SLM 201 (PRINCIPLES OF SOIL SCIENCE)

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1.1   ROCK DEFINITION, TYPE AND PROPERTIES

Soil is formed from the weathering and disintegration of rock. A rock is an indefinite mixture of naturally occurring substances, mainly minerals. There are three major types of rock, namely; 1) igneous, 2) sedimentary and 3) metamorphic rock.

The usual classification of rocks by geologist is genetic, relating to their modes of origin, and includes the following groups:

(a) *Igneous rocks* formed from ascending hot liquid material arising deep in the Earth called magma which crystallizes into the solid state as the temperature falls.

(b) *Sedimentary rocks* formed as a result of accumulation and compaction of:

(i) pre-existing rock fragments disintegrated throughout erosive processes;

(ii) organic debris such as shells;

(iii) materials dissolved in surface of groundwater and later precipitated in conditions of oversaturation;

(c) *Metamorphic rocks* formed from any pre-existing rock subjected to increases in pressure or temperature or both.

Table 1: Major Rock Types, Their Origin and Properties

Rock type	Origin	Example	Properties
Igneous	Cooling of magma	Granite basalt	Light colour coarse grained dark colour fine grained
Sedimentary	Deposition and compaction	Shale Sandstone limestone	Any colour, fined grained Any colour, coarse grained Light coloured, shells or

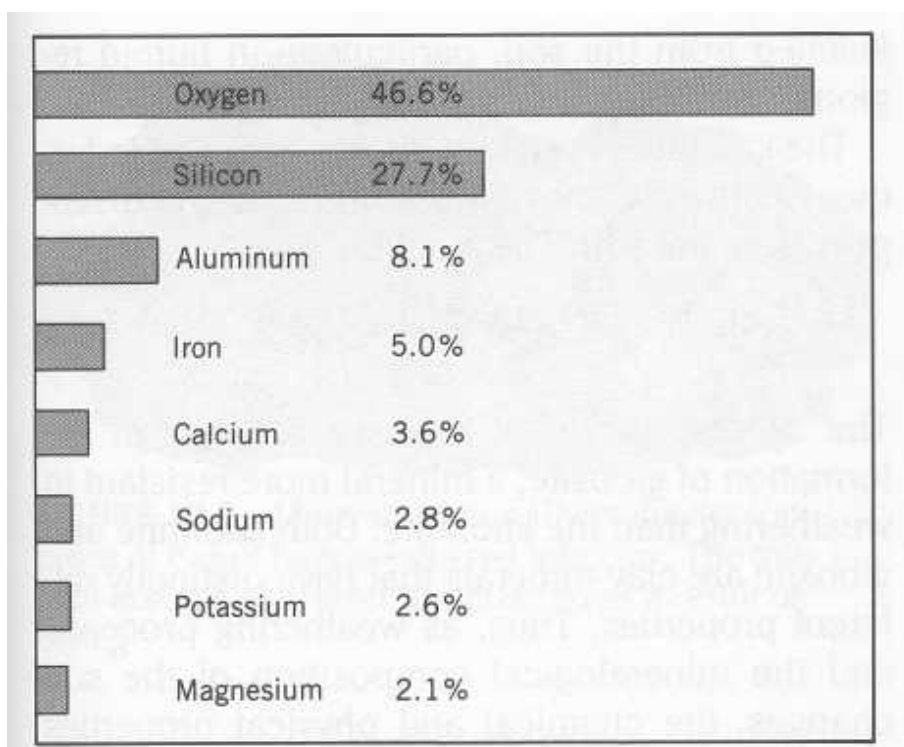|			caco$_3$ present
Metamorphic	Charge in igneous or sedimentary	Slate marble	Any colour, hardened shale Any colour, changed limestone

The rate and extent of rock weathering depends on; 1) the chemical composition of the minerals that comprise the rock or sediment, 2) the type, strength, and durability of the material that holds the mineral grains together, 3) the extent of rock flaws or fractures, 4) the rate of leaching through the material, and 5) the extent and type of vegetation at the surface

1.2   ROCK MINERALS

The earth is composed of various kinds of elements. These elements are in solid form in the outer layer of the earth and in hot and molten form in the interior. The elements in the earth's crust are rarely found exclusively but are usually combined with other elements to make various substances. **These substances are recognized as minerals.** Minerals are therefore naturally occurring organic and inorganic substances, having an orderly atomic structure and a definite chemical composition and physical properties. A mineral is composed of two or more elements. But, sometimes single element minerals like sulphur, copper, silver, gold, graphite etc are found.

Approximately 98 percent of the mass of the earth's crust is composed of eight elements like oxygen, silicon, aluminium, iron, calcium, sodium, potassium and magnesium (see table below) and the rest is constituted by titanium, hydrogen, phosphorus, manganese, sulphur, carbon, nickel and other elements.

Table 2: Elements of the Earth Crust

	Element	Percentage
	Oxygen	46.6%
	Silicon	27.7%
	Aluminum	8.1%
	Iron	5.0%
	Calcium	3.6%
	Sodium	2.8%
	Potassium	2.6%
	Magnesium	2.1%

Oxygen and silicon, compose 75 percent of it. Many of the elements important in the growth of plants and animals occur in very small quantities. Obviously, these elements and their compounds are not evenly distributed throughout the earth's surface. In some places, for example, phosphorus minerals (apatite) are so concentrated that they are mined; in other areas, there is a deficiency of phosphorus for plant growth.

Most elements of the earth's crust have combined with one or more other elements to form the minerals. The minerals generally exist in mixtures to form rocks, such as the igneous rocks, granite and basalt. The mineralogical composition of igneous rocks, and the sedimentary rocks, shale and sandstone, are given in Table below;

Table 3: Mineralogical Composition of Rocks

Mineral Constituent	Origin	Igneous Rock, %	Shale, %	Sandstone, %
Feldspars	Primary	59.5	30.0	11.5
Amphiboles and Pyroxenes	Primary	16.8	—	a
Quartz	Primary	12.0	22.3	66.8
Micas	Primary	3.8	—	a
Titanium minerals	Primary	1.5	—	a
Apatite	Primary or secondary	0.6	—	a
Clays	Secondary	—	25.0	6.6
Iron oxides	Secondary	—	5.6	1.8
Carbonates	Secondary	—	5.7	11.1
Other minerals	—	5.8	11.4	2.2

Data from Clarke, 1924.

Primary minerals weathers to form secondary minerals. The weathering products of some minerals are given in the table below;

Table 4: Weathering Product of Some Minerals

Table 3.2. Weathering products of some minerals

primary mineral	formula	secondary minerals
Olivine	$(Mg,Fe)_2SiO_4$	Fe (hydr)oxides
Pyroxenes	$Ca(Mg,Fe)_3(Al,Fe)_4(SiO_3)_{10}$	Fe (hydr)oxides, smectite
Amphiboles	$Ca_3Na(Mg,Fe)_6(Al,Fe)_3(Si_4O_{11})_4(OH)_2$	Fe (hydr)oxides, smectite
Feldspars	$(K,Na)Si_3O_8 - CaAl_2Si_2O_8$	Kaolinite, gibbsite, allophane
Serpentine	$(Mg,Fe)_3Si_2O_5(OH)_4$	(smectite), Fe-compounds
Muscovite	$KAl_2(AlSi_3O_{10})(OH)_2$	illite
Biotite	$K(Fe,Mg)_2(AlSi_3O_{10})(OH)_2$	vermiculite
Chlorite	$(Mg,Fe,Al)_6(Si,Al)_4O_{10}(OH)_8$	vermiculite, smectite
Garnet	$(Fe^{++},Al,Mn,Ca)_3Al_2Si_3O_{12}$	Fe, Al, Mn (hydrous) oxides
Apatite	$Ca_5(PO_4)_3(OH,F,Cl)$	Fe-phosphate, Al-phosphate
Volcanic glass		allophane

1.3   ROCK WEATHERING

Two classifications of weathering processes exists; **1) *Mechanical* or *physical* weathering, and 2) *chemical* weathering.** However, chemical and physical weathering often goes hand in hand. For example, cracks exploited by mechanical weathering will increase the surface area exposed to chemical action. Furthermore, the chemical action at minerals in cracks can aid the disintegration process.

.1.3.1 Mechanical or physical weathering

***Mechanical* or *physical* weathering** involves the breakdown of rocks and soils through direct contact with atmospheric conditions such as heat, water, ice and pressure.

(i) Thermal expansion

Thermal expansion, also known as onion-skin weathering, exfoliation, insolation weathering or thermal shock, often occurs in areas, like deserts, where there is a large diurnal temperature range. The temperatures soar high in the day, while dipping greatly at night. As the rock heats up and expands by day, and cools and contracts by night, stress is often exerted on the outer layers. The stress causes the peeling off of the outer layers of rocks in thin sheets. Though this is caused mainly by temperature changes, thermal expansion is enhanced by the presence of moisture.

(ii) Freeze thaw weathering

This process can also be called frost shattering. This type of weathering is common in mountain areas where the temperature is around freezing point. Frost induced weathering, although often attributed to the expansion of freezing water captured in cracks, is generally independent of the water-to-ice expansion. It has long been known that moist soils expand or frost heave upon freezing as a result of water migrating along from unfrozen areas via thin films to collect at growing ice lenses. This same phenomena occurs within pore spaces of rocks. They grow larger as they attract liquid water from the surrounding pores. The ice crystal growth weakens the rocks which, in time, break up. The phenomenon is caused by the almost unique property of water in having its greatest density at 4 C, so ice is of greater volume than water at the same temperature. When water freezes, then it expands and puts its surroundings under intense stress.

Freeze induced weathering action occurs mainly in environments where there is a lot of moisture, and temperatures frequently fluctuate above and below freezing point—that is, mainly alpine and periglacial areas. An example of rocks susceptible to frost action is chalk, which has many pore spaces for the growth of ice crystals. When water that has entered the joints freezes, the ice formed strains the walls of the joints and causes the joints to deepen and widen. This is because the volume of water expands by 9% when it freezes.

When the ice thaws, water can flow further into the rock. When the temperature drops below freezing point and the water freezes again, the ice enlarges the joints further. Repeated freeze-thaw action weakens the rocks which, over time, break up along the joints into angular pieces. The angular rock fragments gather at the foot of the slope to form a talus slope (or scree slope). The splitting of rocks along the joints into blocks is called block disintegration. The blocks of rocks that are detached are of various shapes depending on rock structure.

(iii) Pressure release

In pressure release, also known as unloading, overlying materials (not necessarily rocks) are removed (by erosion, or other processes), which causes underlying rocks to expand and fracture parallel to the surface. Often the overlying material is heavy, and the underlying rocks experience high pressure under them, for example, a moving glacier. Pressure release may also cause exfoliation to occur.

Intrusive igneous rocks (e.g. granite) are formed deep beneath the earth's surface. They are under tremendous pressure because of the overlying rock material. When erosion removes the overlying rock material, these intrusive rocks are exposed and the pressure on them is released. The outer parts of the rocks then tend to expand.

The expansion sets up stresses which cause fractures parallel to the rock surface to form. Over time, sheets of rock break away from the exposed rocks along the fractures. Pressure release is also known as "exfoliation" or "sheeting"

(iv) Hydraulic action

This is when water (generally from powerful waves) rushes into cracks in the rock face rapidly. This traps a layer of air at the bottom of the crack, compressing it and weakening the rock. When the wave retreats, the trapped air is suddenly released with explosive force. The explosive release of highly pressurized air cracks away fragments at the rockface and widens the crack itself.

(v) Salt weathering

Salt crystallization causes disintegration of rocks when saline solutions seep into cracks and joints in the rocks and evaporate, leaving salt crystal behind. These salt crystals expand as they are heated up, exerting pressure on the confining rock. Salt crystallization may also take place when solutions decompose rocks (for example, limestone and chalk) to form salt solutions of sodium sulfate or sodium carbonate, of which the moisture evaporates to form their respective salt crystals.

The salts which have proved most effective in disintegrating rocks are sodium sulfate, magnesium sulfate, and calcium chloride. Some of these salts can expand up to three times or even more. It is normally associated with arid climates where strong heating causes strong evaporation and therefore salt crystallisation.

1.3.2   Chemical weathering

Chemical weathering involves the direct effect of atmospheric chemicals, or biologically produced chemicals (also known as *biological* weathering).

Living organisms may contribute to mechanical weathering (as well as chemical weathering). Lichens and mosses grow on essentially bare rock surfaces and create a more humid chemical microenvironment. The attachment of these organisms to the rock surface enhances physical as well as chemical breakdown of the surface microlayer of the rock. On a larger scale seedlings sprouting in a crevice and plant roots exert physical pressure as well as providing a pathway for water and chemical infiltration. Burrowing animals and insects disturb the soil layer adjacent to the bedrock surface thus further increasing water and acid infiltration and exposure to oxidation processes.

The most common form of biological weathering is the release of chelating compounds, i.e acids, by plants so as to break down aluminum and iron containing compounds in the soils beneath them. Decaying remains of dead plants in soil may form organic acids which, when dissolved in water, cause chemical weathering.

Chemical weathering involves the change in the composition of rocks, often leading to a 'break down' in its form. This is done through a combination of water and various chemicals to create an acid which directly breaks down the material. This type of weathering happens over a period of time.

(i) Dissolution

Rainfall is acidic because atmospheric carbon dioxide dissolves in the rainwater producing weak carbonic acid. In unpolluted environments, the rainfall pH is around 5.6. Acid rain occurs when gases such as sulphur dioxide and nitrogen oxides are present in the atmosphere. These oxides react in the rain water to produce stronger acids and can lower the pH to 4.5 or even 3.0. Sulfur dioxide, SO_2, comes from volcanic eruptions or from fossil fuels, can become sulfuric acid within rainwater, which can cause solution weathering to the rocks on which it falls.
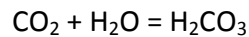
(ii) carbonation

One of the most well-known solution weathering processes is carbonation, the process in which atmospheric carbon dioxide leads to solution weathering. Carbonation occurs on rocks which contain calcium carbonate such as limestone and chalk. This takes place when rain combines
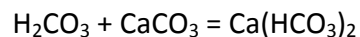
with carbon dioxide or an organic acid to form a weak carbonic acid which reacts with calcium carbonate (the limestone) and forms calcium bicarbonate. This process speeds up with a decrease in temperature and therefore is a large feature of glacial weathering.

The reactions as follows:

$$CO_2 + H_2O = H_2CO_3$$

carbon dioxide + water -> carbonic acid

$$H_2CO_3 + CaCO_3 = Ca(HCO_3)_2$$

Carbonic acid + calcium carbonate = calcium bicarbonate

Carbonation on the surface of well-jointed limestone produces a dissected limestone pavement which is most effective along the joints, widening and deepening them.

(iii) Hydration

Mineral hydration is a form of chemical weathering that involves the rigid attachment of H+ and OH- ions to the atoms and molecules of a mineral. When rock minerals take up water, the increased volume creates physical stresses within the rock. For example iron oxides are converted to iron hydroxides and the hydration of anhydrite forms gypsum.

(iv) Hydrolysis

Hydrolysis is a chemical weathering process affecting Silicate minerals. In such reactions, pure water ionizes slightly and reacts with silicate minerals. An example reaction:

$$Mg_2SiO_4 + 4H^+ + 4OH^- = 2Mg^{2+} + 4OH^- + H_4SiO_4$$
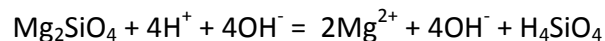
olivine (forsterite) + four ionized water molecules = ions in solution + silicic acid in solution

This reaction results in complete dissolution of the original mineral, assuming enough water is available to drive the reaction. However, the above reaction is to a degree deceptive because pure water rarely acts as a H^+ donor. Carbon dioxide, though, dissolves readily in water forming a weak acid and H^+ donor.

$$Mg_2SiO_4 + 4CO_2 + 4H_2O = 2Mg^{2+} + 4HCO_3^- + H_4SiO_4$$
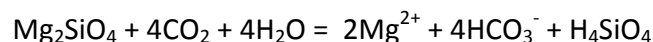
olivine (forsterite) + carbon dioxide + water = Magnesium and bicarbonate ions in solution + silicic acid in solution

This hydrolysis reaction is much more common. Carbonic acid is consumed by silicate weathering, resulting in more alkaline solutions because of the bicarbonate. This is an important reaction in controlling the amount of CO_2 in the atmosphere and can affect climate. Aluminosilicates when subjected to the hydrolysis reaction produce a secondary mineral rather than simply releasing cations.

$$2KAlSi_3O_8 + 2H_2CO_3 + 9H_2O = Al_2Si_2O_5(OH)_4 + 4H_4SiO_4 + 2K^+ + 2HCO_3^-$$
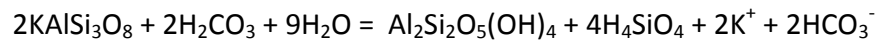
Orthoclase (aluminosilicate feldspar) + carbonic acid + water = Kaolinite (a clay mineral) + silicic acid in solution + potassium and bicarbonate ions in solution

(v) Oxidation

Within the weathering environment chemical oxidation of a variety of metals occurs. The most commonly observed is the oxidation of Fe^{2+} (iron) and combination with oxygen and water to form Fe^{3+} hydroxides and oxides such as goethite, limonite, and hematite. This gives the affected rocks a reddish-brown coloration on the surface which crumbles easily and weakens the rock. This process is better known as 'rusting'.

2.0 SOIL FORMATION

Everything that affects formation, development, characteristics and geographic distribution of soils belongs to soil-forming factors. That´s why the Earth´s soil cover is very diverse. Impact of soil-forming factors results in soil forming processes.

2.1 Three Functions of Soil

  (i) Provides a medium for plant growth

  (ii) Regulates and partitions water flow through the environment

  (iii) Serves as an environmental filter

2.2 Factors of Soil Formation

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

$S = f(Cl, O, R, P, T, …)$

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.

(i)   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.

(a)   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.

(b)   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 . Chemical and biological reaction rates double for every 10 ºC increase

(ii)   ORGANISMS AND VEGETATION (BIOSPHERE)

(a) 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 **t**hus 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.

(b)   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.

(iii)   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)

(a)   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.

(b)   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.

(c)   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 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).

(d)   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.

(iv)   PARENT MATERIALS

Parent materials can vary from solid rock to deposits like alluvium and boulder clay. It has been defined as 'the initial state of the soil system', the mass (consolidated material) from which the

soil is formed. The parent material can influence or impact soil in a number of ways namely; 1) c olour, 2) texture, 3) mineral composition, 4) permeability/drainage, 5) pH, 6) innate soil fertility

Parent materials are classified into two groups namely; a) Residual parent materials b) Transported parent materials

(a) Residual parent materials – Soils develop from underlying bedrock (Igneous, sedimentary, metamorphic) insitu. The parent material differ as widely as the rocks.

Type of rock strongly influences type of soil i.e

- – Limestone → clayey soils,
- – Sandstone → coarse, acidic soils,
- – Granite → coarse, acidic soils,
- – Slate, shale → clayey soils

(b) Transported parent materials – soils develop on transported parent materials, the materials are distinguished according to its mode of transportation viz;

- Colluvial debris
- Alluvial deposits
- Marine sediments
- Lacustrine sediments
- Eolian deposits
- Glacial deposits

(i) Colluvial

Massive to moderately well stratified, nonsorted to poorly sorted sediments with any range of particle sizes from clay to boulders and blocks that have reached their present position by direct, gravity-induced movement. They are restricted to products of mass-wasting whereby the debris is not carried by wind, water, or ice (except snow avalanches).

(ii) Alluvial deposits

Transport by **water** on land produces alluvial, terrace and footslope deposits, during flooding, water spreads and slows, and fine sediment is deposited. Usually very fertile soils

and important for agriculture, forestry, wildlife, poor choice for homes and other urban development

(iii)   Marine sediments

marine deposits under the **sea.** Sediments build up over time, Exposed by changes in elevation of earth's crust, materials are gravely, sandy, clayey depending on area.

(iv)   Lacustrine sediments

Descriptive of materials that have either settled from suspension in bodies of standing fresh water or have accumulated at their margins through wave action. Sediments generally consisting of either stratified fine sand, silt, and clay deposited on the lake bed; or moderately well sorted and stratified sand and coarser materials that are beach and other near-shore sediments transported and deposited by wave action.

(v)   `Eolian deposits

Descriptive of materials transported and deposited by wind. Sediments, generally consisting of medium to fine sand and coarse silt particle sizes, that are well sorted, poorly compacted and may show internal structures such as cross-bedding or ripple laminae, or may be massive. Individual grains may be rounded and show signs of frosting.

(vi)   Glacial deposits

a thick mass of flowing/moving ice, They are generated in areas favored by a climate in which seasonal snow accumulation is greater than seasonal melting i.e (1) polar regions (2) high altitude/mountainous regions

(v)   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).

Table 5: Weathering Stage 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

2.3   SOIL FORMING PROCESSES

Soil forming processes are determined by climate and organisms (both plants and animals) acting on the local geological surface materials over time under the influence of the slope of the land and human activities. The interaction between these factors initiates a variety of processes. **Elementary processes of soil formation (EPP)** are the processes that results in the transformation of the mineral mass of soil forming rocks into horizons with differing composition and properties which are typical for various stages of soil formation and reflecting the totality of phenomena governing the evolution of soils.

The EPP despite their differences can be divided into three (4) main groups;

1. Accumulation

2. Transfer process in soil (Migration)

3. Transformation

4. losses from the soil

Accumulation

The main addition into the soil are organic matter from the surface vegetation with the element they contain, addition by rainfall or runoff, e.g. soluble compound and sediments they contain, and addition of particles carried by wind.

Transfer process in soil (Migration)

Transfer processes involve the vertical and horizontal transport of materials within the soil. These include the following;

- Movement of ions and substances by water moving up and down the profile
- Transfer effected by microbial action, earthworm and other micro fauna
- Transfer effected by cycling of ions by plant

Transformation

Transformation processes within the soil are; Organic compounds that are formed during inorganic matter decomposition

- Organic compounds that are formed during inorganic matter decomposition
- Weathering of primary minerals
- Formation of secondary minerals

losses from the soil

Losses from the soil indicate the removal of substances out from the soil surface by erosion and also washing off of soluble materials down into the deeper layers of the soil by leaching
All these processes combined together leads to differentiation of soil different horizon.

Depending on the complexities of these processes and the restrictiveness of their area of occurrence, these processes can still be classified into **simple** and **complex processes**.
Simple processes include **humification, eluviation, illuviation, leaching, mineralization, amminization, ammonification, nitrification and Denitrification.** The processes are noted not only for their simplicity but also for operating in specific part of the soil profile.

The complex processes of pedogenesis are combinations of many of the simple processes depending on the prevailing bioclimatic conditions of a particular area, two or more of the

simple processes may operate together. Examples of complex processes of pedogenesis are **Lateritization, Podzolization, Calcification, Salinization, Gleization and Solonization.**

2.3.1 SIMPLE PROCESSES OF PEDOGENESIS

humification, mineralization, eluviation, illuviation, leaching, ammonification, nitrification and Denitrification.

1. 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.

- Humification occurs in three phases, namely:

1. rapid initial decomposition of primary plant residues;

2. slow decomposition of plant structural components;

3. alteration of soil organic carbon, Humic Substance genesis.

Rapid initial decomposition of primary plant residues

The first phase in humification involves the decomposition of labile, plant-derived carbon (litter, or 'primary resources'). Decomposition of terrestrial plant litter is the mechanism by which carbon and nutrients are returned to either the atmosphere or to a plant available state. Degrading organisms **immobilize** nutrients such as nitrogen, phosphorus, or calcium in their biomass during this first phase of decomposition.

Degradation rate is fast, and bacteria are active at this stage. The utilizable soluble compound at this stage are plant labile C, sugar, amino acids, organic acids.

Slow decomposition of plant structural components

Degradation rate slows once labile sources of carbon have been utilized, and primarily **lignin-encrusted cellulose and hemicelluloses remain**. Many fungi and eubacteria, such as actinomycetes, are important during this phase of humification. They are capable of producing a variety of cellulolytic and lignolytic enzymes.

Alteration of soil organic carbon, Humic Substance genesis

Microbial contribution to this phase is not only through metabolic activity, but also through incorporation of metabolites and biomass components into **Humic Substances (HS)**. Microbial metabolites and cell components can condense abiotically to form **Humic Substances (HS)**.

For example, phenolic compounds that are released into the soil from litter or microbial activity are highly reactive and spontaneously undergo nonenzymatic chemical reactions to form more complex molecular structures.

Humic Substances (HS)

Humified matter represents more than half soil total organic carbon and can be classified in **two main** types,

1. **humic** and

2. **non-humic** substances

Humic Substances –

- These are generally classified into **three** main groups according to the solubility of each fraction in water adjusted to different acid alkaline (pH levels) conditions:

  (1) HUMIN,

  (2) HUMIC ACIDS (HAs), and

  (3) FULVIC ACIDS (FAs).

Non-humic substances – non-humic substances e.g. carbohydrates, proteins, peptides, amino acids, lipids, waxes and organic acids of low molecular weight

(1) HUMIN – They are the fraction of humic substances which are not soluble in alkali (high pH) and are not soluble in acid (low pH). **Humins are not soluble in water at any pH**
Humin complexes are considered macro organic (very large) substances because their molecular weights (MW) range from approximately 100,000 to 10,000,000.

- **Functions of Humin**
1. improve the soil's water holding capacity,
2. improve soil structure,
3. maintain soil stability,
4. function as an cation exchange system, and generally improve soil fertility.
- Because of these important functions humin is a key component of fertile soils.

2. HUMIC ACIDS (HAs) --- comprise a mixture of weak aliphatic (carbon chains) and aromatic (carbon rings) organic acids which are **not soluble in water under acid conditions but are soluble in water under alkaline conditions.**

   Humic acids consist of that fraction of humic substances that are precipitated from aqueous solution when the pH is decreased below 2. The molecular size of humic acids (HAs) range from approximately **10,000 to 100,000**

   Functions of Humic Acid

1. Humic acid (HA) polymers readily bind clay minerals to form stable organic clay complexes
2. It readily form salts with inorganic trace mineral elements.
3. These trace elements are bound to humic acid molecules in a form that can be readily utilized by various living organisms.
4. As a result humic acids (HAs) function as important ion exchange and metal complexing (chelating) systems.

3. FULVIC ACIDS (FAs)---They are a mixture of weak aliphatic and aromatic organic acids which are **soluble in water at all pH conditions** (acidic, neutral and alkaline). Their composition and shape is quite variable. The size of fulvic acids (HFs) are smaller than humic adds (HAs), **with molecular weights which range from approximately 1,000 to 10,000.**

   Fulvic acids (FAs) have an oxygen content twice that of humic acids (HAs). They have many carboxyl (COOH) and hydroxyl (COH) groups, thus fulvic acids (FAs) are much more chemically reactive.

   The exchange capacity of fulvic acids (FAs) is more than double that of humic acids (HAs). This high exchange capacity is due to the total number of carboxyl (COOH) groups present.

   Functions

- Because of the relatively small size of fulvic acid (FA) molecules they can readily enter plant roots, stems, and leaves. As they enter these plant parts they carry trace minerals from plant surfaces into plant tissues.

- **Fulvic acids (FAs) are key ingredients of high quality foliar fertilizers. Foliar spray applications containing fulvic acid (FA) mineral chelates**

- Fulvic acids (FAs) are the most effective carbon containing chelating compounds known. They are plant compatible, thus non toxic, when applied at relatively low concentrations.

ELUVIATION,

It is the mobilization and translocation of certain constituent's viz. **Clay, Fe, humus, CaCO**, 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 (A or E horizon). Translocation depends upon relative mobility of elements and depth of percolation.

ILLUVIATION,

The process of deposition of soil materials (removed from the eluvial horizon) in the lower layer (or horizon of gains having the property of stabilizing translocated clay materials) is termed as Illuviation. The horizons formed by this process are termed as illuvial horizons (B-horizons, especially Bt) The process leads to textural contrast between E and Bt horizons, and higher fine: total clay ratio in the Bt horizon.

LEACHING,

Wherever rainfall exceeds evaporation and there is free downward movement of water through the soil pore system, soluble minerals are leached or removed from the soil profile. Continual leaching tends to impoverish the upper mineral horizons by removal of basic cations (cations are ions having a a positive electrical charge e.g. Ca_{2+}. Leaching is most active in sandy soils with high porosity and is least in fine-textured soils such as clays which have restricted pore spaces.

MINERALIZATION -- This process of decomposition and liberation of mineral materials from organic material.

Aminization and ammonification are performed by heterotrophic microorganisms and nitrification is brought about mainly by autotrophic soil bacteria. **Heterotrophs require organic carbon compounds for their energy source. Autotrophic organisms get their energy from the oxidation of organic salts and their carbon from the carbon dioxide in the air.**

AMINIZATION

Simple proteins are hydrolyzed (bonds are broken and water molecules added) to form amines and amino acids. The process can be represented as follows:

$$\text{proteins} = R\text{-}NH_2 + CO_2 + \text{energy} + \text{other products}$$
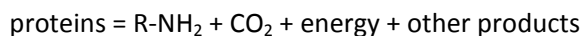
AMMONIFICATION

The **amines and amino acids released by aminization are used by other soil heterotrophs and broken down further to ammoniacal compounds**. This process, called ammonification, can be represented as follows:

$$R\text{-}NH_2 + HOH = NH_3 + R\text{-}OH + \text{energy}$$
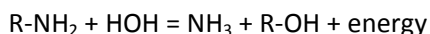
The ammonia produced by this process may be used in several ways in the soil.

1. Be converted to nitrite and nitrate by the process of nitrification.

2. Be used directly by plants.

3. Be used by soil microorganisms.

4. Be tied up by certain types of soil clays.

NITRIFICATION

The conversion of ammonium to nitrate is called nitrification. It is an oxidation process and releases energy for the use of soil microorganisms.

The conversion is a two step process in which ammonium is first converted to **nitrite (NO_2) and nitrite is converted to nitrate (NO_3).**

The conversion of ammonium to nitrite is performed by an obligate autotrophic bacterium known as **NITROSOMONAS**.

This process can be represented as follows:

$$2\ NH_4^+ + 3\ O_2 = 2\ NO_2^- + 2\ H_2O + 4\ H^+$$
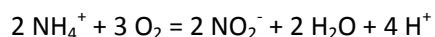
The conversion of nitrite to nitrate is also performed by a number of soil microorganisms but is performed mainly by another group of obligate **autotrophic bacteria known as NITROBACTER**. This process can be represented as follows:

$$2\ NO_2^- + O_2 = 2\ NO_3^-$$
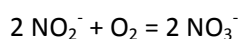
The resulting NO_3^- is highly mobile in soils and is easily lost from the soil with water that moves downward laterally through a soil profile. This NO_3^- is **a potential pollutant** if it reaches surface and ground water supplies.

It is important to remember that these **processes require molecular oxygen; that is, they take place most rapidly in well-aerated soils. Poor aeration due to soil wetness or lack of pore space will slow or stop the mineralization process.** Also, since living soil microorganisms are responsible for the processes, the rate of reaction is very dependent on soil environmental conditions. These include soil **temperature, soil moisture, soil pH, tillage system, cropping system, and the presence of other nutrients**. This has practical implications when organic wastes are soil-applied as fertilizers. The rate at which N is released by organic wastes is dependent on the soil environment as well as the characteristics of the waste itself.

DENITRIFICATION.

Denitrification is the biological reduction of soil nitrates (NO_3) under anaerobic conditions. **If nitrates are present in the soil and the soil is waterlogged, nitrates can be transformed into N_2 and N_2O gases**. These gases can then escape to the atmosphere.

2.3.2   SPECIFIC SOIL FORMING PROCESS

The basic pedologic processes discuss above provide a framework for later operation of more specific processes such as; **Lateritization, Podzolization, Calcification, Salinization, Gleization.**

1. **Laterization**

Laterization is a pedogenic process common to soils found in tropical and subtropical environments. High temperatures and heavy precipitation result in the rapid weathering of rocks and minerals, mineralisation of organic matter is complete and rapid. Movements of large amounts of water through the soil cause eluviations and leaching to occur. Almost all of the by products of weathering, very simple small compounds or nutrient ions, are translocated out of the soil profile by leaching if not taken up by plants for nutrition. The two exceptions to this process are iron and aluminum compounds. Desilication (the removal of combined silica in preference to the Fe and Al which generally accumulate and oxidize to sesquoxide (trivalent oxides)) occurs. Iron oxides give tropical soils their unique reddish coloring. Heavy leaching also causes these soils to have an acidic pH because of the net loss of base cation. Depending on the water situation, the process of lateritization may produce stratified or unstratified soil profile, in the former case, a marked or fluctuating water facilitates the formation of an indurated (hard) zone above the water table. A mottled zone in the area of fluctuation and a pallied zone under the water table or zone of complete or permanent saturation.

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2. Podzolization

Podzolization is associated with humid cold mid-latitude climates and coniferous vegetation. Decomposition of coniferous litter and heavy summer precipitation create a soil solution that is strongly acidic. This acidic soil solution enhances the processes of eluviation and leaching causing the removal of soluble base cations and aluminum and iron compounds from the A horizon. This process creates a sub-layer in the A horizon that is white to gray in color and composed of silica sand.
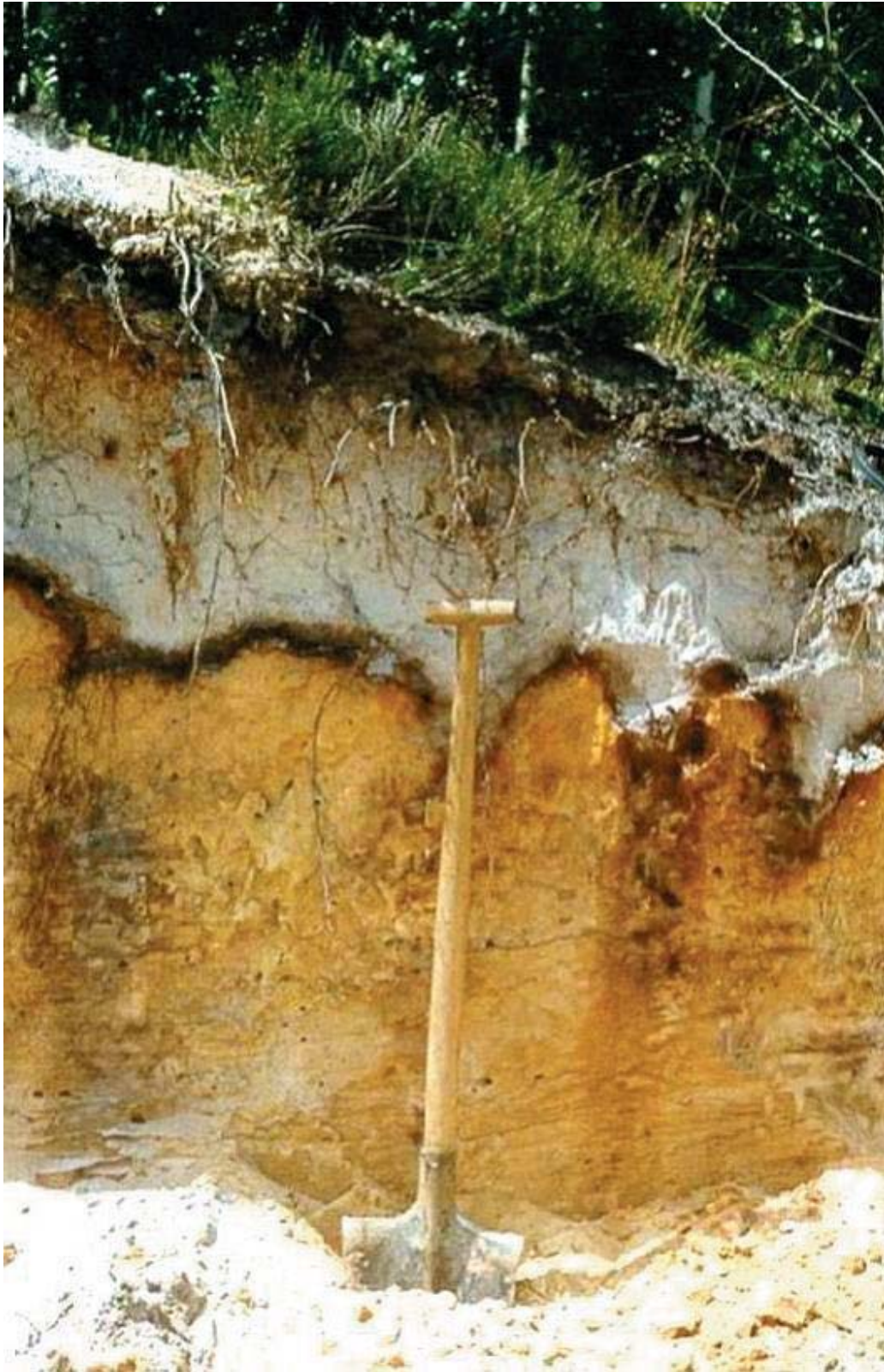
Podzols (from the Russian, pod, meaning under, and zola, meaning ash)
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3. Calcification

Calcification occurs when evapotranspiration exceeds precipitation causing the upward movement of dissolved alkaline salts from the groundwater. At the same time, the movement of rain water causes a downward movement of the salts. The net result is the deposition of the translocated cations in the B horizon. In some cases, these deposits can form a hard layer called **caliche**. The most common substance involved in this process is calcium carbonate. Calcification is common in the prairie grasslands.

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4. *Salinization*

Salinization is a process that functions in the similar way to calcification. It differs from calcification in that the salt deposits occur at or very near the soil surface. Salinization also takes place in much drier climates.

Saline soil occur where saline groundwater comes near to the surface or where the evapo-transpiration is considerably higher than precipitation, at least during a large part of the year. Salts dissolved in the soil moisture remain behind after evaporation of the water and accumulate at or near the surface. Their morphology, characteristics and limitations to plant growth depend on the amount, depth and composition of the salts.

SOLONCHAKS: Strongly saline soil (from the Russian, sol, meaning salt and chak, meaning salty area).

5. Gleization

Gleization is a pedogenic process associated with poor drainage. This process involves the accumulations of organic matter in the upper layers of the soil. In lower horizons, mineral layers are stained blue-gray because of the chemical reduction of iron.

iron can be present **in bivalent (Fe^{2+}, the ferrous ion) and in trivalent forms (Fe^{3+}, the ferric ion respectively).**

Fe^{2+}, the ferrous ion

Fe^{3+}, the ferric ion (commonly called rust

The ferrous form is soluble, whereas the ferric form is not. The more oxidized the soil becomes, the more the ferric forms dominate

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2.4   SOIL PROFILE

The vertical cross section of the soil is called the soil profile. It is seldom uniform in depth, and typically consists of a succession of more-or-less distinct layers or strata. Such layers results from the pattern of deposition, or sedimentation, as can be observed in wind-deposit (aeolian) soils and particularly in water deposit (alluvial) soils. If however the layers form in place by internal soil forming (pedogenic) processes, they are called horizons.

2.4.1 Master Horizon and its Designation

The top layer, A horizon, is the zone of major biological activity and is therefore generally enriched with organic matter and often darker in colour than the underlying soil. The B horizon

is where some of the materials migrating from the A horizon (such as clay or carbonates) tends to accumulate. Under the B horizon lays the C horizon, which is the soil parent material. In the case of a residual soil formed in place from the bedrock, the C horizon consists of the weathered and fragmented rock material. In other case, the C horizon consists of alluvial, Aeolian or glacial sediment.

The A, B, C sequence of horizons is clearly recognizable in some cases as for example in a typical zonal soil such as podsol. In other cases, no clearly discernable B horizon may be discernible, the soil is then characterised by A,C profile, in still some other cases, as in some very recent alluvium, hardly any profile differentiation is apparent. The character of the profile depends primarily on the climate, and secondarily on the parent materials, the vegetation, the topography and time.

Fig. 4.7 Comparison of a soil profile with a weathering profile in glacial till. (After W. H. Scholtes, R. V. Ruhe, and F. F. Riecken.)
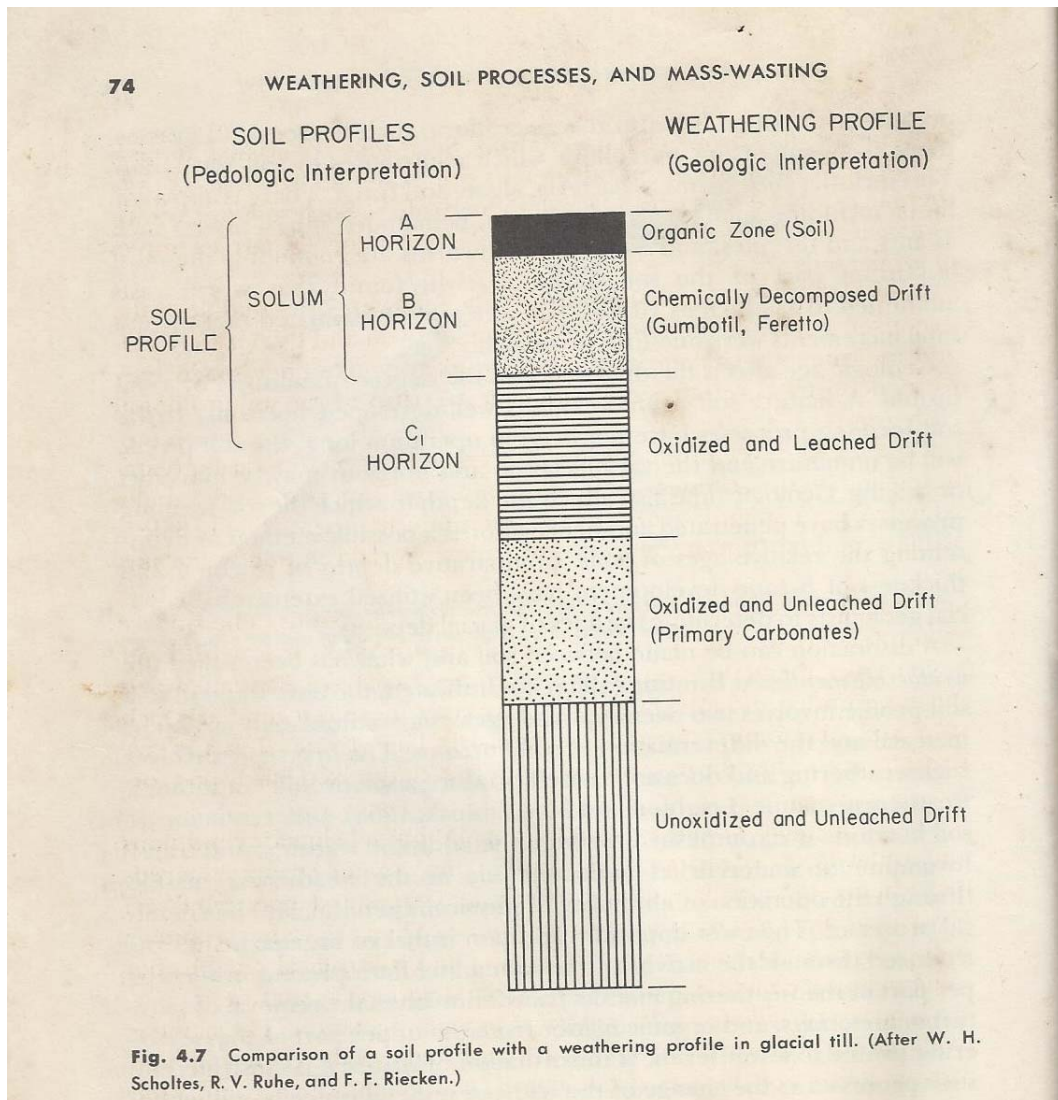
Master Horizon Designation

01 – Undecomposed litter

02 – Partly decomposed litter

A1 – Zone of humus accumulation

A2 (E)– Zone of strongest leaching (eluviation), also called albic horizon

A3 – Transitional to B horizon, has properties more like A than B

B1 – Transitional, more like B than A horizon (sometime absent)

B2 -- Zone of maximum accumulation of clay particles, Fe and Al oxides and humus

   (illuviation). It is referred to as argillic horizons

B3 – Transitional to C horizon

C – Unconsolidated rock

R – Consolidaterd rock

An important aspect of soil formation and profile development are the twin processes of eluviation and illuviation, wherein clay and other substances emigrate from the overlying eluvial A horizon and accumulate the underlying illuvial B horizon. In arid regions, salts such as calcium sulphate and calcium carbonate dissolved from upper part of the soil, may precipitate at some at some depth to form a cemented pan.

2.4.2 Suffixes to the Master Horizon Symbols

Lower case letters can be added as suffixes to the master horizon symbols. Some of the symbols and what they stand for are as follows

a – highly decomposed

e – OM of intermediate decomposition

g – Shinning gleying, evident by mottled appearance of the horizon due to the precipitation of Fe compounds. It is an indication of poor drainage.

h – illuvial accumulation of humus

i – slightly decomposed organic matter

k – accumulation of carbonates

m – strong cementation

p – Plow larger or a mechanically disturbed zone

q – accumulation of silica

s – illuvial accumulation of sesquoxide

t – illuvial accumulation of clay i.e argillic

y – accumulation of gypsum

3.0   SOIL PHYSICAL PROPERTIES

Soil is a heterogeneous, polyphasic, particulate, dispersed and porous system. The three phases of ordinary nature are represented in the soil as follows;

1. The solid phase which constitute the *soil matrix*

2. The liquid phase consists of soil water, which always contain dissolved substances so that it should properly be called the *soil solution;* and

3. The gaseous phase, the *soil atmosphere*.
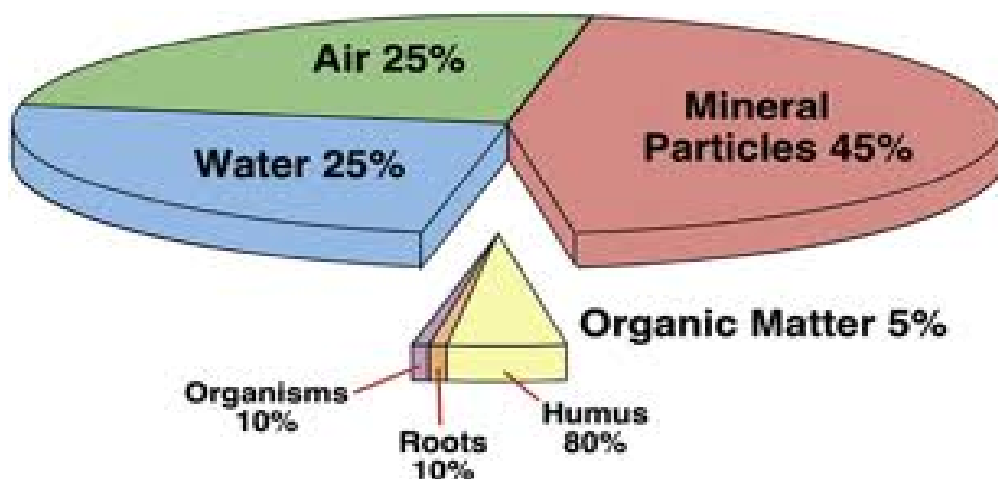
The solid matrix of the soil includes particles which vary in chemical and mineralogical composition as well as size, shape, and orientation. It also contains amorphous substances, particularly organic matter which is attached to the mineral grains and often binds them together to form aggregates. The organisation of the solid components of the soil determines the geometric characteristics of the pore spaces in which water and air are transmitted and retained. Soil air and water vary in composition, both in time and space. The relative proportion of the three phases in soil varies continuously, and depends upon such variables as weather, vegetation and management

3.1   SOIL TEXTURE

This is a stable and an easily determined soil characteristic and refers to the size range of particles in the soil and carries both qualitative and quantitative connotations. Qualitatively it

represents the 'feel' of the soil materials, whether coarse and gritty or fine and smooth. Quantitatively it denotes the measured distribution of particles sizes or the proportion of the various sizes range of particles which occur in a given soil. It is therefore estimated by **feeling and manipulating a moist sample, or more accurately by laboratory analysis**

By feel, Soil texture depends on the amount of each size of particle in the soil. Sand, silt and clay are names that describe the size of individual particles in the soil. **Sand** are the largest particles and they feel "gritty." **Silt** are medium sized, and they feel soft, silky or "floury." **Clay** is the smallest sized particles, and they feel "sticky" and they are hard to squeeze.

It is more or less conventional to define soil materials as particles **smaller than 2 mm** in diameter. Larger particles are generally referred to as gravel, and still larger rock fragments, several centimeters in diameters are variously called stones, cobbles, or if very large, boulders. Soil particle sizes are traditionally characterised into three different conveniently separable size range known as textural fractions or separates namely; **Sand (2 to 0.05 mm) USDA, Sand (2 to 0.02 mm) ISSS, Silt (0.05 to 0.002 mm) and Clay (<0.002 mm).** Where gravel and stones occupy enough of the soils volume to influence soil physical processes significantly, their volume fraction and size range should be reported along with the specification of the finer soil materials. One way of measuring particle-size distribution is to use a pipette or the hydrometer method and read the result off in Textural triangle to get the textural class name

Unfortunately there is yet no universally accepted scheme for classification of particle size i.e USDA (United State Department of Agriculture) standard differs from that of ISSS (International Soil Science Society) as well as that promulgated by ASTM (American Society for Testing Materials). The classification followed by soil engineers often differs from that of agricultural soil scientist; the same terms are used to designate different size ranges, an inconsistency which can result in considerable confusion.

The texture of soils is usually expressed in terms of the percentages of sand, silt, and clay. To avoid quoting exact percentages, 12 textural classes have been defined. Each class, named to identify the size separate or separates having the dominant impact on properties, includes a

range in size distribution that is consistent with a rather narrow range in soil behaviour. The loam textural class contains soils whose properties are controlled equally by clay, silt and sand separates. Such soils tend to exhibit good balance between large and small pores; thus, movement of water, air and roots is easy and water retention is adequate.
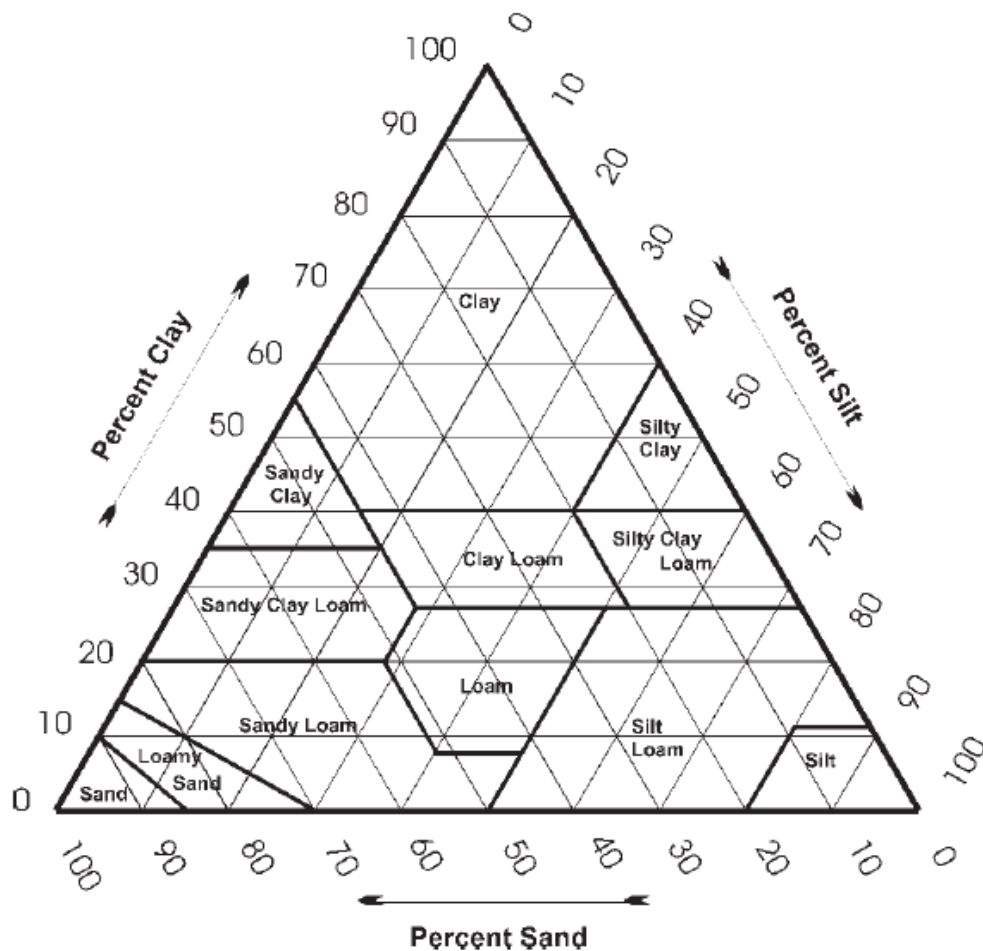
Soil Textural Triangle

3.1.1 Importance of soil texture

Capacity of soils to store water and nutrients increases when their clay percentage increases since clay has a large surface area/unit volume and they can absorb large amounts of nutrients

and water. A soil with sandy texture has difficulty in retaining water and thus nutrients are not made available to the plants growing in such a soil. A clayey soil is poorer in aeration (due to stagnation of water) thus debilitating the plant growth. Loamy soil (a mixture of clay and sand) is the best textured soil for crop cultivation since it has all the beneficial aspects not found in the sandy and clayey soil. Thus we can see how texture affects the plant growth.

3.2   SOIL STRUCTURE

This is the arrangement and organization of the particles in the soil. It is strongly affected by changes in climate, biological activity and soil management practices and vulnerable to destructive forces of a mechanical and physiochemical nature. Soil structure is important in that it determines the total soil porosity, as well as the shapes of individual pores and their size distribution. Hence soil structure affects the retention and transmission of fluids in the soil, including infiltration and aeration. It influences mechanical properties of soil and phenomena such as germination, root growth, tillage, overland traffic, and erosion

Agriculturists are interested in having the soil, at least in its surface zone, in a loose and highly porous and permeable condition. Engineers on the other hand desire a dense and rigid structure so as to provide maximal stability and resistance to shear and minimal permeability.

3.2.1 Types of soil structure

Three broad categories of soil structure are single grained, massive and aggregated.

1. Single grained structure – here soil particles are entirely unattached to each other, the structure is completely loose i.e coarse granular soils or unconsolidated deposits of desert dust. Such soils were labelled structureless in older literature.

2. Massive --- here soils are tightly packed in large cohesive blocks as is sometimes the case with dried clay

3. Aggregated -- An intermediate condition in which the soil particles are associated in quasi-stable small clods known as aggregates or peds

The soil separates can become aggregated together into discrete structural units called "peds". These peds are organized into a repeating pattern that is referred to as soil structure. Between the peds are cracks called "pores" through which soil air and water are conducted. Soil structure is most commonly described in terms of the shape of the individual peds that occur within a soil horizon (Schuler and Wood, 1992). The types of Soil Structures are;

1. Granular – roughly spherical, like grape nuts. Usually 1-10 mm in diameter. Most common in A horizons, where plant roots, microorganisms, and sticky products of organic matter decomposition bind soil grains into granular aggregates shown below

2. Platy – flat peds that lie horizontally in the soil. Platy structure can be found in A, B and C horizons. It commonly occurs in an A horizon as the result of compaction.

3. Blocky – roughly cube-shaped, with more or less flat surfaces. If edges and corners remain sharp, we call it angular blocky. If they are rounded, we call it subangular blocky. Sizes commonly range from 5-50 mm across. Blocky structures are typical of B horizons, especially those with high clay content. They form by repeated expansion and contraction of clay minerals

4. Prismatic – larger, vertically elongated blocks, often with five sides. Sizes are commonly 10-100mm across. Prismatic structures commonly occur in fragipans.

5. Columnar – the units are similar to prisms and are bounded by flat or slightly rounded vertical faces. The tops of columns, in contrast to those of prisms, are very distinct and normally rounded.

3.3 Soil Bulk Density

Bulk density is a measure of the weight of the soil per unit volume (g/cc), usually given on an oven-dry (110° C) basis. It depends greatly on the mineral make up of soil and the degree of compaction. The density of quartz is around 2.65g/cm³ but the bulk density of a mineral soil is normally about half that density, between 1.0 and 1.6g/cm³. Soils high in organics and some friable clay may have a bulk density well below 1g/cm³

Bulk density determination is important in calculating soil moisture movement within a profile and rates of clay formation and carbonate accumulation. It can determined by core, clod or radiation methods and more recently by excavation method

1. Core method; core samples which are taken by driving a metal corer into the soil at the desired depth and horizon are collected on the field. The samples are then oven dried and weighed.

Bulk density = mass of oven dry soil/core volume

$$\rho = \frac{M_s}{V_t}$$

Bulk density = $\dfrac{\text{dry weight}}{\Pi r^2 h}$

The bulk density of soil is inversely related to the porosity of the same soil. The more pore space in a soil the lower the value for bulk density. Soil densities vary over a wide range, and typically between 0.1 g·cm^{-3} for light peats and 1.8 g·cm^{-3} for very dense compacted mineral soils with little pore space.

2. Clod method; the bulk density of clods, or peds, can be calculated from their mass and volume. The volume may be determined by coating a clod of known weight with a water-repellent substance and by weighing it first in air, then again while immersed in a liquid of known density, making use of Archimedes' principle. The clod or ped must be sufficiently stable to cohere during coating, weighing and handling. The clod method is applied commonly by pedologists or paleopedologists.

3. Radiation method; this is an in situ method, transmitted or scattered gamma radiation is measured; and with suitable calibration, the density of the combined gaseous-liquid-solid components of a soil mass is determined. Correction is then necessary to remove the components of density attributable to liquid and gas that are present. The radiation method is an in situ method

4. Excavation method; More recent and has found use in tillage research where surface soil is often too loose to allow core sampling, or where abundant stones preclude the use of core samplers

3.4   SOIL CONSISTENCE

Consistence is a description of a soil's physical condition at various moisture contents as evidenced by the behavior of the soil to mechanical stress or manipulation. Descriptive adjectives such as hard, loose, friable, firm, plastic, and sticky are used for consistence. Soil consistence is of fundamental importance to the engineer who must move the material or compact it efficiently. The consistence of a soil is determined to a large extent by the texture of the soil, but is related also to other properties such as content of organic matter and type of clay minerals.

3.5   SOIL POROSITY

Soil porosity refers to that part of a soil volume that is not occupied by soil particles or organic matter. Pore spaces are filled with either air, other gases, or water. Large pores (macropores) allow the ready movement of air and the drainage of water. They are also large enough to accommodate plant roots and the wide range of tiny animals that inhabit the soil. Large pore spaces permit fast infiltration and percolation of water through a soil or soil horizon. Small pores (micropores) exhibit attractive forces strong enough to hold water in the pore. They are the water retention system of the soil which provides water storage for plant roots. During precipitation, macropores conduct water into the soil where it fills the micropores. At field capacity all pores small enough to retain water against the pull of gravity are filled.

Clay soils have numerous micropores and hold large quantities of water, but since they have few macropores they produce very slow infiltration rates. The pores in the clays may be so small and hold water so tenaciously that the water is not available to plants. Sandy soils with numerous macropores but few micropores have higher infiltration and percolation rates but a lower water-holding capacity than other soil textures. A lower water-holding capacity can mean less available water for plant roots. For revegetation purposes, plants perform best in intermediate soil textures (loams) where soils contain mixtures of micro- and macropores

3.5.1   Determination of Soil Total Porosity

Soil porosity was determined from bulk density using the following formula;

Porosity = {1 - (Db/Dp)} x 100

Where; Db = The dry bulk density (g cm^{-3})

       Dp = Particle density (taken to be 2.65 (g cm^{-3})

- In calculating porosity, it is better to calculate particle density of the soil instead of assuming it to be 2.65 g/cm3. This makes it meaningful and not a constant factor of Bulk density value

-

3.5.2   Determination of Micro porosity

$$(\textit{Mic}) \ = \ \underline{\text{wet weight (Ww) g - dry weight (Wd) g}} \ \ X \ \ 100$$

$$\text{Volume of soil in the cylinder } (V_c) \text{ cm}^{-3}$$

3.5.3   Determination of Macro porosity

Macro porosity =   {TP – Microporosity} x 100

3.6   SOIL WATER

Soil water is usually derived from rainfall or some other form of overland flow. The amount of water that enters the soil is a function of soil structure which in turn is a function of texture and higher-order structure i.e. aggregation (e.g. well-aggregated, porous soils allow greater infiltration). Larger pores created by root channels and animal burrows (ant tunnels, crayfish tunnels and worm holes) as well as other types of macropores also greatly facilitate movement of water into and through the soil profile. This is why these soil animals are usually regarded as highly beneficial in soil and why some scientists often use the abundance of earthworms as an indicator of a "healthy soil", though it should be remembered that highly fertile, productive soils do not always contain large numbers of earthworms and other soil animals.

Water moves through soil by mass flow (i.e. fluid movement through larger channels) and by capillary action, the slower movement through the highly tortuous network of very small channels and pores. Most soil water is held in this system of capillary pores and channels and can be held quite firmly through physical interactions with the solid components of the soil matrix. Matric water is held in small (usually microscopic) pores and is often adsorbed onto particle surfaces that can be held rather tightly. Because these interactions with the soil matrix essentially bind the soil water, energy on the part of plants and microbes is required to extract

water. These forces are usually given as a matric potential or pressure and they indicate how tightly the water is held. These values are usually expressed as a negative value because they are a measure of the amount of suction required to extract water from the soil. It is this adsorption of water to soil solids that explains why plants can wilt in a soil having substantial clay content, even when some matric water is still present. Beyond a certain point, the water is held too tightly for the plants to extract and the so-called wilting point is reached. An often-stated value for the matric potential at the permanent wilting point is approximately -15 bars (-1.5 mPa). Plants exposed to lower matric (more negative) potentials may wilt and not recover. It should be stressed that the matric potential is a function of soil texture and that fine-textured soils (silts and clays) can contain moisture at very low matric potentials because of the much greater numbers of small particles (greater surface area) and the finer pore network which adsorbs and retains water. Thus in clayey soils, water is not as biologically available as it is in coarser-textured soils. A corollary, however, is that coarse-textured soils do not retain as much water because the larger particles do not adsorb water as well and the larger pores drain more rapidly. Thus, nutrients and other chemicals may be retained in a clay soil whereas they may leach, i.e. move downward, with water more rapidly through sandy soils. These textural concerns are of extreme importance to our understanding of the movement of materials into ground water.

Water occupying the pore network in soil is often called the soil solution. It is never pure water but rather it is a solution containing dissolved salts and gases. This is the solution which is most available to plants and microbes and provides some of the nutrients needed for growth. Many nutrients actually reach the roots of plants through mass flow of the water to the vicinity of the roots. Too much water in the soil disrupts the ideal balance between air-filled and water-filled pores, i.e. soil aeration (discussed below). Salts dissolved in the soil solution give rise to an osmotic pressure or potential which can be measured. The osmotic potential can influence the movement of water into and out of cells and therefore it is important that the soil solution not be allowed to become too saline (salty) as it might in a soil experiencing increased salinity due to prolonged irrigation and poor drainage. Under low osmotic (high negative) potentials (salty or sugary solutions), water is actually drawn out of cells resulting in physiological impairment,

that is, being dehydrated beyond some critical point. This is the basis of preserving foods by adding high concentrations of salt.

Osmotic and matric potentials are the two major factors controlling the availability of water to soil life and of these two, it is the matric potential which most affects soil microbes. As a soil dries, life in the soil must work harder to extract water against ever increasing matric and osmotic potentials. For these reasons, adequate soil moisture is required for maximum rates of plant growth and to maintain the vital soil microbial processes essential for life on earth.

3.6.1   Determination of gravimetric moisture content

Soil gravimetric moisture content was determined from soil bulk density using the formula:

$\%MC = W_2 - W_3 / W_3 - W_1 \times 100$

Where; %MC = Moisture content detemination

   W_1 = weight of a moisture can (g)

   W_2 = weight of moisture can with moist soil (g)

   W_3 = weight of moisture can with oven dried soil (g)

3.7   SOIL TEMPERATURE

Every biological process needs an optimum temperature to get accomplished. A maize seed germinates only at a temperature range of 7-10°C and most of the soil microbes function best at 25-35°C Soil gets heated up mainly due to solar radiation and its temperature is highly influenced by the amount of solar radiation received by the soil which in turn depends on the climate of the region. The amount of sunlight reaching the soil again depends on the slope of the land. For example, if the land is slope, then the amount of solar radiation striking the unit volume of land decreases as the slope increases.

This is not the case in the leveled land where each unit area of land receives equal proportion of solar radiation. Soil temperature also depends on the vegetation cover of the land. A barren land gets heated up faster and cools up at a rapid rate whereas in a land covered with

vegetation which acts as an insulation barrier, soil temperature remains near optimum. Soil colour also influences the temperature as seen in darker soils that absorb more solar radiation than a light coloured soil.

4.0 SOIL CHEMISTRY

The division of soil science dealing with the chemical constitution, properties, and reactions of soils.

4.1 Soil pH

Soil pH is probably the most commonly measured soil chemical property and is also one of the more informative. Like the temperature of the human body, soil pH implies certain characteristics that might be associated with a soil. Since pH (the negative log of the hydrogen ion activity in solution) is an inverse, or negative, function, soil pH decreases as hydrogen ion, or acidity, increases in soil solution. Soil pH increases as acidity decreases.

A soil pH of 7 is considered neutral. Soil pH values greater than 7 signify alkaline conditions, whereas those with values less than 7 indicate acidic conditions. Soil pH typically ranges from 4 to 8.5, but can be as low as 2 in materials associated with pyrite oxidation and acid mine drainage. In comparison, the pH of a typical cola soft drink is about 3.

Soil pH has a profound influence on plant growth. Soil pH affects the quantity, activity, and types of microorganisms in soils which in turn influence decomposition of crop residues, manures, sludges and other organics. It also affects other nutrient transformations and the solubility, or plant availability, of many plant essential nutrients. Phosphorus, for example, is most available in slightly acid to slightly alkaline soils, while all essential micronutrients, except molybdenum, become more available with decreasing pH. Aluminum, manganese, and even iron can become sufficiently soluble at pH < 5.5 to become toxic to plants. Bacteria which are important mediators of numerous nutrient transformation mechanisms in soils generally tend to be most active in slightly acid to alkaline conditions.

4.1.2 Soil pH

The most important property of soil as related to plant nutrition is its **hydrogen ion activity**, or **pH** (the term "reaction" is also used, especially in older literature). Soil reaction is intimately associated with most soil-plant relations. Consequently, the determination of pH has become almost a routine matter in soil studies relating directly or indirectly to plant nutrition. Knowledge of soil acidity is useful in evaluating soils because pH exerts a very strong effect on the solubility and availability of many nutrient elements. It influences nutrient uptake and root growth, and it controls the presence or activity of many micro-organisms.

The pH scale is based on the ion product of pure water. Water dissociates very slightly:

$$H_2O \leftrightarrow H^+ + OH^-$$

$$K_w = [H^+] * [OH^-] = 10^{-14} \text{ at } 23°C$$

where K_w is the ion product for water and indicates the activity of each component in moles per liter of solution. Since $[H^+] = [OH^-]$ in pure water at 23°C, each is equal to $(10^{-14})^{1/2} = 10^{-7}$.

The pH of a solution is defined as the negative log (base 10) of the H ion activity or the log of the reciprocal of $[H^+]$:

$$pH = -\log_{10} [H^+] = \log_{10} (1/[H^+])$$

For example, a hydrogen ion activity of 1/10,000 (or 10^{-4}) mol/L would equal pH 4. Water with equal numbers of H^+ and OH^- (hydroxyl) is neutral at pH 7 at 23°C . pH values below 7 are increasingly acid with excess H^+ or hydrogen ions. At 100°C the pH of pure water is 6.0 and at 0°C is 7.5 (i.e., temperature affects pH).

Carbon dioxide dissolves in water to form carbonic acid. Otherwise-pure water in equilibrium with CO_2 at its standard atmospheric concentration of 0.033% (330 ppmv) will have a pH of 5.72. CO_2 concentration may be as high as 10% in poorly aerated soil pores; water in equilibrium with this air would have a pH of 4.45, although other components of soil solution can raise or lower it.

4.1.2 Soil pH

Three soil pH ranges are particularly informative: a pH <4 indicates the presence of free acids, generally from oxidation of sulfides; a pH <5.5 suggests the likely occurrence of exchangeable Al; and a pH from 7.8-8.2 indicates the presence of $CaCO_3$.

The fundamental property of any acid in general (and therefore of a soil acid) is that of supplying protons, and therefore the H^+ ion activity of a system is fundamentally its proton supplying power. In an analogous fashion, the redox potential (E_h) of a system is its electron supplying power.

Hydrogen ions in solution are in equilibrium with those held on soil particle surfaces (i.e., on exchange sites). The soil pH as actually measured represents the **active** (in solution) hydrogen ion concentration.   The total acidity of the soil includes both active and "reserve" (or **exchangeable**) acidity. Thus, two soils with the same pH may have much different amounts of reserve acidity and one may be more difficult to neutralize than another.

Exchangeable aluminum also contributes to soil acidity. When an Al^{3+} ion is displaced from an exchange site into the soil solution, it hydrolyzes, splitting water and releasing a hydrogen ion to solution:

$$Al^{3+} + H_2O = AlOH^{2+} + H^+$$
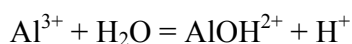
4.1.3 Measuring Soil pH

Measuring pH

There are many factors that affect soil reaction as measured in the laboratory. The pH of many soils tends to increase as the sample is diluted with water. Such pH changes may be caused by variables such as carbon dioxide partial pressure, salt concentration, hydrolysis, and solubility of soil constituents. Various soil :water ratios have been proposed for pH determinations. These range from very dilute suspensions (1:10 soil:solution ratio) to soil pastes. The general effect

sees the pH of most soils increasing with dilution, and becoming constant at about a soil:water ratio of 1:5.

There is no standard procedure for measuring soil pH. Some of the details that vary from one laboratory to the next are: soil:solution ratio, use of a salt solution (e.g., 0.01 M $CaCl_2$) rather than water, method of mixing, time of standing before reading, etc. Soil may be weighed, or measured as a volume (McLean 1982). Therefore, when reporting sol pH, it is essential to include at least a brief summary of the procedure followed.

The exact placement of the pH electrode in the sample may be important. When placed in the settled sediment of a suspension of soil of appreciable cation exchange capacity (CEC), a lower pH is generally measured compared to the measurement obtained in the supernatant solution (called the suspension effect). However, the sediment pH can be lower than, equal to, or higher than that of the supernatant depending on the soil and existing conditions. For example, if the soil has a net positive charge and more OH than H ions are dissociated from the soil, the sediment may have a higher pH than the supernatant.

Soil factors in the field that influence soil reaction include degree of ***base saturation, type of colloid, carbon dioxide partial pressure, oxidation potential, soluble salts, and so on.*** In addition to these factors the measured pH may vary because of the manner in which the sample is handled in the laboratory before and during the determination. Acquaintance with these variables is necessary for intelligent measurement and interpretation of soil reaction.

pH can be determined using either **colorimetric** or **electrometric** methods. The choice of method depends upon the accuracy required, the equipment available, or convenience. Many organic dyes are sensitive to pH, the color of the dye changing more or less sharply over a narrow range of H-ion activity. These methods tends to be slower, less precise, and obscured from view by soil particles and organic matter. Hence, they are used mostly in the field where pH is to be approximated.

The electrometric method involves a glass electrode that is sensitive to H^+: there is an exchange of ions between solution (H^+) and glass (Na^+) (Westcott 1978). A reference electrode that

produces a constant voltage is also required. The electrode pair produces an electromotive force (emf or voltage) that is measured by a millivoltmeter. The relation between emf and pH is governed by the **Nernst equation**

4.2 Cation exchange

Cat ions are positively charged ions, i.e Ca^{2+}, Al^{3+}, H^+, Mg^+, Fe^{3+}, K^+, NH_4^+ and are sourced mainly from the decomposition of Organic Matter and Weathering - Breakdown of mineral matter. Silicate clays and organic matter typically possess net negative charge because of cation substitutions in the crystalline structures of clay and the loss of hydrogen cations from functional groups of organic matter. Positively-charged cations are attracted to these negatively-charged particles, just as opposite poles of magnets attract one another.
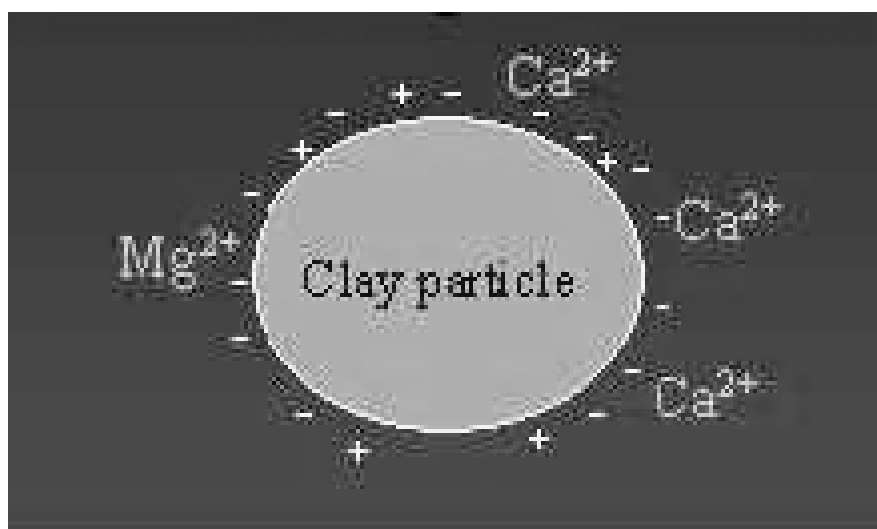
Cation exchange is the ability of soil clays and organic matter to adsorb and exchange cations with those in soil solution (water in soil pore space). A dynamic equilibrium exists between

adsorbed cations and those in soil solution. Cation adsorption is reversible if other cations in soil solution are sufficiently concentrated to displace those attracted to the negative charge on clay and organic matter surfaces.

4.3   Cation Exchange Capacity

This is the quantity of cation exchange measured per unit of soil weight. In other words, it is the total number of cation adsorption sites per unit weight of soils. Organic colloids exhibit much greater cation exchange capacity than silicate clays. Various clays also exhibit different exchange capacities. Thus, cation exchange capacity of soils is dependent upon both organic matter content and content and type of silicate clays.

4.3.1   Importance of Cation exchange capacity

1. exchangeable cations such as calcium, magnesium, and potassium are readily available for plant uptake and

2. cations adsorbed to exchange sites are more resistant to leaching, or downward movement in soils with water.

4.3.2   Factors Affecting Cation Exchange Capacity

1. Soil texture - The CEC of soils increases when their percentage of clay increases i.e. when their texture becomes finer.

2. Soil humus content - Since a negatively charged humic micelle attracts cations and holds them, the CEC increases when the percentage of humus increases.

3. Nature of clay - The CEC depends on the nature of clay minerals present in it since each mineral has its own capacity to exchange and hold cations. The CEC of a soil dominated by vermiculite is much higher than the CEC of another soil dominated by kaolinite.

4. Soil reaction - When the pH of soil increases, more H + ions dissociate from the clay minerals especially kaolinite, thereby the CEC of soil dominated by kaolinite also increases.

4.4   Base Saturation

Closely related to cation exchange capacity is the base saturation, which is the fraction of exchangeable cations that are base cations (Ca, Mg, K and Na). The higher the amount of exchangeable base cations,

the more acidity can be neutralised in the short time perspective. Thus, a site with high cation exchange capacity takes longer time to acidify (as well as to recover from an acidified status) than a site with a low cation exchange capacity (assuming similar base saturations).

4.4.1 Calculating Cation Exchange Capacity and the Percent Base Saturation

The cation Exchange Capacity (CEC) is a term used to describe the holding capacity of a particular soil for positively-charged elements (cations). It may also be described as the capacity for a soil to exchange cations for another. In simple terms the higher the clay content, the higher the CEC since clay particles have the greatest surface area per unit volume of soil and, therefore, can hold the most cations. The CEC is a part of the soil test which is calculated from the levels of potassium, magnesium, calcium, sodium and hydrogen which were extracted in the soil analysis. These cations are the ones of greatest concentration in most soils. The calculations will be explained below:

Each of these elements has its individual atomic weight as found on a Periodic Table.

Calcium = 40 Sodium = 23

Magnesium = 24 Hydrogen = 1

Potassium = 39

From the atomic weight the equivalent weight can be calculated. Calcium and magnesium have two valences or positive charges. Sodium, potassium, and hydrogen each have one positive charge. By dividing the atomic weight by the number of valences we determine the equivalent weight. Therefore,

Calcium = 20 Sodium = 23

Magnesium = 12 Hydrogen = 1

Potassium = 39

By getting the equivalent weights we now have each element in equal terms or in other words 20 ppm of calcium can displace 12 ppm of magnesium on the soil complex.

However, the CEC is reported in meq/100 gm. The equivalent weights are reported as equivalents per gram. Therefore, the equivalent weights must be multiplied by ten to be converted to meq/100 gm. Now that we have these values we can calculate the CEC from the ppm of each of these elements on the soil test. The following will be used for an example:

A soil test shows we have 379 ppm of potassium, 133 ppm of magnesium, 1600 ppm of calcium, 22 ppm of sodium and a pH of 8.0. (Since the pH is above 7.0, we have no hydrogen reported, but if the pH was 6.0, you would use the meq/100 gm of hydrogen given on Midwest Lab's soil analysis in the calculation.

Potassium 379/390 = .972

Magnesium 133/120 = 1.11

Calcium 1600/200 = 8.0

Sodium 22/230 = .1

Hydrogen 0

CEC = .97 + 1.11 + 8.0 + .1 = 10.2 meq/100 gm .

As you can see, the largest factor in the calculation is that of calcium. Soil exchange sites contain mostly calcium; however, in higher pH soils large amounts of free calcium may also be present in soil solution which may also be extracted in the test reflecting a higher holding capacity than what the texture analysis would actually indicate. Midwest Laboratories, Inc. procedures try to reduce the effects of this excess calcium, but the CEC may still reflect a somewhat inflated number as compared to what would be indicated by the texture.

The Percent Base Saturation is then calculated from these values by dividing the milliequivalents of each by the CEC as shown below:

Potassium .97/10.2 = 9.6%

Magnesium 1.11/10.2 = 10.9%

Calcium 8.0/10.2 = 78.6%

Sodium .1/10.2 = .9%

The Percent Base Saturation could be defined as the relative availability of each of these cations.

4.5 Lime requirement

Lime requirement is the amount of a base (in practice, lime or calcium carbonate) needed to neutralize enough of the exchangeable acidity to raise soil pH to a desired value that is more suitable for crop growth.

In most soils it has been noticed that pH tends to increase with depth. This is because the upper horizons receive maximum leaching by rainfall, and by dissolved carbonic acid and organic acids

which remove metal cations (eg., Ca^{++}, K^+, Mg^{++}) and replace them with H^+ ions. Lower horizons are not so strongly leached and, in fact, in dryer areas may accumulate calcium and other materials removed from the upper soil.

5.0   SOIL ORGANISMS

Soil organisms play an important part in soil development. In addition to the roots of higher plants, the soil is inhabited by a wide variety of plant and animal life. In fact, a soil does not usually develop until the inorganic material is "invaded" by various kinds of organisms. The total weight of soil organisms (excluding higher plants) in the upper 30 centimeters of fertile agricultural soils is impressive, as much as 7000 kilograms per hectare. This is, however, only about 0.1 percent of the weight of the soil that the organisms occupy.

Average weight of organisms in the upper 30 centimeters of soil (in kilograms/hectare).

Organism	Low	High
Bacteria	560	1120
Fungi	1680	2240
Actinomyces	895	1680
Protozoa	225	450
Algae	225	335
Nematodes	28	55
Other worms and insects	895	1120

Total	4508	7000

5.1   Higher plants

Higher plants must be considered the principal soil organisms. Roots of trees and other higher plants penetrate crevices in rock, expand them, and split the rocks by the tremendous forces they exert when growing, thereby performing an important function in the continuing process of soil formation. Roots exude many kinds of organic acids and other substances that hasten the solution of soil minerals and make them available to plants. Living roots give off CO_2 which raises the carbonic acid content of the soil solution and increases the rate at which soil minerals dissolve. When leaves fall, or when plants die, their organic matter becomes incorporated into the soil, contributing to its fertility. Channels remaining after dead roots have decayed serve as pathways for the movement of soil water.

5.2   Bacteria

Bacteria are present in all soils. Estimates of the weight of living and dead bacteria in forest soils have been as great as 6300 kilograms per hectare, whereas fertile agricultural soils typically contain less than 20% of this. Bacteria in the soil decompose organic matter produced by plants (and other organisms), releasing the minerals they contain, which then become available for another cycle of plant growth. Some kinds of bacteria, such as those in the genus *Azotobacter*, convert molecular nitrogen from the atmosphere into nitrogenous compounds that can be used by plants. Other species live in symbiotic association with the roots of certain kinds of plants and perform the same function.

5.3   Nitrogen-Fixing Bacteria

Nitrogen-fixing bacteria supply nitrogen to the plant while the plant supplies carbohydrates to the bacteria. Some bacteria play roles in soil formation and degradation, but these processes are poorly understood. Not all soil bacteria are helpful. Many species are **pathogenic**, causing crop diseases and large economic losses. Other harmful types oxidize **ferrous iron** to the slightly soluble **ferric form**, thereby contributing to the formation of soil horizons called **hardpans**, which are extremely rich in iron and quite hard. Hardpans may prevent the drainage of excess water and minhibit root penetration.

5.4   Fungi

Fungi perform many of the same functions that bacteria carry out in decomposing organic matter and in cycling nutrients. Many soil-borne fungi are also pathogenic. **Saprophytic fungi** decompose the dead organic matter of soils. When temperature, moisture, oxygen, and soil acidity are in the right balance, these fungi decompose protein and cellulose, lignin, and other carbohydrates, changing them to humus. Even the bark of trees, which is highly resistant, is decomposed by fungi. Many species of fungi live in such close symbiotic association with roots (particularly of woody plants) that organic and inorganic materials pass back and forth. Such an association is known as a **mycorrhiza**. Water and minerals move from the fungi to the roots, and carbohydrates and other organic materials move in the other direction.

Some mycorrhizal fungi actually penetrate the interior of the roots; others remain on the outside. The importance of these fungi is becoming increasingly apparent. Most species of pines cannot survive without them, and there is increasing evidence that they may be associated with the roots of practically all plant species.

5.5   Algae

Most types of algae are aquatic, but several kinds, mainly the blue-green algae (division Cyanophyta) and the green algae (division Chlorophyta), are also found in the soil, where they hasten the weathering of soil minerals. Algae normally occur near the soil surface, as they require light for photosynthesis.

5.6   Lichen

Lichens are made up of an alga and a fungus living in a symbiotic relationship. They occupy a special niche in the process of soil formation. They become established on rocks, which they very slowly dissolve, and also serve as traps for dust, which accumulates and may form the soil in which other plants become established. Organic acids are leached from living and dead lichens, increasing the rate of solution of the rocks to which they are attached. In making possible the growth of other plants, lichens hasten their own demise, for they soon become covered with litter and die.

5.7   Arthropods

Arthropods (phylum Arthropoda) are among the more obvious soil fauna. Into this large phylum fall the crayfish, mites, ants, centipedes, millipedes, sow bugs, insects, and numerous other species that physically cultivate the soil and contribute to its organic matter when they die. They are numerous and have a short life span and thus contribute to the soil hundreds of kilograms of organic matter per acre. Ants, in particular, cultivate great amounts of soil. In parts of Germany, the red wood ant has been introduced into many forests to compensate for centuries of soil degradation due to timber cutting. Ants also loosen and aerate the soil and destroy many kinds of insects that are harmful to plants. Many ants and termites are severe pests in tropical areas. The large termite mounds in Brazil seriously interfere with cultivation. The leaf-cutting ant is one of the most destructive of tropical pests. Mites and myriapods play an important role in the breakdown and decomposition of forest litter.

The segmented worms (phylum Annelida), commonly called earthworms, and round nonsegmented eelworms, or nematodes (class Nematoda of the phylum Aschelminthes), are both important in agricultural soils. An early study of earthworms by the famous biologist Charles Darwin concluded that about 0.25 to 0.5 centimeters of subsurface soil may be brought to the surface each year by earthworms and deposited as castings.

5.8   Nematode

Nematodes are economically important because some of them are pathogenic to crop plants and others distribute parasitic fungi and increase the extent of root-rot diseases. Some species kill bark beetles and other insect pests; some give rise to serious disorders in mammals and birds. A number of vertebrate animals, including many species of burrowing mammals, must be considered part of the soil fauna. Their effect is mainly one of cultivation, although their excrement contributes organic matter to the soil. Some, however, also destroy crops.

Badgers, gophers, moles, voles, foxes, shrews, mice, ground squirrels, woodchucks, and even some species of birds, such as woodcocks, frequent the soil for all or a part of their lives. When present in great numbers they may become serious pests. Moles, for example, do considerable

damage to turf grasses and certain farm crops. Beavers, through their dam building over the millennia, have had a profound impact on the forest soils of boreal areas.

6.0 SOIL CLASSIFICATION

Soil classification is the arrangement of individual soils into groups or classes on the basis of common characteristics to enable the determination of soil special variability and for land use planning.

6.1 Types of Classification - Generally there are three types of classification, namely; 1) Genetic classification, 2) Technical classification and 3) Taxonomic classification

1. Genetic classification - This method tends to separate soil on the basis of kind, extend and degree of soil evolution, e.g the Russian classification by Dockuchev

2. Technical classification - The grouping of soils for management or other interpretative purposes e.g. land capability classification, land suitability classification and the land use classification of the southwest Nigeria by Smyth and Montgomery, 1962

3. Taxonomic classification- this entails the grouping of soils on the basis of distinguishing characteristics. Although many soil classification systems exist; however, two system are widely used: The USDA Soil Taxonomy and the FAO/UNESCO legend. The French system (ORSTROM) is also commonly used in France and in Francophone Africa.

6.2 USDA Soil Taxonomy '7th approximation'-

This is an American system of soil classification developed by United State Department of Agriculture and the National Cooperative Soil Survey. Soils are classified from the most general categories to the most detailed, the most general categories is called the soil order. All world soils are placed into twelve orders. Soils are classified on the basis of diagnostic horizons, that are different from the O, A, B, C, horizon. A diagnostic horizon has a unique feature that reflects the soil development processes acting at a site. For instance, a mollic epipedon (a type of diagnostic horizon) is a organic-rich horizon typical of a mollisol soil. Categories in this system of classification are;

6.2.1 Hierarchy of Categories in the Soil Taxonomy

There are six levels in the hierarchy of categories: Orders (the highest category), suborders, great groups, subgroups, families and series (the lowest category) (USDA, 1978).

1. Orders - There are ten orders, differentiated on gross morphological features by the presence or absence of diagnostic horizons or features which show the dominant set of soil-forming processes that have taken place. The ten orders and their major characteristics are shown in Table below.

BRIEF CHARACTERISTICS OF SOIL ORDER

ORDER	CHARACTERISTICS
Vertisol	Vertisols are dark black soils rich in expandable clay minerals. The clay readily swells upon wetting and shrinks when dried. Though found in every type of climate, they are often found in steppe and wet/dry tropical climates where the soil develops deep cracks as it dries. Surface fragments fall into the cracks and are "swallowed" when the soil swells upon wetting. The soil then develops an "inverted profile" with organic material that is typically located near the surface of the profile is now found at depth.
Entisol	Soils lacking horizons because their parent material has only recently accumulated. Formed where the parent material is quartz sand, in which horizons do not easily form, have wide geographic distribution and can be found in any climate and under any vegetation. Entisols and Inceptisols are often found on floodplain, delta deposits, or steep slopes where parent material has difficulty accumulating.

Inceptisol	Inceptisols are soils just starting to show horizon development because the soil is quite young. You can see the differentiation of layers in an inceptisol formed on colluvium. Inceptisols, like Entisols, are found in any type of environment and are commonly found forming in alluvium on floodplains and delta deposits.
Aridisol	They are soils of arid and semiarid environments where moisture is scarce. They are typically light in color as there is little vegetation to add organic matter to the soil profile. A negative moisture balance in these soils inhibits eluviation. Calcification and salinization are important soil forming processes acting in these soils. Soil horizons are weakly developed and sodium is often high in concentration making them alkaline. The coarse texture of aridisols makes it difficult to retain much moisture. Aridisols can be quite fertile soil if irrigation is properly used. Used improperly, a salt crust can form on the soil. Most aridisols are used for grazing.
Spodosol	Spodosol soil is commonly found in cool, moist environments under coniferous forest vegetation. Surface litter composed of pine needles breaks down in the presence of water to form a weak organic acid. Acidic soil water removes base ions in solution to create an acidic soil. Easily dissolved materials are leached from surface layers leaving behind the most resistant material like quartz, creating an ashy-gray near-surface layer. Layers at depth are stained with iron and aluminum oxides.
Ultisol	Ultisols share many of the same properties as Oxisols. Highly weathered soils, they are often red/yellow in color reflecting the oxidation of iron and aluminum. Found in the moister portions of the Humid Subtropical climate, they have a illuvial clay layer which distinguishes them from Oxisols which do not.
Mollisol	Mollisols are among the most fertile soils on the Earth. Born under grassland vegetation, these soils are well-known for their dark brown to black organic rich surface layers. These soils have a granular structure and soft consistency when dry. Mollisols are rich in calcium and others nutrients, and generally posses high moisture retention. Calcium nodules are found near the base of the soil as calcium carbonate precipitates out of soil water. Mollisols are found in the drier portions of the humid continental climate through the steppe climate.

Alfisol	Alfisols are soils developed under temperate forests of the humid midlatitudes. Eluviation is moderate and base status is fairly high in these soils. Common to the humid continental and humid subtropical climates, these soils are well-developed and contain a subsurface layer of clay called an *argillic horizon*. Some alfisols are found in the wet/dry tropical climate of Africa, South America, Australia, and Southeast Asia. Having a favorable moisture balance and good fertility, they are very productive soils for agriculture. In fact, they have been successfully used for farming in China and Europe for thousands of years.
Oxisol	Oxisol soil is found in warm, rainy climates under broadleaf, evergreen vegetation like that found in the rain forest. Chemical weathering (especially oxidation) in the presence of warm temperatures combined with heavy rainfall creates a soil rich in iron and aluminum oxides called "sesquioxides". A rich diversity of decomposers, rapid uptake by vegetation, and heavy precipitation quickly removes nutrients from the soil. What is left is a nutrient poor soil, not well-suited for agriculture. Cleared of vegetation, the exposed surface is easily eroded.
Histosol	They have a very high content of organic matter in the dark upper layer of the profile. Found in many different environments from the tundra to the tropics, Histosols form in places where organic matter is slow to decompose and thus accumulates over time such as bogs and swamps. They are often "mined" for peat which is dried and burned as fuel.
Andisol	Andisols are soils developing in parent material containing at least fifty percent volcanic ash. The layers of ash can be seen in this Andisol. Naturally fertile soils, they support a dense natural cover in moist climates. Andisols occur around individual volcanoes created from andesite-rich magma. They are common on the volcanic islands and mountains of 'The Ring of Fire', that encircles the Pacific Ocean from North America through Japan.
Gelisol	Gelisol soil develops on permafrost and common to the tundra. These soils consist of mineral or organic material, or both and have experienced cryoturbation (frost churning) due to annual freeze-thaw cycles

2. **Suborders** - It is the next level of generalization. It permits more statements to be made about a given soil. In addition to morphological characteristics other soil properties are used to

classify the soil. The suborder focusses on genetic homogeneity like wetness or other climatic factors. There are 47 suborders within the 10 orders. The names of the suborders consist of two syllables. The first connotes the diagnostics properties; the second is the formative element from the soil order name. For example, an Ustalf is an alfisol with an ustic moisture regime (associated with subhumid climates).

3. Great groups - The great group permits more specific statements about a given soil as it notes the arrangement of the soil horizons. A total of 230 great groups (140 of which occur in the tropics) have been defined for the 47 suborders. The name of a great group consists of the name of the suborder and a prefix suggesting diagnostic properties. For example, a Plinthustalf is an ustalf that has developed plinthite in the profile. Plinthite development is selected as the important property and so forms the prefix for the great group name.

3. Subgroups - There are three kinds of subgroups:

(i) The typical subgroup which represents the central concept of the great group, for example Typic Paleustalfs.

(ii) Intergrades are transitional forms to other orders, suborders or great groups, for example Aridic Paleustalfs or Oxic Paleustalfs.

(iii) Extragrades have some properties which are not representative of the great group but do not indicate transitions, for example, Petrocalcic Paleustalf.

4. Families - The grouping of soils within families is based on the presence or absence of physical and chemical properties important for plant growth and may not be indicative of any particular process. The properties include particle size distribution and mineralogy beneath the plough layer, temperature regime, and thickness of rooting zone. Typical family names are clayey, kaolinitic, isohyperthermic, etc. There are thousands of families.

5. Series - The soil series is the lowest category. It is a grouping of soil individuals on the basis of narrowly defined properties, relating to kind and arrangement of horizons; colour, texture, structure, consistence and reaction of horizons; chemical and mineralogical properties of the horizons. The soil series are given local place names following the earlier practice in the old systems in naming soil series. There are tens of thousands of series.

Classes of Soil Moisture Regimes

Aquic soil moisture regime.—The aquic (L. *aqua*, water) soil moisture regime is a reducing regime in a soil that is

virtually free of dissolved oxygen because it is saturated by water. Some soils are saturated with water at times while

dissolved oxygen is present, either because the water is moving or because the environment is unfavorable for micro-organisms (e.g., if the temperature is less than 1 C); such a regime is not considered aquic.oIt is not known how long a soil must be saturated before it is said to have an aquic soil moisture regime, but the duration must be at least a few days, because it is implicit in the concept that dissolved oxygen is virtually absent.

Aridic and torric (L. *aridus*, dry, and L. *torridus*, hot

and dry) soil moisture regimes.—These terms are used for the same moisture regime but in different categories of the taxonomy Soils that have an aridic (torric) soil moisture regime o normally occur in areas of arid climates. A few are in areas of semiarid climates and either have physical properties that keep them dry, such as a crusty surface that virtually precludes the infiltration of water, or are on steep slopes where runoff is high. There is little or no leaching in this soil moisture regime, and soluble salts accumulate in the soils if there is a source. The limits set for soil temperature exclude from these soil moisture regimes soils in the very cold and dry polar regions

and in areas at high elevations. Such soils are considered to havem anhydrous conditions (defined earlier).

Udic soil moisture regime.—The udic (L. *udus*, humid) The udic soil moisture regime is common to the soils of

humid climates that have well distributed rainfall…. the amount of stored moisture plus rainfall is approximately equal to, or exceeds, the amount of evapotranspiration

Ustic soil moisture regime.—The ustic (L. *ustus*, burnt; implying dryness) soil moisture regime is intermediate between the aridic regime and the udic regime. Its concept is one of moisture that

is limited but is present at a time when conditions are suitable for plant growth. The concept of the ustic soil moisture regime is not applied to soils that have permafrost (defined above).

Xeric soil moisture regime.—The xeric (Gr. *xeros*, dry) soil moisture regime is the typical moisture regime in areas of Mediterranean climates, where winters are moist and cool and summers are warm and dry. The moisture, which falls during the winter, when potential evapotranspiration is at a minimum, is particularly effective for leaching. In areas of a xeric soil moisture regime, the soil moisture control section, in normal years, is dry in all parts for 45 or more consecutive days in the 4 months following the summer solstice and moist in all parts for 45 or more consecutive days in the 4 months followingthe winter solstice.

Classes of Soil Temperature Regimes

Following is a description of the soil temperature regimes used in defining classes at various categorical levels in this taxonomy.

Cryic (Gr. *kryos*, coldness; meaning very cold

soils).—Soils in this temperature regime have a mean annual temperature between 0 and 8 C but do not have permafrost.

Gelic (L. *gelare*, to freeze).— Soils in this temperature regime have a mean annual soil temperature at or below 0 C

Frigid.—A soil with a frigid soil temperature regime is warmer in summer than a soil with a cryic regime, but its mean annual temperature is between 0 and 8 C

Mesic.—The mean annual soil temperature is 8 C or higher but lower than 15 C,

Thermic.—The mean annual soil temperature is 15 C or higher but lower than 22 C,

Hyperthermic.—The mean annual soil temperature is 22 C or higher,

Isofrigid.—The mean annual soil temperature is lower than 8 oC.

Isomesic.—The mean annual soil temperature is 8 C orhigher but lower than 15 C.

Isothermic.—The mean annual soil temperature is 15 oC orhigher but lower than 22 C.

Isohyperthermic.—The mean annual soil temperature is

22oC or higher.

6.3   FAO Classification system

The Food and Agriculture Organization of the United Nations (FAO) developed a supra-national classification, also called World Soil Classification, which offers useful generalizations about soil pedogenesis in relation to the interactions with the main soil-forming factors.

The FAO soil map was a very simple classification system with units very broad, but was the first truly international system, and most soils could be accommodated on the basis of their field descriptions. The FAO soil map was intended for mapping soils at a continental scale but not at local scale.

FAO Soil Units

- Acrisol
- Andosols
- Arenosols
- Cambisols
- Chemozems
- Ferralsols
- Fluvisols
- Gleysols
- Greyzems
- Gypsisols
- Histosols
- Kastanozems
- Lithosols
- Luvisols
- Nitosols
- Phaeozems
- Planosols
- Podzols
- Podzoluvisols

- Rankers
- Regosols
- Rendzinas
- Solonchaks
- Solonetz
- Vertisols
- Yemosols