^+ is added to the soil solution through the decomposition of organic matter or through acidic materials Ca^{2+} is replaced from the exchange complex by H^+.

Cations are adsorbed on negatively charged colloids forming a double layer. According to the **Electro-kinetic theory of ion exchange,** 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 and hence the double layer is diffused and so it is called **diffuse double**

layer. Due to these oscillations, some of the cations move away from the surface of the clay micelle (particle) and leave a negative charge on the colloid. Another cation of same species or different species present in the soil solution (electrolyte) will move towards the colloid and adsorbed on its surface. The electrolyte cation is now adsorbed on the clay colloid and the surface cation remains in solution as an exchanged ion. This phenomenon is called ion (cation) exchange. In soil, clay carries more negative charge. Silt has little and sand has no negative charge. Organic colloids also contribute negative charge to ion exchange. But the quantity of organic colloids present in soil will be very low and so its contribution to ion exchange will also be low.

Cation exchange capacity (CEC)

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 cations.

or

The sum total of the exchangeable cations that a soil can adsorb is called as cation exchange capacity. It is also defined as "the amount of cationic species bound at pH 7.0". Some authors consider pH 4.0 as the appropriate point.

Unit of expression

CEC is expressed as milliequivalents of cations per 100 grams of soil (meq /100g soil). After 1982, in the metric system the term equivalent is not used but moles are the accepted chemical unit. The recent unit of expression of CEC is centi moles of protons per kilo gram soil **[cmol (p+) kg^{-1} soil]**. One meq/100 g is equal to one cmol (p+) kg^{-1} soil.

Replacing power of cations

The replacing power of cations varies with the **type of ion**, **size, degree of hydration**, **valence**, **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: Li < Na < K < Rb < Cs < H and for divalent cations: Mg < Ca < Sr < Ba. In case of mixture of monovalent and divalent cations as they exist in normal soils the replacing power increases in the following order: Na < K < NH_4 < Mg < Ca < H. This means Na is more easily replaced than K and K more easily than NH4.

- Hydrated cations (Na) have lower replacing power than unhydrated cations (Ca)
- Ions with high positive charge (Al_3^+) are adsorbed strongly than Na^+

- Ions present in high concentration in the soil solution are adsorbed easily than ions in low concentration.
- Higher the degree of dissociation of the associated anion lower will be the adsorption. Na from NaOH is adsorbed easily than Na from NaCl. (The degree of dissociation of Cl is higher than that of OH)

Average CEC (pH 7) for some US soils based on USDA Soil Taxonomy

Soil Taxonomy order	**CEC (cmol$_c$/kg)**
Ultisols	3.5
Alfisols	9
Spodosols	9.3
Entisols	11.6
Mollisols	18.7
Vertisols	35.6
Histosols	128

CEC of different textural classes.

Sand	0-5 C mol (P^+) / Kg.
Sandy	loam 5-10 ”
Loam	10-15 ”
Clay loam	15-30
Clay	30.0

CEC of important clay minerals

Kaolinite	7-10 C mol (P^+) / Kg
Montmorillonite	80-100 ”
Vermiculite	100-150 ”
Illite	25-30 ”
Chlorite	25-30 ”
Fe & Al oxides	5.0 ”
Humus	200-400 ”

Factors influencing CEC

1. Soil Texture

CEC increases with fineness of the soil particles. This means increasing clay content will increase the CEC. As the particle size decreases the surface area increases per unit volume. This naturally increases the net charge and the CEC. Clay soils with high CEC can retain large amounts of cations and reduce the loss of cations by leaching. Sandy soils, with low CEC, retain smaller quantities of cations and therefore cations are removed from soil by leaching.

So, nutrients in cationic form like potassium must be applied frequently in smaller quantities (split applications) in sandy soil to reduce the loss of nutrients by leaching whereas in clay soils, it can be applied in one or two splits as K is retained due tohigh CEC and not leached.

2. Organic matter

In general CEC increases with increase in organic matter content. The pH dependent charges in the organic matter cause variation in CEC. When pH of the soil solution increases the CEC will also increase. The CEC of organic matter ranges from 200 to 400 [**cmol (p+) kg^{-1}**].

3. Nature of clay

CEC of clay minerals vary soil dominated with montmorillonite and vermiculite have higher CEC than those dominated with kaolinite, chlorite or illite. The CEC of clay minerals range from 10 to 150 [**cmol (p+) kg^{-1}].**

4. Soil reaction

In general CEC increases with increase in soil pH. As the pH increases the pH dependent charge increases. In Humus most CEC is pH dependent.

Role of Cation Exchange

Importance of exchangeable cations on plant nutrients is discussed below

Cation exchange reaction is considered as the second most important reaction next to photosynthesis. 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.

Cation exchange reactions are very important chemical reactions for the availability of plant nutrients in the soil. The capacity of soil to exchange cations is the best single index of soil fertility. Plant roots, when they come in contact with colloidal particles, absorb exchangeable cations directly by inter exchange or contact exchange between the root hairs and colloidal complex.

(a) Nature and content of exchangeable bases

The nature and content of exchangeable bases in a soil have an important bearing on its general properties. In all normal fertile soils the total exchangeable bases (Ca, Mg, K, Na) constitute about 80 to 90% of the cation adsorbing capacity. Exchangeable hydrogen is usually under 20%. In these soils, calcium forms the predominant exchangeable base, constituting 60 to 80% of the total exchangeable cation.

The predominance of exchangeable calcium give rise to Ca- clay which imparts a neutral reaction to the soil. The pH value varies from 6.5 to 7.5. When the proportion of exchangeable hydrogen (H) is high it gives rise to acid soil. In such soils, exchangeable calcium is correspondingly low, and in highly acid soils it is almost absent. In such cases the clay is saturated with hydrogen cations (H^+) and forms H-clay. Acid soils are less fertile. It is called base unsaturated soil.

When exchangeable sodium form more than 10 to 15% of the total exchangeable cation it gives rise to alkaline soils. The pH value of such soils is usually greater than 8.0. When the proportion of exchangeable sodium exceeds these limits (or saturates the colloidal complex), the clay is turned into a Na-clay.

The soil is now highly alkaline and the pH value ranges from 9 to 12 Alkaline soils are also less fertile. Soils with a high calcium base saturation are in the most satisfactory physical and nutritional condition. A calcium dominated soil is granular in structure and ensure good drainage and aeration.

(b) Type of colloid

Type of colloid affects the cation exchange. Montmorillonite colloid hold the calcium ion with greater tenacity than Kaolinite at a given base saturation. As a result, Kaolinite will liberate calcium much more readily than Montmorillonite.

(c) Associated ions

Presence of exchangeable calcium in excessive quantities in a soil will limit the availability of potassium to plants. In same manner, high exchangeable potassium may depress the availability of magnesium.

(d) Adsorption of cations

Colloidal clay (humus) hold in varying amount of plant nutrients (calcium, magnesium, potassium, nitrogen, phosphorus and most of the micronutrients) which are available to plant.

(e) Property of base exchange

Base exchange (cation exchange) property checks leaching losses of available nutrients. On application of potassium sulphate fertilizer in the soil, potassium ions are held on the surface of colloids by cation exchange process. Subsequently, exchangeable potassium ions are directly available to plants.

Influence on Soil Fertility

Cation exchange capacity is the best index of soil fertility. By cation exchange, hydrogen ions from the root hairs and microorganisms replace nutrient cations from the exchange complex. The nutrient cations are forced into the soil solution where they can be assimilated by the adsorptive surface of roots and soil organisms, or they may be removed by drainage water.

(i) Cation saturation and Soil fertility

Soil with a high calcium base saturation are the most satisfactory physical and nutritional condition. A calcium dominated soil is granular in structure and porous. Calcium-dominated clay ensures good aeration and good drainage, thus increases fertility of the soils.

Base unsaturated soils are acidic in nature due to exchangeable hydrogen. These soils are less fertile. Base saturated soils with dominant sodium cations are alkaline in nature. Alkaline soils are not fertile due to de-flocculation, stickiness, hard to work, poor drainage and poor aeration.

(ii) Cation exchange and Soil fertility

Due to the property of cation exchange (base exchange) the soluble inorganic fertilizer nutrients are not washed away from the soil. For example, ammonium sulphate fertilizer is added to the soil, ammonium ions are held on the surface of colloids by cation exchange. Ammonium ions are taken up by plants. This process checks nutrient losses by leaching and make the soil fertile. The cations Ca, Mg, K, and NH_4 are held on the colloidal surfaces and are readily available to plants.

(iii) Influence of complementary adsorbed cations and soil fertility

The order of strength of adsorption, when the ions are present in equivalent quantities, is as follow

$$Al^{3+} > Ca^{2+} > Mg^{2+} > K^{+} = NH_4^{+} > Na^{+}$$

Consequently, a nutrient cation such as K^+ is less tightly held by the colloids if the complementary ions are Al^{3+} and H^+ (acid soils) than if they are $Mg^{++} + Na^+$ (neutral to alkaline soils). The loosely held K^+ ions are more readily available for absorption by plants or for leaching in acid soils.

There are also some nutrient “antagonisms”, which in certain soil cause inhibition of uptake of some cations by plants. Thus, potassium uptake by plants is limited by high levels of calcium in some soils. Likewise, high potassium levels are known to limit the uptake of magnesium even when significant quantities of magnesium are present in the soil.

(iv) The exchangeable K is a major source of plant K.

(v) The exchangeable Mg is often a major source of plant Mg.

(vi) The amount of lime required to raise the pH of an acidic soil is greater as the CEC is greater.

(vii) Cation exchange sites hold Ca^+, Mg^+, K^+, Na^+, and NH_4^+ ions and slow down their losses by leaching.

(viii) Cation exchange sites hold fertilizer K^+ and NH_4^+ and greatly reduce their mobility in soils.

(ix) Cation exchange sites adsorb many metals (Cd^{2+}, Zn^{2+}, Ni^{2+}, and Pb^{2+}) that might be present in wastewater adsorption removes them from the percolating water, thereby cleansing the water that drains into groundwater.

Base saturation

The relative percentage of base forming cations in the total CEC is termed as per cent base saturation.

or

The percentage of CEC that is satisfied by the base forming cations is called base saturation percentage.

$$\%\ \text{Base saturation} = \frac{\text{Exchangeable base forming cations (C mol/kg)}}{\text{CEC (C mol/kg)}} \times 100$$

Aluminium and hydrogen are considered as acid forming ions. Calcium, Magnesium, Potassium and Sodium are considered as base forming ions. Base forming substances are called as besoids and acid forming substances are called as acidoids. The percentage of sodium in the total CEC is called Exchangeable sodium percentage

$$\text{ESP} = [(\text{Na/CEC}) \times 100]$$

These parameters are considered while deriving fertilizer prescription and amendments for problem soils.

A knowledge of base saturation percentage is useful in many ways.

1. It helps in determining the quantity of lime required to raise the pH of acid soils.
2. It indicates the proportion of plant nutrients in CEC. It is an index of soil fertility.
3. Degree of saturation of a particular cation in CEC indicates the ease with which the cation can be released for plant nutrition. For example if calcium saturation is more Ca can be very easily replaced from the exchange complex.
4. For a fertile soil it is considered that the base saturation percentage should be more than 80.

Anion exchange

Anion adsorption: Adsorption of negative ion (anions) E.g. Cl^-, NO_3^-, SO_4^{2-}, and $H_2PO_4^-$ on positively charged sites of clay and organic matter is known as anion adsorption.

Source of positive charge

1. Isomorphous substitution: Low valence cations replaced by high valence cations.
2. Surface and exposed broken bonds of clay lattice: OH group in certain acid soils.

$$\underset{\text{(Soil solids)}}{\underset{\text{No charge}}{Al - OH}} + \underset{\text{(soil solution)}}{H^+} \longrightarrow \underset{\text{(Soil solids)}}{\underset{\text{Positive charge}}{Al - OH_2^+}}$$

3. Complex aluminium and iron hydroxy ions in acid soils.

$$\underset{\text{No charge}}{Al\,(OH)_3 + H^+} \longrightarrow \underset{\text{Positive charge}}{Al\,(OH)_2^+ + H_2O}$$

4. pH dependent charges are important for anion exchange of organic matter

Anion exchange

Colloids are positive and the exchange is among negatively charged anions.

or

Replacement of one anion by another anion on the positively charged colloids is called anion exchange.

Positive charges are due to OH^- of iron and aluminium, 1:1 clays and allophone (amorphous clays). Anion exchange is pH dependent. Lower the pH greater is the anion exchange.

$$\text{Clay } NO_3^- + \text{solution } Cl^- \longrightarrow \text{Clay } Cl^- + \text{Solution } NO_3^-$$

Equivalent quantities of NO_3^- and Cl^- are exchanged. The reaction is reversible. Plant nutrients are released for plant absorption. The adsorption and exchanges of some anions like phosphate, molybdate, and sulphate is complex because of specific reactions between the anions and soil constituents. For example, the H_2PO_4 ion may react with the protonated hydroxyl group rather than remain as an easily exchanged anion.

$$\underset{\text{(Soil solid)}}{Al^- OH_2^+} + \underset{\text{(In soil solution)}}{H_2PO_4^-} \longrightarrow \underset{\text{(Soil solid)}}{Al^- H_2PO_4^-} + \underset{\text{(Soil solution)}}{H_2O}$$

This reaction actually reduces the net positive charge on the soil colloid. Also H_2PO_4 is held very tightly by the soil solids and is not readily available for plant uptake. Despite these complexities anion exchange is an important mechanism for interactions in the soil and between the soil and plant. Together with cation exchange it largely determines the ability of soil to provide nutrients to plants promptly.

Anion exchange capacity

"The sum total of exchangeable anions a soil can adsorb is known as anion exchange capacity". It is expressed as cmol / kg or m.eq./ 100 g soil. The capacity of adsorption and exchange of anions varies with the type of clay mineral, soil reaction, and the nature of anion. Kaolinitic minerals have a greater anion adsorbing and exchange capacity than montmorillonitic and illitic clays because the exchange is located at only a few broken bonds. The capacity for holding anions increases with the increase in acidity. The lower the pH the greater is the adsorption. All anions are not adsorbed equally readily. Some

anions such as H_2PO_4 are adsorbed very readily at all pH values in the acid as well as alkaline range. Cl and SO_4 ions are adsorbed slightly at low pH but none at neutrality, while NO_3 ions are not adsorbed at all. The affinity for adsorption of some of the anions commonly present in soil is of the order: $NO_3 < Cl < SO_4 < PO_4$. Hence at the pH commonly prevailing in cultivated soils, nitrate, chloride and sulphate ions are easily lost by leaching.

Soils with Kaolinite dominant clay have higher anion exchange capacity than montmorillonite or illite. The relative order of anion exchange is

$$OH > H_2PO_4 > SO_4 > NO_3 > Cl$$

Importance of anion exchange

1. The phenomenon of anion exchange is important for the release of fixed P in the soil. In acid soils the phosphorus is fixed as insoluble Al-Phosphate. Liming the acid soils release fixed P. Here the OH ion replaces H_2PO_4 from Al(OH).

$$Al(OH)_2\ H_2PO_4 + OH \longrightarrow Al(OH)_3 + H_2\ PO_4$$

2. Similarly the availability of other nutrients like NO_3, SO_4 and Cl are Influenced by anion exchange.

$$\boxed{\text{Soil Colloid}}\ NO_3 + Cl \longrightarrow \boxed{\text{Soil Colloid}}\ Cl + NO_3$$

Soil solution Soil solution

3. The phosphate ions, therefore, react with the hydrous oxides also and get fixed as in the case of silicate clay, forming insoluble hydroxy - phosphates of iron and aluminium.

$$\underset{\text{(Soluble)}}{Al\ (OH)_3 + H_2PO_4} \longrightarrow \underset{\text{(Insoluble)}}{Al\ (OH)_2.\ H_2PO_4 + OH^-}$$

If the reaction takes place under conditions of slight acidity it is reversible, and soluble phosphate is again liberated when hydroxy phosphate comes in contact with ions. If the reaction takes place at a low pH under strongly acid conditions, the phosphate (ions) are irreversibly fixed and are totally unavailable for the use of plants.

Significance of ion exchange

Next to photosynthesis ion exchange is the most important reaction in the world. Plants take up their food material from the soil through ion exchange only. Plant roots which are in contact with the soil solution exchange the nutrients from soil solution for H+ ions present on the surfaces of root hairs.

Effect on soil fertility

A soil is considered to be fertile when the base saturation percentage is more than 80. Each percent of humus contributes about 2 C mol /kg of CEC. Montmorillonite contributes about 1 C mole and Kaolinite contributes about 0.08 C mol/kg for every one percent.

Availability of applied nutrients

When fertilizers are applied to supply plant nutrients elements like K, Ca, Mg and NH_4 dissolve in soil solution. These nutrients in soil solution are exchanged for other cation like H^+ present in the exchange complex. If there is no cation exchange the applied nutrients would be lost in drainage water.

Similar is the case with anion radicals like PO_4, NO_3, SO_4 etc. Soils with high CEC can adsorb higher amounts of nutrients. Hence, in clay soils we can apply larger quantities of fertilizers in a single dose. Sandy soils have very low CEC and in such soils fertilizers should be applied in splits.

Effect of adsorbed cations

When the soil exchange complex has calcium the soil will have desire able physical properties. The activity of soil microorganisms, ammonification and nitrification processes also are determined by the cations of exchange complex.

Toxic ions

When the exchange complex had adsorbed metals like Cadmium, Nickel and lead they are toxic to the crop plants.

Effect on soil pH

Clays with H are acidic and with Na are alkaline. The acidic and alkaline nature of soil has its own effect on soil properties.

Lecture 29: Soil organic matter: composition, properties and its influence on soil properties

Soil Organic Matter

Substances containing carbon are organic matter. Soil organic matter consists of decomposing plant and animal residues. It also includes substances of organic origin either leaving or dead. Soil organic matter plays an important role in deciding / maintaining soil physical conditions.

It also influences soil chemical properties especially cation exchange capacity. Organic matters supply the energy sources for soil micro organisms. Soil development is another aspect which is influenced by the soil organic matter.

Plant tissue is the major source. Animals are considered as the secondary sources. They attack original plant tissues, contribute waste products and leave their own bodies after death.

Composition of organic residues

Plant residues contain 75% moisture and 25% dry matter. This 25% is made up of Carbon (10-12%), Oxygen (9-10%) , Hydrogen (1.5-2.5%) , N(1-2%) and mineral matter (1-3%).

Composition of plant tissues

Carbohydrates

Celluloses 20-50%

Hemicellulose 10-30%

Starch, Sugar 1-5%

Proteins 1-15%

Fats, waxes, tannins 1-10%

Lignins 10-30%

Inorganic residues (mineral matter)

1. Water insolubles

Proteins, Peptides, Nitrogenous Peptones and S containing materials

2. Water solubles

NO_3, NH_4 compounds, Soil organic residues, Non Nitrogenous Carbohydrates (celluloses, Hemicellulose, Starch, Sugar etc).

3. Ether solubles

Fats, oils, waxes, resins and Lignins

The organic matter is also classified on the basis of their rate of decomposition

1. **Rapidly decomposed:** Sugars, starches, proteins etc.
2. **Less rapidly decomposed:** Hemicelluloses, celluloses etc.
3. **Very slowly decomposed:** Fats, waxes, resins, lignins etc.

Decomposition of soil organic matter

Different organic residues contain different organic compounds. There is great variation in the rate of decomposition of organic residues. Sugars, starches and simple proteins are very rapidly decomposed. On the other hand Fats, waxes and lignins are very slowly decomposed. Hemicellulose, celluloses and protein are intermediate. Even though the composition may vary the end products are more or less the same.

The general reactions taking place during decomposition are

1. Enzymatic oxidation of the bulk with the release of CO_2, water, energy and heat
2. Essential elements are released (N, P, S etc) and immobilized by a series of reactions.
3. Formation of compounds which are resistant to microbial action.

Molecules very resistant to microbial action is formed either through modification of compounds or by microbial synthesis.

Under aerobic conditions the products formed are

CO_2, NH_4, NO_3 , H_2PO_4, SO_4, H_2O and essential plant nutrients like Ca, Mg, Fe, Cu, Zn etc.

Under anaerobic conditions

CH_4, organic acids like lactic, propionic, butyric, NH_4, various amine residues ($R\text{-}NH_2$) H_2S, ethylene ($CH_2{=}CH_2$) and humic substances.

A. Decomposition of soluble substances

When glucose is decomposed under aerobic conditions the reaction is as under

$$\text{Sugar} + \text{Oxygen} \longrightarrow CO_2 + H_2O$$

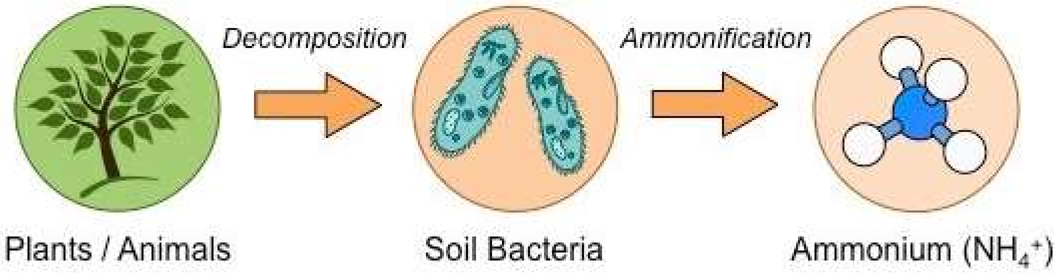
Under partially oxidized conditions,

Sugar + Oxygen ⟶ Aliphatic acids (Acetic, formic etc.) or Hydroxy acids, (Citric, lactic etc.) or Alcohols (ethyl alcohol etc.)

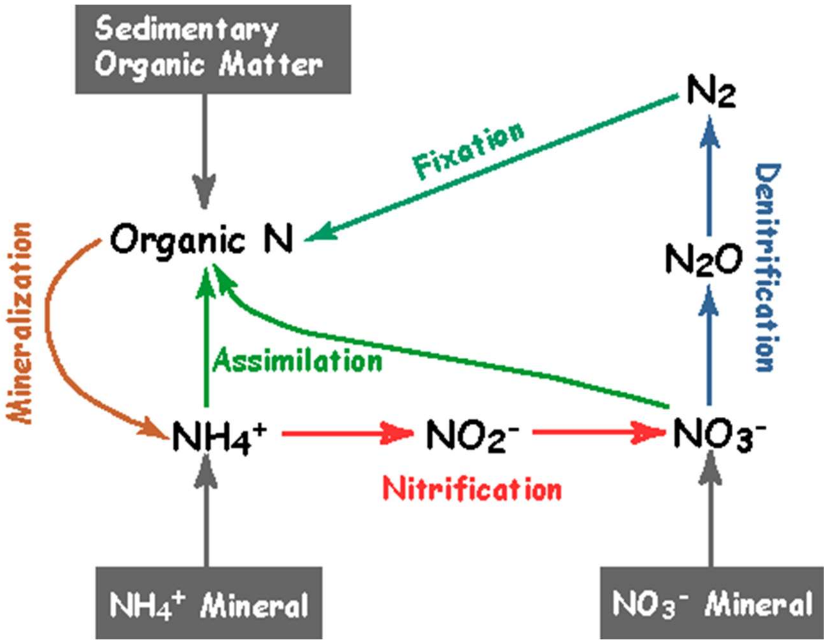
Some of the reactions involved may be represented as under:

i) Ammonification

The transformation of organic nitrogenous compounds (amino acids, amides, ammonium compounds, nitrates etc.) into ammonia is called ammonification. This process occurs as a result of hydrolytic and oxidative enzymatic reaction under aerobic conditions by heterotrophic microbes.



ii) Nitrification



The process of conversion of ammonia to nitrites (NO_2) and then to nitrate (NO_3^-) is known as nitrification. It is an aerobic process by autotrophic bacteria.

$$\underset{\text{Ammonia}}{NH_3} \xrightarrow{\text{Nitrosomons}} \underset{\text{Nitrite}}{NO_2} \xrightarrow{\text{Nitrobacter}} \underset{\text{Nitrate}}{NO_3^-}$$

The net reactions are as follows

$$NH_4 + O_2 \longrightarrow NO_2^- + 2H^+ + H_2O + \text{energy}$$

$$NO_2 + O_2 \longrightarrow NO_3^- + \text{energy}$$

iii) Denitrification

The process, which involves conversion of soil nitrate into gaseous nitrogen or nitrous oxide, is called denitrification. Water logging and high pH will increase N loss by denitrification.

$$\text{Nitrate} \xrightarrow{\text{Pseudomonas/Bacillu}} N_2 \text{ gas}$$

2. Under anaerobic conditions

$C_6H_{12}O_6$ (Glucose) - Lactic acid, butyric acid Ethyl alcohol are formed. Protein and other N compounds are converted into elemental N.

B. Decomposition of Insoluble Substances

i) Breakdown of Protein

During the course of decomposition of plant materials, the proteins are first hydrolyzed to a number of intermediate products.

Aminization

The process of conversion of proteins to amino acids.

Ammonification

The process of conversion of amino acids and amides to ammonia.

ii) Breakdown of cellulose

The decomposition of the most abundant carbohydrates.

$$\text{Cellulose} \xrightarrow[\text{(cellulase)}]{\text{Hydrolysis}} \text{Cellobiose} \xrightarrow[\text{(cellobiase)}]{\text{Hydrolysis}} \text{Glucose} \xrightarrow{\text{Oxidation}} \text{Organic acids} \xrightarrow{\text{Oxidation}} CO_2 + H_2O$$

iii) Breakdown of Hemicellulose

Decompose faster than cellulose and are first hydrolyzed to their components sugars and uronic acids. Sugars are attacked by microbes and are converted to organic acids, alcohols, carbon dioxide and water. The uronic acids are broken down to pentose and CO_2. The newly synthesized hemicelluloses thus form a part of the humus.

iv) Breakdown of Starch

It is chemically a glucose polymer and is first hydrolyzed to maltose by the action of amylases. Maltose is next converted to glucose by maltase. The process is represented as under:

$$(C_6H_{10}O_5)n + nH_2O \longrightarrow n\,(C_6H_{12}O_6)$$

C. Decomposition of ether soluble substances

$$\text{Fats} \longrightarrow \text{Glycerol + fatty acids}$$

$$\text{Glycerol} \longrightarrow CO_2 + \text{water}$$

D. Decomposition of lignin

Lignin decomposes slowly, much slower than cellulose. Complete oxidation gives rise to CO_2 and H_2O.

Sulphur containing organic compounds

Converted to $SO_4^{-2} + H^+$ + energy by sulphur oxidizing bacteria.

Phosphorus containing organic compounds

Various micro organisms mineralize phospholipids and other organic Phosphorus compounds in the presence of phosphates enzymets H_2PO_4 and HPO_4^{-2} depending on soil pH.

Mineralization

The biological conversion of organic forms of C, N, P and S to inorganic or mineral forms is called mineralization.

Immobilization

The conversion of inorganic forms of C, N, P and S by the soil organism into organic forms is called Immobilization.

Factors affecting decomposition

1. **Temperature**: Cold periods retard plant growth and organic matter decomposition. Warm summers may permit plant growth and humus accumulation.
2. **Soil moisture**: Extremes of both arid and anaerobic conditions reduce plant growth and microbial decomposition. Near or slightly wetter than field capacity moisture conditions are most favorable for both processes.
3. **Nutrients**: Lack of nutrients particularly N slows decomposition.
4. **Soil pH**: Most of the microbes grow best at pH 6 to 8, but are severely inhibited below pH 4.5 and above pH 8.5.
5. **Soil Texture**: Soils higher in clays tend to retain larger amounts of humus.
6. **Other Factors**: Toxic levels of elements (Al, Mn, B, Se, Cl), excessive soluble salts, shade and organic phytotoxins in plant materials.

Role of organic matter

Organic matte creates a granular condition of soil which maintains favorable condition of aeration and permeability. Water holding capacity of soil is increased and surface runoff, erosion etc., are reduced as there is good infiltration due to the addition of organic matter. Surface mulching with coarse organic matter lowers wind erosion and lowers soil temperatures in the summer and keeps the soil warmer in winter.

Organic matter 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. Fresh Organic matter supplies food for earthworms, ants and rodents and makes soil P readily available in acid soils. Organic acids released from decomposing organic matter help to reduce alkalinity in soils; organic acids along with released CO_2 dissolve minerals and make them more available.

Humus (a highly decomposed organic matter) provides a storehouse for the exchangeable and available cations. It acts as a buffering agent which checks rapid chemical changes in pH and soil reaction.

Lecture 30: Humic substances – fractionation, nature and properties, Theories of humus formation

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 they are 1. Decomposition and 2. Synthesis.

1. Decomposition

Chemicals in the plant residues are broken down by soil microbes into simpler compounds or monomers like CO_2, NO_3, CH_4, H_2O, etc., Other simpler organic compounds that result from the breakdown take part immediately in the second of the humus-forming processes, biochemical synthesis. These simpler chemicals are metabolized into new compounds in the body tissue of soil microbes. 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. These monomers undergo polymerization by which polyphenols and polyquinones are formed. These high molecular weight compounds interact with N-containing amino compounds and forms a significant component of resistant humus. Colloidal clays encourage formation of these polymers. Generally two groups of compounds that collectively make up humus, the humic group and the non humic group.

Soil organic matter fractions

There are two organic matter fraction they are, Humic matter, Non humic matter. When soil is extracted with alkali the humic substances go into solution. The insoluble portion forms the non humic matter.

Humic group

This group makes up about 60-80% of the soil organic matter. They are most complex. They are most resistant to microbial attack. Humic substances have aromatic ring type structures.

These include polyphenols and poly quinones. These are formed by decomposition, synthesis and polymerization. The humic substances are classified based on resistance to degradation and solubility in acids and alkalis into

- Humic acid
- Fulvic acid
- Humin

Non humic group

This group makes upto 20-30% of the organic matter in soil. These are less complex and less resistant to microbial attack as compared to humic substances. They are polysaccharides, polymers having sugar like structures and polyuronides. These include proteins, carbohydrates, lignins, fats, waxes, resins, tannins and some compounds of low molecular weight.

Fulvic acid	Humic acid	Humin
Low molecular weight	Medium molecular weight	High molecular weight
Soluble in acid and alkali	Soluble in alkali and insoluble in acid	Insoluble in both acid and alkali
Most susceptible to microbial attack	Intermediate in resistance to degradation	Most resistant to microbial attack

Types of humus in soil

Humus occurs in soils in many types, differentiates in regard to morphology and fractional composition. A type of humus is it a morphological form of naturals accumulation of humic substances in profile or on the surface of soil, conditioned by general direction of soil forming process and humification of organic matter.

A types of humus in terrestrial enviroment are following:

1. Mor
2. Moder
3. Mull

Mor is a type of humus, which occur largely in coniferous forest soils and the moorlands soils. This humus arise under conditions of low-biological activity in soil. The mineralization of organic matter proceed slowly and create layers, which maintain a structure of vegatable material. Acidophilic fungi and low active invertebrates participates in transformations of plant residues. Under these circumstances forms a litter of large thickness. C/N ratio of mor humus is always more than 20, or even 30-40, whereas pH is acid.

Moder is a transitional form of humus between mull and moder, characteristic for sod-podzolic soils, loesses and mountain grassland soils. The organic horizons with moder humus consist of low-thicknessed litter (2-3 cm), which gradually, without bounds, pass on to humus-accumulative horizons. Moder is a type of medium humified humus. Acidophilic fungi and arthropodan participates in transformations of plant residues. C/N ratio equal 15-25. Produced mineral-organic complexes are labile and weakly bounded with mineral portion of soil.

Mull is a type of humus characteristic for chestnut soils, phaeozems, rendzinas and others soils. This type of humus arise under grass vegetation. Mull is a well humified organic matter, which is produce in very biologically active habitat. This type of humus is characterized by neutral pH, C/N ratio nearing to 10 and ability to creation stable mineral-organic complexes. Mull is a type of humus which occurs in soils under cultivation.

According to Kononova, the types of humus are divide as follows

First type of humus is characteristic for podzolic soils, grey brown soils and lateric soils under forest communities. In this humus predominate humic acids, thus humic acid/fulvic acid ratio is below 1. Humic acid indicate small extent of aromatic rings condensation and they are approximate to fulvic acids. Considerable hydrophilic properties of humic acids favor to creation of chelates with polyvalent cations and ability to displacement deep into profile of soil. Considerable mobility of this humus favor process of podsolization.

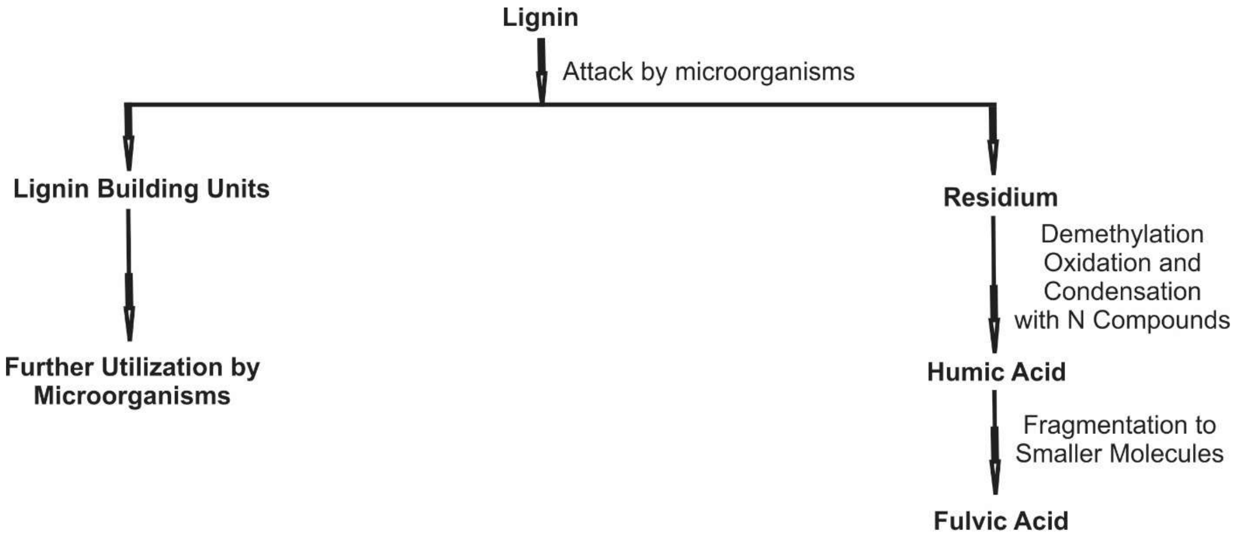
Second type of humus is characteristic for phaeozems, rendzinas, black earths and brown soils. Humic acid/fulvic acid ratio is upper than 1, Extent of aromatic rings condensation is high in humic acids, which cause their hydrophobic properties and inability to creation of chelates. Humic acids are strongly connected with mineral portion of soil in this type of humus.

Third type of humus is characteristic for semidesert soils. In this humus predominatefulvic acids fraction, whereas arise of humic acids is limited. Beyond this, humic acids are largely bounded with mineral portion of soil.

Theories on humus formation

Pathway 1 - The lignin theory (Waksman (1932)

For many years it was thought that humic substances were derived from lignin (pathway 1). According to this theory, lignin is incompletely utilized by microorganism and the residuum becomes part of the soil humus. Modification in lignin include loss of methoxyl (OCH_3) groups with the generation of o-hydroxyphenols and oxidation of aliphatic side chains to form COOH groups. The modified material is subject to further unknown changes to yield first humic acids and then fulvic acids. This pathway, illustrated on the picture , is exemplified by Waksman's lignin-protein theory.



Lignin Theory of Humus formation (Waksman 1932)

The following evidence was cited by Waksman in support of the lignin theory of humic acid formation:

1. Both lignin and humic acid are decomposed with considerable difficulty by the great majority of fungi and bacteria.
2. Both lignin and humic acid are partly soluble in alcohol and pyridine.
3. Both lignin and humic acid are soluble in alkali and precipitated by acids.
4. Both lignin and humic acid contain OCH_3 groups.
5. Both lignin and humic acid are acidic in nature.
6. When lignins are warmed with aqueous alkali, they are transformed into methoxyl-containing humic acids.
7. Humic acids have properties similar to oxidized lignins.

Although lignin is less easily attacked by microorganisms than other plant components, mechanisms exist in nature for its complete aerobic decomposition. Otherwise undecomposed plant remains would accumulate on the soil surface and the organic matter content of the soil would gradually increase until CO_2 was depleted from the atmosphere. The ability of soil organisms to degrade lignin has been understimated in some quarters and its contribution to humus has been exaggerated.

In normally aerobic soils lignin may be broken down into low-molecular-weight products prior to humus synthesis. On the other hand, the fungi that degrade lignin are not normally found in excessively wet sediments. Accordingly, it seems logical to assume that modified lignins may make a major contribution to the humus of peat, lake sediments, and poorly drained soils.

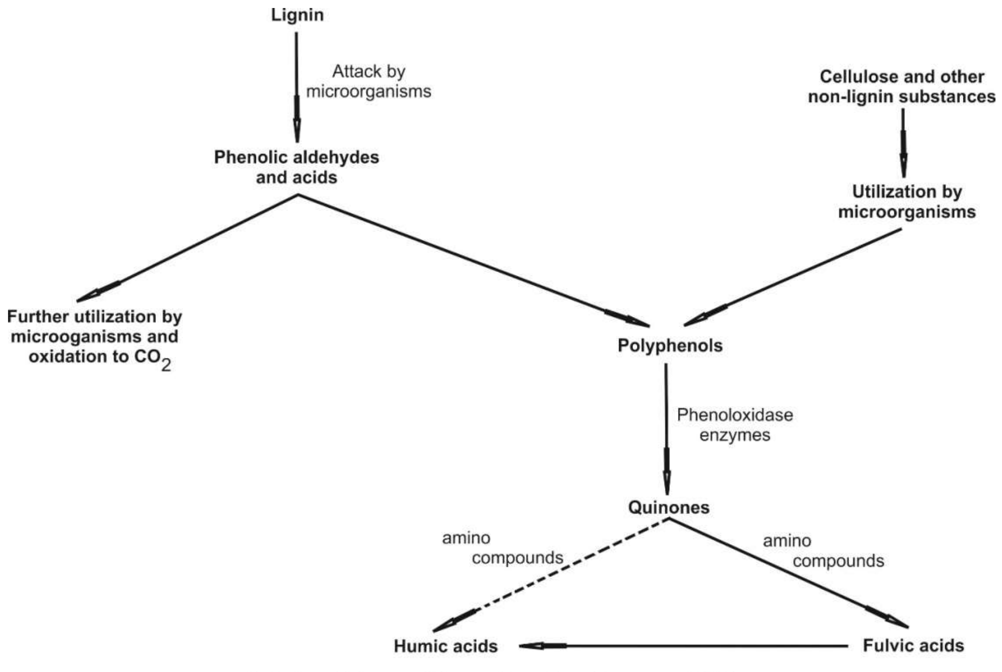
Pathway 2 and 3 - The polyphenol theory

In pathway 3 lignin still plays an important role in humus synthesis, but in a different way. In this case phenolic aldehydes and acids released from lignin during microbiological attack undergo enzymatic conversion to quinones, which polymerize in the presence or absence of amino compounds to form humic like macromolecules.

Pathway 2 is somewhat similar to pathway 3 except that the polyphenols are synthesizedby microorganisms from nonlignin C sources (E.g. **Cellulose**). The polyphenols are then enzymatically oxidized to quinones and converted to humic substances. As noted earlier, the classical theory of Waksman is now considered obsolete by many investigators. According to current concepts quinones of lignin origin, together with those synthesized by microorganisms, are the major building blocks from which humic substances are formed.

The formation of brown-colored substances by reactions involving quinones is not rareevent, but is a well-known phenomenon that takes place in melanine formation, such as in the flesh of ripe fruits and vegetables following mechanical injury and during seed coat formation.

Possible sources of phenols for humus synthesis include lignin, microorganisms, uncombined phenols in plants and tannins. Of these, only the first two have received serious attention.



Polyphenol Theory of Humus formation (Stevenson 1982)

Flaig's concept

1. Lignin, freed of its linkage with cellulose during decomposition of plant residues, is subjected to oxidative splitting with the formation of primary structural units (derivatives of phenylpropane).
2. The side-chains of the lignin-building units are oxidized, demethylation occurs, and the resulting polyphenols are converted to quinones by polyphenoloxidase enzymes.
3. Quinones arising from the lignin (and from other sources) react with N-containing compounds to form dark-colored polymers.

Kononova concept

The role of microorganisms as sources of polyphenols has been emphasized by Kononova. She concluded that humic substances were being formed by cellulose-decomposing myxobacteria prior to lignin decomposition.

The stages leading to the formation of humic substances were postulated to be:

1. Fungi attack simple carbohydrates and parts of the protein and cellulose in the medullary rays, cambrium, and cortex of plants residues.
2. Cellulose of the xylem is decomposed by aerobic myxobacteria. Polyphenols synthesized by the myxobacteria are oxidized to quinones by polyphenoloxidase enzymes, and the quinones subsequently react with N compounds to form brown humic substances.

3. Lignin is decomposed. Phenols released during decay also serve as source materials for humus synthesis.

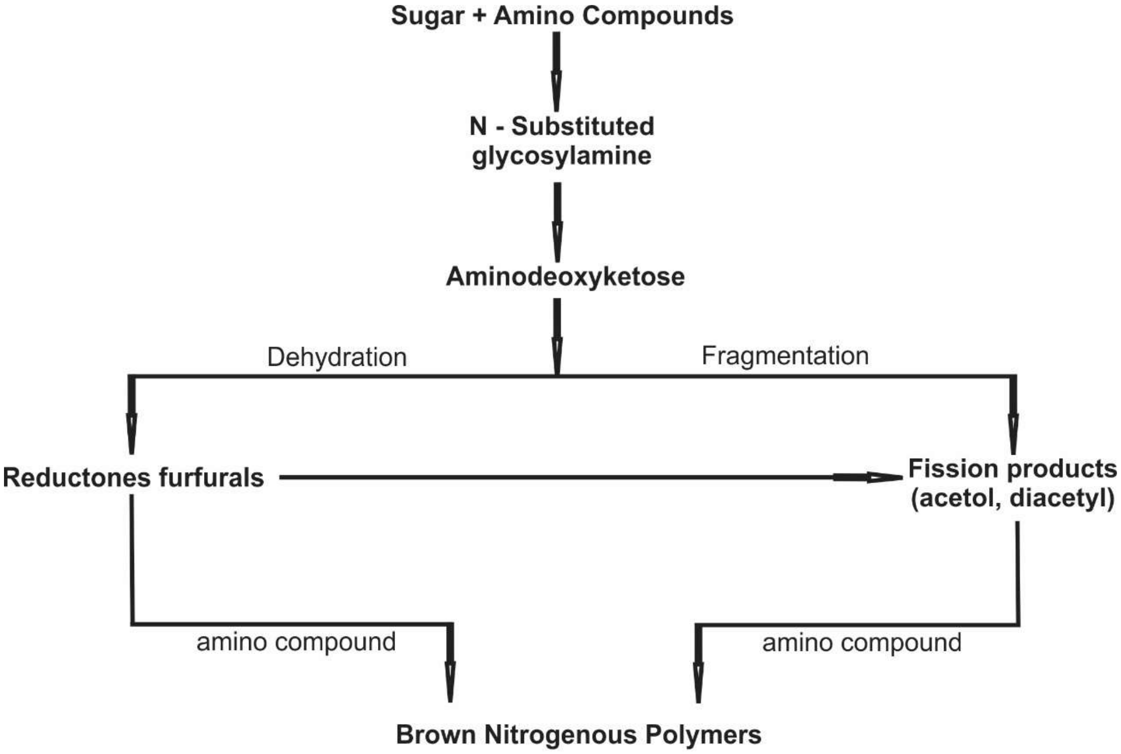
Pathway 4 - Sugar-amine condensation

The notion that humus is formed from sugars (pathway 4) dates back to the early days of humus chemistry. According to this concept reducing sugars and amino acids, formed as by-products of microbial metabolism, undergo non enzymatic polymerization to form brown nitrogenous polymers of the type produced during dehydratation of certain food products at moderate temperatures.

A major objection to this theory is that the reaction proceeds rather slowly at the temperatures found under normal soil conditions. However, drastic and frequent changes in the soil environment (freezing and thawing, wetting and drying), together with the intermixing of reactants with mineral material having catalytic properties, may facilitate condensation. An attractive feature of the theory is that the reactants (sugars, amino acids etc.) are produced in abundance through the activities of microorganisms.

The initial reaction in sugar-amine condensation involves addition of the amine to the aldehyde group of the sugar to form the n-substituted glycosylamine. The glycosylamine subsequently undergoes to form the N-substituted-1-amino-deoxy-2-ketose. This is subject to fragmentation and formation of 3-carbon chain aldehydes and ketones, such as acetol, diacetyl etc.; dehydration and formation reductones and hydroxymethyl furfurals.

All of these compounds are highly reactive and readily polymerize in the presence of amino compounds to form brown-colored products.



Sugar - Amine Condensation Theory (Stevenson 1982)

Properties of Humus

The tiny colloidal particles are composed of C, H, and O_2. 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. Humus has a very favorable effect on aggregate formation and stability. Impart black colour to soils. Cation exchange reactions are similar to those occurring with silicate clays

Clay - Humus Complex

It is evident that various clay minerals can bind humus substances through some linkages known as 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.

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.

Maintenance of Humus (Soil organic matter)

Maintenance of humus at a higher level is difficult. This is due to the reason that the loss of carbon from the soil increases as the organic matter content is raised. No attempt should be made to increase the organic matter content over that the soil plant-climate control mechanism can permit. There is a strong linkage between soil Nitrogen and soil organic matter.

To maintain adequate level of organic matter in the soil the N level should be maintained by inclusion of legumes in crop rotation and judicious application of N fertilizers. Loss of N from the soil is also to be minimized. To maintain the organic matter level continuous addition of organic materials is essential.

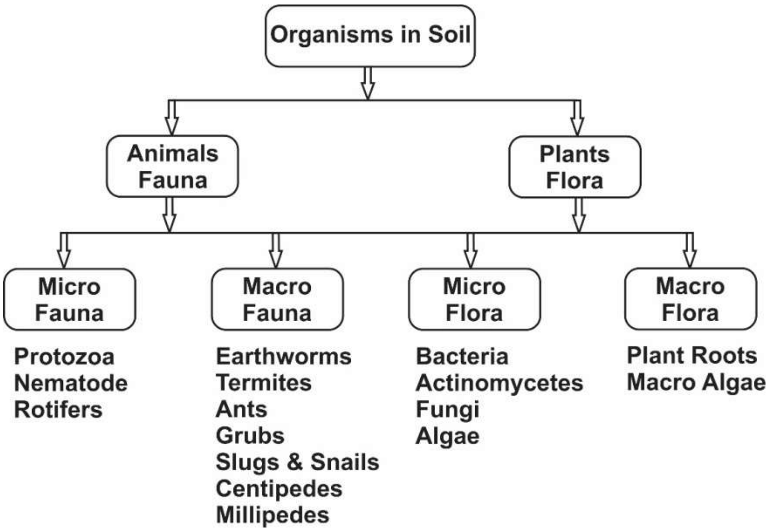
These include animal manure, organic wastes and crop residues. Incorporation of green manure will add to the organic carbon level of soil. Ensuring vigorous crop growth by removing constraints in crop production will result in addition of root and top residues to the soil. Conservation tillage (minimum tillage) should be followed to the extent possible. This will reduce the decay of residues. Keeping the land fallow to encourage natural vegetation is also advisable.

Lecture 31: Soil Biology- Soil organisms: macro and micro organisms, their beneficial and harmful effects, Soil enzymes

The soil organisms include both plants (flora) and animals (fauna) that are responsible for the degradation of organic materials and synthesis of newer products in soils. Most of the soil organisms belong to plant life and however, the role of animals is important during the early stages of organic matter decomposition.

The soil organisms which can be seen only with the aid of a microscope are called micro organisms and organisms of varying size from worms to ants, snails, rodents and mammals are classified as macro organisms.

Classification of microorganisms



Beneficial roles of soil organisms

Soil organisms are vital to the cycle of life earth. The soil flora and fauna play important roles in decomposing organic materials. During decomposition, soil organisms release essential plant nutrients in inorganic forms that can be absorbed by plant roots or leached from the soil. Without the activities of soil organisms, organic materials would accumulate as waste on the soil surface. They incorporate plant and animal residues into the soil and digest them, returning carbon dioxide to the atmosphere where it can be recycled through higher plants.

They create humus which is vital to good physical and chemical soil conditions. They even help control soil colour through oxidation-reduction reactions.

A soil microorganism plays an important role in completing the carbon cycle. Soil organisms also help in moving nutrients through soil layers and in fixing atmospheric nitrogen. Animals, particularly earthworms, are responsible for mechanically incorporating residues into the soil, and they leave open channels through which water and air can flow.

Bacteria

Bacteria are single-celled organisms, and are the most numerous numbers of the soil, ranging from 100 million to 3 billion in a gram. They are capable of rapid reproduction by binary fission in favourable conditions

One bacterium is capable of producing 16 million or more in just 24 hours. Most soil bacteria live in close contact with to plant roots and are often referred to as rhizobacteria. The majority of the beneficial soil-dwelling are aerobic bacteria and those that are anaerobic cause putrefaction of dead organic matter. Aerobic bacteria are most active in a soil that is moist and neutral soil pH, and where there is plenty of food (from organic matter) is available.

The important roles that bacteria play are

Nitrification

It is important process where the bacterium transforms nitrogen in the form of ammonium and then into nitrates, which are available to growing plants.

Nitrogen fixation

In this process, nitrogen fixing bacteria convert nitrogen from the atmosphere into nitrogen-containing organic substances. This is carried out by free-living nitrogen fixing bacteria in the soil such as Azotobacter, or by symbiotic bacteria such as rhizobia which form colonies in nodules on the roots of peas, beans and related species. The process of nitrogen fixation contributes additional nitrogen into biological cycle.

Denitrification

In this process nitrates and nitrites in the soil converted into nitrogen gas or nitrous oxide or nitric oxide. While nitrogen fixation converts nitrogen from the atmosphere into organic compounds, denitrification process returns an approximately equal amount of nitrogen to the atmosphere. Denitrifying bacteria are mostly anaerobic including *Achromobacter* and *Pseudomonas*.

Actinobacteria

Actinobacteria helps in the decomposition of organic matter and in humus formation, and their presence is responsible for the sweet "early" smell which in association with a good healthy soil. They require plenty of air and a pH between 6.0 and 7.5, but are more tolerant of dry conditions than most other bacteria and fungi.

Viruses

Viruses are sub-microscopic organisms with a relatively simple structure, a DNA or RNA (ribonucleic acid) core and an encasing protein coat. Viruses are *intra* cellular parasites that do not respire or metabolize and *must* have a biological host to reproduce. Soils may harbor many viruses, but it is believed that they do not directly affect soil characteristics.

Little is known about pathogenic turfgrass viruses. However St. Augustinegrass decline and centipedegrass mosaic have been attributed to panicum mosaic virus. The virus is spread through sap on mower blades. Mowing when the turfgrass is dry has proven to reduce the spread of the virus. To best control this pathogen, use resistant St. Augustinegrass varieties. Fortunately, many of the improved varieties currently on the market are resistant.

Fungi

A gram of garden soil contains around one million fungi, such as yeasts and moulds. Fungi have no chlorophyll, not able to photosynthesise and they can't use atmospheric carbon dioxide as a source of carbon. The most important saprotrophic fungi live on dead or decaying organic matter, breaking it down into forms are available to higher plants.

Some fungi are able to break down cellulose and lignin. Many fungi are parasitic, often causing disease to their living host plant, although some have beneficial relationships with living plants.

Mycorrhizae

Fungi that live symbiotically with living plants, that is beneficial to both, are known as *Mycorrhizae* (from *myco* meaning fungal and *rhiza* meaning root). The mycorrhiza obtains the carbohydrates from the root, in return provides the plant with nutrients including nitrogen moisture. Later the plant roots will also absorb the mycelium into its own tissues.

Beneficial mycorrhizal associations are to be found in many of our edible and flowering crops.These include at least 80% of the *Brassica* and *Solanum* families (including tomatoes and potatoes), as well as the majority of tree species, especially in forest and woodlands.

Larger organisms

Larger soil-inhabiting animals and insects include earthworms, grubs, slugs, mole crickets, ants and mites. These organisms mostly affect turf by churning the soil while feeding, which aids aggregation, water movement, aeration and thatch degradation.

Not only is the soil composed of sand, silt, clay, air and water, it is also a living, "breathing," dynamic environment containing a broad diversity of organisms. These serve vital functions in the rootzone - most are beneficial but some are detrimental. Our understanding of soil micorbiology is in the early stages. As we learn more, there's little doubt we'll be able to put this knowledge to use to grow better turf.

Harmful effects of soil organisms on higher plants

In general, the microflora causes the most devastating effect on higher plants. Although bacteria and actinomycetes contribute their quota of plant diseases, fungi are responsible for most of the common soil-borne disease of crop plants. Soils are easily infested with disease organisms. These are transferred to the soil through farm implements, plant and even through manure from animals that were fed with infected plants.

Another harmful effect by soil microbes to higher plants is by competition for available nutrients. Soil organisms can quickly absorb essential nutrients into their own bodies, so that the slowly growing higher plants can use only what is left. The completion is usually greatest for total N and similar competitions occur for P, K, Ca and even the micro nutrients.

Under poor drainage, the soil micro organisms may deplete the oxygen supply. The shortage of oxygen may affect plants adversely in two ways. First, the plant roots require certain minimum amount of oxygen for normal growth and nutrient uptake. Second oxidized forms of several elements, including N, S, Fe and Mn, will be chemically reduced by further microbial action. In the case of nitrogen and sulfur, some of the reduced forms are gaseous and may be lost to the atmosphere. In soils that are quite acid, the reduction of iron and manganese may produce soluble forms of these elements in toxic quantities.

Importance of Soil Microorganisms:

1. Decomposition of resistant components of plant and animal tissue

2. Role in microbial antagonism
3. Participate in humus formation
4. Predator of nematodes
5. Surface blooming reduces erosion losses
6. Improve soil structure
7. Involved soil structure
8. Maintenance of biological equilibrium

Important characters of soil microorganisms

- **Aerobes** - Organisms that require free oxygen (O_2) for respiration.
- **Anaerobes** - Organisms that do not require free oxygen, but use oxygen derived from various compounds for respiration. Anaerobes are associated with saturated soil conditions.
- **Autotrophs** - Organisms (plants and other types) that use the sun to produce their own energy and obtain carbon from atmospheric carbon dioxide (CO_2) or the breakdown of inorganic substances.
- **Endophytes** - Root inhabiting fungi that form beneficial relationships with plants; have been shown to improve stress tolerance and insect resistance.
- **Facultative anaerobes** - Organisms that can survive in the presence or absence of oxygen.
- **Facultative parasites** - Organisms that can survive as saprophytes but may attack plants if a suitable host is available.
- **Heterotrophs** - Organisms that obtain energy by consuming soil organic materials.
- **Parasites** - Disease-causing organisms that feed on living plants.
- **Saprophytes** - Organisms that live off of dead organic matter.
- **Symbiosis** - A relationship in which soil organisms and plants each benefit from the presence of the other.

Soil Enzymes

Soil enzymes are a group of enzymes whose usual inhabitants are the soil and arecontinuously playing an important role in maintaining soil ecology, physical and chemical properties, fertility, and soil health. These enzymes play key biochemical functions in the overall process of organic matter decomposition in the soil system. They are important in catalyzing several vital reactions necessary for the life processes of micro-organisms in soils and the stabilization of soil structure, the decomposition of organic wastes, organic matter formation, and nutrient cycling, hence playing an important role in agriculture.

All soils contain a group of enzymes that determine soil metabolic processes which, in turn, depend on its physical, chemical, microbiological, and biochemical properties. The enzyme levels in soil systems vary in amounts primarily due to the fact that each soil type has different amounts of organic matter content, composition, and activity of its living organisms and intensity of biological processes. In practice, the biochemical reactions are brought about largely through the catalytic contribution of enzymes and variable substrates that serve as energy sources for microorganisms.

Kinds of Enzymes

1. Constitutive

Always present in nearly constant amounts in a cell (not affected by addition of any particular substrate genes always expressed.) (E.g. **Pyrophosphatase**).

2. Inducible

Present only in trace amounts or not at all, but quickly increases in concentration when its substrate is present. (E.g. **Amidase**)

Both enzymes are present in the soil.

Origin of Soil Enzymes

1. Microorganisms -Living and dead
2. Plant Roots and Plant Residues
3. Soil Animals

State of Enzymes in Soil

Role of Clays

a. Most activity associated with clays.

b. Increases resistance to proteolysis and microbial attack

c. Increases the temperature of inactivation.

Role of Organic Matter

a. Humus material provides stability to soil nitrogen compounds

b. Enzymes attached to insoluble organic matrices exhibit pH and temperature changes.

c. Inability to purify soil enzymes free of soil organic matter (bound to O.M)

Role of O.M. - Clay Complex

a. Lignin + bentonite (clay) protect enzymes against proteolitic attack, but not bentonite alone.

b. Enzymes are bound to organic matter which is then bound to clay.

Role of soil enzymes

Soil enzymes increase the reaction rate at which plant residues decompose and release plant available nutrients. The substance acted upon by a soil enzyme is called the substrate. For example, glucosidase (soil enzyme) cleaves glucose from glucoside (substrate), a compound common in plants. Enzymes are specific to a substrate and have active sites that bind with the substrate to form a temporary complex.

The enzymatic reaction releases a product, which can be a nutrient contained in the substrate. Sources of soil enzymes include living and dead microbes, plant roots and residues, and soil animals. Enzymes stabilized in the soil matrix accumulate or form complexes with organic matter (humus), clay, and humus-clay complexes, but are no longer associated with viable cells.

It is thought that 40 to 60% of enzyme activity can come from stabilized enzymes, so activity does not necessarily correlate highly with microbial biomass or respiration. Therefore, enzyme activity is the cumulative effect of long term microbial activity and activity of the viable population at sampling.

However, an example of an enzyme that only reflects activity of viable cells is dehydrogenase, which in theory can only occur in viable cells and not in stabilized soil complexes. Enzymes respond to soil management changes long before other soil quality indicator changes are detectable. Soil enzymes play an important role in organic matter decomposition and nutrient cycling. Some enzymes only facilitate the breakdown of organic matter (E.g. **Hydrolase** and **glucosidase**), while others are involved in nutrient mineralization (E.g. **Amidase, urease, phosphatase** and **sulfates**).

With the exception of phosphatase activity, there is no strong evidence that directly relates enzyme activity to nutrient availability or crop production. The relationship may be indirect considering nutrient mineralization to plant available forms is accomplished with the contribution of enzyme activity.

Functions of soil enzymes

Enzyme	Organic Matter Substances Acted On	End Product	Significance	Predictor of Soil Function
Beta glucosidase	Carbon compounds	Glucose (sugar)	Energy for microorganisms	Organic matter decomposition
FDA hydrolysis	Organic matter	Carbon and various nutrients	Energy and nutrients for microorganisms,	Organic matter decomposition nutrient cycling

			measure microbial biomass	
Amidase	Carbon and nitrogen compounds	Ammonium (NH_4)	Plant available NH_4	Nutrient cycling
Urease	Nitrogen (urea)	Ammonia (NH_3) and carbon dioxide (CO_2)	Plant available NH_4	Nutrient cycling
Phosphatase	Phosphorus	Phosphate (PO_4)	Plant available P	Nutrient cycling
Sulfatase	Sulfur	Sulfate (SO_4)	Plant available S	Nutrient cycling

Specific problems that might be caused by poor function

Absence or suppression of soil enzymes prevents or reduces processes that can affect plant nutrition. Poor enzyme activity (E.g. **Pesticide degrading enzymes**) can result in an accumulation of chemicals that are harmful to the environment; some of these chemicals may further inhibit soil enzyme activity.

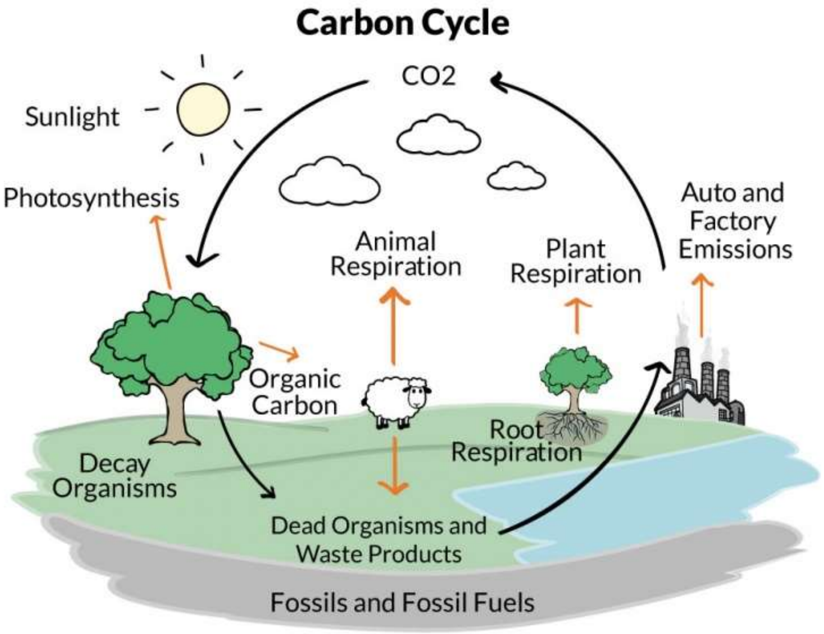
Application of Soil Enzymes

- Correlation with soil fertility.
- Correlation with microbial activity.
- Correlation with biochemical cycling ofvarious elements in soil (C, N, S).
- Degree of pollution (heavy metals, SO).
- To assess the successional stage of anecosystem.
- Forensic purposes.
- Rapid degradation of pesticides.
- Disease studies.

Lecture 32: Soil carbon sequestration and carbon trading

Carbon cycle

The carbon cycle is the process in which carbon moves between different parts of the Earth and its living organisms. This makes it really important to life on our planet, as it allows the carbon to be continually reused and recycled.



Carbon is a common constituent of all organic matter (plant and animal residues). Carbon is continually being fixed into organic form by photosynthetic organisms under the influence of light and once bound, the carbon becomes unavailable for use in the generation of new plant life. Therefore, it is essential for the carbonaceous materials to be decomposed and returned to the atmosphere for the survival of the higher organisms.

The decomposition of plant and animal remains in soil constituents a basic biological process in that carbon (C) is recirculated to the atmosphere as CO_2, nitrogen (N) is made available as NH_4 and NO_3 and other associated nutrient elements like P, S, Fe, Mn, Cu and Zn etc, appear in plant available forms. In the process, part of the nutrient elements is assimilated by micro organisms and incorporated into microbial tissues (Soil biomasses).

The conversion of organic forms of C, N, P and S into inorganic or mineral forms is called mineralization and the conversion of inorganic forms of those elements to their organic forms is known as immobilization. The carbon cycle revolves about CO_2 and its fixation and regeneration.

Chlorophyll containing plants utilize the gas as their sole carbon source, and the carbonaceous matter thus synthesized serves to supply the animal world with preformed organic carbon. Upon the death of the plant or animal, microbial world with performed organic carbon. Upon the death of the plant or animal, microbial metabolism assumes the dominant role in the cyclic sequence.

The dead tissues undergo decay and are transformed into microbial cells and a large amount of heterogeneous carbonaceous compounds together known as humus or as the soil organic fraction. The cycle is completed and carbon made available with the final decomposition and production of CO_2 from humus and the rotting tissues.

C: N ratio

The ratio between the nitrogen content in the microbes and in the organic residues and to the carbon content is called as C:N ratio. When fresh plant residues are added to the soil they are rich in carbon and poor in N. This results in wider C:N ratio (40:1) decomposition of the organic matter in the soil changes to humus resulting in a narrow C:N ratio (10:1).

When materials high in carbon are added to the soil the microbial population increase due to the plentiful supply of food material. A lot of CO_2 is released. During this process the micro organisms utilize the soil N for their body build up and there is a temporary block of N. When the decomposition of fresh organic residues reaches to the stage where the C:N ratio is 20:1 there is an increase in the availability of N.

Important points

- The C: N ratio of cultivated soils ranges from 8:1 to 15:1 and average ratio is 10:1 to 12:1.
- Legumes and Farm manure: 20:1 - 30:1.
- Straw : 100:1
- Saw Dust : 400:1
- In micro organisms: 4:1 to 9:1
- There exists a relationship between the organic matter and N content. Soil carbon has a definite proportion of the organic matter. Hence the C to N ratio in soils is fairly constant.
- The C:N ratio is lower in soils of arid regions than humid regions.
- C: N ratio is smaller in subsoils.

Importance of C:N ratio

It is the intimate relationship between organic matter and nitrogen contents of soils. The ration of the weight of organic carbon to the weight of total nitrogen in a soil or organic material is known as C:N ratio. The importance of C:N ratio in controlling the available nitrogen, total organic matter and rate of organic materials decomposition is recognized in developing appropriate soil management practices.

Ratio in soils

The C: N ratio of soil is one of its characteristics equilibrium values, the figure for humus being roughly 10:1 although values from 5:1 to 15:1 are generally found in most arable soils. The critical ratio (10:1) is a reflection of the dynamic equilibrium that results from the dominating presence of a microbial population, the ratio being similar to the average chemical composition of microbial cells.

As a rule microbial protoplasm contains 5 to 15 parts of carbon to 1 part of nitrogen, but 10:1 is a reasonable average for the predominant aerobic of fractions resistant to further decomposition can modify the C:N equilibrium value of humus. Such C:N ratio of soils can vary with the climates i.e. rainfall, temperature etc.

The C:N ratio is generally lower in warmer (arid soil conditions) regions than that of cooler ones (humid soil conditions) inspite of having the same rainfall under both the soil conditions. The ratio is also narrower for sub soils as compared to surface soil horizons.

Ratio in plants and microbes

The carbon and nitrogen (C:N) ratio in plant material is variable and ranges from 20:1 to 30:1 to legumes and farm yard manure to as high as 100:1 in certain straw residues. On the other hand C:N ratio of the bodies of micro organisms is not only more constant but much narrower between 4:1 and 9:1. Bacterial tissue in general is somewhat richer in protein than fungi and consequently has a narrow C:N ratio.

Therefore, it is usually found that most of the applied fresh organic materials in soils carry large amounts of carbon with relatively very small amounts of total nitrogen. As a result, the value C:N ratio for soils are in between those of higher plants and the microbes.

Effects of the C:N ratio

The C:N ratio in soil organic matter is important for two major reasons

The keen competition for available nitrogen results when organic residues of high C:N ratio are added to soils and because this C:N (10:1) is relatively constant in soils, the maintenance of carbon and hence soil organic matter is dependent to no small degree on the level of soil nitrogen.

So the C: N ratio obviously has practical implications on the availability of nitrogen in soils as well as in plants. As for example, large amount of fresh organic materials having wide C: N ratios (50:1) are incorporated into the soil under favourable soil conditions for decomposition.

A rapid change will found. The heterotrophic micro organisms like bacteria, fungi and actinomycetes become active and increases their population with the production of large amounts of CO_2. Under these conditions, nitrate nitrogen (NO_3-N) disappears from the soil because of the urgent needs by the micro organisms, and for the time being, little or no nitrogen is available to plants. As the decomposition precedes, the C:N ratio on the organic materials decreases with the loss of carbon and conservation of nitrogen.

Reasons for the stabilization of C:N ration to a constant value

As the decay process proceeds, both carbon and nitrogen are now subject to loss as CO_2 and nitrates respectively. It is only a question of time until their percentage rate of disappearance from the soil becomes more or less the same, *i.e*, the percentage of the total carbon being lost. At this point the C:N ratio becomes more or less constant 10:1 to 12:1 always being somewhat greater than that of the ratio in microbial tissue.

C:N ratio and the level of organic matter

Since carbon and nitrogen are reduced to almost a definite ratio, the amount of soil nitrogen largely determines the amount of organic carbon present when stabilization occurs. Thus, the greater the amount of nitrogen present in the original organic material, the greater will be the possibility of an accumulation of organically bound carbon.

Since a definite ratio (1:17) exists between the organic carbon and the soil humus, the amount of organic matter to be maintained in any soil is largely conditional on the amount of organic nitrogen present. The ratio between nitrogen and organic matter is thus constant (organic matter: nitrogen, 20:1 for most soils).

2. Carbon sequestration

Carbon sequestration is a method by which carbon-rich gases are absorbed from the atmosphere, converted into other products, which then stored in plants, soil, oceans, or geological formations. This process has occurred naturally since billions of years, and in recent times humans are trying to utilize it to reduce the amount of carbon dioxide released by industrial activities.

Role of Carbon Sequestration

Carbon occurs in various forms on the planet; in the air, soil, plants and animal matter, and the oceans. In the atmosphere, it occurs in the form of greenhouse gases like carbon dioxide. These gases form an envelope around the surface of the Earth, and keep it warmer than it would otherwise be, by trapping infrared radiations, thus helping support life.

This is similar to a greenhouse, in which heat is trapped inside to enhance plant growth. Carbon dioxide is released by plants and animals when they breathe and during their decomposition, as well as from soil and oceans. However, if its concentration in the atmosphere becomes excessive, it can raise the Earth's temperatures to dangerous levels.

To maintain the balance of carbon (called the carbon cycle), carbon dioxide must be absorbed from the air and stored in other forms on Earth, in 'sinks'. Natural carbon sinks are plants, soil, and oceans.

However, due to increasing pollution levels, the carbon balance has been thrown off balance, leading to an excess accumulation of carbon dioxide in the air. There are fears that this may lead to climate change, by warming up the Earth. Therefore, in order to control the level of carbon dioxide and prevent climate change, more emphasis needs to be given to carbon sequestration.

Carbon Sequestration Methods

Plants absorb carbon dioxide from the air via pores in their leaves, called stomata, and perform photosynthesis to convert this gas into carbon, stored in their bodily tissues like roots, stems, and leaves.

Animals which consume plants, called herbivores, transfer the stored carbon into the soil via their excreta. Plants and animals, on dying, are decomposed by soil microbes, which also release carbon. Soil carbon is stored in the form of decomposing plant and animal fragments, microbes such as nematodes, and fungi, and in the form of a mineral-carbon composite called humus.

In peat bogs, plant matter is decomposed in the absence of air, leading to more carbon being stored in it than is released to the atmosphere. Over time, large carbon deposits form coal reserves, such as those used by man. Most of the carbon dioxide is sequestered in oceans by plankton and aquatic plants which absorb it via photosynthesis. When these die, they sink to the bottom and decompose, releasing carbon deposits.

Artificial Processes

Carbon sequestration can be enhanced by changing land use practices, especially the pattern of cropping, such as, by growing cover plants in the field between two seasons, or between rows of the main crop, by no till farming, use of compost and manure, crop rotations, and slope protection by planting trees. Fast growing trees can be grown specifically for trapping carbon dioxide on barren lands.

Since it's desirable that they should not release the carbon back to the atmosphere by decomposition, long-lived trees, especially those that live above 100 years, are preferred. Protection of existing forests, restoration of wetlands, better management of grasslands, reducing grazing, and reducing chances of forest fires can go a long way in maintaining the natural rate of carbon sequestration.

Carbon dioxide produced in power plants can be trapped, liquified, and injected into aging oil fields, which pressurizes more oil to enter the wells, a process called Enhanced Oil Recovery (EOR). Earlier, water was injected for the same purpose. Liquefied carbon dioxide can be injected deep inside mineral deposits, such as magnesium and calcium silicates, so that it reacts to form solid carbonates which remains underground.

This process is still in the planning stages, and would increase power plant energy consumption by 60 - 180%. Carbon dioxide gas can be injected deep into oceans, where the high pressure causes it to get liquefied or solidified and slowly dissolve into the water, or stay stabilized.

Another method for ocean sequestration is to provide aquatic plankton with artificial nutrients (such as iron filings), so that they use it to perform photosynthesis, while absorbing enhanced amounts of carbon dioxide from the air. Burying dead trees and other organic material deep within the ground is another sequestration method to retain the carbon inside the ground, which otherwise would have been released as carbon dioxide gas after conventional decomposition.

Advantages

Carbon sequestration prevents climate change by removing excess amounts of carbon dioxide from the air. The gas can be easily liquefied and transported by pipelines, which makes deep injection convenient. Along with removing excess pollutants from the air, deep injection of carbon dioxide also enhances extraction of fuels like oil and methane from their deposits.

Renewable energy sources like solar and wind energy are not expected to completely replace coal in the near future. Still, it may be possible to reduce emissions by 80 to 85%, using carbon sequestration, even while using fossil fuels. There are no reports yet of carbon dioxide leaking out from where it was injected, and this is unexpected to happen for 1,000 years.

3. Carbon trading

Carbon trading is an approach used to control carbon dioxide (CO_2) pollution by providing economic incentives for achieving emissions reductions. It is sometimes called **cap**

and trade or **carbon emissions** trading.Carbon trading is administered by a central authority such as a government or international organization which sets a limit or cap on the amount of CO_2 that can be emitted.

Companies or other groups are issued permits that require them to hold allowances (or credits) in order to emit an equivalent amount of CO_2. The total amount of allowances and credits cannot exceed the cap, limiting total emissions to that level. Companies that need to increase their allowance must buy credits from those who pollute less.

The transfer of allowances is referred to as a trade. The buyer therefore pays to pollute, while the seller is financially rewarded for reducing CO_2 emissions. In theory, those that can easily reduce emissions most cheaply will do so.

History

Carbon trading began in response to the Kyoto Protocol, signed by 180 countries in 1997. The Kyoto Protocol, signed by 180 countries in 1997, called for 37 industrialized countries to reduce their greenhouse gas emissions between the years 2008 to 2012 to levels that are 5% lower than those of 1990.

Article 17 of the Kyoto Protocol established emissions trading by allowing countries that have emission units to spare (emissions permitted to them but unused) to sell this excess capacity to countries that are over their emissions limits. In effect, this created a new commodity in the form of emissions and created a carbon market.

Since CO_2 is the principal greenhouse gas, emissions trading effectively became carbon trading. The units which may be transferred under Article 17 emissions trading, each equal to one tonne of CO_2 equivalent.

An assigned amount unit (AAU) issued by an Annex I Party on the basis of its assigned amount pursuant to Articles 3.7 and 3.8 of the Protocol. A removal unit (RMU) issued by an Annex I Party on the basis of land use, land-use change and forestry (LULUCF) activities under Articles 3.3 and 3.4 of the Kyoto Protocol.

An emission reduction unit (ERU) generated by a joint implementation project under Article 6 of the Kyoto Protocol. A certified emission reduction (CER) generated from a clean development mechanism project activity under Article 12 of the Kyoto Protocol. Transfers and acquisitions of these units are to be tracked and recorded through the registry systems under the Kyoto Protocol.

Carbon Market

Carbon emissions trading has been steadily increasing in recent years. According to the World Bank's Carbon Finance Unit, 374 million metric tonnes of carbon dioxide equivalent (tCO_2e) were exchanged through projects in 2005, a 240% increase relative to 2004 (110 $mtCO_2e$).

In 2008, the carbon market was valued at $47 billion, while in 2009 the World Bank estimated its value at $126 billion. On June 9, 2005, twenty three multinational corporations from the G8 Climate Change Roundtable released a statement advocating market-based solutions to climate change.

The group, including Ford, Toyota, British Airways, BP and Unilever, called on governments to establish "clear, transparent, and consistent price signals" through "creation of a long term policy framework" that would include all major producers of greenhouse gases. By December 2007 this group had grown to encompass 150 global businesses.

Carbon Credit

This concept was brought to help reduce carbon dioxide, the biggest contributor to greenhouse gases leading to global warming. A carbon footprint is measured of each company or individual or country using various carbon calculators, or a central authority is assigned for this purpose.

As per the calculations, a certain number of credits are assigned to each company for a certain period of time. If the company is successful in controlling its carbon emissions under the assigned credit value, it is permitted to sell those credits in the carbon market for monetary gain.

On the other hand, those companies or countries who fail to control their emissions under the required limit, need to purchase these carbon credits as a offset for the extra carbon they emitted. In a nutshell, having one carbon credit permits the holder to emit one ton of carbon dioxide.

Under the Kyoto Protocol, there are three exchange mechanisms for carbon credits - Joint implementation, Clean Development Mechanism (CDM), and International Emissions Trading (IET). These mechanisms allow developed countries or companies with high carbon footprints to acquire these credits, but all differ according to the clauses of buying and selling them. These mechanisms aim at achieving carbon neutrality, which means there should be net zero carbon footprint.

What is a 'Carbon Trade'

Carbon trading is an exchange of credits between nations designed to reduce emissions of carbon dioxide.Carbon trading is also referred to as carbon emissions trading. Carbon emissions trading accounts for most emissions trading.

Why We Have the Carbon Trade

When countries use fossil fuels, and produce carbon dioxide, they do not pay for the implications of burning those fossil fuels directly. There are some costs that they incur, like the price of the fuel itself, but there are other costs, not included in the price of the fuel.

These are known as externalities. In the case of fossil fuel usage, often these externalities are negative externalities, meaning that the consumption of the good has negative effects on third parties. These externalities include health costs, (like the contribution that burning fossil fuels makes to heart disease, cancer, stroke, and lung diseases) and environmental costs, (like environmental degradation, pollution, climate change and global warming).

Interestingly, research has found that, often, the burdens of climate change most directly effect countries with the lowest greenhouse emissions. So, if a country is going to burn fossil fuels, and produce these negative externalities, the thinking is that they should pay for them.

The carbon trade originated with the 1997 Kyoto Protocol, with the objective of reducing carbon emissions and mitigating climate change and future global warming. At the time, the measure devised was intended to reduce overall carbon dioxide emissions to roughly 5% below 1990 levels by between 2008 and 2012.

How It Works

Basically, each country has a cap on the amount of carbon they are allowed to release. Carbon emissions trading then allows countries that have higher carbon emissions to purchase the right to release more carbon dioxide into the atmosphere from countries that have lower carbon emissions.

The carbon trade also refers to the ability of individual companies to trade polluting rights through a regulatory system known as cap and trade. Companies that pollute less can sell their unused pollution rights to companies that pollute more. The goal is to ensure that companies in the aggregate do not exceed a baseline level of pollution and to provide a financial incentive for companies to pollute less.

Carbon Trading Implementation Mechanisms

Emission Trading (ET)

Countries whose emissions are less than their assigned amount can sell the excess amount to countries whose emissions have exceeded their assigned amount

The Assigned amounts can be defined as a tradable allowances, or commodity, and this free market is known as the “**CARBON MARKET**."

Clean Development Mechanism (CDM)

Developed countries can fund emission reduction projects (E.g. Solar energy, wind energy and other green technologies) in developing nations that did not sign Kyoto Protocol.

In exchange, the developed countries earn legally recognized emission credits called CERs (Certified Emission Reduction) to offset their emission obligations.

Joint implementation (JI)

Developed countries can implement emission reduction projects in another developed or developing country and earn Emission Reduction Units (ERUs). ERUs can be used to meet the carbon allowance or can be sold in the market.

Advantages of Carbon trading

i. Reduction in green house gas emission

Stringency in the cap or the upper threshold limit is contributing to lower emission over the years.

ii. Source of revenue for developing nations

Developing nations can earn revenue by selling carbon credits to countries with more fossil fuel demand.

iii. Supports a free market system

The carbon trade market is without any economic intervention and regulation by government except to regulate against force or fraud.

iv. Impetus for Alternative sources of energy or green technology

Threshold limits encourages industries to harness alternative sources of energy and invest in green technology globally or in indigenous research.

Disadvantages of Carbon Trading

i. Right to pollute

Industries in the ratified nations are purchasing legal rights to pollute the atmosphere

ii. Slow process

Industries are opting the easy way– purchase more allowances than implementing greener technologies.

iii. Lack of centralized system or global framework

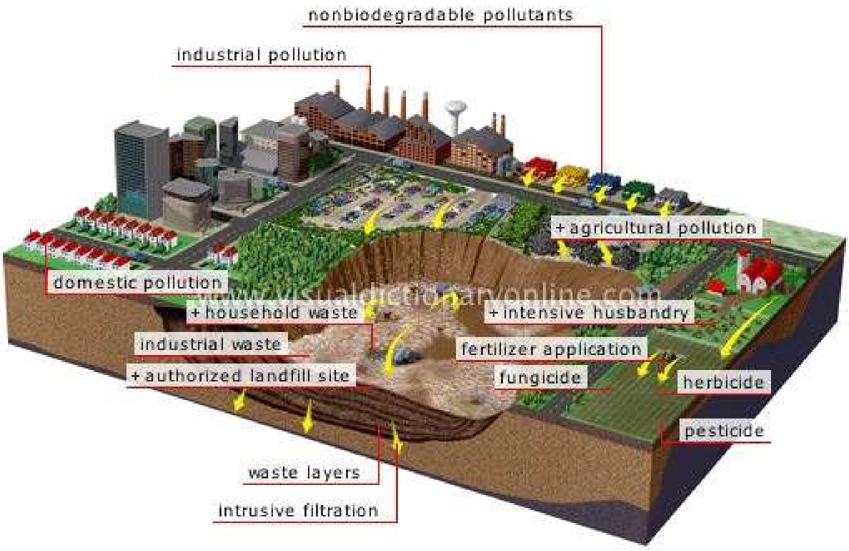
Absence of a centralized and accepted global standards/act are missing.

iv. No effective carbon reduction in the atmosphere

Leads to carbon reduction in one place and results in carbon emission in some other place.

Lecture 33 & 34: Soil pollution - behaviour of pesticides and inorganic contaminants, Prevention and mitigation of soil pollution

Soil contamination or soil pollution as part of land degradation is caused by the presence of chemicals or other alteration in the natural soil environment. It is typically caused by industrial activity, agricultural chemicals, or improper disposal of waste.



Source if pollution

Soil pollution usually originates from the development of industry, intensive agriculture associated with modern systems of cultivation like use of high analysis chemical fertilizers, use of pesticides, use of sewage, sludge, city composts and other industrial wastes etc.

The contaminants produced by industrial processes 'reach the soil by three ways

(a) Through the air

Gaseous and particulate contaminants are emitted from chimneys and exhausts or are deflated from spoil heaps and are blown away by the wind ultimately to sink or be washed by rain to the soil environment.

(b) Through drainage system

Industrial effluents and water drainage from spoil and rubbish heaps either washes direct on to nearby fields or enters the local streams and rivers and ultimately enter into the soil.

(c) Through direct mechanical or gravitational effects on the soil

Solid materials like refuses from mines, slurries or sewage sludges etc. Can be dumped directly on to the agricultural land and thereby creates soil pollution. Once pollutants enter and are incorporated into the soil, their concentrations in soil are continuously increasing and accumulating as toxic to all forms of life like plants, micro organims, human beings etc.

Kinds of soil Pollution

There are generally five different kinds of pollution namely

i. Pesticides

ii. Inorgnic contaminants or pollutants

iii. Organic wastes

iv. Fertilizers and other salts

v. Radioactive materials

i. Pesticide pollution

Pesticides are a cause of pollution, affecting land and water in particular. The problem is huge and growing. Pesticides decrease biodiversity in the soil because they do not just kill the intended pest; they often kill many of the other small organisms present. Soil fertility is affected in other ways, too. When pesticides kill off most of the active soil organisms, the complex interactions which result in good fertility break down.

Plants depend on millions of bacteria and fungi to bring nutrients to their rootlets. When these cycles are disrupted plants become more dependent upon exact doses of chemical fertilizers at regular intervals. Even so, the fantastically rich interactions in healthy soil cannot be fully replicated by the farmer with chemicals.

Pesticides, legally termed economic poisons, include insecticides, fungicides, herbicides, rodenticides and nematocides and many other chemical like disinfectants, antibiotics, defoliants, chemosterilants and juvenile hormones etc. Many toxic pesticides of group organo phosphates are active in the environment only a few days or a week or so.

A major concern particularly about chlorinated hydrocarbons like DDT, aldrin, heptachlor, etc. Is their persistence in soil and their movement into water streams through soil erosion and entry in the food chain of various wild life.

Persistence of various pesticides in Soil

Name of Pesticides	Approximate persistence in soils
Heptachlor	9 years
Aldrin, Dieldrin	9 years
DDT	10 years
BHC (Benzene hexachloride)	11 years
Chlorodane	12 years
Diuron	19 months
Simazine	17 months
Atrazint	18 months

Monuron	36 months
2,4-D	14-30 days

ii. Inorganic contaminants

The tract elements concentrated in an irregular way are usually metals especially the heavy metals (density > 6 g cc^{-1}). Some are essential for life process but all are toxic to organisms at higher concentrations. Seventy six (76) trace elements all of which are potential pollutants.

There are number of inorganic contaminants including heavy metals like, Hg, Cd, Pb, As, Ni, Cu, Zn, Mn, etc. Among these Cd and As are extremely poisonous. Mercury, Pb and Ni are moderately poisonous and Cu, Zn and Mn are relatively lower in toxic.

Effects of heavy metal pollution in the soil-water-air-organisms ecosystem

Different beneficial microbial processes in soil and interaction are affected or sometimes altered by the presence of high concentrations of different heavy metals like Cd, Pb, Hg, As, Ni etc. in soil. Heavy metals are disturb the nutrient balance in soil and thereby affect the bio-chemical processes and crop yields. Such as exchange reactions, chelation by organics, absorption by colloidal oxides and hydroxides of iron and manganese etc.

Concentration (mg kg^{-1}) of Heavy metals in soil and plants

Heavy metal	Soils (mg kg^{-1})	Plants (mg kg^{-1})
Arsenic (As)	0.1-40	0.1-5
Cadmium (Cd)	0.1-7	0.2-0.8
Lead (Pb)	2-200	0.1-10
Nickel (Ni)	10-1,000	1

Zinc, Copper, Manganese and Nickel

These elements affected the pH, organic matter content, and redox status of the soil. The tendency of these cations is to form 'chelate' in the presence of organic matter which influences their behavious in soils. The relative strength of chelation is generally Copper > Nickel > Zinc > Manganese.

Lead

The maximum amount of lead was found to be concentrated on the soil surface. Liming reduces the availability of lead and its uptake by plants. Therefore, lead does not create so much of problem in the phytotoxicity.

Arsenic

Arsenic behaves in soils like that of phosphate. Arsenic present in an anionic form (AsO_4^{3-}) and it is adsorbed by oxides and hydrous oxides of Fe and Al. The toxicity of arsenic can be minimized with the application of sulphate salts to Zn, Fe and Al which can form their insoluble arsenate compounds unavailable to plants.

Cadmium

Cadmium is extremely phytotoxic as well as poisonous to human beings. Besides its phytotoxic effect, high levels of Cd in soils may disturb the nutrient balance in soil. Cadmium is most mobile in acid soil whereas in alkaline soil, Cd is immobile.

Cadmium seemed less strongly associated with high molecular weight organics and hydrous oxides of Fe and Al and thus it is readily accumulated by plants grown in Ca contaminated soils.

The phytotoxic effect of Cd can be reduced with the application of organic matter and also with the application of micronutrient fertilizer like Zn and Cu which may interact with Cd rendering unavailable to plants.

Mercury

Metallic mercury is first oxidised by the following chamical reaction in the sediment layer of lakes and streams. Mercury is not readily absorbed by plants from soils unless it is in the methyl mercury form.

iii. Organic wastes

Sewage effluents, agricultural wastes like animal and crop residues etc. cause a serious soil pollution. Sludge also contributes soil pollution through the supply of live viruses, viable intestinal worm eggs, high concentrations of soluble salts, various heavy metals etc. Different organic waste materials are rich source for building up soil fertility if it is recycled in agricultural lands after compositing but the main problem is its disposal because of its bulk nature and the huge cost of transport.

iv. Fertilizers and other soluble salts

The accumulation of excess application of chemical fertilizers or through the application of irrigation water particularly in arid or semi-arid regions. Fertilizers used to increase the growth of crops also increase algal growth in surface waters into which they are washed and accelerated algae and water plant growth, is called eutrophication.

Most often increased amounts of phosphorus and nitrate nitrogen are the cause of eutrophication. On the other hand, the major source of phosphorus contamination inn surface waters comes from direct dumping of wastes (Sewage, animal wastes and industrial wastes) and from eroded suspended soils from urban and agricultural lands. In high concentrations salts are unwanted because they reduce or inhibit plant growth, make drinking water unpalatable and interfere in many other uses of water.

Radioactive pollution

The main concern with the radio- activity is the fall out of radio- active materials produced from explosions and hazards associated with waste materials from the nuclear power plants, nuclear reactors and wastes from the medical and other research laboratories.

The most important and long lived ratio-active elements in soils arestrontium-90 (half life = 28 years) and cesium -137 (half life =30 years). .

These radio-active elements cold be decontaminated by continuous cropping, removal of surface soil, deep ploughing, leaching, use of fertilizers and amendments. The use of chelates like EDTA, FEDTA, DTPA in conjunction with algal and fungal cells is most effective in reducing radioactivity from the soil.

Effects of Soil Pollution

Sewage and industrial effluents which pollute the soil ultimately affect human health. Various types of chemicals like acids, alkalis, pesticides, insecticides, weedicides, fungicides, heavy metals and industrial discharges affect soil fertility by change of physical, chemical and biological properties. Radioactive fallout on vegetation is the source of radio-isotopes which enter the food chain in the grazing animals.

Control of soil Pollution

Effluents and solid wastes should be properly treated before discharging them on the soil.Reducing chemical fertilizer and pesticide use. Materials such as glass containers, plastic bags, paper, cloth etc. can be reused at domestic levels rather than being disposed, reducing solid waste pollution. From the wastes, recovery of useful products should be done. Biodegradable organic waste should be used for generation of biogas. Microbial degradation of biodegradable substances is also one of the scientific approaches for reducing soil pollution.