

MEKDELA AMBA UNIVERSITY



COLLEGE OF NATURAL AND COMPUTATIONAL SCIENCES

DEPARTMENT: BIOLOGY

Course Title: Soil Science

Course Code: Biol-M2051

Handout for Soil Science

CHAPTER 1

INTRODUCTION

Concepts of soils

The term “soil” has many definitions depending upon who is using the term. For example:

- To a farmer: soil is that portion of the earth’s surface which he can plough and grow crops on to provide him with food and fiber for his own needs and that of his animals;
- To an engineer: soil provides the foundation for all construction activities: roads, highway, buildings, runways, embankment of canals and drains;
- To an archaeologist: soil is a record of the past;
- To a pedologist: soil is a natural body;
- To an environmentalist: reflecting interest in the role of soils in hydrologic and other environmental cycles, regard soils as ‘living filter’ or ‘geologic kidney’;

Soil is, therefore, defined as a three-dimensional, dynamic, natural body occurring on the surface of the earth and which is a medium for plant growth and whose characteristics have resulted from the forces of climate and living organisms acting upon parent material, as modified by the relief over a period of time.

Soil science is the study of soil as a natural resource on the surface of the earth including soil formation, classification and mapping; physical, chemical, biological, and fertility properties of soils; and these properties in relation to the use and management of soils. It is an integrative science that interlinks knowledge of the atmosphere, biosphere, lithosphere and hydrosphere.

Why we study soils?

The major source of the three most basic human necessities, food, shelter, and clothing, can be traced back to soils. The soil is considered as a renewable resource, but if it is not managed properly, its ability to produce food, shelter and clothing material is greatly diminished. Improper management of the soil resource may lead to soil erosion, organic matter decline, soil salinization, and soil pollution. All of these problems impair the soil's productivity. We must understand our soil resource so that we recognize potential problems and seek solutions to any

problem.

Approaches in soil study

A. Pedological: involves certain aspects, such as origin of the soil, its classification and description. Consider soils as a natural entity (a bio-chemically weathered and synthesized product of nature) and it places minor emphasis on its immediate practical utilization.

B. Edaphological: it is the study of soils from the standpoint of higher plants. Consider various properties of soils as they relate to plant production. Determines reasons for variations in the productivity of soils and tries to find means of conserving and improving the productivity.

Pedology: the science that deals with the formation morphology, and classification of soil bodies as landscape components.

Edaphology: the science that deals with the influence of soils on living things, particularly plants, including human use of land for plant growth.

Types of soil in Ethiopia

Reading assignment

Available information indicates the existence of 18 major soil associations in Ethiopia according to FAO (UNESCO) description. Considering the total land area Lithosols, Nitosols, Cambisols, Regosols, Vertisols and Fluvisols cover approximately 17.1%, 12.2%, 11.6%, 10.9%, 10.0%, and 8.3% of the area of the country, respectively. When only arable lands are considered, Nitosols, Cambisols, and Vertisols are the major soil types occupying nearly 23%, 19%, and 18% of the total arable area, respectively.

CHAPTER 2

Soil formation

Weathering: physical and chemical

The weathering processes can be grouped into two main categories:

1. Physical weathering /disintegration/
2. Chemical weathering /decomposition/

Physical weathering

The disintegration of the parent material into smaller pieces with no change in mineral structure or composition is called Physical weathering. The smaller pieces retain most of the characteristics of the parent material. Only the grain sizes become smaller and smaller, but compositions are not changed and no new materials are formed.

Physical processes of weathering.

1. Abrasion

Water carrying suspended rock fragments has a scouring action on surfaces. The grinding action of glaciers, gravels, pebbles and boulders moved along a constantly abraded by fast-flowing streams.

2. Wetting and drying

Water penetrates into rocks and reacts with their constituent minerals.

3. Freezing and thawing

When water is trapped in the rock crevices and pores of a rock and is subjected to repeated freezing and thawing results in forces of expansion and contraction. The water near the rock surface, exposed to the cold air, freezes first, exerting pressure on the water remaining in the rock crevices and pores. This produces the mechanical effect of a wedge on the rocks. This process is called frost wedging. The pressure resulting from the growth of ice crystals within rocks is sufficiently high to crack the rocks apart. Water expands about 9 percent with freezing.

4. Thermal expansion and contraction of minerals

Rocks are composed of different kinds of minerals. When heated up by solar radiation, each different mineral will expand and contract a different amount at a different rate with surface temperature fluctuations. With time, the stresses produced are sufficient to weaken the bonds along grain boundaries, and thus flaking of fragments.

5. Release of confining pressure

Many of the rocks at the earth's surface were formed at considerable depth and later were uncovered by erosion of the overlying rock material. Deeply buried rocks are subjected to high, and equal, pressure on all sides. The vertical pressure lessens progressively as more and more of the overlying rocks are removed by erosion. This release of pressure allows the compressed rock to expand in the vertical direction, forming curved sheets parallel to the earth's surface. The formation of curved sheets of rock by release of pressure is called Exfoliation. Exfoliation results in the formation of rounded features called exfoliation domes.

6. Crystallization /salt crystal growth/

In arid environments, water evaporates at the surface of rocks and crystals form from dissolved minerals. Overtime, the crystals grow (they expand their volume) and exert a force great enough to separate mineral grains and break up rocks.

7. Animals and plants /living organisms/

The growth of plant and tree roots into rock crevices pries apart the rock in a process called root wedging. Burrowing animals and extensions of plant roots help in the formation of regolith by physically displacing rock particles and increasing the volume of openings in the regolith. These new openings allow more oxygen and water to penetrate the regolith, increasing the effects of chemical weathering.

In general, the main forces of physical weathering involved in all of the above processes are water, ice, wind, and temperature changes as well as plants and animals.

Chemical Weathering

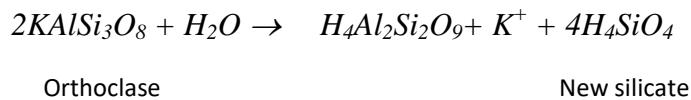
The decomposition of the original parent material, forming new minerals and soluble substances, is called chemical weathering. Chemical weathering consists of a number of reactions in which the ions that make up minerals combine with the ions of the atmosphere and hydrosphere. *In chemical weathering /decomposition/* definite chemical changes take place, soluble materials are released, and new minerals are synthesized or are left as resistant end products. The most common reactions in chemical weathering involve addition of water, carbon dioxide, and oxygen to rocks and sediments at the earth's surface.

Chemical processes of weathering

1. Hydrolysis

It is the reaction between minerals and the H^+ and OH^- ions formed during dissociation of water. Hydrolysis is perhaps the most important way by which chemical breakdown occurs. When water molecules at the mineral surface dissociate into H^+ and $(OH)^-$ ions, the mobile H^+ ions penetrate the crystal lattice, creating a charge imbalance, which causes cations such as Ca^{2+} , Mg^{2+} , K^+ and Na^+ to diffuse out. Hydrolysis is important in the weathering of a wide range of minerals, including the feldspars and micas.

Example:



Fate of K^+ thus released:

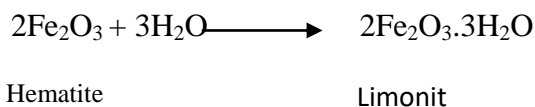
- Adsorption by soil colloids
- Uptake by plants
- Removal in drainage water

The H_4SiO_4 is soluble and can be:

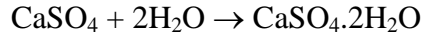
- Removed slowly in drainage water
- Recombined with other compounds to form secondary minerals such as the silicate clays.

2. Hydration

Hydration involves the rigid attachment of H and OH ions to the compound being attacked. In most cases, these ions become an integral part of the mineral crystal lattice. A good example of hydration is formation of limonite from hematite as follows:



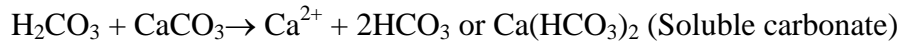
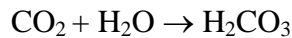
The evaporite mineral anhydrite, CaSO_4 , is converted by hydration to gypsum, $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$, by the following reaction when water is available:



This transformation involves an increase in volume to accommodate the water in the mineral structure; this change commonly causes distortion and deformation in the resulting gypsum crystals. When the products of hydration dry out because of varying weather conditions, dehydration may occur.

3. Carbonation and other acidic processes

Weathering is accelerated by the presence of acids, which increase the activity of hydrogen ions in water. When CO_2 dissolves in water, the carbonic acid produced hastens the chemical dissolution of some minerals such as calcite in limestone or marble as:



The regolith around plant roots is enriched in carbon dioxide by plant respiration. As rain soaks through this regolith, large quantities of carbonic acid are formed. The decomposition of minerals by reaction with carbonic acid is called **Carbonation**. The decomposition of carbonic acid to form the hydrogen and bicarbonate ions also frees some H^+ ions for hydrolysis reaction as:



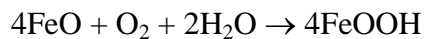
Other acids much stronger than carbonic acid are also present in most humid region soils. They include:

- Very dilute inorganic acids such as HNO_3 and H_2SO_4
- Some organic acids.

4. Oxidation

Oxidation is the combination of oxygen ions with cations such as iron, magnesium, calcium, sodium and potassium. Oxidation is particularly important in rocks that contain iron, an element that is easily oxidized. In some minerals, Fe is present in the reduced ferrous form (Fe^{2+}). If Fe^{2+} is oxidized to the ferric (Fe^{3+}) state, other ionic adjustments must be made because a three-valent ion is replacing a two-valent one. The adjustments result in a less stable mineral, which is then subject to both disintegration and decomposition.

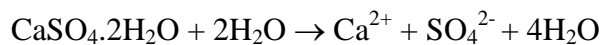
In other cases, ferrous iron may be released from the mineral and is almost simultaneously oxidized to the ferric form. An example of this is the hydrolysis of olivine and the release of ferrous oxide, which may be immediately oxidized to ferric oxide (Goethite).



When ions such as Fe^{2+} are removed or are oxidized within the minerals, the rigidity of the mineral structure is weakened and the mechanical breakdown is made easier.

5. Dissolution

Water is capable of dissolving many minerals by hydrating the cations and anions until they become dissociated from each other and surrounded by water molecules. Gypsum dissolving in water provides an example:



Soil profiles description, Major horizons and its nomenclature

Soil horizon is a layer of soil or soil material approximately parallel to the land surface and differing from adjacent genetically related layers in physical, chemical, and biological properties or characteristics such as colour, structure, texture consistence, kinds and number of organisms present, degree of acidity or alkalinity, and so on. The distinct characteristics are produced by soil-forming processes. The layering or horizon development eventually gives rise to natural bodies called soils. Each soil is characterized by a given sequence of these horizons. This sequence, i.e., a vertical exposure of a soil consisting of the horizons extending to the parent material is called a soil profile.

In all systems, there are two main kinds of horizons, organic and mineral, which are distinguished by their organic matter content. Master horizons (major horizons) are designated by capital letters, such as O, A, E, B, C, and R.

1. Organic Horizons

An organic horizon must have:

- Greater than 30 % organic matter when the mineral fraction has 50 % or more clay, or
- Greater than 20 % organic matter when the mineral fraction has no clay, or
- More than a proportionate amount of organic matter (between 20 and 30 %) if the clay content is intermediate.

Organic horizons are dominated by organic materials derived from the accumulation of dead plant and animal matter (litter) at varying stages of decomposition that have accumulated on the mineral soil surface. They can be divided into two based on the stage of decomposition of the debris:

O₁- organic horizon wherein the original forms of the plant and animal residues can be recognized by the naked eye. The organic material is not decomposed by microorganisms. The sources of organic matter can be distinguished.

O₂- the original forms of the plant and animal residues /debris/ cannot be recognized. The source of organic matter cannot be distinguished.

2. Mineral Horizons

A-horizons: Mineral horizons that formed at the surface or below an O layer, that exhibit obliteration of all or much of the original rock or depositional structure (in the case of transported materials). **A-horizons** show one or more of the following:

- An accumulation of humified organic matter intimately mixed with the mineral fraction and not dominated by characteristic properties of the E or B horizons or,
- Properties resulting from cultivation, pasturing or other similar kinds of disturbance.

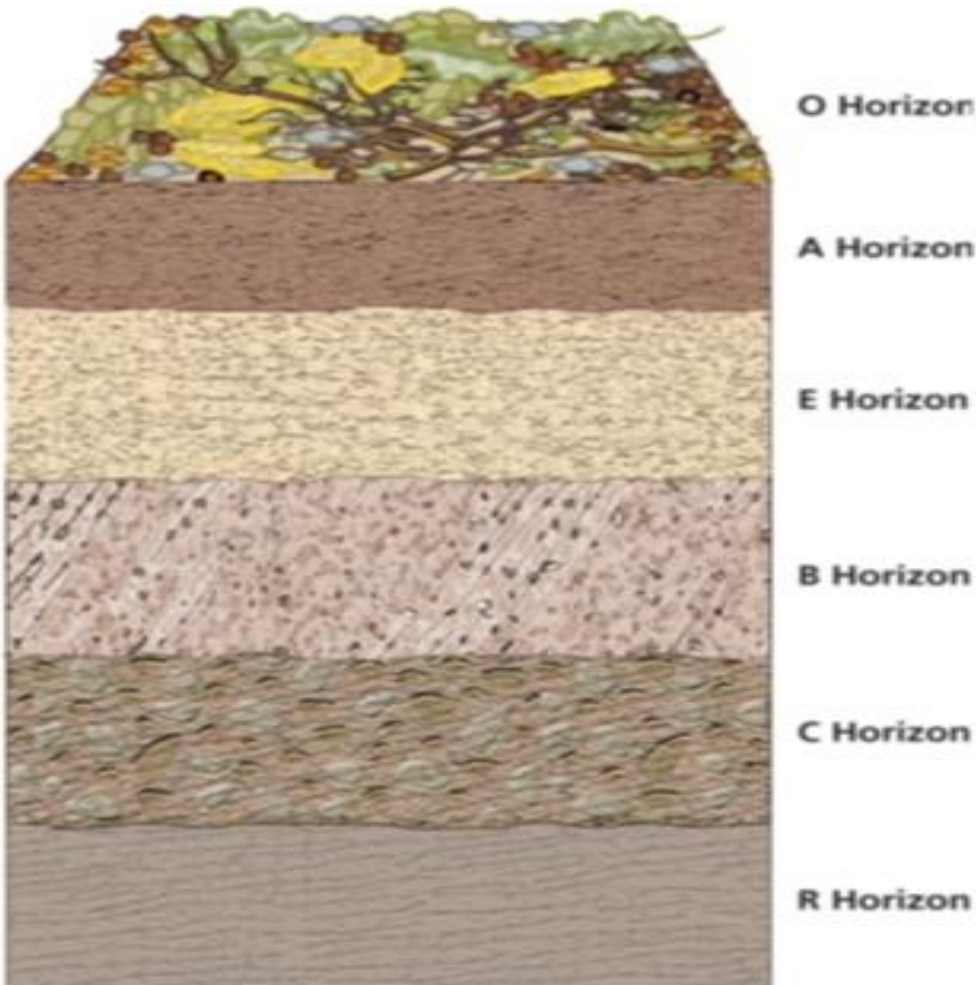
E-horizons: Mineral horizons in which the main feature is loss of silicate clay, iron, aluminum, or some combination of these, leaving a concentration of sand and silt particles and lighter colours. The horizons exhibit obliteration of all or much of the original rock structure. They are characterized as zones of maximum leaching or eluviations (eluvial from the Latin words ex or e, meaning out, and luv, meaning washed).

B-horizons: Horizons in which the dominant feature(s) is one or more of the following:

- An illuvial concentration of silicate clay, iron, aluminium, carbonates, gypsum, or humus
- Removal of carbonates
- A residual concentration of sesquioxides or silicate clays, alone or mixed, that has formed by means other than solution and removal of carbonates or more soluble salts
- Coatings of sesquioxides adequate to give darker, stronger, or redder colours than overlying and underlying horizons but without apparent illuviation of iron
- An alteration of material from its original condition that obliterates original rock structure, that form silicate clay, liberates oxides, or both, and that forms a granular, blocky, or prismatic structure
- Any combination of these.

C-horizons: Mineral horizons that are little altered by soil forming processes. They lack properties of O, A, E, or B-horizons. The designation C is also used for saprolite, sediments, or bedrock not hard enough to qualify for R. The material designated as C may be like or unlike the material from the A, E, and B-horizons are thought to have formed.

R Layers: Consolidated bedrock (hard bedrock), such as granite, basalt, quartzite, sandstone, or limestone. Small cracks, partially or totally filled with soil material and occupied by roots, are frequently present in the R layers.



Soil forming factors

Five soil-forming factors are generally recognized: parent material, organisms, climate, topography, and time.

1. Parent materials

The rock or sediment on which soils develop is its parent material. It is a starting point for the process of soil formation. In the early stages of their formation, soils are not much different from their parent materials. With time, fewer and fewer features of the parent material persist, and ultimately the soil takes on an identity of its own. The nature of parent material is a base line that must be known in order to assess the amount of soil formation that has taken place. Some parent materials rapidly and with little energy are converted to soil, whereas others are resistant to weathering. The role of parent material is most apparent from studies of soils formed under similar conditions on different parent materials or lithosequences, which provide a basis for understanding the role of parent material as a factor.

2. Climate

Soils tend to show a strong geographical correlation with climate, especially at the global scale. Energy and precipitation strongly influence physical and chemical reactions on parent material. Climate also determines vegetation cover which in turn influences soil development. Precipitation also affects horizon development factors like the translocation of dissolved ions through the soil. As time passes, climate tends to be a prime influence on soil properties while the influence of parent material is less.

Climate, interacting with vegetation, also affects soil chemistry. Pine forests tend to dominate cool, humid climates. Decomposing pine needles in the presence of water creates a weak acid that strips soluble bases from the soil leaving it in an acidic state. Additionally, pine trees have low nutrient demands so few soil nutrients are taken back up by the trees to be later recycled by decaying needle litter. Broadleaf deciduous trees like oak and maple have higher nutrient demand and thus continually recycle soil nutrients keeping soils high in soluble bases.

Climate is perhaps the most influential of the four factors acting on parent material because it determines the nature and intensity of the weathering that occurs over large geographical areas. The principal climatic variables influencing soil formation are **effective precipitation** and **temperature**, both of which affect the rates of chemical, physical, and biological processes.

3. Living Organisms/Biota

Organic matter accumulation, biochemical weathering, profile mixing, nutrient cycling and aggregate stability are enhanced by the activity of organisms in the soil. Vegetative cover reduces natural soil erosion rates, thereby slowing down the rate of mineral surface soil removal. Organic acids produced from certain types of plant leaf litter bring iron and aluminum in to solution by complexation and accelerate the down ward movement of these metals and their accumulation in the B horizon.

4. Relief (Topography)

Topography influences soil formation primarily through it's associated with water and temperature relations. Important elements of topography affecting soil formation are the slope (gradient) and aspect (slope azimuth).

Soils within the same general climatic area developing from similar parent material and on steep hillsides typically have thin developing horizons because less water moves down through the profile as a result of rapid surface runoff and because the surface erodes quite rapidly. Similar materials on gently sloping hillsides have more water passing vertically through them than do materials on steeper slopes. The profile on gentle slopes generally is deeper, the vegetation more luxuriant, and the organic matter level higher than in similar materials on steep topography.

Materials lying in landlocked depressions, on the other hand, receive runoff waters from surrounding higher areas. Such conditions favor greater production of vegetation but slower decomposition of dead plant remains because of oxygen deficiency in waterlogged/saturated soil; this results in soils with large amounts of organic accumulations. If the accumulating water dissolves salts from surrounding soils, the depression may become a salt marsh or it may develop toxic salt conditions with no plants at all. When soils on the watershed are strongly acid, iron may leach from them and be deposited in depressions to form bog iron (Limonite). Alkaline soils on sloping topography in humid regions may result in lime being eroded into depressions and lead to the formation of marl.

5. Time

Time acts on soil formation in two ways:

1. The value of a soil-forming factor may change with time (e.g. climatic change, new parent material).
2. The extent of a pedogenetic reaction depends on the time for which it has operated.

The length of time required for a soil to develop the distinct layers called genetic horizons depends upon many interrelated factors of climate, nature of the parent material, the organisms, and topography. Horizons tend to develop most rapidly under warm, humid, forested conditions where there is adequate water to move colloids. Under ideal conditions, a recognizable soil profile may develop within 200 years; under less favorable circumstances, the time may be extended to several thousand years. Soil development proceeds at a rate determined by the effects of time plus the intensities of climate and biota, further modified by the effect of land relief (topography) on which the soil is situated and the kind of parent material from which it is developing.

Pedological Processes

The change from minerals or rocks to soil involves changes over centuries of weathering. The many changes can be summarized as:

A. Transformation

Transformations occur when soil constituents are chemically or physically modified or destroyed and others are synthesized from the precursor materials. Many transformations involve weathering of primary minerals; disintegrating and altering some to form products recombine into new materials that include additional types of silicate clays and hydrous oxide of iron and aluminum. Other important transformations involve the decomposition of organic residues and the synthesis of organic acids, humus, and other products.

B. Translocations

Translocations involve the movement of inorganic and organic materials laterally within a horizon or vertically from one horizon up or down to another. Water, either percolating down with gravity or rising up by capillary action, is the most common translocation agent. The materials moved within the profile include dispersed fine clay particles, dissolved salts, and dissolved organic substances. Translocations of materials by soil organisms also have a major influence on soil genesis.

C. Additions

- Inputs of materials for the developing soil profile from outside sources are considered addition.
- The input of organic matter from fallen plant leaves is common example. Water (rainfall, irrigation)
- Nitrogen from bacterial fixation (biological nitrogen fixation) are some example

D. Losses

Materials are lost from the soil profile by leaching to ground water, erosion of surface materials or other forms of removal. Erosion, a major loss agent, often removes the finer particles (humus, clay and silt), leaving the surface horizon relatively sandier and less rich in organic matter than before.

Soil Horizon Differentiation

Horizons formed because of differences in weathering with depth, amounts of humus accumulated and colloids Translocated by water to deeper depths, and losses of colloids from the profile in percolating water. Usually a horizon is separated from others because:

- The horizon has accumulated more humus and is dark-coloured.
- The horizon has had some of its clay and humus moved to greater depths and is a leached horizon.
- The horizon has accumulated or produced more clay than is in other horizons.
- The horizon has accumulated some secondary minerals (calcite, gypsum, and iron oxides), forming horizons such as lime zones or hardpans.

CHAPTER 3

Soil classification

Fundamental of soil classification

Throughout history people have used various systems to name and classify soils. From the time crops were first cultivated; humans noted differences in soils, classified and grouped them according to their suitability for different uses such as *black cottonsoils*, *rice soils*, or *olive soils*. Other soil names still commonly used today have geological connotations suggesting the parent material from which soils have been formed: *limestone soils*, *piedmont soils*, and *alluvial soils*. But all such groupings are not helpful to organize our scientific knowledge of soils, or for defining the soils as natural bodies and their relationship with each other.

The main purposes of soil classification are to:

- ❖ Organize knowledge about groups of pedons,
- ❖ Provide a framework for establishing relationship among soils and their environment,
- ❖ Establish groupings of soils for which useful and reliable interpretations can be made, and
- ❖ Foster global communication about soils among soil scientists and all people concerned with the management of land and conservation of soil resources.

Concept of individual soil: compared to most sciences, the organized study of soils is rather young. Russian scientist V.V. Dokuchaev and his associates first conceived the idea that soils exist as a natural bodies. Later in 1920, C. F. Marbut of US Department of Agriculture (USDA) accepted this concept and developed a classification based on this principle. The natural body concept of soils recognizes the existence of individual entities, each of which we call a soil.

Pedon and polypedon: - there are no sharp demarcations between one soil and another. Rather, there is gradation in properties as one move from one soil to another, which can be compared to gradual change in wavelengths of light as we move from one color to other in a light spectrum. Consequently, it is necessary to characterize a soil in terms of an imaginary 3- dimensional unit called Pedon, which can be defined as the smallest sampling unit that displays the full range of properties or characteristics of particular soil pedons occupy from about 1 to 10m² of land area.

However, a soil unit in a landscape usually consists of a group of very similar pedons, closely associated in the field together. Such a group of similar pedons, or Polypedons, is recognized as a landscape component termed as soil.

Diagnostic horizons: these are the horizons with characteristics typical to that horizon located either at soil surface or in sub-surface.

(A) Diagnostic surface horizons: - the diagnostic horizons that occur at the soil surface are called epipedons. On the basis of their characteristics the following epipedons are recognized:

1. **Mollicepedon:** it is a mineral surface horizon noted for its dark color associated with its accumulated organic matter (more than 0.6% organic throughout), for its thickness (more than 25 cm) and high base saturation (more than 50%)
2. **Umbricpedon:** it has the same general characteristics as the mollicepedons except the percentage base saturation is less than 50%
3. **Ochricpedon:** it is a mineral horizon that is either thin, light in color, or low in organic matter to be either a mollic or umbric horizon.
4. **Melanicpedon:** it is a mineral horizon that is very black in color due to its high organic matter (more than 6% organic carbon), more than 30cm thick and light in weight
5. **Histicpedon:** it has a 20-60cm thick layer of organic soil materials overlying a mineral horizon

(B) Diagnostic sub- surface horizons: - these are the diagnostic horizons found below the soil surface. The more commonly found sub-surface horizons are:

1. **Argillic horizon:** it has a sub-surface accumulation of high-activity silicate clays which have moved down wards from upper horizons. The clay is usually found as coating on the surface of sub- surface structures termed as clay skins or argillans.
2. **Natric horizon:** likewise argillic horizon it also has clay accumulation but the clays are accompanied by more than 15% exchangeable sodium on colloidal complex. It has columnar or prismatic soil structural units
3. **Kandic horizon:** it has the accumulation of Fe and Al oxides and has low activity of silicate clays

4. **Oxic horizon:** it is a highly weathered sub-surface horizon that is very high in Fe and Al oxides. CEC is less than 16cmols/kg clay and the horizon is at least 30cm deep
5. **Spodic horizon:** it is an illuvial horizon characterized by accumulation of colloidal organic matter and Al oxide (with or without Fe oxides).
6. **Sombric horizon:** it is also an illuvial horizon, dark in colour because of high organic matter accumulation and has low degree of base saturation
7. **Albic horizon:** it is a light colored eluvial horizon that is low in clay and oxides of Al and Fe

Distribution and origin of soil orders.

There are six categories of classification in soil taxonomy: (1) order = 12, the highest (broadest category), (2) suborder = 68, (3) great group = 444, (4) subgroup = ~2500, (5) family = ~8000, (6) series = ~ 23,000 (in U.S.)

The names of classification units in soil taxonomical classes convey a concept of soil characters or genesis. Therefore, the name automatically describes the general kind of soil being described. The classification of world soils, therefore, follows an order similar to the scientific classification of organisms.

1. Soil orders: - Each of the world soils are assigned to one of the 12 soil orders, largely on the basis of soil properties that reflect a major course of development, with the considerable emphasis on the presence or absence of the diagnostic horizons. The names of orders are combinations of **formative elements**, which generally define the characteristics of the soil, and the ending **soils** (from the Latin solum, soil). For instance, soils of the order Aridisols (from the Latin aridus, dry) are characteristically dry soils in arid regions.

2. Suborders: - within each soil order, soils are grouped into suborders on the basis of soil properties that reflect major environmental controls on current soil forming processes. E.g *Ustolls* are dry Mollisols. Likewise soils in *Udult* suborder are moist Ultisols.

3. Great group: - the great groups are subdivisions of suborders; so far about 444 great groups are recognised. They are defined largely by presence or absence of diagnostic horizons and their

arrangement. For instance, *Argiudolls* are Mollisols of the **Udolls** suborder characterised by an argillic horizon.

4. Subgroups: - are subdivisions of a great group. Approximately about 2500 subgroups are recognised. The central concept of a great group makes subgroups, termed a typic subgroup. For instance, *TypicHapludoll* subgroup typifies the **Hapludoll** great group. Other subgroups may integrate characteristics with the main concept of the great group e.g a *hepludool* with restricted drainage may be known as **AquicHapludool** subgroup.

5. Families: - within a subgroup, soils fall into a particular family if, at a specified depth, they have similar physical and chemical properties affecting the growth of plant roots. The properties considered are: type of mineral and its activities in terms of retention of ions, temperature, depth etc... A *TypicArgiudoll* with loamy texture, having a mixture of moderately active clay *minerals* and with annual soil temperature (at 50cm) between 8 and 15⁰cis classed as *TypicArgiudoll loamy*, mixed, active, mesic family.

6. Series: - the series category is the most specific unit of classification. It is the subdivision of family and each soil series is defined by a specific range of soil properties involving primarily the kind, thickness, and arrangement of horizons. Features such as hardpan, calcium carbonate accumulation, striking colour characteristics etc... help in series identification.

Systems of soil classification

USDA systems for soil classification

According to the United States Department of Agriculture (USDA) system of soil classification, world soils are classified into twelve soil orders. They are Entisols, Inceptisols, Spodosols, Aridisols, Andisols, Mollisols, Vertisols, Alfisols, Oxisols, Ultisols, Gelisols and Histosols.

Entisols are the youngest soils that have little or no evidence of the development of pedogenic horizons. Such soils are common on sloppy lands where there is not stable condition for horizon differentiation.

Inceptisols are soils that are at inception stage of horizon differentiation. They have A and C horizons. Such soils are common on sloppy lands where there is relatively stable condition for horizon differentiation than Entisols.

Spodosols are soils that are developed under coniferous forest. They have elluvial (E) and Illuvial (B) horizons besides to A and C horizons.

Aridisols are soils of the dry areas that have higher concentration of salts in their horizons. There is little or no leaching of nutrients due to rainfall scarcity.

Andisols are soil developed on volcanic ash parent materials. These soils are found in the rift valley.

Mollisols are very fertile soils developed under grassland condition on parent materials containing calcite.

Vertisols are soils developed under lowland conditions and contain montmorillonite clay. These soils are known by extreme stickiness and plasticity when wet and develop deep wedge shaped cracks when dry.

Alfisols are red well drained soils that contain oxides of aluminum and iron. These soils are characterized with subsurface zone of clay accumulation. These soils develop in medium rainfall areas and are very suitable for agricultural production and road construction.

Oxisols contain very high amounts of oxides of iron and aluminum and develop under high rainfall humid and hot climatic condition. These soils are extremely red in color.

Ultisols are the most weathered soils that develop under high rainfall humid and hot climatic condition. They are considered to be the oldest soils.

Gelisols are soils developed under frigid conditions in the North Pole areas.

Histosols develop on peat parent materials on marshy lands. These soils contain very high amount of organic matter.

FAO systems for soil classification

Soil	Characteristic
Acresols	Low base status soils with argillic horizon
Andisols	Soils formed in volcanic ash that have dark surface
Cambisols	Soils with slight colour, structure, or consistency change due to weathering
Ferralsols	Highly weathered soils with sesqui-oxide rich clays
Fluvisols	Water deposited soils with little alteration
Lithosols	Shallow soils over hard rock
Luvisols	Medium to high base saturation soils with argillic horizon

Nitrosols	Soils with low C.E.C clay in argillic horizons
Regosols	Thin soils over unconsolidated material
Vertisols	Self mulching, inverting soils, rich in smectite clay

CHAPTER 4

Physical properties of soil

Soil texture

Texture is defined as the relative proportions of the different particle size fractions, specifically referred to as sand, silt, and clay. The distribution of particle sizes determines the soil texture, which can be assessed in the field or by a particle-size analysis in the laboratory. A field analysis is carried out in the following way: a small soil sample is taken, water is added to the sample, and it is kneaded between the fingers and thumb until the aggregates are broken down. The guidelines to determine the particle class are as follows:

Sand: Sand particles are large enough to grate against each other and they can be detected by sight. Sand shows no stickiness or plasticity when wet.

Silt: Grains cannot be detected by feel, but their presence makes the soil feel smooth, soapy, and only very slightly sticky.

Clay: A characteristic of clay is the stickiness. If the soil sample can be rolled easily and the sample is sticky and plastic when wet (or hard and cloddy when dry) it indicates high clay content. Note that high organic matter content tends to smooth the soil and can influence the feeling for clay.

Soil texture is typically permanent and an intrinsic property of the soil. It is often used to characterize the physical make-up of the soil, having a bearing on soil behavior.

1. United States Department of Agriculture (USDA)

- (a) Sand
 - Very coarse sand: 1-2 mm
 - Coarse sand: 0.5-1.0 mm
 - Medium sand: 0.25-0.5 mm
 - Fine sand: 0.1- 0.25 mm
 - Very fine sand: 0.05-0.1 mm
- (b) Silt: 0.002-0.05 mm
- (c) Clay: < 0.002 mm

2. United States Public Roads Administration (USPRA)

- (a) Sand
 - Coarse sand: 0.25-2.0 mm
 - Fine sand: 0.05- 0.25 mm
- (b) Silt: 0.002 – 0.05 mm
- (c) Clay: < 0.002 mm

3. International Society of Soil Science (ISSS)

- (a) Sand
 - Coarse sand: 0.2 –2.0 mm
 - Fine sand: 0.02 –0.2 mm
- (b) Silt: 0.002 – 0.02 mm
- (c) Clay: < 0. 002 mm

4. British Standards Institution (BSI)

- (a) Sand
 - Coarse sand: 0.6 –2.0 mm
 - Medium sand: 0.2 – 0.6 mm
 - Fine sand: 0.06 – 0.2 mm
- (b) Silt
 - Coarse silt: 0.02-0.06 mm
 - Medium silt: 0.006 – 0.02 mm
 - Fine silt: 0.002 – 0.006 mm
- (c) Clay: < 0.002 mm

Mechanical analysis is the procedure for determining the particle size distribution of a soil sample. The first step in this procedure is to disperse the soil sample in an aqueous suspension. The primary soil particles, often naturally aggregated, must be separated and made discrete by removal of cementing agents (such as organic matter, calcium carbonate, or iron oxides) and deflocculating the clay. Organic matter can be removed by oxidation with hydrogen peroxide; Calcium carbonate can be dissolved by addition of hydrochloric acid.

Textural triangle

Soil texture in the laboratory is determined using a texture triangle. Textural triangle is a diagram that shows the relationship between the textural class name of a soil and its particle size distribution. For example, a particle size distribution of 33 % clay, 33 % silt, and 33 % sand would result in the soil texture class 'clay loam'.

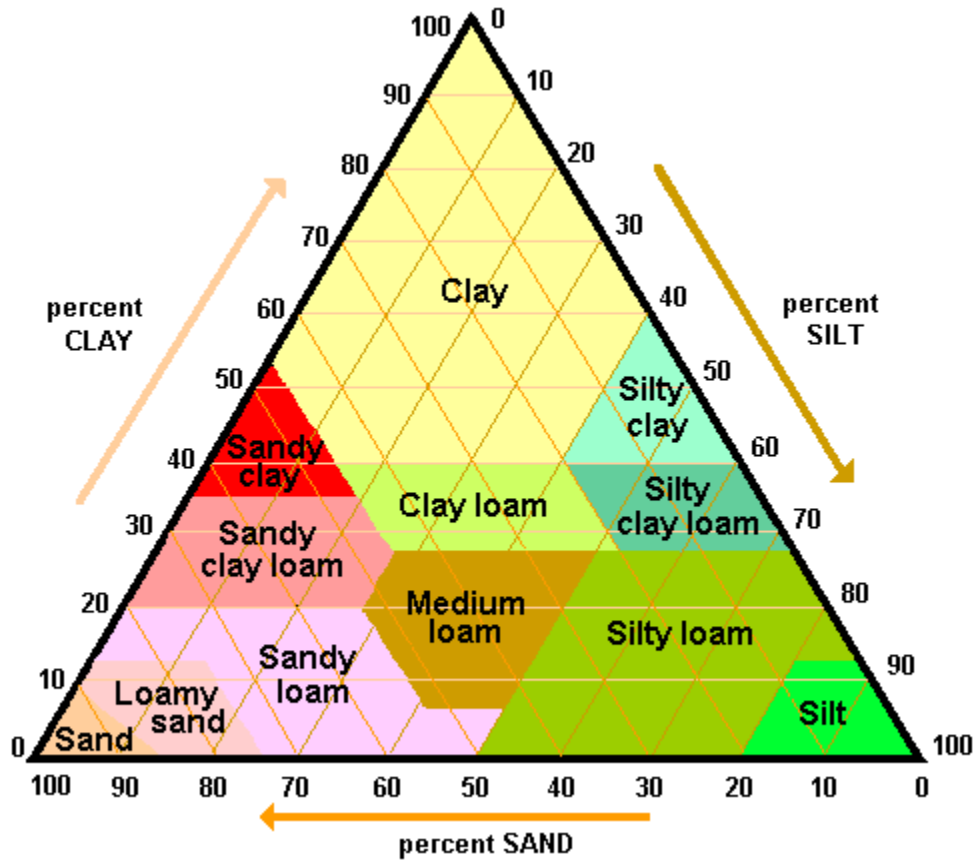


Figure of Textural triangle.

Example: naming a textural class of soil containing 33 % clay 33 % silt and 34 % sand

1. Take a point on percent clay at 33 % and draw a line parallel to the base of the triangle.
2. Take a point showing percent silt 33 and draw a line parallel to the left side of the triangle.
3. Take percent sand 34 and draw a line parallel to the right side of the triangle.

They must meet at the same point. This gives the textural class of the soil, which is ‘clay loam’ in the present case.

Significance of soil texture

The fine and medium-textured soils (e.g. clay loams, silty clay loams, sandy clay loams) are favorable from an agricultural viewpoint because of their high available retention of water and exchangeable nutrients. In fine pores the water is strongly adsorbed and not available for plants, i.e. cohesion and adhesion water occupy the micropore space and they are retained in soil by forces that exceed gravity. In medium-sized pores the available water content is high, whereas in macropores water is more weakly held and percolation is high (gravitational water). In silty soils the distribution of macropores, medium-sized, and fine pores is optimal leading to more available water content.

In general, coarse-textured soils permit rapid infiltration because of the predominance of large pores, while the infiltration rates of finer-textured soils is smaller because of the predominance of micropores. Other factors, like the compaction of the soil, management practices, vegetation, and saturation of the soil have also a significant impact on infiltration and have to be considered.

Soil texture has an impact on soil temperature. Fine-textured soils hold more water than coarse-textured soils, which considering the differences in the specific heat capacity results in a slow response of warming up of fine-textured soils compared to coarse-textured soils.

Soil Structure

Physically a soil is a mixture of inorganic particles, decaying organic materials, water and air. The inorganic primary particles of various sizes (sand, silt or clay fraction) generally cluster together to form complex and irregular patterns of secondary particles which are called aggregates or peds. Structure refers to the arrangement of sands, silts and clays into stable (cemented) aggregates. Aggregates are secondary units or granules composed of many soil particles held together by cementing agents, such as organic substances, iron oxides, carbonates, clays, and/or silica. A soil structure is the product of processes that *aggregate, cement, compact* or *unconsolidate* soil material. In essence, soil structure is a physical condition that is distinct from that of the initial material from which it is formed, and can be related to processes of soil formation. The peds or aggregates are separated from the adjoining peds by surfaces of *weakness*.

3.2.1. Classification of soil structure

Soil may be either:

1. **Structured**, or
2. **Structureless**.

Structureless soils have no recognizable natural aggregates. Structureless material may be:

1. **Single grained**

Single grained material results when individual soil particles show little or no tendency to adhere to other particles, an example being coarse-textured sandy soils.

2. **Massive**

Massive materials result when the soil is composed of finer particles such as silts and especially clays. The natural cohesiveness of these size separates binds the individual particles together. The massive materials show little or no tendency to break apart under light pressure into smaller structural units.

Classification of soil structure is based on:

1. **Type (form) of structure:** This is determined by the shape and arrangement of peds.
2. **The Size (class) of structure:** as differentiated by the size of the peds.
3. **The grade of structure:** as determined by the distinctiveness (differential between cohesion within peds and adhesion between peds) and durability of the peds. It relates to the degree of aggregation or the development of soil structure. In the field a

classification of grade is based on a finger test (durability of peds) or a crushing of a soil sample.

1. Type of soil structure

Based on the shape and arrangement of peds or aggregates soil structure is classified into four principal types-*Platelike*, *prismlike*, *blocklike* and *spheroidal* types.

- i. Platelike*-the aggregates are arranged in relatively thin horizontal plates, or leaflets. It is often formed from parent materials and can also result due to compaction by heavy tractor on clayey soils. It requires force to separate horizontal soil layers. Frost heaving, fluctuating water tables, compaction, and thin layering of different textured allivium or lacustrine material can aid the formation of plates.
- ii. Prismlike*-vertically oriented aggregates, occurring commonly in subsurface horizons (B-horizons) of arid and semiarid regions. The prisms having rounded tops are called **columnar** structure, and the prisms having angular tops and relatively flat horizontally are called **prismatic** structures.
- iii. Blocklike*-the aggregates look like blocks having irregularly six-faced and more or less equal in three dimensions. Block like structures in which the cubes of the blocks have sharp edges and distinct rectangular faces are called **angular blocky** and the ones in which some rounding of the cubes occur are called **subangular blocky**. The block like structure types are common in B-horizons, particularly in humid regions, and may also occur in A-horizons. Root penetration, soil drainage and aeration are some of the factors that influence the formation of these structure types.
- iv. Spheroidal*- these are characteristic of surface (A) horizons high in organic matter, grasslands and subject to wide and rapid changes. When the peds are relatively nonporous they are called **granules** and the porous granules are termed **crumbs**.

2. Class of soil structure

The various sizes of peds are designated by class of soil structure as very fine, fine, medium, coarse and very coarse.

Table of Classification of soil structure considering size and type of peds

Size	Angular and subangular blocky structure [mm] diameter	Granular and crumb structure [mm] diameter	Platy structure [mm] width	Prismatic and columnar structure [mm] diameter
Very fine	< 5	< 1	< 1 (very thin)	< 10
Fine	5 – 10	1 - 2	1 - 2 (thin)	10 - 20

Medium	10 – 20	2 - 5	2 - 5	20 - 50
Coarse	20 – 50	5 - 10	5 - 10 (thick)	50 - 100
Very coarse	> 50	> 10	> 10 (very thick)	> 100

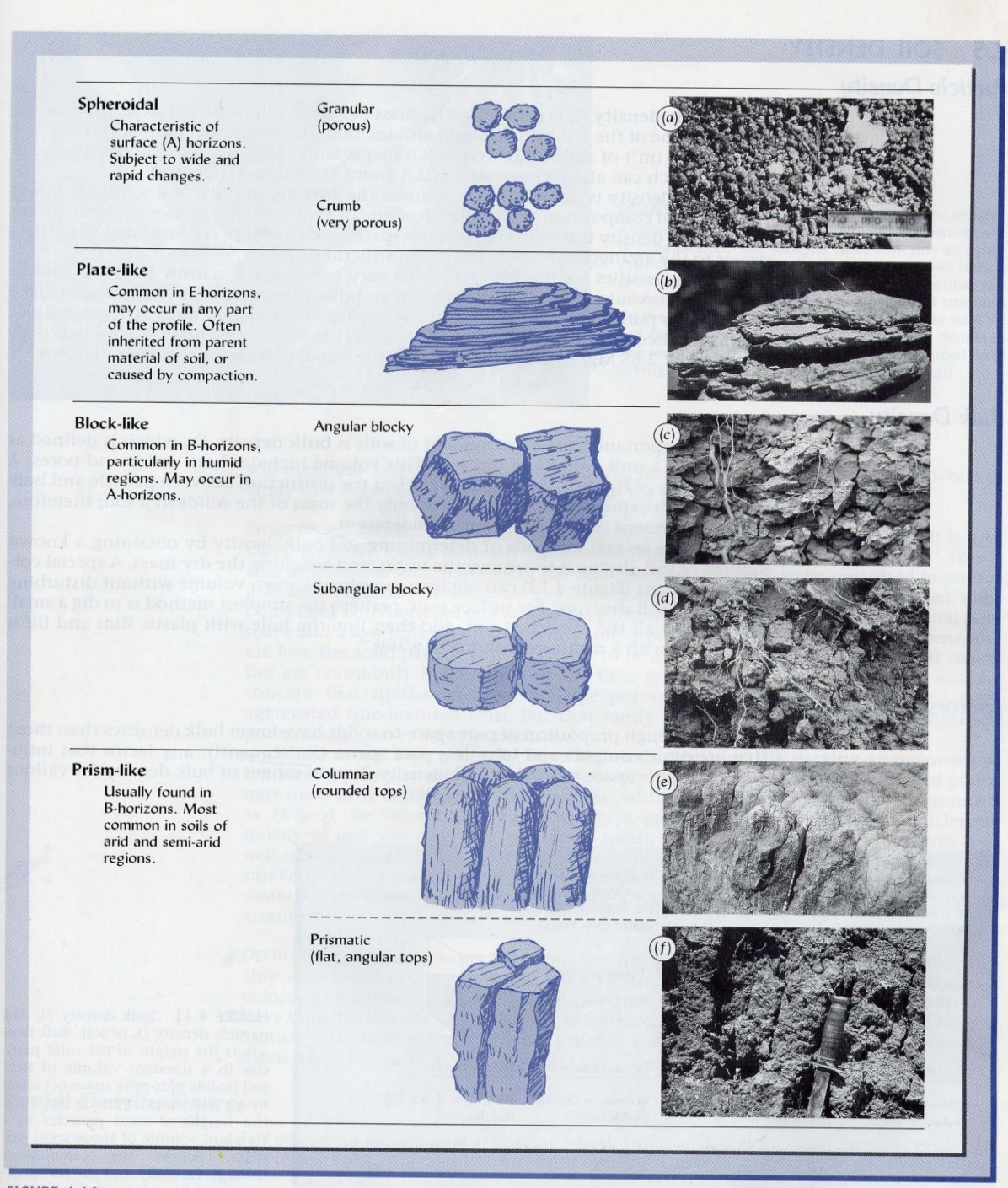


FIGURE 4.10 The various structure types (classes) found in soil horizons. (a) Spheroidal structure. (b) Plate-like structure. (c) Angular blocky structure. (d) Subangular blocky structure. (e) Columnar structure. (f) Prismatic structure.

Figure of Structure types

3. Grade of soil structure

The degree of aggregation of soil structure is the distinctness or durability with which the peds are observed, viz. weak, moderate and strong. The description of each grade is presented below in the table.

Table of Classification of soil structure considering grade and type of peds

Grade	Abbreviation	Description
Structureless	0	No observable aggregation or no orderly arrangement of natural lines of weakness
Weak	1	Poorly formed indistinct peds
Moderate	2	Well-formed distinct peds, moderately durable and evident, but not distinct in undisturbed soil
Strong	3	Durable peds that are quite evident in undisplaced soil, adhere weakly to one another, withstand displacement, and become separated when soil is disturbed

Effects of soil structure on soil conditions and properties

Soil structure affects the pattern of pores. This in turn affects the pore size distribution, which greatly influences water movement, heat transfer, and aeration. Moreover, structure modifies the influence of soil texture with regard to water and air relationships and the ease of root penetration. Activities such as, timber, grazing, tillage, trafficking, drainage, liming and manuring impact soils largely through their effect on soil structure, especially in surface horizons.

Factors affecting soil structure

A number of factors affect soil structure in different ways. The most important ones are the following:

1. Vegetation

During growth, plant roots pushing through the soil tend to compress soil particles into small aggregates and break the larger aggregates present in the soil. Root exudates help to bind the soil particles together into aggregates. The dehydration of the soil in the vicinity of the root system as water is absorbed by the plants causes local shrinkage and formation of surface fracture.

2. Soil Fauna and microorganisms

The beneficial effects on soil aggregation and its stability originate from the integrated activity of soil fauna, microorganisms and vegetation.

- Fungi and actinomycetes cause mechanical binding of the aggregates by the mycelia they produce.

- The metabolic processes of the microorganisms synthesize complex organic molecules, which have cementation effect. Thus, the microbial products like bacterial gum and microbial polysaccharides play a major role in soil aggregate stabilization.
- Products of microbial decomposition of organic materials added to soil such as humic acids, colloidal proteins and cellulose materials also produce stabilizing effect.

3. Cation effects

The cations adsorbed by the soil colloids greatly affect aggregate formation. If Na is the dominant cation adsorbed, the soil particles do not come together to form structural aggregates but they repel each other causing dispersed condition. The divalent cations like Ca^{2+} and Mg^{2+} , on the other hand form electropositive links between the electronegative soil particles causing the soil colloids to come together and form small aggregates called floccules.

4. Organic Matter

Organic matter stimulates the formation and stabilization of granular and crumb type aggregates. Organic compounds, such as polysaccharides, chemically interact with the clays and form bridges between the individual soil particles and bind them together in water-stable aggregates.

5. Clay particle and clay-organic interaction

Three distinct groups of colloidal materials have cementation and aggregating effects-clay particles themselves, inorganic colloids of iron and aluminum oxides and organic colloids.

Clay-to-clay particle interaction takes place due to electrostatic and van der Waal's forces leading to the formation of secondary particles. The linkage established by polyvalent exchangeable cations on the clay particles influence the formation of aggregates. Also, polyvalent cations serve as bridges to form clay-organic complexes also leading to the formation of stable aggregates.

Soil Consistence

Consistence refers to the resistance of the soil to mechanical stress or manipulation at various moisture contents. It indicates the cohesion among soil particles and adhesion of soil to other substances or the resistance of the soil to deformation. Whereas soil structure deals with the arrangement and form of peds, consistence deals with the strength and nature of the forces between particles. Consistence is described for three moisture levels: **wet**, **moist**, and **dry**. The *stickiness* describes the quality of *adhesion* to other objects and the *plasticity* the capability of being moulded by hands (malleability). *Wet* consistence is when the moisture content is at or slightly more than *field capacity*. *Moist* consistence is when the soil moisture content is between field capacity and the permanent wilting point. *Dry* consistence is when the soil moisture content

is below permanent wilting point. When recording consistence it is important to record the moisture status as well.

Table of Classification of consistence (Buol et al., 1997).

Moisture status	Consistence	Description
Wet	Non-sticky	Almost no natural adhesion of soil material to fingers
	Slightly sticky	Soil material adheres to only one finger
	Sticky	Soil material adheres to both fingers
	Very sticky	Soil material strongly adheres to both fingers
	Non-plastic	No wire is formable by rolling material between the hands
	Slightly plastic	Only short (< 1cm) wires are formed by rolling material between the hands
	Plastic	Long wires (>1cm) can be formed and moderate pressure is needed to deform a block of the moulded material
Moist	Very plastic	Much pressure is needed to deform a block of the moulded material
	Loose	Soil material is non-coherent
	Very friable	Aggregates crush easily between thumb and finger
	Friable	Gentle pressure is required to crush aggregates
	Firm	Moderate pressure is required to crush aggregates
	Very firm	Strong pressure is required to crush aggregates
Dry	Extremely firm	Aggregates cannot be broken by pressure
	Loose	Non-coherent
	Soft	Breaks under slight pressure between thumb and forefinger to a powdery mass
	Slightly hard	Breaks under moderate pressure.
	Hard	Breaks with difficulty.
	Very hard	Very resistant to pressure; cannot be broken between thumb and forefinger.
Extremely hard	Extreme resistance to pressure; cannot be broken in the hand.	

Compaction, Volume and relations in soil

Particle density

Particle density is the mass of a unit volume of soil solids. It refers to the density of the soil solids or the average density of the individual soil grains.

$$\text{Particle density } (\rho_s) = \frac{\text{Mass of soil solids}}{\text{Volume of soil solids}}$$

Most mineral soils have particle density between the narrow limits of 2.60 g cm^{-3} and 2.75 g cm^{-3} . This occurs because quartz, feldspar, and the colloidal silicates with densities within this range usually make up the major portion of mineral soils. When unusual amounts of heavy minerals such as magnetite, garnet, epidote, zircon, tourmaline, and hornblende are present, the particle density may exceed 2.75 g cm^{-3} . It should be emphasized that the fineness of the particles of a given mineral and the arrangement of the soil solids have nothing to do with the particle density.

Organic matter weighs much less than an equal volume of mineral solids having a particle density of 1.2 to 1.5 g cm^{-3} . Consequently, the amount of this constituent in a soil markedly affects the particle density. This accounts for the fact that surface soils usually possess lower particle densities than do subsoils.

Some mineral topsoils high in organic matter may drop as low as 2.4 g cm^{-3} or even below in particle density. Nevertheless, for general calculations the average arable surface soil may be considered to have a particle density of about 2.65 g cm^{-3} .

2. Bulk density of mineral soils

Bulk density is defined as the mass (weight) of a unit volume of dry soil. This volume includes both solids and pores. It is the density for a volume of soil as it exists naturally, includes any air space and organic materials in the soil volume.

$$\text{Bulk density } (\rho_b) = \frac{\text{Mass of oven dry soil}}{\text{Total volume of soil}}$$

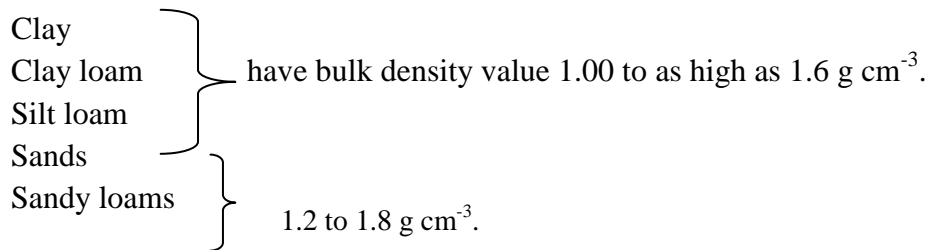
Moisture is not included in the sample weight. The bulk soil volume is assumed not to have changed by drying; only the water has been removed leaving empty pores. This is not true for soils with considerable amounts of swelling clays.

Factors that affect bulk density

Structure of the soil

Soils that are loose and porous will have low weights per unit volume (bulk density) and those that are more compact will have high values.

Example: Particles of sandy soils generally lie in close contact, such soils have high bulk densities. The low organic matter content of sandy soils further encourages the high bulk density value. On the other hand, particles of the fine-textured surface soils, such as silt loams, clay loams and clays, ordinarily do not rest so close together. This occurs because these surface soils are comparatively well granulated, a condition encouraged by their relatively high content of organic matter.



Very compact subsoils, regardless of texture, may have bulk densities as high as 2 g cm⁻³. There is a tendency for the bulk density to rise with depth. This apparently results from:

- Lower content of organic matter
- Less aggregation and root penetration
- Compaction caused by the weight of the overlying layers.

Crop and soil management practices are also likely to influence bulk density, especially of the surface layers. For instance, addition of farm manure in large amounts tends to lower bulk density whereas intensive cultivation operates in the opposite direction.

Soil aeration, renewal of soil air.

Pore space of mineral soils (voids)

The pore space of a soil is that portion of the soil volume occupied by air and water. Voids consist of that portion of the soil volume not occupied by soil solids, either mineral or organic. Under field conditions, the pores are occupied at all time by air and water. ‘Tortuous pathways’ best describe soil pores. Soil particles have irregular shapes and thus leave the spaces of pores between them very irregular in shape, size and direction. Sands, for example, have large and continuous pores whereas clays have more total pore space. The amount of this pore space is determined largely by the arrangement of the solid particles. In particles that lie close together as in sands or compact subsoils, the total porosity is low whereas in those particles arranged in porous aggregates, the total pore space per unit volume will be high.

Calculation of porosity

$$\text{Percent solid space} = \frac{\text{Bulk density}}{\text{Particle density}} \times 100 = \frac{\text{Volume of solids } (V_s)}{\text{Total volume } (V_t)}$$

Since % pore space + % solid space = 100 and pore space = 100 - % solid space, then % pore space:

$$\text{Porosity } (f) = 1 - \frac{\rho_b}{\rho_s} \quad \text{and } \% f = \left(1 - \frac{\rho_b}{\rho_s} \right) \times 100$$

CHAPTER 5

Chemical properties of soil

Weathering of rocks and minerals

1. Igneous Rocks

Igneous rocks are formed from the cooling of molten materials, which have been pushed up from the center of the earth. The fastest-cooled, expelled, igneous rocks have a glassy amorphous texture; those less rapidly cooled have small crystals in the rock mass. When the solidification of the molten magma takes place deep in the earth's crust, the rocks formed are called Plutonic or Intrusive igneous rocks and since the magma cools more slowly, these rocks are comprised of large crystals. Examples are granites, diorites and gabbros. Granite and Rhyolite are composed mainly of feldspars ($\approx 50\%$), quartz ($\approx 30\%$) and iron and magnesium-rich minerals (20%).

If the cooling and solidification takes place on the surface of the earth, the rocks formed are generally called Volcanic, Extrusive or Effusive igneous rocks. These rocks are the ones with glassy texture as there is not enough time for the development of crystals. Examples are rhyolites, andesites and basalt. Glassy (amorphous, non crystalline) are obsidian (dense) and pumice and scoria (porous). Since the minerals in basalt weather more easily than those in granite, a fine-textured soil will develop from basalt. Also, soils formed on granite develop less fertile soils, whereas soils formed on basalt develop soils with more bases and therefore a higher fertility.

2. Sedimentary Rocks

These are composed of the weathering products of igneous and metamorphic rocks and are formed after deposition by wind and/or water. Therefore, the sedimentary rocks were at one time rock and mineral particles of soluble substances that became consolidated or cemented into hard mass. Cycles of geologic uplift, weathering, erosion, and subsequent deposition of eroded materials in rivers, lakes and seas have produced superimposed strata of sediments. Under the weight of the overlying sediments, deposits gradually consolidate to form sedimentary rocks.

3. Metamorphic Rocks

Metamorphic rocks formed when igneous and sedimentary rocks are subjected to intense heat and great pressure. Changes in mineralogy and rock structure generally render the metamorphosed rock more resistant to weathering, but they weather to produce similar soils.

Soil Colloids

Genesis of Soil Colloids

Soil colloids are the colloidal complex of soils that consists chiefly of clay and humus and plays an important role in ion exchange and fertility. The soil colloids are the most active portion of the soil and largely determine the physical and chemical properties of a soil.

Inorganic colloids (clay minerals, hydrous oxides) usually make up the bulk of soil colloids. Colloids are particles less than 0.001 mm in size and the clay fraction includes particles less than 0.002 mm in size. Therefore, all clay minerals are not strictly colloidal. **The organic colloids** include highly decomposed organic matter generally called humus. Organic colloids are more reactive chemically and generally have a greater influence on soil properties per unit weight than the inorganic colloids. Humus is amorphous and its chemical and physical characteristics are not well defined.

Clay minerals are usually crystalline (although some are amorphous) and usually have a characteristic chemical and physical configuration. Both inorganic and organic colloids are intimately mixed with other soil solids. Thus, the bulk of the soil solids are essentially inert and the majority of the soil's physical and chemical character is a result of the colloids present.

One of the most important properties of colloids is their ability to adsorb, hold, and release ions. Colloids generally have a net negative charge as a result of their physical and chemical composition. This negative charge is balanced by thousands of cations. Thus, colloids can be viewed as huge anions surrounded by a swarm/group of rather loosely held cations.

General Statement: Most clay fraction and humus are referred to collectively as the colloidal fraction, because of their small size and colloid-like properties. Some of the important functional properties of these colloids may include high specific surface area, electrostatic charges (cations and anions); adsorption of water, thereby the overall chemical and physical properties of the soil.

Origins: Mineral colloids came from various stages of weathering of primary materials or inherited from sediments of old clay formation; whereas humus came from various stages of organic residue decomposition.

Structures of clays minerals

The most important silicate clays are known as phyllosilicates (Greek phyllon, leaf) because of their leaf like or planar structure. They are composed of two kinds of horizontal sheets.

1. Tetrahedral sheet: this kind of sheet consists of two plans of oxygen with mainly silicon in the spaces between the oxygen. The basic building block for the tetrahedral sheet is a unit composed of one silicon atom surrounded by four oxygen atoms and it is called a tetrahedron.

2. Octahedral sheets: the building block is composed of six oxygen/hydroxide atoms coordinating with a central aluminum or magnesium atom, forming the shape of an eight-sided geometric solid, or octahedron.

The silica and alumina sheets join together to form layer silicate clay minerals. Based on the proportion of the tetrahedral and octahedral making the mineral, layer silicate clay minerals in soils are structurally of three general classes: These include:

- The 1:1 type,
- The 2:1 type, and
- The 2:1:1 (or 2:2) type minerals

Origin of the net negative charge in soils

Inorganic colloids

- Crystalline silicate clays
 - ✓ 1:1 layer silicate clays (Example: kaolinite)
 - ✓ 2:1 layer silicate clays (Example: fine grained mica, smectite, vermiculite, chlorite)
- Noncrystalline silicate clays
 - ✓ Allophane and imogolite
- Nonsilicate colloids
 - ✓ Gibbsite (Al-oxide)
 - ✓ Geotite and hematite (Fe-oxides).

Organic Colloids

Humus is a temporary intermediate product left after considerable decomposition of plant and animal remains. It's temporary because the organic substances remaining continue to decompose slowly. The humus is often referred to as an organic colloid. It consists of various chains and loops of linked carbon atoms. Humus is amorphous, dark brown to black, nearly insoluble in water, but mostly soluble in dilute alkali (NaOH or KOH) solution. On the basis of solubility in acids and alkalis, humus is thought to be made up of three classes:

- a. Fulvic acid – lowest in molecular weight and lightest in color, soluble in both acid but not in alkali.
- b. Humic acid –medium in molecular weight and color, soluble in alkali but insoluble in acid.

c. Humin -highest in molecular weight, darkest in color and insoluble in both acid and alkali.

Cation exchange capacity (CEC)

Cation exchange capacity is the sum total of the exchangeable cations that a soil can absorb. CEC is expressed in centimoles of charge per kilogram ($\text{cmol}_c \text{ kg}^{-1}$). CEC is associated with the amount of negative charges the colloid has. The excess negative charges in the crystal lattice caused by isomorphous substitution are called exchange sites. These sites attract and hold cations somewhat loosely. Other cations in solution can compete with and often replace originally adsorbed cations.

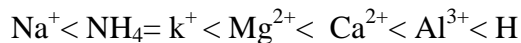
Any electrically charged colloidal surface area will attract mobile substances of opposite charge in the soil water. The dominant residual charge on most soil colloids is negative. These negatively charged sites attract positively charged ions in the soil water. Positively charged ions are called cations. Adsorbed cations resist removal by leaching water but can be replaced /exchanged/ by other cations in solution by mass action (competition for the negative charge /site/ because of the large number of ions present/. This exchange of one positive ion by another is called cation exchange. The cations most numerous on exchange sites in soils are:

- Calcium, hydrogen, potassium, magnesium, sodium, and aluminum.

The strength of adsorption increases as:

1. The valence of the cation increases
2. The cation's hydrated size is smaller
3. The strength of the site's negative charge increases

Thus, adsorption strength of cations increases approximately in this order:



Cation exchange capacity (CEC) is defined as the sum of positive (+) charges of the adsorbed cations that a soil can adsorb at a specific pH. Conversely, the CEC is equivalent to the sum of the negative charges of the cation exchange sites. It is the amount of exchangeable cations per unit weight of dry soil. It is measured in centimoles of cations per kilogram of soil ($\text{cmol}_c \text{ kg}^{-1}$).

The CEC of soils is affected mainly by the amount and type of clay and the amount and degree of decomposition of the organic matter.

Buffering Capacity of Soils

Buffering in soil is defined as the resistance of the soil to variations in pH and is chiefly due to the colloidal humus and clay fractions. That a soil does not undergo marked fluctuations in pH following additions of acids or bases various activities under those particular pedological conditions. A large change in pH means a radical modification in soil environment, which severely affects the availability of plant nutrients and alters the pedogenetic processes.

Soil acidity and alkalinity

Soil acidity (pH) is known as the master variable because it controls the nature of many chemical and microbial processes in the soil. Soil pH values range from 1 to 14. A pH value of 1 indicates the presence of high concentration of H⁺ ions (10⁻¹ moles/liter) and low concentration of OH⁻ ions (10⁻¹³ moles/liter). A pH value of 14 indicates the presence of very low concentration of H⁺ ions (10⁻¹⁴ moles/liter) and very high concentration of OH⁻ ions (100 = 1 mole/liter). However, the extreme pH values for soils are between 3.5 and 10.5.

Sodic (Alkali) soils contain exchangeable sodium in their exchange complex in concentrations great enough to interfere with the growth and production of most crop plants. They are characterized by an ESP of greater than 15, an ECe at 25 °C less 4mmhos/cm and a pH reading usually ranging between 8.5 and 10.

Waterlogged soil and Soil conservation status in Ethiopia

Oxygen availability in the soil is regulated by three factors: (1) soil micro-porosity (as affected texture and structure), (2) soil water content (as it affects the proportion of porosity that is field with air), and (3) O₂ consumption by respiring organisms (including plant roots and microorganisms). Typically, poor aeration becomes a series impediment to plant growth when more than 80 to 90% of the soil pore space is filled with water (leaving less than 10 to 20% of the pore space filled with air). The high soil water content not only leaves little pore space for air storage, but, more important, the water blocks the pathways by which gases could exchange with the atmosphere. Compaction can also cut off gas exchange, even if the soil is not very wet and

has a large percentage of air-filled pores. For respiration to take place in the roots there is a need to make more O₂ available. Poor soil aeration becomes a serious impediment to plant growth when more than 80-90% of the soil pore space is filled with water.

When water occupies all or nearly all of the soil pores, the soil is called water saturated or water logged. In such conditions only hydrophytes can survive. Such plants (for example rice) have hollow structures (*aerenchyma tissues*) in their stems and roots to transport oxygen while submerged in water. It can amend by the application of drainage system, which can support to drains the surface and subsurface accumulated water. Regarding to soil conservation both physical and biological soil water conservation measures is applicable in our country which is used to alleviate soil; erosion and enhance soil fertility and productivity.

CHAPTER 6

Soil water

Movement of water in soil

Water is present in almost every soil profile, but the amount varies with time and depth as a result of supply and demand by its environment. If the supply of water at the soil surface (e.g., irrigation, rainfall, dew, flood) during a certain period exceeds the extraction of water from a given soil profile (e.g., evaporation from soil surface, water uptake by plant roots, deep percolation), the excess supply during that period is stored in the soil profile. If the excess supply surpasses the storage capacity of the soil, water discharges to lower regions (e.g., drainage). The supply at the soil surface can also be smaller than the sum of transpiration and evaporation. In that case, the topsoil dries out, while upward transport of water from the subsoil (capillary rise) may take place.

Generally, three types of water movement within the soil are recognized:

- Saturated flow in which all the pores, large and small, are filled with water,
- Unsaturated flow in which the macropores are filled with air, leaving the micropores to accommodate water movement,
- Vapor movement which occurs in relatively warm and dry soils.

Loss of soil water and hydrological cycle

Evapotranspiration loss of soil water and its control

Vapor losses from soils occur in two ways:

- a. By the evaporation of water at the soil surface, and
- b. By transpiration from the leaf surfaces of plants.

The combined loss resulting from these two processes is termed evapotranspiration (ET). This process is responsible for most of the water removal from soils under field conditions. The evaporation component of ET may be viewed as a 'waste' of water from the standpoint of plant productivity. However, at least some of the transpiration component is essential for plant growth

in that it provides water plants need for cooling, nutrient transport, turgor maintenance and photosynthesis.

Potential evapotranspiration rate tells how fast water vapor would be lost from a densely vegetated plant-soil system if soil water content was not limiting or supply at optimal rate. The PET is largely dictated by climatic variables, such as temperature, relative humidity, cloud cover, and wind speed. These climatic variables influence the vapor pressure gradient between a wet soil, leaf, or body of water and the atmosphere. Besides, soil and plant parameters also play an important role.

- a. **Radiant energy:** water has got high heat of vaporization (540 calories of energy are required to vaporize 1 g of water). The primary source of this energy is the sun. If the soil moisture is ample, there exists a close relationship between evaporation and the absorption of this radiant energy. In arid regions the sparse cloud cover permits a high proportion of the solar radiation to strike the earth, maximizing the opportunities for evapotranspiration. In contrast, in regions characterized by cloudy days a smaller proportion of the solar radiation strikes the soil and plants, the evaporative potential is relatively low.
- b. **Atmospheric vapor pressure:** controls evaporation from soils and plants. Low atmospheric vapor pressure results in rapid evaporation rate. Dry atmosphere, therefore, has high evaporative demand whereas cloudy days have low evaporative demand.
- c. **Temperature:** high temperature increases the vapor pressure at the leaf surface or the surface of moist soil. However, temperature does not have a similar direct effect on the vapor pressure of the atmosphere. As a result, on hot days there is a sharp difference in vapor pressure between leaf or soil surfaces and the atmosphere (a higher vapor pressure gradient) and evaporation proceeds rapidly.
- d. **Wind:** a dry wind will continually sweep away water vapor from a wet surface. The moist air thus moved is replaced by air with lower water content, thereby maintaining the vapor pressure gradient and thus encouraging evaporation loss.

- e. **Soil moisture supply:** climatic variables largely dictate the rate of evapotranspiration under conditions where the plant and soil surfaces are amply supplied with water. Under conditions where water supply is limited, soil water suction will limit the rate of supply of water to the soil and plant surfaces and evapotranspiration losses will decrease from the potential rate. Under such conditions the evapotranspiration is termed actual evapotranspiration. The plant responds to this water deficiency by closing of the leaf stomata and eventually by wilting.

Control of surface evaporation

For arable soils, the most effective practices for controlling surface evaporation are those practices that provide some cover for the soil. This cover can be provided by mulches of different types and by selected conservation tillage practices that leave plant residues on soil surface.

1. Vegetative Mulches

Any material placed on the soil surface primarily for the purpose of reducing loss of water by evaporation or controlling weeds can be designated as mulch. Materials such as sawdust, manure, straw, leaves and crop residues are good examples of mulches. Mulches can be highly effective in controlling evaporation, but they may be expensive and labor-intensive to produce, purchase, transport to the field and apply to the soil. This limits the best use of mulches to small areas such as gardens and for high-value crops, such as fruit trees.

2. Paper and plastic mulches

Specially prepared paper and plastics have been used as mulches. This mulch can be used only with crops planted in rows or in hills. The cost of the cover and the difficulty of keeping it in place limit the use of these materials to high-valued crops.

3. Crop residue and conservation tillage

Plant residues left on the soil surface are effective in reducing evaporation. Conservation tillage practices leave a high percentage of the residues from the previous crop on or near the surface. A conservation tillage practice widely used in sub-humid and semiarid regions is stubble or trash

mulch. In this case the mulching materials have been grown in place and consist of refuses of the previous crops. With this method, the residues are uniformly spread on the soil surface and the land is tilled with special implements that permit much of the plant residue to remain on or near the surface. No tillage is another conservation tillage in which new crops are planted directly into the sod or residues of the previous crop, with no plowing or disking.

Liquid losses of soil water and their control

Two types of liquid losses of water from the soil are recognized:

1. **Percolation**-downward movement of free water
2. **Runoff**-loss of excess water from the soil surface

Losses by percolation occur when the amount of rainfall entering the soil becomes greater than its water-holding capacity. Percolation results in the loss of soluble salts, thus depleting soils of certain nutrients. The losses by percolation are influenced by the amount of rainfall and its distribution by evaporation, by the character of the soil, and by the presence of a crop.

In most soils where the land is sloping or the soil is somewhat impermeable to water, a considerable amount of precipitation is likely to be lost by runoff. Two serious consequences of runoff:

1. Depriving the crops of this water by preventing it from entering the soil,
2. Causing soil erosion which carries away the valuable topsoil.

Control measures for percolation and runoff

- Keeping crops on the land as much as possible to reduce leaching losses of nutrients,
- Increasing the water holding capacity of soils,
- Organic matter maintenance to encourage infiltration,
- Use of an effective vegetative cover,
- Use of mechanical conservation practices such as terraces on sloppy areas,

- Using appropriate irrigation techniques (use of drip and sprinkler irrigation on sandy soils, shallow soils and slowly permeable clay soils).

Chapter 7: Soil organic Matter

Soil organic matter is the organic fraction of the soil that includes plant and animal residues at various stages of decomposition, cells and tissues of soil organisms, and substances synthesized by the soil population.

Sources of soil organic matter

The sources of soil organic matter can be categorized as:

Primary / Original Sources

Plant tissues constitute the original sources of soil organic matter. Plants trap atmospheric CO₂ to produce various compounds like carbohydrates, proteins, oils and so on. When plants die they add large quantities of organic residue to the soil.

Secondary Sources

Animals directly or indirectly live on foods synthesized by plants. Therefore, animals contribute organic matter by their waste products or leave their own bodies as their life cycles are consummated. Certain forms of animal life, especially earthworms, ants, and centipedes, also play an important role in the translocation of plant residues.

The Carbon – Nitrogen (C: N) ratio and its significance

C/N ratio defines the relative quantities of these two elements in:

- Crop residue and other fresh organic materials
- Soil organic matter
- Soil microorganisms

It is a measure of the speed with which organic materials will decay.

This ratio is important in controlling:

- The available nitrogen
- Total organic matter
- The rate of organic decay

The ratio in soils varies between 8:1 to 15:1, the median being between 10:1 and 12:1. Only little variation exists in similarly managed soils. Variations that do occur are generally related to/with climatic conditions (especially, temperature and the amount and distribution of rainfall).

Ratio in plants

The C:N ratio in plants is variable ranging from 20:1 to 30:1 (legumes and farm manures) to 100:1 (straw residues). It may go as high as 400:1 in some plant materials such as sawdust.

Ratio in the bodies of microorganisms

It ranges between 4:1 and 9:1 and more constant.

Most organic residues entering the soil carry large amounts of carbon and comparatively small amounts of total nitrogen. The C/N values for soils are, therefore, in between those of higher plants and the microbes.

Significance of the Ratio

It is important for two major reasons:

1. It determines the competition for available soil nitrogen among the microorganisms and in turn with plants.
2. The soil nitrogen level constrains the maintenance of carbon, and eventually soil organic matter.

Influence of Soil Organic Matter on Soil Properties and Plant Growth

1. Improves soil aggregation which, in turn, influences:
 - Infiltration and percolation rate
 - Movement and retention of soil water
 - Soil aeration
 - Soil temperature
 - Root penetration
2. Assists in the control of surface runoff and erosion
3. High cat-ion adsorption capacity
4. Source of food and energy for microorganisms and plants

5. Effect on soil colour-brown to black-important in areas where there is low radiation.

Factors and practices influencing soil organic matter content

The balance between organic matter production (plant growth) and organic matter destruction (microbial decomposition) is the key factor affecting the level of soil organic matter and nitrogen. These two processes are dictated by a number of environmental and edaphic factors, the most important of which are the following:

1. Climatic factors

- **Temperature:** plant growth and organic matter decomposition respond differently to temperature. At low temperatures (< 25 °C), growth exceeds decomposition, but the opposite is true at temperatures greater than 25 °C. Therefore, under uniform moisture and comparable vegetation conditions, the organic matter and nitrogen level of soils in cooler climates is generally greater than that of warmer climates.
- **Soil moisture/effective precipitation:** Under comparable conditions, the organic matter and nitrogen content of soils increases with an increase in effective precipitation or moisture increases. The reason is related to the high production of plant biomass at high moisture conditions.

2. Natural vegetation: Climate and vegetation usually act together to influence the organic matter and nitrogen contents of the soil. In climatic zones where the natural vegetation includes forests and grasslands, the total organic matter is higher in soils developed under grasslands than under forests. With grasslands, a relatively high proportion of the plant residues consist of root matter that decomposes more slowly and contributes more efficiently than does forest leaf litter.

3. Texture and drainage: these two factors play important role in bringing about marked differences in organic matter and nitrogen content of the soil within a local landscape. All other factors being equal, fine-textured soils (soils high in clay and silt) are generally higher in organic matter and nitrogen than the coarse textured (sandy) soil. The main reasons are:

- Production of more plant biomass
- Generally poor aeration, and

- Protection of some of the carbon by the clay-humus complexes in the fine-textured soils.