

**COURSE
GUIDE**

**SLM 516
ADVANCED SOIL SCIENCE**

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Printed 2020

ISBN: 978-978-970-164-3

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INTRODUCTION

The study of Advanced Soil Science is fundamentally concerned with the soil of any given area. Essentially, soil scientists are interested in the process of soil transformation to the benefit of plants growth within a soil. Soil science knowledge has made enormous contributions towards ensuring that we have a good understanding and control of soils in general.

Soil scientists that focus on soil science (soil fertility) are interested in managing nutrients to improve crop production. They focus on using commercial fertilisers, manures, waste products, and composts to add nutrients and organic matter to the soil.

WHAT YOU WILL LEARN IN THIS COURSE

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This course guide tells you briefly what to expect from reading this course material. The course is developed basically on four different modules and units under these modules, different subheadings as a unit were provided for a proper arrangement of the entire topics within the context of the entire course.

In addition, it provides some general guidelines for the amount of time you are likely to spend on each unit of the course material. The course material also provides guidance in respect of your Tutor-Marked Assignment assigned for each unit.

As mentioned earlier, soil scientists that focus on soil fertility are interested in managing soil nutrients to improve crop production. They focus on using commercial fertilisers, manures, waste products, and composts to add nutrients and organic matter to the soil.

COURSE AIM

The course aims to provide a good understanding of the soil for a better management of agriculture.

COURSE OBJECTIVES

After going through this course, you will be able to learn and explain the following terms:

- Silicate chemistry
- Soil forming rocks and minerals
- Composition of the soil
- Processes and factors of soil formation
- Texture Physical state of soils
- Structures
- Porosity
- Density
- Soil water relationship
- Evapo-transpiration
- Tillage and soil properties
- Factors and processes affecting plant growth
- Soil micro-organisms
- Shifting cultivation and the fallow system.

WORKING THROUGH THIS COURSE

This course has been carefully put together bearing in mind the fact that it is an advance course. Therefore, adequate explanation of the concepts and issues are treated in this work. Diagrams and tables have been used where necessary to enhance your understanding. You are advised to spend good time to study the work and ensure that you attend tutorial sessions if there is any provision for that this would enable you ask questions and compare your knowledge with that of your classmates.

COURSE MATERIALS

The main components of the course are:

1. The Course Guide
2. Study Units
3. References/Further Reading
4. Assignments
5. Presentation Schedule

STUDY UNITS

The course is divided into 14 units. The following are the study units contained in this course:

Module 1

Unit 1	Silicate Chemistry
Unit 2	Soil Forming Rocks and Minerals
Unit 3	Composition of the Soil
Unit 4	Processes and Factors of Soil Formation

Module 2

Unit 1	Texture Physical State of Soils
Unit 2	Structures
Unit 3	Porosity
Unit 4	Density

Module 3

Unit 1	Soil Water Relationships
Unit 2	Evapo- Transpiration

Module 4

Unit 1	Tillage and Soil Properties
Unit 2	Factors and Processes Affecting Plant Growth
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Unit 4	Shifting Cultivation and the Fallow System

PRESENTATION SCHEDULE

The presentation schedule included in the course material gives you the important dates for the completion of tutor-marking assignments and attending tutorials. You are required to submit all your assignments by due date. You should guide against falling behind in your work.

ASSESSMENT

There are two types of assessments in the course. First is the tutor-marked assignments; and the second is a written examination.

In attempting the assignments, you are expected to apply the information, knowledge and techniques gathered during the course. The assignments must be submitted to your course tutor for formal Assessment in accordance with the deadlines stated in the Presentation Schedule and the Assignments File. The works you submit to your course tutor for assessment constitute 30 % of the total course mark.

At the end of the course, you will need to sit for a final written examination of two hours' duration. This examination will constitute 70% of your total course mark.

TUTOR-MARKED ASSIGNMENT

There are three Tutor-Marked Assignments (TMAs) to be submitted in this course. The TMAs constitute 30% of the total score. You are encouraged to work all the questions thoroughly.

Assignment questions for the units in this course are contained in the Assignment File. You will be able to complete your assignments from the information and materials contained in your set books, reading and study units. However, it is desirable that you demonstrate that you have read and researched more widely than the required minimum. You should use other references to have a broad viewpoint of the subject and also to give you a deeper understanding of the subject.

When you have completed each assignment, send it, together with a TMA form, to your tutor. Make sure that each assignment reaches your tutor on or before the deadline given in the Presentation File. If for any reason, you cannot complete your work on time, contact your tutor before the assignment is due to discuss the possibility of an extension. Extensions will not be granted after the due date unless there are exceptional circumstance.

FINAL EXAMINATION AND GRADING

The final examination will be of two-hour duration and have a value of 70% of the total course grade. The examination will consist of questions which reflect the types of self-assessment practice exercises and tutor-marked problems you have previously encountered. All areas of the course will be assessed.

Revise the entire course material using the time between finishing the last unit in the module and that of sitting for the final examination. You might find it useful to review your self-assessment exercises, tutor-marked assignments and comments on them before the examination. The final examination covers information from all parts of the course.

HOW TO GET THE MOST FROM THIS COURSE

In distance learning, the study units replace the university lecturer. This is one of the great advantages of distance learning; you can read and

work through specially designed study materials at your own pace and at a time and place that suit you best.

Think of it as reading the lecture instead of listening to a lecturer. In the same way that a lecturer might set you some reading to do, the study units tell you when to read your books or other material, and when to embark on discussion with your colleagues. Just as a lecturer might give you an in-class exercise, your study units provides exercises for you to do at appropriate points.

Each of the study units follows a common format. The first item is an introduction to the subject matter of the unit and how a particular unit is integrated with the other units and the course as a whole. Next is a set of learning objectives. These objectives let you know what you should be able to do by the time you have completed the unit.

You should use these objectives to guide your study. When you have finished the unit you must re-check whether you have achieved the objectives. If you make a habit of doing this, you will significantly improve your chances of passing the course and getting the best grade. The main body of the unit guides you through the required reading from other sources. This will usually be either from your set books or from a readings section.

Self-assessments are interspersed throughout the units, and answers are given at the end of the units. Working through these tests will help you achieve the objectives of the unit and prepare you for the assignments and the examination. You should do each self-assessment exercises as you come to it in the study unit. Also, ensure to master some major historical dates and events during the course of studying the material.

If you run into any trouble, consult your tutor. Remember that your tutor's job is to help you. When you need help, don't hesitate to call and ask your tutor to provide it. The following is a practical strategy for working through the course:

1. Read this Course Guide thoroughly.
2. Organise a study schedule. Refer to the 'Course overview' for more details. Note the time you are expected to spend on each unit and how the assignments relate to the units. Important information, e.g. details of your tutorials, and the date of the first day of the semester is available from study centre. You need to gather together all this information in one place, such as your diary or a wall calendar. Whatever method you choose to use, you should decide on and write in your own dates for working breach unit.

3. Once you have created your own study schedule, do everything you can to stick to it. The major reason that students fail is that they get behind with their course work. If you get into difficulties with your schedule, let your tutor know before it is too late for help.
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5. Assemble the study materials. Information about what you need for a unit is given in the 'Overview' at the beginning of each unit. You will also need both the study unit you are working on and one of your set books on your desk at the same time.
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On successful completion of the course, you would have developed sufficient critical thinking skills with the material necessary for efficient and effective discussion on issues related to soil survey and land evaluation; you will also be able to proffer solutions to problems with soil classification in the field.

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Printed 2020

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MODDULE 1

Unit 1	Silicate Chemistry
Unit 2	Soil Forming Rocks and Minerals
Unit 3	Composition of the Soil

UNIT 1 SILICATE CHEMISTRY

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2.0	Objectives
3.0	Main Content
	3.1 Silicate Chemistry
	3.2 Colloidal Properties in Relation to Structure
	3.3 Silicates and the Processes of Solution
4.0	Conclusion
5.0	Summary
6.0	Tutor- Marked Assignment
7.0	Reference/Further Reading

1.0 INTRODUCTION

The silicates are the largest, the most interesting and the most complicated class of minerals than any other minerals. Approximately 30% of all minerals are silicates and some geologists estimate that 90% of the Earth's crust is made up of silicates, SiO_4^{4-} based material. Thus, oxygen and silicon are the two most abundant elements in the earth's crust.

Silicates are based on the basic chemical unit SiO_4^{4-} , tetrahedron shaped anionic group. The central silicon ion has a charge of positive four while each oxygen has a charge of negative two (-2) and thus each silicon-oxygen bond is equal to one half the total bond energy of oxygen. This condition leaves the oxygens with the option of bonding to another silicon ion and therefore linking one SiO_4^{4-} tetrahedron to another.

In the extreme case, the tetrahedral are arranged in a regular, orderly fashion forming a three-dimensional network. **Quartz** is such a structure and its formula is SiO_2 . If silica in the molten state is cooled very slowly it crystallizes at the freezing point. But if molten silica is cooled more rapidly, the resulting solid is a disorderly arrangement which is called **glass**, often also called **quartz**.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss the structural classification of the silicates
- explain what are the silicate structures
- examine the colloidal properties in relation to structure
- analyse silicates and their processes of solution.

3.0 MAIN CONTENT

3.1 Silicate Chemistry

Structural Classification of the Silicates

The silicate minerals, as the X-ray crystallographers have revealed, possess structures whose geometry is dominated by the disposition of the silicon and oxygen atoms in space. This disposition follows certain well defined rules, arising in turn from the chemistry and geometry of the atoms themselves. Silicon is a small atom, with a positive electrovalency four and an extremely strong tendency toward a coordination number four. Oxygen is a large atom with a negative electrovalency two and very weak coordination of its own. Hence, in all the silicates we find a silicon atom ensconced in the central pore formed by four oxygen atoms in tetrahedral arrangement. This unit SiO_4 has four negative charges. The modern classification of the silicates, due largely to W. L. Bragg and his associates stems from the manner in which these charges are neutralised.

If silicon itself, or some other atom similarly disposed to a tetrahedral arrangement with oxygen neutralises the charge, then the effect is that the characteristic group SiO_4 is repeated. This may occur once or more than once, with each tetrahedral group. If it occurs once we can speak of adjacent tetrahedral sharing corner oxygen. When it occurs twice then chains or discrete ring structures are produced. When it occurs three times a complete planar sheet is results. Finally, when all four oxygen are shared, three dimensional framework structures appear.

The charges may be neutralised also by a great variety of other cations, acting either independently or in conjunction with repetitions of the SiO_4 grouping. Their incorporation brings into play important limitations due to their varied sizes, valencies and coordination numbers. It is evident that the completed structure must afford room for the various atoms present, and that all valencies must be satisfied, all charges neutralised.

Furthermore, atoms with a strong tendency to certain coordination numbers must find themselves in appropriate positions. The two factors charge and coordination number, are involved in the important generalization known as Pauling's Rule. This has proved of immense importance in deciding between alternative structures.

Introduction to silicate structures

This rule postulates that charges are balanced by the immediate neighbors. Thus, each cation in any structure is surrounded by a group of anions whose total contribution of negative charge closely balances the positive charge of the cation. Simultaneously, the same applies to each anion surrounded by cations.

In the orthosilicates (Olivine, $(\text{MgFe})_2\text{SiO}_4$; Zircon, ZrSiO_4) the whole charge $(\text{SiO}_4)^{4-}$ is neutralized by cations. We therefore have a salt-like structure. The cations concerned, however, are small compared with oxygen and polyvalent. The electrical attractions are therefore very intense, and the structure attained is closely packed. No cases are known in which orthosilicates are so loosely built as to allow for the entry of other atoms. These compounds are for the most part well crystallized and highly insoluble. They can only show colloidal properties in virtue of small particle size, and such occurrences have thus far rarely been found. This may partly be due to the fact that the orthosilicates are only minor constituents of the rocks on the surface of the earth.

In the disilicates we find the first evidence of an extremely strong tendency for the charge on the primary SiO_4 group to be neutralized in part by silicon itself. Two SiO_4 groups become linked giving $(\text{O}_2\text{Si-O-SiO}_2)$ or $(\text{Si}_2\text{O}_7)^{4-}$ having six negative charges. (The hydrated zinc silicate hemimorphite $(\text{OH})_2\text{Zn}_4\text{Si}_2\text{O}_7 \cdot \text{H}_2\text{O}$ belongs to this group.) Compared with the orthosilicates, the negative charge is smaller and less concentrated. These are also well-crystallized minerals, of somewhat rare occurrence. The same may be said also of the trisilicates, in which a ring forms the basis of the structure (Benitoite, $\text{BaTiSi}_3\text{O}_{10}$).

A larger ring structure containing 6 SiO_2 groups linked by six oxygen atoms is also known. In the mineral beryl the twelve units of negative charge are neutralised by two aluminum and three beryllium atoms. These are both small and are grouped between adjacent rings. In the center of each ring there is, therefore, considerable free space. The rings are stacked one above the other giving rise to a system of parallel pores. Beryl, therefore, provides the rare example of a porous structure achieved by the packing of discrete silicate groups. The system of parallel pores, of diameter about 2.5-2.6 Å, should endow it with the

capacity to absorb considerable amounts of gases of small molecular size, such as helium and hydrogen. Water molecules (diameter about 2.9 Å) would be too large for penetration. These conclusions have not as yet been tested by experiment. We could predict also that, since the structure as a whole is electrically neutral, there will be no tendency for small cations to move into the pores and, hence, no capacity for cation exchange. Even if the walls carried a negative charge the lack of interconnections between the pores would make exchange reactions exceedingly sluggish, as Hendricks has pointed out.

These considerations may now be applied to the second great group of the silicates; those in which silicon-oxygen linkages extend indefinitely in one, two, or three dimensions in space. Here we find practically all the common silicate minerals. Elongated chains of composition $[(\text{SiO}_3)^{2-}]_n$ are characteristic of the pyroxenes double chains of composition $[(\text{Si}_4\text{O}_{11})^6]_n$ give rise to minerals of the amphibole group. Sheet structures of composition $[(\text{Si}_2\text{O}_5)^{2-}]_n$ are typical of the micas and the clays. Finally, silicon-oxygen linkages extending in all three dimensions give rise to framework structures based on $(\text{SiO}_2)_n$ of varying degrees of openness. Of these, quartz itself has the closest packing. The feldspars, feldspathoids, ultramarines, and zeolites are progressively more open.

In appreciating these varied structures, certain further peculiarities of the silicates must be borne in mind. In inorganic chemistry it is customary to speak of the isomorphous substitution of one atom for another in a crystal when the two atoms concerned are of comparable size and coordination and of the same valency. Thus, in the group of the alums various trivalent atoms can substitute for aluminum, and various monovalent atoms for potassium. In the silicates, however, size and coordination take precedence over valency. Thus aluminum can substitute for silicon at the center of a tetrahedron of oxygen atoms, in spite of the fact that the valencies are three and four, respectively. The tetrahedral group thus acquires a negative charge. This must be neutralised by appropriate cations, for which space must be found in the final structure.

The three-dimensional framework structures (feldspars, feldspathoids, ultramarines, and zeolites) afford clear cut examples of the substitution of Al^{+++} cation for Si and of the way in which the nature of this balancing cation is limited by Pauling's Rule. These assemblages of linked tetrahedral afford no positions where six oxygen atoms surrounding an octahedral pore are in contact. Atoms having strong six-fold coordination cannot, therefore, be accommodated. Hence, iron and magnesium are excluded. Spaces surrounded by eight or more oxygens are available. The tendency, therefore, is to incorporate cations such as

K, Na, Ca, Sr, Ba, with relatively weak coordination. They readily accommodate themselves with coordination numbers from eight to twelve.

In contradistinction to the three-dimensional frameworks, the silicates based on silicon-oxygen chains or sheets do contain elements with strong sixfold coordination. The three most important are aluminum, magnesium, and iron. Thus aluminum plays a dual role in silicate chemistry; it can either substitute for silicon in tetrahedral coordination or it can form part of a different unit in which it is coordinated with six oxygen in octahedral arrangement. In the micas and clays, which have silica sheets, the necessary octahedral positions are provided by double sheets of relatively closely packed oxygen atoms. The manner in which these units are combined together will be discussed in detail later.

3.2 Colloidal Properties in Relation to Structure

1. **Particle size:** The well-known inverse proportionality between mean particle diameter and specific surface naturally extends over the silicates. In general, however, colloidal properties are well marked in material having one or more dimensions less than 0.1μ (10^{-7} cm.). In the range $0.1-1 \mu$ they are detectable with varying degrees of difficulty according to the methods employed. Particles larger than 1μ are generally regarded as outside the colloidal range, although by highly sensitive methods some of their surface properties can be measured. Small particle size frequently reflects certain aspects of the mineral structure. A slow rate of crystallization, a very facile cleavage or ready abrasion, all of these are consequences of the atomic arrangement, and all favor the formation, or the subsequent production, of small particles. It is believed also that strains introduced into silicate lattices by the presence of somewhat ill fitting atoms are an important cause of limited growth, and hence favor crystals of colloidal size.
2. **Porous structure:** Internal pores or channels can naturally increase the specific surface, especially when they are so numerous that all molecular parts of the system become accessible to the outer environment. Such cases have been termed "eugels", and their uptake of neutral molecules "persorption", indicating that here the surfaces and their reactions have reached the ultimate limit. In some instances, the solid is amorphous, and the pores assume all sizes[^] down to the molecular. The silicate water-softeners known as permutites are of this kind. In other case, the pores form an essential part of the crystal structure-

They are, therefore, precisely delimited as to size, number, orientation within the crystal, and extent of interconnection. The case of beryl was mentioned above. Other well known examples are the zeolites and ultramarines. Another kind of porous structure is found in the swelling clays, where planar channels of variable width provide access to all lattice units.

3. **Charged framework:** The properties of silicate surfaces, whether external or internal, depend enormously upon whether ionisation can occur by simple dissociation in contact with an ionising solvent. Such dissociating or polar surfaces arise in two ways:
 - i. The more important is produced by structures having inherent framework charges. For instance, when Al + K substitutes for Si the potassium atom lies entirely outside the oxygen tetrahedra. In contact with water there is a tendency for the potassium to achieve a certain degree of independence of the silicon-oxygen framework. The extent will depend upon the properties of the solvent, those of the cation, the distribution of charges in the framework, and the dimensions of the pores. This may be contrasted with the case of an anhydrous silica skeleton alone, in which there is no substitution, a completely neutral silicon-oxygen framework, and no mechanism for ionisation. In each case we can envisage an extensive and accessible surface. The two will differ least in respect to properties which do not involve ionisation or the orientation of dipoles. For instance, the adsorption of inert gases at low temperatures might be fairly comparable for equal specific surfaces. On the other hand, the behavior toward water and the electrochemical properties of the solid liquid systems would be entirely different.
 - ii. A second mechanism for production of a charged surface is accidental rather than inherent. In the termination of a lattice the valency relationships of the constituent atoms are sometimes "upset". The number of these accidental charges depends on the degree and nature of the subdivision. This comes to light very clearly in the clay mineral kaolinite. The flat cleavage surfaces of this platy mineral are believed to be chemically inert, having no inherent or accidental charges. They do not contribute to ionisation. The edges of the cleavage flakes, however, do contain broken bonds, which by reaction with ionising solvents produce localized charges. Thus, the electrochemistry of kaolinite is chiefly the electrochemistry of its exposed edges.

In most colloidal silicates two or more of the above factors contribute to the ensemble of the colloidal properties. Thus, as we shall see later, the zeolites and ultramarines combine a porous structure with a high inherent framework charge. The clay mineral attapulgite has a moderately small particle size, a porous structure, and both inherent and accidental charges. Kaolinite has small particle size and an accidental charge. Beryl has a high porosity toward small molecules only, but no appreciable framework charge; and since the particle size is usually large, accidental ionization plays no measurable part. The clays of the montmorillonite group combine all four of the above factors, but their great porosity and high inherent charge dominate the situation. The compact clays of the hydrous mica group have no internal porosity but their inherent charge is high. This, combined with the accidental charge due to small particle size gives the clue to their colloidal properties.

3.3 Silicates and the Processes of Solution

Since we shall be greatly concerned with ionisation and the effects of water molecules on colloidal silicates it is well to consider the restricted meaning of the term 'solubility' as applied to silicates generally. Even the orthosilicates, disilicates, etc., with their discrete groups, differ materially from sparingly soluble salts like barium sulphate. No well-defined silicate ions are found, even in cases where small quantities of material pass into solution. It is generally assumed that, because silicic acid is so weak, hydrolysis plays an important part. Once the silicic acid is released its strong polymerisation tendency operates and further complicates the situation.

In those silicates having indefinitely extending structures, with which we shall chiefly be concerned, the whole concept of solubility loses its precise meaning. Material released from the framework consists of fragments only. They bear no relation to the unit cell, which is a purely geometrical concept. Furthermore, since points of weakness in the silicon-oxygen framework (caused, for instance, by substitution of Al for Si) are distributed at random, the fragments released are not uniform in size or in composition. This is seen very clearly when an electric field is applied to a silicate under mild attack by hydrogen ions. Fragments containing silicon frequently are found migrating both to the anode and the cathode chambers. Some are, therefore, positively charged, others negatively, depending upon the accidental balance of charges at the instant of release.

4.0 CONCLUSION

The silicate minerals consist essentially of silicon-oxygen tetrahedral. The fundamental unit in building of the atomic structure of the silicate minerals is the $[\text{SiO}_4]^{4-}$ unit in which the silicon atom (more strictly cation) is situated at the center of the tetrahedron whose corners are occupied by four oxygen atoms.

5.0 SUMMARY

In this unit, we have learnt that:

- i. The silicate minerals possess structures whose geometry is dominated by the disposition of the silicon and oxygen atoms in space.
- ii. Silicate structures rule postulates that charges are balanced by their immediate neighbors.
- iii. There are three different colloidal properties in relation to structure
- iv. Silicates and the processes of solution is greatly concerned with ionization and the effects of water molecules on colloidal silicates it is well to consider the restricted meaning of the term 'solubility' as applied to silicates generally.

5.0 TUTOR-MARKED ASSIGNMENT

1. What does the silicate minerals structure possess?
2. Briefly discuss the following colloidal properties in relation to structure:
 - a. Particle size
 - b. Porous structure
 - c. Charged framework.
3. Explain the term 'solubility' as applied to silicates generally.

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UNIT 2 SOIL FORMING ROCKS

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Soil Forming Rocks
 - 3.2 How Weathering Drives Soil Formation
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 - 3.4 Factors Affecting Soil Formation
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- 5.0 Summary
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1.0 INTRODUCTION

Weathering is the name given to the process by which rocks are broken down to form soils. Rocks and geological sediments are the main parent materials of soils (the materials from which soils have formed). There is a very wide variety of rocks in the world, some acidic, some alkaline, some coarse-textured like sands, and some fine-textured and clayey. It is from the rocks and sediments that soils inherit their particular texture. When you see rocks in the landscape it is easy to appreciate how long the process of breaking down rocks to form soil takes. In fact, it can take over 500 years to form just one centimeter of soil from some of the harder rocks. Fortunately, in some respects at least, huge amounts of rocks were broken down during the Ice Age over 10,000 years ago and converted into clays, sands or gravels, from which state it was easier to form soils. There are three main types of weathering; physical, chemical and biological. **Physical weathering** is the influence of processes such as freezing and thawing, wetting and drying, and shrinking and swelling on rocks and other sediments, leading to their breakdown into finer and finer particles. **Chemical weathering** is the decomposition of rocks through a series of chemical processes such as acidification, dissolution and oxidation. Some minerals, while stable within solid rock, become less stable on being more exposed to the atmosphere and so begin to alter in the rocks near the surface, destabilizing the rocks. **Biological weathering** is the effect of living organisms on the breakdown of rock. This involves, for example, the effects of plant roots and soil organisms. Respiration of carbon dioxide by plant roots can lead to the formation of carbonic acid which can chemically attack rocks and sediments and help to turn them into soils. There are a whole range of weathering processes

at work near the surface of the soil, acting together to break down rocks and minerals to form soil.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss the three different types of rocks
- explain five different factors responsible for soil formation
- examine how soil is formed through weathering.

3.0 MAIN CONTENT

3.1 Soil Forming Rocks

Soils may be formed in place from rock or formed in weathered rock and minerals that have been transported from where the original rock occurred.

Rocks are generally classified into three types on the basis of their origin and structure.

These are as under:

- (1) Igneous rocks,
- (2) Sedimentary rocks,
- (3) Metamorphic Rocks

1. Igneous rocks:

These are formed by cooling and solidification of molten magma or lava which escapes through fissures from the interior of the earth. The solidification may occur at great depths resulting in the plutonic rocks, or at moderate depth in the earth forming the so-called intrusive rocks or on the surface of the earth forming the extrusive or effusive rocks. Extrusive rocks are formed as a result of volcanic activities. Important igneous rocks involved in the weathering process are granite, basalt, gabbro diabase, syenite, andesite etc. They are composed of primary minerals such as quartz, feldspars and biotite, augite and hornblende.

2. Sedimentary rocks:

These develop due to gradual accumulation and consolidation of weathering product or mineral particles brought by wind or water on the surface of earth. Such rocks are characterised by the presence of distinct sediments or layers in them. Some examples of sedimentary rocks are limestone, shale; conglomerate, clays, quartz and dolomite are the common dominant minerals in them. Sedimentary rocks do not weather as rapidly as the igneous rocks.

3. Metamorphic rocks:

These are formed after transformation of igneous and sedimentary rocks when they are subjected to intense heat and pressure and are influenced chemically by active and gases. Important metamorphic rocks which take part in weathering process are formed from shale' marble (formed from limestone), schist (formed from shale) and quartzite (from sandstone). Dominant minerals are quartz, clays and calcite.

3.2 How Weathering Drives Soil Formation

If you start with a big boulder in the ground and leave it there for a long, long time. Over thousands of years, it will break down because of weathering. Both mechanical and chemical weathering plays an important role in soil formation. These processes will transform that boulder into particles of sands, silt and clays. Soil formation is due to the percolation of water and weathering. This water seepage is what dissolves rocks into soil. As water seeps downwards, it breaks down material. Eventually, you get a set of "soil layers". These soil layers are dependent on how far down water seeps into the ground and the amount of weathering.

Soil formation is a function of the following 5 factors:

- Climate
- Relief
- Organisms
- Parent material
- Time

Climate – temperature and water

Climate, temperature and water affect how fast chemical weathering occurs. Moist regions experience more chemical weathering because water affects the abundance of chemical reactions.

In addition, both temperature and its rate of change are critical in weathering. Again chemical reactions tend to occur more rapidly when temperatures are higher, especially in moist areas.

Organisms – decay and organic matter

As plants die, small organisms break down and decay material into organic matter. During this process, bacteria and plants produce a type of acid. This acid is a big contributor to the breakdown of soil particles. Next, soil with organic material is most important for humans because it supports plant biomass and agricultural crops. This is the topsoil which takes hundreds of thousands of years to form.

Parent material – underlying bedrock

Soil inherits its traits from parent material. For example, feldspars go through a chemical process which converts it into clay. But if you have granite with feldspar, it could develop into clay and sand. As mentioned above, soils can develop from bedrock in one place. But they can also be transported. For example, glaciers, water and wind can move and erode material.

relief – terrain Characteristics

The aspect that a hill faces influences the amount of sunlight it receives. If there's more sunlight, it changes the amount of available water. At the base of a landform, thicker, hummocky soils tend to form. Gravity moves water downward which is where water accumulates.

Time – weathering duration

Finally, all these factors take a long, long period of time to make a noticeable difference. Soils take thousands of years to form. In general, the longer rocks have the chance to weather; it will result in finer particle sizes.

But the reality is that our environment is constantly in flux. Soil formation factors such as water, organisms and relief is never a constant.

3.3 How Soil is Formed

Soil is the thin layer of material covering the earth's surface and is formed from the weathering of rocks. It is made up mainly of mineral particles, organic materials, air, water and living organisms—all of which interact slowly yet constantly most plants get their nutrients from the soil and they are the main source of food for humans, animals and birds. Therefore, most living things on land depend on soil for their existence Soil is a valuable resource that needs to be carefully managed as it is easily damaged, washed or blown away. If we understand soil and manage it properly, we will avoid destroying one of the essential building blocks of our environment and our food security.

The soil profile

As soils develop over time, layers (or horizons) form a soil profile. Most soil profiles cover the earth as 2 main layers—topsoil and subsoil. Soil horizons are the layers in the soil as you move down the soil profile. A soil profile may have soil horizons that are easy or difficult to distinguish.

Most soils exhibit three main horizons:

- **A horizon:** humus-rich topsoil where nutrient, organic matter and biological activity are highest (i.e. most plant roots, earthworms, insects and micro-organisms are active). The A horizon is usually darker than other horizons because of the organic materials.
- **B horizon:** clay-rich subsoil. This horizon is often less fertile than the topsoil but holds more moisture. It generally has a lighter colour and less biological activity than the A horizon. Texture may be heavier than the A horizon too.
- **C horizon:** underlying weathered rock (from which the A and B horizons form).

Some soils also have an **O horizon** mainly consisting of plant litter which has accumulated on the soil surface. The properties of horizons are used to distinguish between soils and determine land-use potential.

3.4 Factors Affecting Soil Formation

Soil forms continuously, but slowly, from the gradual breakdown of rocks through weathering. Weathering can be a physical, chemical or biological process:

- Physical weathering—breakdown of rocks from the result of a mechanical action. Temperature changes, abrasion (when rocks collide with each other) or frost can all cause rocks to break down.
- Chemical weathering—breakdown of rocks through a change in their chemical makeup. This can happen when the minerals within rocks react with water, air or other chemicals.
- Biological weathering—the breakdown of rocks by living things. Burrowing animals help water and air get into rock, and plant roots can grow into cracks in the rock, making it split.

The accumulation of material through the action of water, wind and gravity also contributes to soil formation. These processes can be very

slow, taking many tens of thousands of years. Five main interacting factors affect the formation of soil:

- parent material—minerals forming the basis of soil
- living organisms—influencing soil formation
- climate—affecting the rate of weathering and organic decomposition
- topography—grade of slope affecting drainage, erosion and deposition
- Time—influencing soil properties.

Interactions between these factors produce an infinite variety of soils across the earth's surface.

Parent materials

Soil minerals form the basis of soil. They are produced from rocks (parent material) through the processes of weathering and natural erosion. Water, wind, temperature change, gravity, chemical interaction, living organisms and pressure differences all help break down parent material.

The types of parent materials and the conditions under which they break down will influence the properties of the soil formed. For example, soils formed from granite are often sandy and infertile whereas basalt under moist conditions breaks down to form fertile, clay soils.

Organisms

Soil formation is influenced by organisms (such as plants), micro-organisms (such as bacteria or fungi), burrowing insects, animals and humans. As soil forms, plants begin to grow in it. The plants mature, die and new ones take their place. Their leaves and roots are added to the soil. Animals eat plants and their wastes and eventually their bodies are added to the soil. This begins to change the soil. Bacteria, fungi, worms and other burrowers break down plant litter and animal wastes and remains, to eventually become organic matter. This may take the form of peat, humus or charcoal.

Climate

Temperature affects the rate of weathering and organic decomposition. With a colder and drier climate, these processes can be slow but, with heat and moisture, they are relatively rapid. Rainfall dissolves some of the soil materials and holds others in suspension. The water carries or leaches these materials down through the soil. Over time this process can change the soil, making it less fertile.

Topography

The shape, length and grade of a slope affect drainage. The aspect of a slope determines the type of vegetation and indicates the amount of rainfall received. These factors change the way soils form. Soil materials are progressively moved within the natural landscape by the action of water, gravity and wind (for example, heavy rains erode soils from the hills to lower areas, forming deep soils). The soils left on steep hills are usually shallower. Transported soils include:

- Alluvial (water transported)
- Colluvial (gravity transported)
- Aeolian (wind transported) soils.

Time

Soil properties may vary depending on how long the soil has been weathered. Minerals from rocks are further weathered to form materials such as clays and oxides of iron and aluminum.

4.0 CONCLUSION

In conclusion we can say that the study of Soil forming rocks is essentially hinged on the knowledge of rocks which are generally classified into three types on the basis of their origin and structures namely: Igneous rocks, sedimentary rocks and metamorphic rocks.

5.0 SUMMARY

In this unit we have learnt that:

- a. Rocks are generally classified into three types on the basis of their origin and structure.
- b. Both mechanical and chemical weathering plays an important role in soil formation.
- c. Soil formation is a function of five different factors.
- d. Soils develop over time layers (or horizons) to form a soil profile.

6.0 TUTOR-MARKED ASSIGNMENT

1. Differentiate between igneous rocks and sedimentary rocks.
2. Soil formation as a function of five different factors discuss any two factors.

3. A soil profile may have soil horizons that are easy or difficult to distinguish what is the function of the A horizon in relation to crop production.
4. What are the different types of micro-organisms that influenced soil formation?

7.0 REFERENCES/FURTHER READING

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UNIT 3 MINERAL COMPOSITION OF SOIL

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 The Size Range
 - 3.2 The importance of soil texture
 - 3.3 Silicate Chemistry
 - 3.3.1 Silicate Structures and Structural Formula
 - 3.3.2 General Formula for Silicates
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 References/Further Reading

1.0 INTRODUCTION

The largest component of soil is the mineral portion, which makes up approximately 45% to 49% of the volume. Soil minerals are derived from two principal mineral types. Primary minerals, such as those found in sand and silt, are those soil materials that are similar to the parent material from which they formed. They are often round or irregular in shape. Secondary minerals, on the other hand, result from the weathering of the primary minerals, which releases important ions and forms more stable mineral forms such as silicate clay. Clays have a large surface area, which is important for soil chemistry and water-holding capacity.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss different components of soil
- examine mineral composition of soil.
- explain the effect of elements in the Earth's crust on minerals composition of soil.

3.0 MAIN CONTENT

3.1 The Size Range

Rock fragments and mineral particles in soil vary enormously in size from boulders and stones down to sand grains and very small particles that are beyond the resolving power of an optical microscope ($< 0.2 \mu\text{m}$ in diameter). Particles smaller than $c.1\mu\text{m}$ are classed as colloidal. Particles that do not settle quickly when mixed with water are said to form a colloidal solution or sol; if they settle within a few hours they form a suspension. Colloidal solutions are distinguished from true solutions (dispersions of ions and molecules) by the Tyndall effect. This occurs when the path of a beam of light passing through the solution can be seen from either side at right angles to the beam, indicating a scattering of the light rays.

An arbitrary division is made by size-grading soil into material:

- That passes through a sieve with 2 mm diameter holes – the *fine earth*, and
- That retained on the sieve ($> 2 \text{ mm}$) – *the stones or gravel*, but smaller than
- fragments $> 600 \text{ mm}$, which are called *boulders*.

The separation by sieving is carried out on air-dry soil that has been gently ground by mortar and pestle, or crushed between wooden rollers, to break up the aggregates. Air-dry soil is soil allowed to dry in air at ambient temperatures (between 20 and 40°C).

Particle-size distribution of the fine earth

The distribution of particle sizes determines the soil texture, which may be assessed subjectively in the field or more rigorously by particle-size analysis in the laboratory.

Size classes

All soils show a continuous range of particle sizes, called a frequency distribution, which is obtained by plotting the number (or mass) of particles of a given size against their actual size. When the number or mass in each size class is summed sequentially we obtain a cumulative distribution of soil particle sizes. In practice, it is convenient to subdivide the continuous distribution into several class intervals that define the size limits of the *sand*, *silt* and *clay* fractions. The extent of this subdivision, and the class limits chosen, vary from country to country and even between institutions within countries. The major systems in use are those adopted by the Soil Survey Staff of the USDA,

the British Standards Institution and the International Union of Soil Sciences (IUSS). All three systems set the upper limit for clay at 2 μm diameter, but differ in the upper limit chosen for silt and the way in which the sand fraction is subdivided.

Particle-size analysis in the laboratory

The success of the method relies on the complete disruption of soil aggregates and the addition of chemicals that ensure dispersion of the soil colloids in water. Full details of the methods employed are given in standard texts. The sand particles are separated by sieving; silt and clay are separated using the differences in their settling velocities in suspension. The result of particle-size analysis is expressed as the mass of the individual fractions per 100 g of oven-dry (o.d.) soil (fine earth only). Oven-dry soil is soil dried to a constant weight at 105°C.

3.2 The Importance of Soil Texture

Soil scientists are primarily interested in the texture of the fine earth fraction. Nevertheless, in some soils the size and abundance of stones cannot be ignored because they can have a marked influence on the soil's suitability for agriculture. As the stone content increases, a soil holds less water than a stoneless soil of the same fine-earth texture, so that crops become more susceptible to drought. Conversely, such soils may be better drained and therefore warm up more quickly in spring in cool temperate regions. Large stones on the soil surface act as sinks during daytime for heat energy that is slowly released at night – this is of benefit in cool climate vineyards, where frost in spring and early summer can damage flowering and fruit set. Stoniness also determines the ease, and to some extent the cost of cultivation, as well as the abrasive effect of the soil on tillage implements.

Texture is one of the most stable soil properties and is a useful index of several other properties that determine a soil's agricultural potential. Fine and medium-textured soils, such as clays, clay loams, silty clays and silty clay loams, are generally more desirable than coarse-textured soils because of their superior retention of nutrients and water. Conversely, where rapid infiltration and good drainage are required, as for irrigation or liquid waste disposal, sandy or coarse-textured soils are preferred. In farming terms, clay soils are described as 'heavy' and sandy soils as 'light', which does not refer to their mass per unit volume, but to the power required to draw a plough or other implements through the soil. Because it is easy to estimate, and is routinely measured in soil surveys, texture (and more specifically clay content) has been used as a 'surrogate' variable for other soil properties that are less easily measured, such as the cation exchange capacity.

Texture has a pronounced effect on soil temperature. Clays hold more water than sandy soils, and the presence of water considerably modifies the heat required to change a soil's temperature because:

- its specific heat capacity is three to four times that of the soil solids
- considerable latent heat is either absorbed or evolved during a change in the physical state of water, for example, from ice to liquid or vice versa. Thus, the temperature of wet clay soils responds more slowly than that of sandy soils to changes in air temperature in spring and autumn.

Texture should not be confused with tilth, of which it is said that a good farmer can recognize it with his boot, but no soil scientist can describe it. Tilth refers to the condition of the surface of ploughed soil prepared for seed sowing: how sticky it is when wet and how hard it sets when dry. The action of frost in cold climates breaks down the massive clods left on the surface of a heavy clay soil after autumn ploughing, producing a mellow 'frost tilth' of numerous small granules.

Mineralogy of the sand and silt fractions

Simple crystalline structures

Sand and silt consist almost entirely of the resistant residues of primary rock minerals, although small amounts of secondary minerals (salts, oxides and hydroxides) formed by weathering also occur. The primary rock minerals are predominantly silicates, which have a crystalline structure based upon a simple unit – the silicon tetrahedron, SiO_4^{4-} . An electrically neutral crystal is formed when cations, such as Al^{3+} , Fe^{3+} , Fe^{2+} , Ca^{2+} , Mg^{2+} , K^+ and Na^+ , become covalently bonded to the O atoms in the tetrahedron and the surplus valencies of the O^{2-} ions in the SiO_4^{4-} group are satisfied. An example of this kind of structure is the primary mineral olivine, which has the composition $(\text{Mg, Fe})_2\text{SiO}_4$

Coordination number

The packing of the O atoms, which are the largest of the more abundant elements in the silicates, determines the crystalline dimensions. In quartz, for example, O occupies 98.7% and Si only 1.3% of the mineral volume. The size ratio of Si to O is such that four O atoms can be packed around one Si, and larger cations such as iron (Fe) can accommodate more O atoms. The cation to oxygen radius ratio determines the coordination number of the cation. Many of the common metal cations have radius ratios between 0.41 and 0.73, which means

that an octahedral arrangement of six O atoms around the cation (coordination number 6) is possible. Larger alkali and alkaline earth cations, such as K^+ and Ba^{2+} , that have radius ratios > 0.73 form complexes of coordination number 8 or greater. Aluminium, which has a cation to oxygen radius ratio close to the maximum for coordination number 4 and the minimum for coordination number 6 (0.41), can exist in either fourfold (IV) or sixfold (VI) coordination.

Isomorphous substitution

Elements of the same valency and coordination number frequently substitute for one another in a silicate structure – a process called isomorphous substitution. The structure remains electrically neutral. However, when elements of the same coordination number but different valency are exchanged, there is an imbalance of charge. The most common substitutions are Mg^{2+} , Fe^{2+} or Fe^{3+} for Al^{3+} in octahedral coordination, and Al^{3+} for Si^{4+} in tetrahedral coordination. The excess negative charge is neutralized by the incorporation of additional cations, such as unhydrated K^+ , Na^+ , Mg^{2+} or Ca^{2+} into the crystal structure, or by structural arrangements that allow an internal compensation of charge (see the chlorites).

More complex crystalline structures

Chain structures

These are represented by the pyroxene and amphibole groups of minerals, which collectively make up the ferromagnesian minerals. In the pyroxenes, each Si tetrahedron is linked to adjacent tetrahedra by the sharing of two out of three basal O atoms to form a single extended chain. In the amphiboles, two parallel pyroxene chains are linked by the sharing of an O atom in every alternate tetrahedron. Cations such as Mg^{2+} , Ca^{2+} , Al^{3+} and Fe^{2+} are ionic covalently bonded to the O^{2-} ions to neutralize the surplus negative charge.

Mineralogy of the clay fraction

A large assortment of minerals of varying degrees of crystallinity occurs in the clay fraction of soils. Broadly, these minerals may be divided into the crystalline clay minerals – predominantly phyllosilicates – and other minerals (oxides, hydroxides and salts). Because of their large specific surface areas and surface charges, these minerals are very important sites for physical and chemical reactions in soil.

For many years the small size of clay particles prevented scientists from elucidating their mineral structure. It was thought that the clay fraction consisted of inert mineral fragments enveloped in an amorphous gel of hydrated sesquioxides ($Fe_2O_3 \cdot nH_2O$ and $Al_2O_3 \cdot nH_2O$)* and silicic acid ($Si(OH)_4$). The surface gel was amphoteric, the balance between

acidity and basicity being dependent on the soil pH. Between pH 5 and 8, the surface was usually negatively charged (proton deficient), which could account for the observed cation exchange properties of soil. During the 1930s, however, the crystalline nature of the clay minerals was established unequivocally by X-ray diffraction (XRD). Most of the minerals were found to have a phyllosilicate structure similar to the micas and chlorite. The various mineral groups were identified from their characteristic *d* spacings, as measured by XRD, and their unit cell compositions deduced from elemental analyses. Subsequent studies using scanning and transmission electron microscopes have confirmed the conclusions of the early work.

In addition, within the accessory minerals there are the weathered residues of resistant primary minerals that have been comminuted to colloidal size, and soil minerals synthesized during pedogenesis. The latter mainly comprise Al and Fe hydroxides and oxyhydroxides, which occur as discrete particles or as thin coatings on the clay minerals. The crystallinity of these minerals varies markedly depending on their mode of formation, the presence of other elements as inclusions, and their age. Some, such as the iron hydroxide ferrihydrite, were previously thought to be amorphous, but are now known to form extremely small crystals and to possess short-range order: that is, their structure is regular over distances of a few nanometres, but disordered over larger distances (tens of nanometres).

The crystalline clay minerals

Most clay minerals have a phyllosilicate structure, but a small group – the sepiolite-palygorskite series – has chain structures and another group – the allophanes – forms hollow spherical crystals. Palygorskite and sepiolite are unusual in having very high Si: Al ratios, with Mg occupying most of the octahedral positions. Sepiolite is very rare in soil and palygorskite survives only in soils of semi-arid and arid regions. They are not discussed further.

Under mild (generally physical) weathering conditions, clay minerals may be inherited as colloidal fragments of primary phyllosilicates, such as muscovite mica. Under more intense weathering, the primary minerals may be transformed to secondary clay minerals, as when soil illites, vermiculites and smectites are formed by the leaching of interlayer K from primary micas, or from the weathering of chlorites. Neof ormation of clay minerals is a feature of intense weathering, or of diagenesis in sedimentary deposits, when minerals completely different from the original primary minerals are formed. When the soluble silica concentration in the weathering environment is high, 2:1 layer minerals such as smectites are likely to form. Leaching and removal of silica,

however, can produce kaolinite and aluminium hydroxide. Increased negative charge in the crystal due to isomorphous substitution of Al^{3+} for Si^{4+} in the smectites leads to K^+ being the favoured interlayer cation, with the resultant formation of illite.

3.3 Silicate Chemistry

Silicate minerals are the most common of Earth's minerals and include quartz, feldspar, mica, amphibole, pyroxene, and olivine. Silica tetrahedra, made up of silicon and oxygen, **form** chains, sheets, and frameworks, and bond with other cations to **form** silicate minerals.

3.3.1 Silicate Structures and Structural Formula

The relative abundance of elements in the Earth's crust determines what minerals will form and what minerals will be common. Because Oxygen and Silicon are the most abundant elements, the silicate minerals are the most common. Thus, we will spend some time here on the structure, chemistry, and occurrence of silicate minerals. In order to discuss the silicates and their structures it is first necessary to remember that the way atoms are packed together or coordinated by larger anions, like oxygen depends on the radius ratio of the cation to the anion, R_x/R_z .

Element	Wt%	Atomic%	Volume%
O	46.60	62.55	~94
Si	27.72	21.22	~6
Al	8.13	6.47	
Fe	5.00	1.92	
Ca	3.63	1.94	
Na	2.83	2.34	
K	2.59	1.42	
Mg	2.09	1.84	
Total	98.59	100.00	100

R_x/R_z	C.N.	Type
1.0	12	Hexagonal or Cubic Closest Packing
1.0 - 0.732	8	Cubic
0.732 - 0.414	6	Octahedral
0.414 - 0.225	4	Tetrahedral

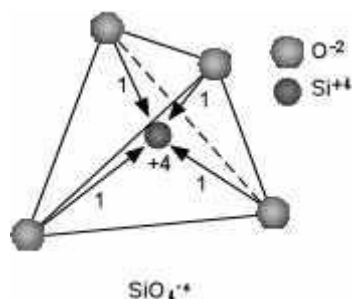
0.225 -0.155	3	Triangular
<0.155	2	Linear

Since oxygen is the most abundant element in the crust, oxygen will be the major anion that coordinates the other cations. Thus, for the major ions that occur in the crust, we can make the following table showing the coordination and coordination polyhedra that are expected for each of the common cations.

Ion	C:N (with oxygen)	Coord. polyhedron	Ionic radius A
K ⁺	8-12	Cubic to closest	1.51 (8) -1.64 (12)
Na ⁺	8-6	Cubic to octahedral	1.18 (8) -1.02 (6)
Ca ⁺²	8-6	Cubic to octahedral	1,12 (8) – 1,00 (6)
Mn ⁺²	6	Octahedral	0.83
Fe ⁺²	6	Octahedral	0.78
Mg ⁺²	6	Octahedral	0.72
Fe ⁺²	6	Octahedral	0.65
Ti ⁺⁴	6	Octahedral	0.61
Al ⁺³	6	Octahedral	0.54
Al ⁺³	4	Tetrahedral	0.39
Si ⁺⁴	4	Tetrahedral	0.26
C ⁺⁴	3	Triangular	0.08

The radius ratio of Si⁺⁴ to O⁻² requires that Si⁺⁴ be coordinated by 4 O⁻² ions in tetrahedral coordination.

In order to neutralize the +4 charge on the Si cation, one negative charge from each of the Oxygen ions will reach the Si cation. Thus, each Oxygen will be left with a net charge of -1, resulting in a SiO₄⁻⁴ tetrahedral group that can be bonded to other cations. It is this SiO₄⁻⁴ tetrahedron that forms the basis of the silicate minerals.



Since Si⁺⁴ is a highly charged cation, Pauling's rules state that it should be separated as far as possible from other Si⁺⁴ ions. Thus, when these SiO₄⁻⁴ tetrahedrons are linked together, only corner oxygens will be

shared with other SiO_4^{-4} groups. Several possibilities exist and give rise to the different silicate groups.

Nesosilicates (island silicates)

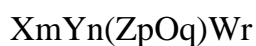
If the corner oxygens are not shared with other SiO_4^{-4} tetrahedrons, each tetrahedron will be isolated. Thus, this group is often referred to as the island silicate group. The basic structural unit is then SiO_4^{-4} . In this group the oxygens are shared with octahedral groups that contain other cations like Mg^{+2} , Fe^{+2} , or Ca^{+2} . Olivine is a good example: $(\text{Mg}, \text{Fe})_2\text{SiO}_4$.

3.3.2 General Formula for Silicates

Based on these basic structural units, we can construct a general structural chemical formula for the silicates. But one substitution in particular tends to mess things up a bit. This is Al^{+3} , the third most abundant element in the Earth's crust. Al^{+3} has an ionic radius that varies between 0.54 and 0.39 depending on the coordination number. Thus, it could either fit in 6-fold coordination with oxygen or 4-fold coordination with oxygen. Because Al^{+3} will go into 4-fold coordination with oxygen, it sometimes substitutes for Si^{+4} . If such a substitution takes place, it creates a charge imbalance that must be made up elsewhere in the silicate structure.

The other common elements in the Earth's crust that enter the silicates do so in other types of coordination. Ions like Al^{+3} , Mg^{+2} , Fe^{+2} , Fe^{+3} , Mn^{+2} , and Ti^{+4} enter into 6-fold or octahedral sites. Larger ions like Ca^{+2} , and Na^{+1} , are found in octahedral coordination or 8-fold, cubic coordination sites. Very large cations like K^{+1} , Ba^{+2} , and sometimes Na^{+1} are coordinated by 12 oxygens in 12-fold coordination sites.

We can thus write a general structural formula for the silicates as follows:



where X represents an 8 to 12-fold coordination site for large cations like K^+ , Rb^+ , Ba^{+2} , Na^+ , and Ca^{+2} .

Y represents a 6-fold (octahedral) site for intermediate sized cations like Al^{+3} , Mg^{+2} , Fe^{+2} , Fe^{+3} , Mn^{+2} , and Ti^{+4} .

Z represents the tetrahedral site containing Si^{+4} , and Al^{+3} . the ratio p:q depends on the degree of polymerisation of the silica (or alumina) tetrahedrons, or the silicate structural type as discussed above.

O is oxygen, and W is a hydroxyl (OH-1) site into which can substitute large anions like F-1 or Cl-1.

The subscripts m, n, and r depend on the ratio of p to q and are chosen to maintain charge balance.

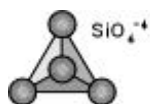
This is summarised in the table below. In this table note that there is very little substitution that takes place between ions that enter the X, Y, and Z sites. The exceptions are mainly substitution of Al⁺³ for Si⁺⁴, which is noted in the Table, and whether the X site is large enough to accept the largest cations like K⁺¹, Ba⁺², or Rb⁺¹.

Site	C.N.	Ion
Z	4	Si ⁺⁴
Z	4	Al ⁺³
Y	6	Al ⁺³
Y	6	Fe ⁺³
Y	6	Fe ⁺²
Y	6	Mg ⁺²
Y	6	Mn ⁺²
Y	6	Ti ⁺⁴
X	8	Na ⁺¹
X	8	Ca ⁺²
X	8-12	K ⁺¹
X	8-12	Ba ⁺²
X	8-12	Rb ⁺¹

Nesosilicates (island silicates)

Next is a systematic look at the most common rock forming minerals, starting with the common nesosilicates. Among these are the olivines, garnets, Al₂SiO₅ minerals, staurolite, and sphene.

The nesosilicates or island silicates are based on the isolated SiO₄⁻⁴ tetrahedral groups. In the olivines, the remaining corner oxygens form octahedral groups that coordinate Mg⁺² and Fe⁺² ions.



Olivines

The olivines consist of a complete solid solution between Mg₂SiO₄ (forsterite, FO) and Fe₂SiO₄ (fayalite, Fa). There is limited substitution of the following end members:

Ca₂SiO₄ - larnite

Mn₂SiO₄ - tephroite

CaMgSiO₄ - monticellite (which is commonly found in metamorphosed dolomites)

4.0 CONCLUSION

In conclusion we can say that the study of minerals composition of the soil is essentially hinged on the knowledge of Sand and silt consist almost entirely of the resistant residues of primary rock minerals, although small amounts of secondary minerals (salts, oxides and hydroxides) formed by weathering also occur. The primary rock minerals are predominantly silicates, which have a crystalline structure based upon a simple unit – the silicon tetrahedron, SiO₄⁴⁻.

5.0 SUMMARY

In this unit we have learnt that:

- a. Rock fragments and mineral particles in soil vary enormously in size from boulders and stones down to sand grains and very small particles that are beyond the resolving power of an optical microscope (< 0.2 µm in diameter).
- b. Sand and silt consist almost entirely of the resistant residues of primary rock minerals, with small amounts of secondary minerals (salts, oxides and hydroxides) which are formed by weathering.
- c. A large assortment of minerals of varying degrees of crystallinity occurs in the clay fraction of soils.
- d. Oxygen and silicon are the most abundant elements and the silicate minerals are the most common.

6.0 TUTOR-MARKED ASSIGNMENT

1. Why it is that presence of water considerably modifies the heat required to change a soil's temperature?
2. Silicate minerals are the most common of Earth's minerals list five among them.

7.0 REFERENCES/FURTHER READING

Robert E. White (2006). *Principles and practice of soil science*.

GDC Handwara (n.d.). Lecture Notes on Silicate Structures and Structural Formula.

MODULE 2

UNIT 1 PROCESSES AND FACTORS OF SOIL FORMATION

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main content
 - 3.1 Soil Formation Processes
 - 3.2 Factors of Soil Formation
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 References/Further Reading

1.0 INTRODUCTION

The processes of soil formation have four major processes. These processes take place simultaneously but at different rates. Equilibrium is rarely attained because the processes are dynamic. The resulting soil reflects the most predominant of these processes. Thus a soil that suffers erosion loses or leaching losses may have low level of soluble elements like calcium, Magnesium, potassium and sodium. Continuous addition of plant or animal residue may on the other hand leads to accumulation of organic matter, creating a unique soil known for example as organic soil. There can be movement of materials in and out of the soil. Dissolved minerals and clay can be translocated from the top soil to the subsoil, while burrowing animals can remove soils from the subsoils to the top and vice versa. Weathering of minerals leads to the formation of new minerals. For example, the weathering of talc, may lead to the formation of smectite in areas with impeded or poor drainage. The transformation process will lead to the formation of new products.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss soil formation processes
- explain factors of soil formation
- parent material is the initial mineral substance that forms a soil
- examine the effect of topography on soil formation.

3.0 MAIN CONTENT

3.1 Soil Formation Processes

Soil development begins with a parent material that has a surface layer altered by vegetation and weathering. For example, a young Coastal Plain soil has relatively uniform material throughout, and is altered only by a dark-stained surface layer that has been formed by vegetation. A more mature soil, on the other hand, shows evidence of the removal and transport of surface-layer clay to a subsurface layer called the B horizon. In an even older soil, chemical weathering and leaching have removed silicon, causing a change in the suite of clay minerals. A senile soil is excessively weathered and dominated by very resistant iron and aluminum oxide minerals. The rate that a young Coastal Plain soil becomes a senile soil depends not on its chronological age but on how rapidly minerals are transported and transformed within the profile.

Human activity frequently alters the process of pedogenesis. Once human activity ends, soil formation can continue as before if no radical change in the soil-forming factors occurred in the interim. Because fine material leaches selectively faster than coarse material, differences between human-altered and undisturbed soils in the ratio of fine to coarse clay may be apparent in a relatively short span of time (one hundred years in a humid environment).

The processes of soil formation have four major processes. The four processes are:

- i. Addition
- ii. Removal or subtraction
- iii. Transfer or Translocation
- iv. Transformation or Synthesis

These processes take place simultaneously but at different rates. Equilibrium is rarely attained because the processes are dynamic. The resulting soil reflects the most predominant of these processes. Thus a soil that suffers erosion losses or leaching losses may have low level of soluble elements like calcium, Magnesium, potassium and sodium. Continuous addition of plant or animal residue may on the other hand leads to accumulation of organic matter, creating a unique soil known for example as organic soil. There can be movement of materials in and out of the soil. Dissolved minerals and clay can be translocated from the top soil to the subsoil, while burrowing animals can remove soils from the subsoils to the top and vice versa. Weathering of minerals leads to

the formation of new minerals. For example, the weathering of talc, may lead to the formation of smectite in areas with impeded or poor drainage. The transformation process will lead to the formation of new products.

3.2 Factors of Soil Formation

The five soil forming factors are: parent material, topography, climate, biological activity, and time. Soil formation begins with a parent material derived from weathering of either the native rock or material transported to the site. The concerted effect of climate and biological activity then transforms parent material by producing the physical and chemical energy to alter minerals and vertically redistribute material through the soil profile. The effect of climate and biological activity is modified by topography.

Parent material

Parent material is the initial mineral substance that forms a soil. It is derived from the weathering of rocks and may reside at the site of its origin or be transported from somewhere else to its current location. A soil formed from parent material found at the site of its origin is called a residual or sedentary soil. Bedrock weathering in place produces a stony, massive material called saprolite. As physical and some chemical weathering occur, the saprolite becomes denser than the underlying bedrock. The texture and original rock structure remain, but the material is soft enough to dig with a hand shovel. As chemical weathering converts primary minerals to secondary minerals, particles are redistributed vertically. As material is both added and removed, a soil develops. A residual soil will retain many of the characteristics of the underlying bedrock. Also the texture, mineralogy, pH, and other characteristics of the soil may be a direct result of the saprolite below.

Material can be eroded from one place and transported to another where it becomes parent material for a soil at the new site. Often weathering occurs before the material is transported to the new site. In this case, the soil may have few features in common with the underlying rock. Transported material can bury an existing soil at the new site. Once a depositional episode is completed, time zero for the new soil's formation begins. Several forces can supply energy for the transportation of parent material: ice, wind, water, and gravity.

Topography (relief)

Topographic relief, or the slope and aspect of the land, has a strong influence on the distribution of soils on a landscape. Two aspects of topography that influences soil formation are the altitude and the slope. Position on a slope influences the soil depth through differences in

accumulation of erosion debris. Slope affects the amount of precipitation that infiltrates into soil versus that which runs off the surface. Aspect, or the direction a slope is facing, affects soil temperature. In northern hemisphere sites, south facing slopes are warmer than those facing north. Differences in moisture and temperature regimes create microclimates that result in differences in vegetation with aspect. Differences in weathering, erosion, leaching, and secondary mineral formation also can be associated with relief.

Increase in altitude leads to decrease in ambient temperature and consequently leading to decrease in soil temperature. An increase of 1000m in altitude leads to a decrease of about 6_ in ambient temperature. This change in temperature creates a micro-climatic condition that differs from the immediate environment. This also creates some unique conditions for the formation of soils that differs from those of the immediate environment.

Climate

Climate arguably has the greatest effect on soil formation. It not only directly affects material translocation (leaching or erosion, for example) and transformation (weathering), but also indirectly influences the type and amount of vegetation supported by a soil. The two most important aspect of climate that has direct bearing on the process of soil formation are precipitation (total amount, intensity and distribution) and temperature (soil temperature). Precipitation is the main force in moving clay and organic matter from the surface to a depth within the profile. When a soil is at field capacity, the addition of more water will result in drainage either downward or laterally. Drainage water carries with it dissolved and suspended clay particles and soluble minerals that collect at a new location within the soil profile. As a result, soils often show an increase in clay with depth resulting in an accumulation of clay and some mineral elements at certain depth of the soil. While the total amount of precipitation affects the amount of dissolved substances passing through the profile, the intensity of precipitation (amount of rainfall per unit time) affects the erosive potentials of the rain. Erosion affects the process of soil formation by reducing the effective solum depth in the upper and middle slope of the toposequence while increasing the soil depth at lower slope and valley bottom. Differences in rainfall distribution pattern accounts for regional variation in soil characteristics. These variations lead to zonation of soil types.

Temperature affects the rate of biochemical and biophysical reactions taking place in the soil environment. As we know, the rate of reaction is directly proportional to the soil temperature. Most soil reactions are enzymatic reaction depending mainly on temperature. There are optima

temperatures for the soil reactions above or below which these reactions will be negatively affected. The rate of chemical weathering as well as biological decomposition of plant and animal remains in the soil environment is determined by the amount of soil moisture and soil temperature. Diurnal and seasonal changes in temperature cause particles to expand and contract unevenly, breaking them apart. Heat and moisture are active agents of chemical weathering, the conversion of one mineral into another. Climate affects the type and amount of vegetation in a region. A warm, humid climate produces the most vegetative growth; however, microbial decomposition is also rapid.

The net effect is that tropical and subtropical soils are generally low in organic content. In contrast, organic matter tends to be highest in a cool damp environment where decomposition is slow. Temperature and the amount of water moving through a profile affect all of the following:

- The amount and characteristics of organic matter;
- The depth at which clay accumulates;
- The type of minerals present;
- Soil pH (humid climates tend to produce more acidic soil than do arid climates);
- Soil color;
- Iron, aluminum, and phosphorus distributions within a soil profile; and
- The depth to calcium carbonate and/or salt accumulation.

Biological activity

Biological activity and climate are active forces in soil formation. Soil pedogenesis involves a variety of animals, plants, and microorganisms. Ants, earthworms, and burrowing animals, for example, mix more soil than do humans through plowing and construction. Plant roots remove mineral nutrients from subsoil and redeposit them at the surface in leaf litter. Growing roots open channels through soil where rainwater can wash clay and organic matter down along these channels. Soil microbes decompose plant and animal debris, releasing organic acids. This biochemical activity is the catalyst for a great deal of the oxidation/reduction and other chemical reactions in soil.

The distribution of organic matter in a forest soil is different from that of grassland. The surface soils of forests tend to have concentrated organic matter, which quickly decreases with depth. Grassland soils tend to accumulate organic matter to a greater depth than do forest soils. It is important for pedologist to note that the dark staining from the humic fraction of organic matter can persist in a buried soil. Thus, ancient buried surface soils may be recognized in the field by color alone.

The distribution of iron and aluminum throughout a profile also differs between forest and grass land soils. In forests, due to the greater rainfall, clays and organics drain downward, leaving behind resistant minerals. As a result, iron and aluminum in B horizons in forest soils are found in higher concentrations than in grassland soils.

Time

Soils develop over time. Soil formation is a dynamic process, where a steady state is slowly approached but only rarely reached. The rate at which a soil forms is related more to the intensity of other soil forming factors than to chronological age.

Soil formation processes

Soil development begins with a parent material that has a surface layer altered by vegetation and weathering. For example, a young Coastal Plain soil has relatively uniform material throughout, and is altered only by a dark-stained surface layer that has been formed by vegetation. A more mature soil, on the other hand, shows evidence of the removal and transport of surface-layer clay to a subsurface layer called the B horizon. In an even older soil, chemical weathering and leaching have removed silicon, causing a change in the suite of clay minerals. A senile soil is excessively weathered and dominated by very resistant iron and aluminum oxide minerals. The rate that a young Coastal Plain soil becomes a senile soil depends not on its chronological age but on how rapidly minerals are transported and transformed within the profile.

Human activity frequently alters the process of pedogenesis. Once human activity ends, soil formation can continue as before if no radical change in the soil-forming factors occurred in the interim. Because fine material leaches selectively faster than coarse material, differences between human-altered and undisturbed soils in the ratio of fine to coarse clay may be apparent in a relatively short span of time (one hundred years in a humid environment).

The processes of soil formation have been summarised into four major processes and represented schematically in the above figure. The four processes are addition, removal or subtraction, transfer or translocation and transformation or synthesis. These processes take place simultaneously but at different rates. Equilibrium is rarely attained because the processes are dynamic. The resulting soil reflects the most predominant of these processes. Thus a soil that suffers erosion or leaching losses may have low level of soluble elements like calcium, Magnesium, potassium and sodium. Continuous addition of plant or animal residue may on the other hand leads to accumulation of organic matter, creating a unique soil known for example as organic soil. There

can be movement of materials in and out of the soil. Dissolved minerals and clay can be translocated from the top soil to the subsoil, while burrowing animals can remove soils from the subsoils to the top and vice versa. Weathering of minerals leads to the

Formation of new minerals. For example, the weathering of talc, may lead to the formation of smectite in areas with impeded or poor drainage. The transformation process will lead to the formation of new products.

4.0 CONCLUSION

The processes of soil formation have four major processes. The four processes are addition, removal or subtraction, transfer or translocation and transformation or synthesis. These processes take place simultaneously but at different rates. Equilibrium is rarely attained because the processes are dynamic. The resulting soil reflects the most predominant of these processes. Thus a soil that suffers erosion loses or leaching losses may have low level of soluble elements like calcium, Magnesium, potassium and sodium. Continuous addition of plant or animal residue may on the other hand leads to accumulation of organic matter, creating a unique soil known for example as organic soil. There can be movement of materials in and out of the soil. Dissolved minerals and clay can be translocated from the top soil to the subsoil, while burrowing animals can remove soils from the subsoils to the top and vice versa.

5.0 SUMMARY

In this unit, we have learnt that:

- i. There are five different soil forming factors.
- ii. Soil formation begins with a parent material derived from weathering of either the native rock or material transported to the site.
- iii. Parent materials are the initial mineral substance that forms a soil.
- iv. Topography has a strong influence on the distribution of soils on a landscape.
- v. Climate has the greatest effect on soil formation.

6.0 TUTOR-MARKED ASSIGNMENT

1. List all the five different soil forming factors you have learned
2. Parent materials are the initial mineral substance that forms a soil discuss.

3. Differentiate between topography and climate as a soil forming factors.
4. Which of the two factors above have the greatest effect on soil formation?

7.0 REFERENCE/FURTHER READING

Dr. Soretire, A. A. & Dr. Azeez, J.O. (n.d.). Lecture Note on Principles of Soil Science.

Robert E. White (2006). *Principles and Practice of Soil Science*.

UNIT 2 PHYSICAL STATE OF SOILS

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Texture
 - 3.2 Structure
 - 3.3 Porosity
 - 3.4 Inherent Factors Affecting Bulk Density and Available Water Capacity
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 Reference/Further Reading

1.0 INTRODUCTION

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

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss the relative proportions of particles of various sizes such as sand, silt and clay in the soil.
- explain the percentages of sand, silt and clay in a soil
- examine the effect of texture in regard to moisture and air relationships in the soil
- analyse the porosity of a soil in relation to the total volume of the soil material.

3.0 MAIN CONTENT

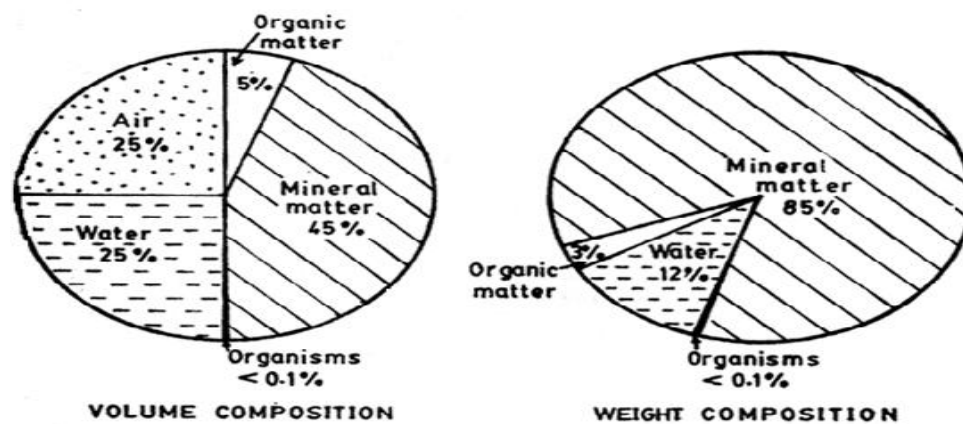
3.1 Texture

Texture refers to the relative proportions of particles of various sizes such as sand, silt and clay in the soil. The proportions of the separates in classes commonly used in describing soils are given in the textural triangle shown in Fig. 3. In using the diagram, the points corresponding to the percentages of silt and clay present in the soil under consideration are located on the silt and clay lines respectively. Lines are then projected inward, parallel in the first case to the clay side of the triangle and in the second case parallel to the sand side. The name of the compartment in which the two lines intersect is the class name of the soil in question. For example soil containing 15% clay, 20% silt and 65% sand is sandy loam and a soil containing equal amounts of sand, silt and clay is clay loam.

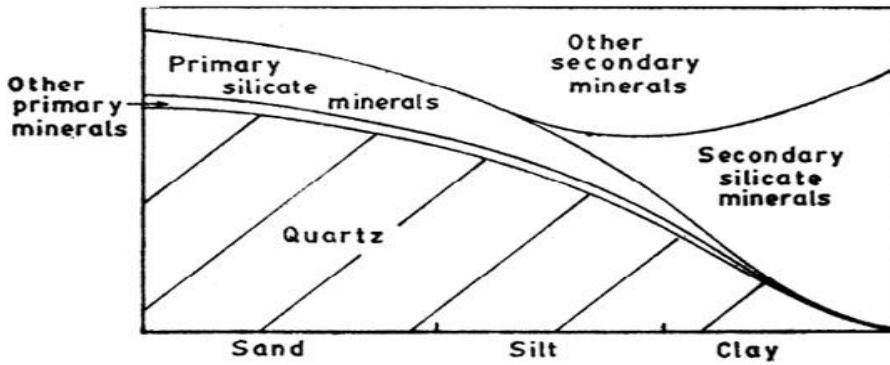
The percentages of sand, silt and clay in a soil could be determined in a soil laboratory by two standard methods - hydrometer method and pipette method (Black *et al.*, 1965a). Both methods depend on the fact that at any given depth in a settling suspension the concentration of the particles varies with time, as the coarser fractions settle at a faster rate than the finer (Fig.4).

In the field, soil texture could be estimated by the following methods:

(i) **Feel method:** In this method, the soil is moistened with water and rubbed between the thumb and fingers. The way the wet soil “slicks out” gives a good idea of the clay content. The sand particles are gritty, the silt has a floury or talcum - powder feel when dry and is only moderately plastic and sticky when wet. Accuracy of this method depends largely on experience.



Volume and weight composition of a soil (Percentage of air and water varies according to moisture saturation of soil)



Mineralogical composition of soil (area within the figure denotes the relative abundance of minerals)

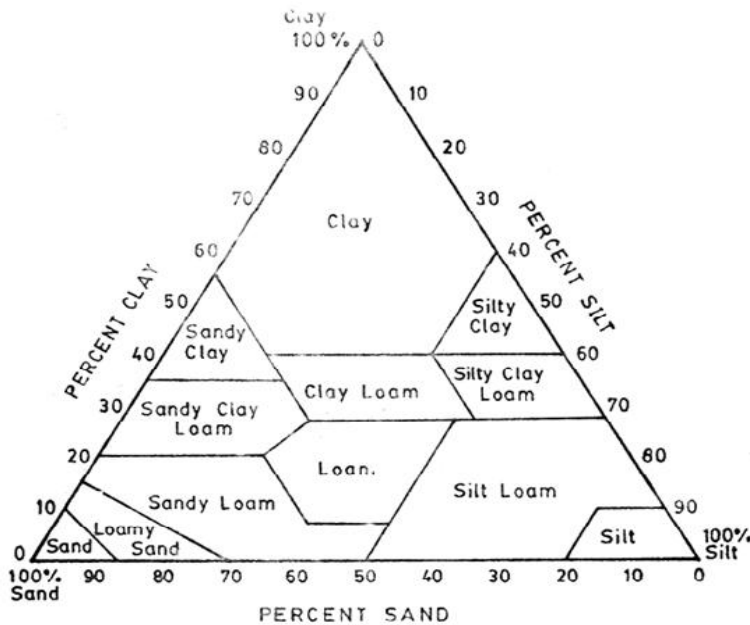


Fig. 3: Soil Textural Triangle

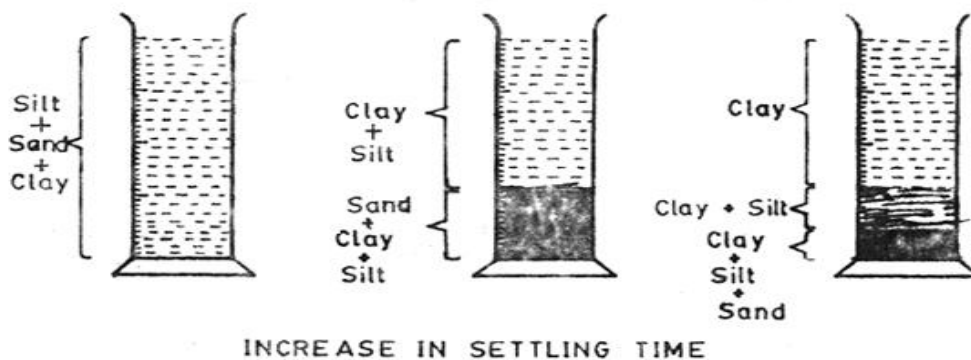


Fig. 4: The Settling of Particles in a Soil Suspension

(ii) **Ball and ribbon method:**

The procedure of this method as described by Coche and Laughlin (1985) is as follows: Take a handful of soil and wet it so that it begins to stick together without sticking to the hand. A ball of about 3 cm diameter is made and put down. If it falls apart it is **sand**. If it sticks together roll the ball into a sausage shape 6 – 7 cm long. If it does not remain in this form it is **loamy sand**. If it remains in this shape, continue to roll until it reaches 15 – 16 cm long. If it does not remain in this form, it is **sandy loam**. If it remains in this shape, try to bend the sausage into a half circle and if it doesn't, it is a loam. If it does, bend the sausage to form a full circle and if it doesn't it is **heavy loam**. If it does with slight cracks in the sausage, it is **light clay**. If it does without any cracks, it is a **clay**.

(iii) **Ball throwing method:**

The texture of the soil can be inferred by the way a ball of soil acts when it is thrown at a hard surface such as a wall or a tree (Fig.5). The steps to be followed in this method as described by Coche and Laughlin (1985) is as follows: Throw a ball of soil to a tree or wall 3 m away. If the soil is good only for splatter shots when either wet or dry, it has a coarse texture (loamy sand). If there is a “shot gun” pattern when dry and it holds its shape against medium range target when wet, it has a moderately coarse texture (sandy loam). If the ball shatters on impact when dry and clings together when moist but does not stick to the target it has a medium texture (loam, sandy clay loam, silty clay loam). If the ball holds its shape for long - range shots when wet and sticks to the target but is fairly easy to remove it has a moderately fine texture (clay loam). If the ball sticks well to the target when wet and becomes a very hard missile when dry, it has a fine texture (clay).

The usual mechanical analysis of soils in the laboratory gives the percentages of the three size fractions, sand, silt and clay. For special uses, the same methods of laboratory analyses (pipette method or hydrometer method) can provide a much more detailed analysis giving further breakdown of the relative amounts of soil particles for more size classes in the form of a table or graph. The data in the graphical form is given as a particle - size frequency curve (PSF curve). PSF curves for selected soils are shown in Fig. 1.9. The vertical axis represents the cumulative percentage of occurrence of the various particles sizes and the horizontal axis represents the logarithms of the particle size. The vertical axis in the left hand side relate to the percentages of particles passing through sieves of a particular size and the vertical axis in the

right hand side relate to the percentages of particles not passing through sieves of a particular size.

The more vertical the PSF curve or part of the curve, the more uniform the particle size; a vertical line represents a perfect uniform particle size. The more inclined the curve or part of it, the greater the difference between the particle sizes (i.e. smaller porosity and higher compaction). The inflexion point of the curve shows the most frequent particle size by weight. Fine textured soils have their curves towards the right hand side of the graph and the coarse textured soils to the left hand side. From the PSF curves, the percentages of silt, sand and clay can be calculated and using the textural triangular diagram the texture could be determined.

Soil texture is an important soil parameter determining the suitability of a site for aquaculture. A clayey soil stabilises pond bottom besides the fact that it absorbs large quantity of nutrients and release them slowly over a long period to the overlying water. The clayey soil normally holds higher amounts of organic matter than light textured soils and thereby increase the productivity of the pond. It should be noted that too clay a soil (very sticky clay) may not be very satisfactory as it may give rise to fixation of phosphorus and create other physico-chemical biological problems. Such soils may give rise to cracks on draining the ponds, thereby increase seepage losses.

3.2 Structure

The term texture is used in reference to the size of individual soil particles but when the arrangement of the particles is considered the term structure is used. Structure refers to the aggregation of primary soil particles (sand, silt and clay) into compound particles or cluster of primary particles which are separated by the adjoining aggregates by surfaces of weakness. Structure modifies the effect of texture in regard to moisture and air relationships, availability of nutrients, action of microorganisms and root growth. E.g. a highly plastic clay (60% clay) is good for crop product if it has a well-developed granular structure which facilitates aeration and water movement. Similarly, a soil though has a heavy texture, can have a strongly developed structure, thus making it not very satisfactory for aquaculture as a result of this soil allowing high seepage losses.

Structure is defined in terms of grade, class and type of aggregates.

Grade:

Grade of structure is the degree of aggregation and expresses the differential between cohesion within aggregates and adhesion between

aggregates. These properties vary with the moisture content of the soil and it should be determined when the moisture content is normal - not when unusually dry or unusually wet. The four major grades of structure rated from 0 to 3 are listed below.

- 0- Structureless: No observable aggregation or no definite orderly arrangement of natural lines of weakness. Massive if coherent; single grain if non-coherent.
- 1 - Weak: That degree of aggregation characterised by poorly formed indistinct aggregates that are barely observable in place. When disturbed, soil material that has this grade of structure breaks into a mixture of few entire aggregates many broken aggregates and much un-aggregated material.
- 2 - Moderate: Well-formed distinct aggregates that are moderately durable and evident but not distinct in undisturbed soil. When disturbed, they break down into a mixture of many distinct entire aggregates, some broken aggregates and little none aggregated material.
- 3 - Strong: Durable aggregates that are quite evident in undisturbed soil that adhere weakly to one another. When removed from the profile the soil material consists very largely of entire aggregates and includes few broken ones and little or no none aggregated material.

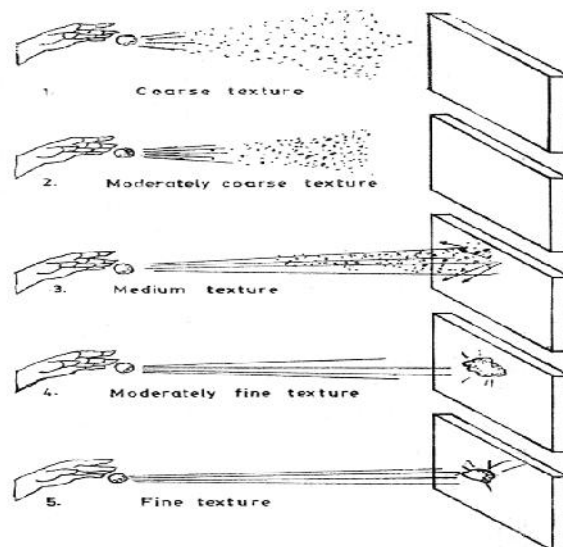
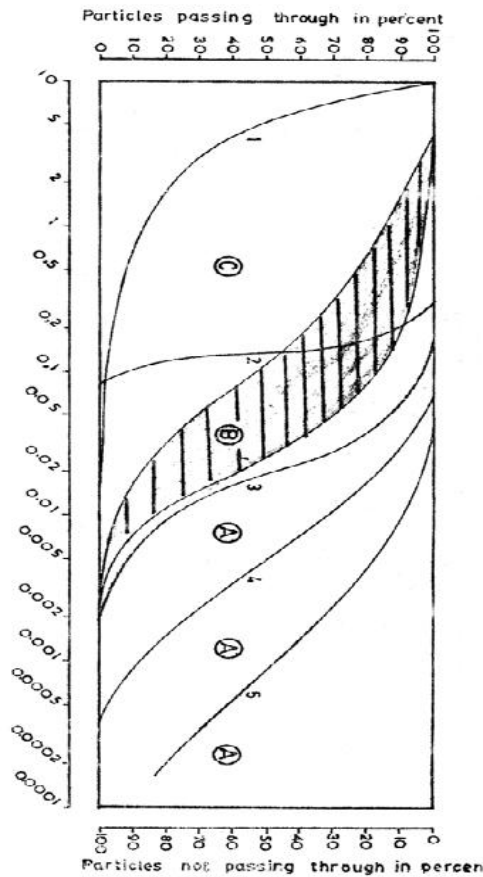


Fig.5: Field Methods of Soil Texture Estimation (Ball Throwing Method)



Diameter of particle (mm)

- | | |
|--|---|
| <p>1 Gravel and sand (old alluvium)</p> <p>2 Sand</p> <p>3 Silt</p> <p>4 Calcareous clayey soil (marl)</p> | <p>A Soil suitable for pond bottom if coefficient of permeability is less than 5×10^{-6} m/s</p> <p>B Soil suitable for building dikes without impermeable clay core</p> <p>C Soil suitable for pond bottom or dike only after modification of soil using amendments.</p> |
|--|---|

Fig. 6: Particle Size Frequency (PSF) Curves for Selected Soils

Class:

The **Class** of structure describes the average size of individual aggregates and **type** describe their form or shape. The various class divisions are: very fine or very thin, fine or thin, medium, coarse or thick and very coarse or very thick.

Water movement and drainage are poor in soils having blocky, prismatic, columnar and platy structures. These structured soils especially the platy type are most suitable for aquaculture.

3.3 Porosity

Soil texture is determined by the relative proportion of sand, silt, clay and small rocks (pebbles) found in a given sample. **Sand** is gritty to the touch and the individual grains or particles can be seen with the naked eye. It is the largest of the three size classes of soil particles. **Silt** is smooth and slippery to the touch when wet and the individual particles are much smaller than those of sand. These individual particles can only be seen with the aid of a microscope. **Clay** is sticky and plastic-like to handle when wet. The individual particles are extremely small and can only be seen with the aid of an electron microscope.

Soils are made of particles of different types and sizes. The space between particles is called **pore space**. Pore space determines the amount of water that a given volume of soil can hold. **Porosity** refers to how many **pores**, or holes, a soil has. The porosity of a soil is expressed as a percentage of the total volume of the soil material. Porosity is an important measurement in areas where drinking water is provided by groundwater reserves.

Porosity or **pore space** refers to the volume of soil voids that can be filled by water and/or air. It is inversely related to bulk density. Porosity is calculated as a percentage of the soil volume:

$$\frac{\text{Bulk density}}{\text{Particle density}} \times 100 = \% \text{ solid space}$$

$$100\% - \% \text{ Solid Space} = \text{Percent Pore Space}$$

Loose, porous soils have lower bulk densities and greater porosities than tightly packed soils. Porosity varies depending on particle size and aggregation. It is greater in clayey and organic soils than in sandy soils. A large number of small particles in a volume of soil produce a large number of soil pores. Fewer large particles can occupy the same volume of soil so there are fewer pores and less porosity.

Compaction decreases porosity as bulk density increases. If compaction increases bulk density from 1.3 to 1.5 g/cm³, porosity decreases from 50 percent to 43 percent. Aggregation also decreases porosity because larger pores are present as compared to single clay and silt particles that are associated with smaller pores. Pores of all sizes and shapes combine

to make up the total porosity of a soil. Porosity, however, does not tell us anything about the size of pores.

3.3 Density

Bulk density is an indicator of soil compaction and soil health. It affects infiltration, rooting depth/restrictions, available water capacity, soil porosity, plant nutrient availability, and soil microorganism activity, which influence key soil processes and productivity. It is the weight of dry soil per unit of volume typically expressed in grams/cm³. Total volume of surface soil is about 50% solids, mostly soil particles (45%), and organic matter (generally < 5%); and about 50% pore space which are filled with air or water. When determining bulk density, the amount of soil moisture must be determined. Available water capacity is the amount of soil moisture available to plants, varies by texture (Figure 2), and is reduced when compaction occurs. Bulk density can be managed, using measures that limit compaction and build soil organic matter.

3.4 Inherent Factors Affecting Bulk Density and Available Water Capacity

Inherent factors that affect bulk density such as soil texture cannot be changed. Bulk density is dependent on soil organic matter, soil texture, the density of soil mineral (sand, silt, and clay) and their packing arrangement. As a rule of thumb, most rocks have a density of 2.65 g/cm³ so ideally, a silt loam soil has 50% pore space and a bulk density of 1.33 g/cm³. Generally, loose, well aggregated, porous soils and those rich in organic matter have lower bulk density. Sandy soils have relatively high bulk density since total pore space in sands is less than silt or clay soils. Bulk density typically increases with soil depth since subsurface layers are more compacted and have less organic matter, less aggregation, and less root penetration compared to surface layers, therefore contain less pore space. Available water capacity is affected by soil texture, presence and abundance of rock fragments, soil depth and restrictive layers.

Bulk density can be changed by management practices that affect soil cover, organic matter, soil structure, compaction, and porosity. Excessive tillage destroys soil organic matter and weakens the natural stability of soil aggregates making them susceptible to erosion caused by water and wind. When eroded soil particles fill pore space, porosity is reduced and bulk density increases. Tillage and equipment travel results in compacted soil layers with increased bulk density, most notably a “plow pan” (Figures 3 and 4). Tillage prior to planting temporarily decreases bulk density on the surface but increases at the depth of tillage. Subsequent trips across the field by farm equipment, rainfall

events, animals, and other disturbance activities will also compact soil. Long-term solutions to soil compaction center on decreasing soil disturbance and increasing soil organic matter. A soil's available water capacity is also affected by organic matter and compaction. Organic matter increases a soil's ability to hold water, both directly and indirectly. Compaction increases bulk density and reduces total pore volume, consequently reducing available water holding capacity.

The following measures increase organic matter, and reduce compaction, which improve bulk density and porosity:

- Practices that increase organic matter such as continuous no-till, cover crops, solid manure or compost application, diverse rotations with high residue crops and perennial legumes or grass used in rotation
- Minimize soil disturbance and avoid operating equipment when soils are wet
- Use designated roads or rows for equipment
- Reduce the number of trips across a field
- Subsoil to disrupt existing compacted layers; and
- Use multi-crop systems involving plants with different rooting depths to help break up compacted soil layers.

Water-filled pore space and porosity

When determining bulk density, water-filled pore space and porosity can also be calculated. When water-filled pore space exceeds 60%, several important soil processes are impacted. Soil respiration and nitrogen cycling (ammonification and nitrification) increase with increasing soil moisture. Lack of aeration also interferes with a soil organism's ability to respire and cycle nitrogen. In dry soils, these processes decline because of lack of soil moisture. As soil water-filled pore space exceeds 80%, soil respiration declines to a minimum level and denitrification occurs resulting in loss of nitrogen as gas, emission of potent greenhouse gases, yield reduction, and/or increased N fertilizer expense.

4.0 CONCLUSION

The percentages of sand silt and clay in a soil could be determined in a soil laboratory by two standard methods. Both methods depend on the fact that at any given depth in a settling suspension the concentration of the particles varies with time, as the coarser fractions settle at a faster rate than the finer.

5.0 SUMMARY

In this unit, we have learnt that:

- In the field, soil texture could be estimated by three different methods namely: Feel method, Ball and ribbon method and ball throwing method.
- Structure is defined in terms of grade, class and type of aggregates.
- Porosity refers to how many pores, or holes, a soil has.
- Bulk density is an indicator of soil compaction and soil health.

6.0 TUTOR-MARKED ASSIGNMENT

1. List and explain all the methods use in a field to estimate soil texture.
2. Briefly define soil porosity.
3. What do you understand by total volume of surface soil?
4. What are the procedure of Ball and Ribbon method as described by Coche and Laughlin?

7.0 REFERENCES/FURTHER READING

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MODULE 3

UNIT 1 SOIL WATER RELATIONSHIP

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Soil Water Relationship
 - 3.2 Water Holding Capacity
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 References/Further Reading

1.0 INTRODUCTION

Water-holding capacity is controlled primarily by soil texture and organic matter. Soils with smaller particles (silt and clay) have a larger surface area than those with larger sand particles, and a large surface area allows a soil to hold more water. In other words, a soil with a high percentage of silt and clay particles, which describes fine soil, has a higher water-holding capacity. Organic matter percentage also influences water-holding capacity. As the percentage increases, the water-holding capacity increases because of the affinity organic matter has for water.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss soil permeability classification system
- explain range of water holding capacity for different soil textures
- examine the ability of a soil to hold water.

3.0 MAIN CONTENT

3.1 Soil Water Relationship

Soil acts as a sponge to take up and retain water. Movement of water into soil is called infiltration and the downward movement of water within the soil is called percolation, permeability or hydraulic conductivity. Pore space in soil is the conduit that allows water to

infiltrate and percolate. It also serves as the storage compartment for water.

Infiltration rates can be near zero for very clayey and compacted soils, or more than 10 inches per hour for sandy and well aggregated soils. Low infiltration rates lead to ponding on nearly level ground and runoff on sloping ground. Organic matter, especially crop residue and decaying roots, promotes aggregation so that larger soil pores develop, allowing water to infiltrate more readily.

Permeability also varies with soil texture and structure. Permeability is generally rated from very rapid to very slow (Table1). This is the mechanism by which water reaches the subsoil and rooting zone of plants. It also refers to the movement of water below the root zone. Water that percolates deep in the soil may reach a perched water table or groundwater aquifer. If the percolating water carries chemicals such as nitrates or pesticides, these water reservoirs may become contaminated.

Table 1: Permeability Classification System

<i>Permeability class</i>	<i>Rate (inches/hour)</i>
Very rapid	Greater than 10
Rapid	5 to 10
Moderately rapid	2.5 to 5
Moderate	0.8 to 2.5
Moderately slow	0.2 to 0.8
Slow	0.05 to 0.2
Very slow	Less than 0.05

Infiltration and permeability describe the manner by which water moves into and through soil. Water held in a soil is described by the term **water content**. Water content can be quantified on both a gravimetric (g water/g soil) and volumetric (ml water/ml soil) basis. The volumetric expression of water content is used most often. Since 1 gram of water is equal to 1 milliliter of water, we can easily determine the weight of water and immediately know its volume. The following discussion will consider water content on a volumetric basis.

Saturation is the soil water content when all pores are filled with water. The water content in the soil at saturation is equal to the percent porosity. **Field capacity** is the soil water content after the soil has been saturated and allowed to drain freely for about 24 to 48 hours. Free drainage occurs because of the force of gravity pulling on the water. When water stops draining, we know that the remaining water is held in the soil with a force greater than that of gravity. **Permanent wilting point** is the soil water content when plants have extracted all the water they can. At the permanent wilting point, a plant will wilt and not

recover. Unavailable water is the soil water content that is strongly attached to soil particles and aggregates, and cannot be extracted by plants. This water is held as films coating soil particles. These terms illustrate soil from its wettest condition to its driest condition.

Several terms are used to describe the water held between these different water contents. **Gravitational water** refers to the amount of water held by the soil between saturation and field capacity. **Water holding capacity** refers to the amount of water held between field capacity and wilting point. **Plant available water** is that portion of the water holding capacity that can be absorbed by a plant. As a general rule, plant available water is considered to be 50 percent of the water holding capacity. The volumetric water content measured is the total amount of water held in a given soil volume at a given time. It includes all water that may be present including gravitational, available and unavailable water.

The relationship between these different physical states of water in soil can be easily illustrated using a sponge. A sponge is just like the soil because it has solid and pore space. Obtain a sponge about 6 x 3 x 1/2 inch in size. Place it under water in a dishpan, and allow it to soak up as much water as possible. At this point, the sponge is at saturation. Now, carefully support the sponge with both hands and lift it out of the water. When the sponge stops draining, it is at field capacity, and the water that has freely drained out is gravitational water. Now, squeeze the sponge until no more water comes out. The sponge is now at permanent wilting point, and the water that was squeezed out of the sponge is the water holding capacity. About half of this water can be considered as plant available water. You may notice that you can still feel water in the sponge. This is the unavailable water.

Water in the form of precipitation or irrigation infiltrates the soil surface. All pores at the soil surface are filled with water before water can begin to move downward. During infiltration, water moves downward from the saturated zone to the unsaturated zone. The interface between these two zones is called the **wetting front**. When precipitation or irrigation cease, gravitational water will continue to percolate until field capacity is reached. Water first percolates through the large pores between soil particles and aggregates and then into the smaller pores.

Available water is held in soil pores by forces that depend on the size of the pore and the surface tension of water. The closer together soil particles or aggregates are, the smaller the pores and the stronger the force holding water in the soil. Because the water in large pores is held with little force, it drains most readily. Likewise, plants absorb soil

water from the larger pores first because it takes less energy to pull water from large pores than from small pores.

Use of soil water estimates on a percentage volume basis does not allow for any practical interpretation. Therefore, water is usually converted from a percentage volume basis to a depth basis of inches of water/foot of soil (Table 2.).

Table 2: Estimated Soil Water for Three Soil Textures

	~Inches of water/foot of soil~		
	Sand	Loam	Silty clay loam
Saturation	5.2	5.8	6.1
Field Capacity	2.1	3.8	4.4
Permanent wilting point	1.1	1.8	2.6
Oven dry	0	0	0
Gravitational	3.1	2	1.7
Water holding capacity	1	2	1.8
Plant available	0.5	1	0.9
Unavailable	1.1	1.8	2.6

The table values are derived from laboratory analysis of soil samples. Some of this information is also published in the Soil Survey. Other techniques have been developed to estimate soil water if laboratory data is not available. Generally, field capacity is considered to be 50 percent of saturation and permanent wilting point is 50 percent of field capacity.

3.2 Water Holding Capacity

Water holding capacity designates the ability of a soil to hold water. It is useful information for irrigation scheduling, crop selection, groundwater contamination considerations, estimating runoff and determining when plants will become stressed. Water holding capacity varies by soil texture (Table 3.).

Table 3: Range of Water Holding Capacity for Different Soil Textures

Textural class	Water holding capacity, inches/foot of soil
Coarse sand	0.25 - 0.75
Fine sand	0.75 - 1.00
Loamy sand	1.10 - 1.20
Sandy loam	1.25 - 1.40
Fine sandy loam	1.50 - 2.00
Silt loam	2.00 - 2.50
Silty clay loam	1.80 - 2.00
Silty clay	1.50- 1.70
Clay	1.20 - 1.50

Medium textured soils (fine sandy loam, silt loam and silty clay loam) have the highest water holding capacity, while coarse soils (sand, loamy sand and sandy loam) have the lowest water holding capacity. Medium textured soils with a blend of silt, clay and sand particles and good aggregation provide a large number of pores that hold water against gravity. Coarse soils are dominated by sand and have very little silt and clay. Because of this, there is little aggregation and few small pores that will hold water against gravity. Fine textured clayey soils have a lot of small pores that hold much water against gravity. Water is held very tightly in the small pores making it difficult for plants to absorb it.

Since soil texture varies by depth, so does water holding capacity. A soil may have a clayey surface with a silty B horizon and a sandy C horizon. To determine water holding capacity for the soil profile, the depth of each horizon is multiplied by the available water for that soil texture, and then the values for the different horizons are added together. These determinations are shown for two soils in Table 4.

Table 4: Calculation of Water Holding Capacity for a Soil Profile

<i>Depth from soil surface</i>	<i>Depth of layer</i>	<i>Soil texture</i>	<i>Water holding capacity</i>	<i>Available water</i>	<i>Available water</i>
<i>Inches</i>	<i>Feet</i>		<i>In/ft</i>	<i>In/layer</i>	<i>In/5 ft</i>
Soil A					
0-6.0	0.5	Loamy fine sand	1.2	0.6	
6.0-24	1.5	Loamy fine sand	1	1.5	
24-60	3	Fine sand	0.7	2.1	
Total					4.2
Soil B					
0-12.0	1	Silty clay	1.5	1.5	
12.0-30	1.5	Silty clay loam	2	3	
30-60	2.5	Loamy sand	1.1	2.7	
Total					7.2

Water relations are greatly affected by cultural practices, but the effect is largely indirect. For instance, tillage breaks down aggregates, decreasing the number of large pores. This would cause a decrease in infiltration rate and percolation, the water content at field capacity would increase, and gravitational water would decrease. If compaction causes an increase in the number of very small pores, unavailable water may increase, and water holding capacity may decrease. As a result, the amount of plant available water would also decrease.

On your own, consider the effect of different crops, crusting and organic matter on water relations and their relationship to other physical properties and processes.

4.0 CONCLUSION

The study of soil water relationship is essentially hinged on the knowledge of movement of water into soil.

5.0 SUMMARY

In this unit we have learnt that:

- i. Soil acts as a sponge to take up and retain water.
- ii. Permeability also varies with soil texture and structure.
- iii. The volumetric water content measured is the total amount of water held in a given soil volume at a given time.
- iv. Water relations are greatly affected by cultural practices

6.0 TUTOR-MARKED ASSIGNMENT

1. Differentiate between these terms considering soil water relationship: Saturation and Field capacity.
2. In a table form, rate permeability in general.
3. In a table form, list all range of water holding capacity for different soil textures.

7.0 REFERENCES/FURTHER READING

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UNIT 2 **EVAPO-TRANPIRATION**

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main content
 - 3.1 Evapo-Transpiration
 - 3.2 Determination of Evapotranspiration
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 References/Further Reading

1.0 INTRODUCTION

In agriculture not all the water applied is used up by the crop and the amount use or the uptake of the irrigated water depends on the evaporation and transpiration of water from the soil surface and leaf surface respectively. This unit you are about to study deals with water use by crop for the physiological processes and water loss through evapo-transpiration.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- define evapotranspiration
- discuss how to determine evapotranspiration
- state the method of determining evapotranspiration.

3.0 MAIN CONTENT

3.1 Evapo-Transpiration

Evaporation can be defined as the process where liquid water is transformed into a gaseous state. Evaporation can only occur when water is available. It also requires that the humidity of the atmosphere be less than the evaporating surface (at 100% relative humidity there is no more evaporation). The evaporation process requires large amounts of energy. For example, the evaporation of one gram of water requires 600 calories of heat energy.

Transpiration is the process of water loss from plants through **stomata**. Stomata are small openings found on the underside of leaves that are

connected to vascular plant tissues. In most plants, transpiration is a passive process largely controlled by the humidity of the atmospheric and the moisture content of the soil. Of the transpired water passing through a plant only 1% is used in the growth process. Transpiration also transports nutrients from the soil into the roots and carries them to the various cells of the plant and is used to keep tissues from becoming overheated. Some dry environment plants do have the ability to open and close their stomata. This adaptation is necessary to limit the loss of water from plant tissues. Without this adaptation these plants would not be able to survive under conditions of severe drought.

It is often difficult to distinguish between evaporation and transpiration. So we use a composite term **evapotranspiration**. The rate of evapotranspiration at any instant from the Earth's surface is controlled by four factors:

- i. Energy availability. The more energy available the greater the rate of evapotranspiration. It takes about 600 calories of heat energy to change 1 gram of liquid water into a gas.
- ii. The humidity gradient away from the surface. The rate and quantity of water vapor entering into the atmosphere both become higher in drier air.
- iii. The wind speed immediately above the surface. Many of us have observed that our gardens need more watering on windy days compared to calm days when temperatures are similar. This fact occurs because wind increases the potential for evapotranspiration. The process of evapotranspiration moves water vapor from ground or water surfaces to an adjacent shallow layer that is only a few centimeters thick. When this layer becomes saturated evapotranspiration stops. However, wind can remove this layer replacing it with drier air which increases the potential for evapotranspiration.
- iv. Water availability. Evapotranspiration cannot occur if water is not available.

On a global scale, most of the evapotranspiration of water on the Earth's surface occurs in the subtropical oceans. In these areas, high quantities of solar radiation provide the energy required to convert liquid water into a gas. Evapotranspiration generally exceeds precipitation on middle and high latitude landmass areas during the summer season. Once again, the greater availability of solar radiation during this time enhances the evapotranspiration process.

Evapotranspiration (ET)

Evaporation is the escape of water in form of vapour from the soil to the atmosphere through wind. Transpiration is the escape of water in form of vapour from the plant stomata in the leaf to the atmosphere through wind. These two processes are joint together and are evapotranspiration which is the escape of water in the form of vapour from the plant and soil surface to the atmosphere through wind. Evapotranspiration is affected by climatic factors such as sunshine, temperature, wind and rainfall. Soil fertility or organic matter, soil Contexture and amount of mulch materials on the soil and factors such as types of crop, growth stage of the crop and size of crop leaf affect evapotranspiration.

3.2 Determination of Evapotranspiration

There are many methods used in determining evapotranspiration but the most commonly used methods are:

1. **Direct methods:** through use of conservation mass principle. The most widely used direct techniques are based on conservation of mass principles.

$$D_s = D_{rz} (f - i)$$

Where D_s = changes in soil moisture within the control volume during the time interval being considered

D_{rz} = Depth of root zone (below soil surface) f and i = soil moisture content by volume at the end (final) and beginning (initial) of the time being considered.

2. Calculate methods of evapotranspiration all method for computing crop evapotranspiration involves the following equation.

$$ET = K_c \times E_{To}$$

Where ET = evapotranspiration (mm/day)

E_{To} = References/Further reading crop evapotranspiration (mm/day)

K_c = crop coefficient or crop factor

Reference crop evapotranspiration may be potential evapotranspiration of crop reference evapotranspiration. Potential evapotranspiration is the maximum rate at which water available can be removed from soil and plant surfaces. Potential evapotranspiration depends on the amount of energy (sunlight) available for evaporation and varies from day to day. Reference crop evapotranspiration is the rate of evapotranspiration from a large area covered by green grass which grow actively, completely shades the ground and which is not short of water. Many methods with

different data requirement and level of sophistication have been developed for computing reference crop evapotranspiration. Some of these methods require daily relative humidity, solar radiation, wind and air temperature data while others need only mean monthly temperature. Measuring the loss of water from an open phase pan of water is a relatively inexpensive and simple way of assessing the evaporating capacities of the atmosphere. The amount of water evaporating from a pan (that is pan evaporation) is determined by measuring the change in water level in the pan and correcting for precipitation has been prevented or negligible.

The principles of the pan ration are as follows:

- i. The pan is installed in the field
- ii. The pan is filled with a known quantity of water (surface area of the pan is known and the water depth is measured).
- iii. The water is allowed to evaporate during certain period of time (usually 24 hours) for example, each morning at 7:00am a measurement is taken. The rain if any is measured simultaneously.
- iv. The amount of evaporation per unit (the differences between the two measured water depths) is calculated; this is then pan evaporation: E_{pan} or EP (in mm / day)
- v. The E_{pan} is multiplied by a pan coefficient K_{pan} or KP , to obtain Reference/Further reading crop evapotranspiration E_{To} daily pan evaporation is routinely measured of weather stations using the formula.

$$E_{To} = K_p \times E_p$$

Where E_{To} = reference crop evaporation

K_p = pan coefficient that account for differences in pan type

E_p = pan evaporation

Example = given the following daily pan evaporation data, calculate reference crop evaporation using 0.8 as pan coefficient.

Date	EP (mm)	E_{To} (mm)
July 1	12.2	12.2
	$12.2 \times 0.8 = 9.76\text{mm/day}$	
July 2	11.2	11.2
	$11.2 \times 0.8 = 8.66\text{mm/day}$	
July 3	8.4	8.4
	$8.4 \times 0.8 = 6.72\text{mm/day}$	
July 4	6.6	6.6
	$6.6 \times 0.8 = 5.28\text{mm/day}$	

July 5	9.7
$9.7 \times 0.8 = 7.76\text{mm/day}$	
July 6	9.7
$9.7 \times 0.8 = 7.76\text{mm/day}$	
July 7	8.1
$8.1 \times 0.8 = 6.48\text{mm/day}$	
July 8	6.6
$6.6 \times 0.8 = 5.28\text{mm/day}$	
July 9	8.6
$8.6 \times 0.8 = 6.88\text{mm/day}$	
July 10	5.8
$5.8 \times 0.8 = 4.64\text{mm/day}$	

$$EP = 88.3 \text{ ETo} = 70.64\text{mm}/10\text{days}$$

$$\begin{aligned} \text{Solution} = \quad \text{ETo} &= K_p \times EP \\ K_p &= 0.8 \\ \text{Daily ETo} &= 0.8 \times \text{daily EP} \end{aligned}$$

$$\begin{aligned} \text{July 1 ETo} &= 0.8 \times 12.2 = 9.76\text{mm/day} \\ \text{Total ETo} &= \text{Addition of the daily ETo} \\ \text{Or EP} \times K_p &= 88.3 \times 0.8 \\ \text{Total ETo} &= 70.64\text{mm}/10\text{days}. \end{aligned}$$

4.0 CONCLUSION

The knowledge of consumptive use, evapotranspiration and reference crop evaporation is important in planning for irrigation. This would guide the farmer to estimate his crop at different stages of their growth and this would enhance water use efficiency.

5.0 SUMMARY

Consumptive use include all water used in physiological processes as well as for direct evaporation from the soil and plant surface. Consumptive use is of three types namely. Daily, seasonal and peak and usually peak consumptive used to calculate irrigation requirement of crop. Evaporation and transpiration are component of consumptive use and because the two cannot be separated, they are jointly called evapotranspiration. The method used in calculating evapotranspiration and reference crop evapotranspiration includes pan evaporation and use of equation.

6.0 TUTOR-MARKED ASSIGNMENT

1. Define evapotranspiration.
2. State the principle used in pan evaporation.

7.0 REFERENCE/FURTHER READING

Larry, G. James. (1993). *Principles of Farm Irrigation*. John Wiley and Sons Inc.

MODULE 4**UNIT 1 TILLAGE AND SOIL PROPERTIES****CONTENTS**

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Tillage
 - 3.2 Advantages and Disadvantages of Tillage
 - 3.3 Types of Soil Tillage
 - 3.4 Effects of Tillage
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 References/Further Reading

1.0 INTRODUCTION

In our previous units, we learnt that sustainable agriculture demands the effective integration of agricultural management technology to produce quality food and fiber while maintaining soil productivity, farm profitability and environmental quality. In this unit, we shall look into the growing interest in developing tillage practices to give greater protection to the soil against soil and water losses. The amount of surface residues and surface roughness, both have an effect. Crop residue management has been developed to leave more of the harvest residues, leaves and roots on or near the surface.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- define soil tillage
- discuss its advantages and disadvantages
- explain and differentiate the types of tillage
- examine the effect of tillage.

3.0 MAIN CONTENT**3.1 Tillage**

What is Soil Tillage?

It is the mechanical manipulation of soil for any purpose, but in agriculture it is usually restricted to the modifying of soil conditions for crop production. The surface forms of soil manipulation performed by the application of mechanical forces to the soil with a tillage tool, such as cutting, shattering, inversion or mixing.

3.2 Advantages and Disadvantages of Tillage

Advantages

- i. Improves infiltration and more efficient use of water.
- ii. Increases hectrage of sloping land that can safely be used for row crops.
- iii. Improves crop yield.
- iv. Reduces soil erosion by water and wind.
- v. Minimises labour, machinery and fuel costs.

Disadvantages

- i. More potential for farm animals (rodents), insects and diseases.
- ii. Skilled manpower is needed.

3.3 Types of Soil Tillage

1. **Clean tillage:** A process of frequent cultivation or plowing to prevent growth of all vegetation, except the particular crop derived during the growing season.
2. **Conservation tillage:** Any tillage sequence which reduces loss of soil or water relative to conventional tillage.
3. **Conventional tillage:** The combined primary and secondary tillage operations normally performed in preparing a seed-bed for a given crop grown in a given geographical area.
4. **Minimum tillage:** The minimum soil manipulation necessary for crop production or meeting tillage requirements under the existing soil and climatic conditions.
5. **Mulch tillage:** Tillage or preparation of the soil in such a way that plant residues or other materials are left to cover the surface, also mulch farming, trash farming, stubble mulch tillage, plow less farming.
6. **Zero tillage:** A procedure whereby a crop is planted directly into a seedbed; not tilled since harvest of the previous crop; also no – tillage system.

3.4 Effects of Tillage

Effects of tillage vary greatly, depending on soil, crop and weather conditions.

- **Surface residues:** The amount of surface residues remaining after one tillage operation varies with the implement. Subsurface implements leave most of the residue on the soil surface to help protect the surface against erosion. Also, the quantity of residue incorporated or left on the surface after tillage also depends on type of crop given.
- **Soil loss:** The amount of residue required to prevent soil erosion depends on the following factors: soil characteristics (texture, organic matter, structure, depth, slope, length), residue characteristics, rainfall characteristics and wind characteristics (velocity, direction to mention but two).
- **Soil temperature:** Soil temperature early in the growing season are generally lower under conservation tillage than under conventional tillage due to the insulating effect of the unincorporated crop residues on the surface.
- **Soil moisture:** Increasing residues on the surface by reducing tillage residues, runoff and soil erosion while increasing infiltration, water conservation increases with maintenance of surface residue cover.
- **Soil microbial activity:** The interaction of soil temperature and moisture with tillage will dramatically influence microbial activity. When the soil is tilled, increased aeration encourages microbial activity and mineralization of organic matter, which essentially releases Nitrogen and other nutrients. Microbial activity is lower early in the season because of lower temperature however, it is slightly higher later in the season because of greater soil moisture.

4.0 CONCLUSION

In this unit, we can safely conclude that tillage have both favourable and unfavourable effects on aggregation. If the soil is not too wet or too dry when the tillage is performed, the short term effect of tillage is generally favourable. Tillage implements break up large clods, incorporate organic matter into the soil, kill weeds and generally create a more suitable seedbed. Soon after plowing, the surface soil is loosened (its cohesive strength is decreased) and total porosity is increased.

5.0 SUMMARY

In this unit, we have learnt that:

- i. Soil tillage is the mechanical manipulation of soil for any purpose but in agriculture it is usually confined to the modifying of soil conditions for crop production.
- ii. Soil tillage improves infiltration, crop yield, minimizes soil erosion.
- iii. Soil tillage requires skilled manpower.
- iv. There are different types of soil tillage namely; clean tillage, conservation tillage, conventional tillage, mulch tillage, minimum tillage and zero tillage.

6.0 TUTOR – MARKED ASSIGNMENT

1. What is soil tillage?
2. List the types of tillage practices you know
3. What is the difference between the following?
 - i. Minimum and zero tillage
 - ii. Conventional and conservation tillage

7.0 REFERENCES / FURTHER READING

Brady, N.C. & Weil, R.R. (2007). *The Nature and Properties of Soils*. (13th ed.). Pearson Education Inc. pp976.

Tisdale, S.L.; Nelson, W.L.; Beaton, J.D. & Havlin, J.L. (2003). *Soil Fertility and Fertilisers* (5th ed.). India, New Delhi: Prentice – Hall pp 634.

UNIT 2 FACTORS AND PROCESSES AFFECTING PLANT GROWTH

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Factors that Affect the Growth of Plants
 - 3.2 Plant Growth Factors
 - 3.3 Hormonal Factors
 - 3.4 Genetic Factors
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 Reference/Further Reading

1.0 INTRODUCTION

Temperature, amount of sunlight and pH of soil are all factors that can affect the growth and development of a plant. Environmental factors can either allow plants to thrive, or leave them unable to grow and die. One important factor is the quality of the soil the plant is growing in. Urban development of land can have a serious impact on the soil, shaping how healthy a plant can be, while, understandably, soil in untouched areas of land would be of far better quality and better suited to grow plants.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss different factors that affect the growth of plant
- explain the role of temperature in the rate of plant growth and development
- examine the effect of some plant hormones on the growth and differentiation of plants
- discuss the importance of water to plant growth.

3.0 MAIN CONTENT

3.1 Factors that Affect the Growth of Plants

The growth of plants involves various changes such as the addition of new cells through cell division, an increase in its size and weight, and an irreversible increase in the volume. Therefore, we can define the term

growth as a permanent and irreversible change in the size of a cell, organ or whole organism usually accompanied by an increase in its dry weight.

3.2 Plant Growth Factors

The plant growth is influenced by a variety of external and internal factors. The former include light, temperature, water, nutrients etc. while the latter includes genetic and hormonal factors

Light

As far as the plants are concerned there are two major effects of light on them, one related with photosynthesis and the other with growth and morphogenesis. The effect of light on the growth, development, and differentiation in plants independent of photosynthesis is called photomorphogenesis.

For example, if a plant is allowed to grow in complete darkness, it shows marked etiolation (i.e. phenomenon exhibited by green plants when grown in darkness). The seedlings grown in dark are pale, elongated with primary leaves unexpanded and yellow. On the other hand, the seedlings grown in light are sturdy, green with shorter internodes, expanded leaves, and upright apex. These effects are in response to light only because nutritional and other environmental conditions were identical in both. Besides this, most of the processes in plants are controlled by light: reproduction, seed germination, the growth of seedling, and differentiation in various tissues and organs.

Temperature

Temperature is a primary factor affecting the rate of plant growth and development. Plant growth is the result of cellular metabolism and processes. Since the metabolic activities of plants are directly affected by the variation in temperature, the growth rate is also influenced. The rate of growth of plant tissues increases with the increase in temperature up to optimum levels, after which it declines as the temperature becomes extreme. The temperatures for optimal growth vary with the type of plant. Some annual flowers and vegetables are extremely sensitive to cold whereas some plants require a cold treatment in order to stimulate flowering in later stages. Germination of seeds in plants is also affected by extreme temperature conditions. Some seeds of winter season crops germinate at 5°C to 25°C while seeds of summer crops germinate at 10°C to 35°C. Thus, in the spring, cool soil temperatures may prohibit seed germination.

Water

We all know the significance of water in plants growth. Water is necessary for virtually every function of plant growth. It is used in photosynthesis and other metabolic processes and then ultimately lost by transpiration. Excessive transpiration or water deficiency in the soil causes water stress in the plants. The plant growth is considerably retarded during water stress, but sometimes excessive water also stops the growth of the plant. Shortage of water damages plant cells, resulting in decreased growth, wilting, and leaf scorch, and eventually leaf drop and root damage. Whereas excess water, reduces the amount of oxygen in the soil resulting in root loss or injury. It can also make the plant more susceptible to many fungal diseases.

Nutrients

Plants require nutrients in the form of inorganic ions for normal growth and development. Almost every metabolic process in plants requires inorganic nutrition. Thus, the growth of plants greatly depends upon the supply of these essential elements. In the absence of these elements, the plant growth gets retarded. They obtain inorganic nutrients from the air, water and soil. For example, carbon and oxygen are obtained from carbon dioxide and hydrogen from water. They also absorb a wide variety of mineral elements from the soil. For example, nitrogen, phosphorus, potassium, calcium, magnesium, iron, zinc, boron etc.

Essential elements perform several functions. They participate in various metabolic processes in the plant cells such as permeability of cell membrane, maintenance of osmotic concentration of cell sap, electron transport systems, buffering action, enzymatic activity and act as constituents of macromolecules and co-enzymes.

3.3 Hormonal Factors

In addition to temperature, water, and inorganic nutrition, the growth and differentiation of plants also depend on some plant hormones. These plant hormones are also known as plant growth regulators. They regulate the distribution and fate of nutrients in different organs of the plant body. Plant growth regulators are broadly divided into two groups based on their functions in a living plant body. One group is involved in growth promoting activities, such as cell division, cell enlargement, pattern formation, tropic growth, flowering, fruiting, and seed formation.

The other group is involved in various growth inhibiting activities such

as dormancy and abscission. They also play an important role in plant responses to wounds and stresses of biotic and abiotic origin. Thus, these hormonal factors play a major role in the growth and differentiation of plants.

3.4 Genetic Factors

The seed of mango germinates to produce a mango plant; it never grows into a palm tree. Similarly, a tomato seed gives rise to a tomato plant. All the information about the form and shape of the plant body and the specific pattern of growth and differentiation in plants are stored as genetic information in the genes located inside the cell of a seed. As the development processes begin the genetic information is passed from genes to RNA to proteins within the cells. Some very specific genes are responsible for the synthesis of specific enzymes (proteins) which then catalyze specific biochemical processes that are necessary for growth and differentiation in plant cells.

Thus, we can conclude that the growth of a plant is dependent on various factors which may be internal or external. This is because these factors are directly or indirectly involved in various morphological, physiological and developmental processes in plants that are necessary for the overall growth and development of plants. So, that is all for now.

4.0 CONCLUSION

The study of factors that affect the growth of plants is essentially hinged on the knowledge of the individual factors in relation to plant growth.

5.0 SUMMARY

In this unit we have learnt that:

- i. There are different factors that affect the growth of plant
- ii. Plants require nutrients in the form of inorganic ions for normal growth and development.
- iii. Temperature is a primary factor affecting the rate of plant growth and development.
- iv. In addition to temperature, water, and inorganic nutrition, the growth and differentiation of plants also depend on some plant hormones.
- v. Water is necessary for virtually every function of plant growth.

6.0 TUTOR-MARKED ASSIGNMENT

Discuss the following three factors as factors that affect the growth of plants:

- i. Temperature
- ii. Nutrient
- iii. Light

7.0 REFERENCE/FURTHER READING

<https://sciencesamhita.com/factors-affecting-plant-growth/>

UNIT 3 SOIL MICRO-ORGANISMS

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Different Types of Soil Organisms
 - 3.2 Important Activities of Soil Organisms
 - 3.3 Decomposition of Organic Matters
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 Reference/Further Reading

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1.0 INTRODUCTION

Microorganisms are very small forms of life that can sometimes live as single cells, although many also form colonies of cells. A microscope is usually needed to see individual cells of these organisms. Many more microorganisms exist in topsoil, where food sources are plentiful, than in subsoil. They are especially abundant in the area immediately next to plant roots (called the rhizosphere), where sloughed-off cells and chemicals released by roots provide ready food sources. These organisms are primary decomposers of organic matter, but they do other things, such as provide nitrogen through fixation to help growing plants, detoxify harmful chemicals (toxins), suppress disease organisms, and produce products that might stimulate plant growth.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss the different types of soil organisms
- explain the important activities of soil organisms
- Examine decomposition of organic matter
- analyse products of decomposition.

3.0 MAIN CONTENT

3.1 Different Types of Soil Organisms

1. **Micro-fauna (protozoa, Nematodes): Macrofauna,** in soil science, animals that are one centimetre or more long but smaller than an earthworm.

Protozoa: Soil protozoa are unicellular organism but larger than bacteria, the size varying from a few microns to a few centimeters. The protozoa exist in all arable soils, but numerically form only a small part of the soil population. A fertile soil may contain 10,000 to more than a million of protozoa per gram of soil. Protozoa tolerate adverse environmental conditions such as moisture, acidity, and climate to a greater extent than other soil microbes. The role of protozoa is not definitely known. As they feed on bacteria and actinomycetes, they probably help to maintain favorable balance of the micro-flora in the soil.

Nematodes: These are abundant soil micro-fauna in soil. They are also called threadworms or round worms. Those that feed on decaying organic matter are called saprophytic, those on earthworms or other nematodes are predatory and those on the roots of higher plants are parasitic. The saprophytic group is abundant in the soils but those belonging to the other groups are agriculturally important in the sense that they cause loss of vigor of the root system and make plant growing in nematodes infested soil liable to diseases.

2. **Macro- flora of the soil
Root of higher plant**

Higher plants are by large the most important group of organism in the whole soil population. These higher plants are the primary producers. The dead plants contribute to the formation of soil organic matter, which in turn, provides food energy and nutrients to microorganisms and higher plants. The roots of higher plant contribute significantly to soil biological activity. Organisms are therefore considered one of the factors of soil formation.

3. **The micro-flora of the soil**

Bacteria: These are the simplest and smallest form of plant. Bacteria are single celled plant, it exceeds 0.005mm in diameter. Under favorable conditions, they multiply by elongating and dividing.

Actinomycetes: They resemble bacteria in that they are unicellular, they are aerobic and sensitive to change in pH. They

are better adapted to drier and hotter conditions than bacteria and fungi but cannot tolerate a pH below 5.0 they help to break down organic matter and so release plant nutrients that it contains. They help in the circulation of soil nitrogen and they are useful in spreading up formation of compost.

Algae: These are simplest form of plant life that contains chlorophyll. They obtain nitrogen and minerals from the soil, carbon from the carbon dioxide of the atmosphere and energy from the sun by photosynthesis. They differ greatly in size and shape but the species commonly found in soil are microscopic in size and are either unicellular or filamentous in structure.

Fungi: These organisms vary in structure and size from the simplest yeast to more complex mushrooms. Fungus contains no chlorophyll and therefore cannot photosynthesize and drive their nutrients from the breakdown of complex organic substances. Fungus can tolerate acidic soil better than bacteria. Some fungi grow in association with the root of higher plants.

3.2 Important Activities of Soil Organisms

The activities of the soil organisms influence the soil as a medium for plant growth in various ways:

Fixation of atmospheric nitrogen

Atmospheric nitrogen changes into nitrogenous compounds in the soil. This process is accomplished by bacteria, which are symbiotic or non-symbiotic. The symbiotic bacteria cause nodules to form on the roots of legumes. The bacteria penetrate the root and grow rapidly within the tissue of the cortex. They derive food from the plant and in turn release nitrogenous compounds into the roots; the plant can utilize these. If the legumes crop is ploughed into the soil, then nitrogenous compounds are released as the nodules decompose. The non-symbiotic nitrogen-fixation bacteria includes azotobacter sp; which are most active in alkaline and mildly acidic soil, and clostridium sp; which are much more acidic-tolerant. These bacteria are capable of fixing nitrogen directly into the soil.

Decay of organic matter

Some soil dwelling organisms also play a part in decomposition of dead organic matter. When large amount of organic matters is added to the soil, bacteria, actinomycetes and fungi are stimulated to divide rapidly and they start to break down the carbon compounds in the organic material. Eventually, humus is formed and this improves the structure and fertility of the soil.

Release of mineral nutrients

Some organisms produce various organic and inorganic acids which are able to act as solvent for most minerals. Algae and lichens have a corroding effect on rocks and minerals.

Development of soil structure

Microorganisms increase the stability of soil aggregates, which enables the structure to resist disruption by raindrops and tillage.

3.3 Decomposition of Organic Matters

Decomposition is the biochemical process through which soil organic matter acted upon by soil microbes to enhance the conversion of the organic nutrients they contain into inorganic nutrients. The decomposition of plant and animal residues and the recycling of their nutrients are probably the most important functions of soil microorganisms. Microbial activity hydrolysis organic compound to produce sulphate, ammonium, phosphate, nitrates, Ca, Mg, and K ions, as well as trace elements of carbon dioxide and water. Where no readily decomposable organic material is available in the soil, microbial activity would be minimal.

However, when fresh organic residue is added to a soil, there is a boost in microbial activity. The microorganisms multiply very rapidly and soon the decomposition of organic residues reaches a peak and considerable heat and carbon dioxide are given out. The active general-purpose (heterotrophic) organisms (actinomycetes, bacteria and fungi) are decomposing the organic tissue and at the same time synthesizing material for their body cells. The organic matter at this stage contains various forms of intermediate products of decomposition as well as living and dead microbial cell, as the readily available energy source (the organic residue) is used up and food supplies diminish, microbial activity slows down and the general purpose organisms reduce to the number which was present just before fresh residues was added. The decomposition product left behind is dark or brown heterogeneous, amorphous and colloidal mass called humus. This is composed essentially of two groups of materials:

- i. Decay- resistant compounds such as Lignin, Lignin complexes, Oil, Fats
- ii. Waxes and synthesised compounds like the new protein-n Deleted: z of microbial cell, polysaccharides, and polyuronides.

Rate of decomposition

Organic compounds vary greatly in their rate of decomposition. They may be listed in terms of ease decomposition as follows:

1.	Sugar, starches, and simple proteins	Rapidly decomposed
2.	Crude proteins	
3.	Hemi cellulose	
4.	Cellulose	
5.	Lignin, fats, waxes, etc.	Very slowly decomposed

These compounds usually begin to decompose simultaneously when fresh plant tissue is added to the soil. The rate at which decomposition occur decreases as we move from the top to the bottom of the list. Thus, sugars and water- soluble proteins are example of readily available energy source for soil organisms.

Simple decomposition products

As the enzymatic changes of the soil organic matter proceed, simple products begin to manifest themselves. Some of these, especially carbon dioxide and water, appear immediately. Others, such as nitrate nitrogen, accumulate only after the peak of the vigorous decomposition has passed and the general-purpose decay organisms have diminished in numbers. The more common simple products resulting from the activity of the soil microorganisms may be listed as:

Carbon	CO ₂	Elemental Carbon
	CO ₃ ²⁻	
	HCO ₃ ⁻	
	CH ₄	
Nitrogen	NH ₄ ⁺	Gaseous nitrogen
	NO ₂ ⁻	
	NO ₃ ⁻	
Sulfur	S	
	HS	
	SO ₃ ⁻	
	SO ₄ ²⁻	
Phosphorus	H ₂ PO ₄ ⁻ , HPO ₄ ²⁻	
Others	H ₂ O, O ₂ , H ₂ , H ⁺ , OH ⁻ , K ⁺ , Ca ²⁺ , Mg ²⁺ , est.	

Product of decomposition

In well-aerated soil, the end products of decomposition are: CO_2 , NH_4 , NO_3 , H_2PO_4 , SO_4 , HO_2 , resistant residues and numerous other essential plant nutrient elements, in smaller quantities. If soil is not well aerated less desirable product results, for example, in anaerobic conditions significant amount of methane (CH_4), also called “swamp gas” are produced, as well as some organic acids (R-COOH), ammonium (NH_4), various amine residues (R-NH_2), the toxic gases hydrogen sulphide (H_2S) and ethylene ($\text{H}_2\text{C}=\text{CH}_2$), plus the resistant humus residue.

4.0 CONCLUSION

The study of soil microorganisms is essentially hinged on the knowledge of the different types of soil microorganisms present in the soil and their role in the development of soil and plant growth.

5.0 SUMMARY

In this unit we have learnt that:

- i. There are different types of soil organisms.
- ii. There are important activities of soil organisms.
- iii. Decomposition is the biochemical process through which soil organic matter acted upon by soil microbes to enhance the conversion of the organic nutrients they contain into inorganic nutrients.
- iv. Organic compounds vary greatly in their rate of decomposition.
- v. In well-aerated soil, the end products of decomposition are: CO_2 , NH_4 , NO_3 , H_2PO_4 , SO_4 , HO_2 , resistant residues and numerous other essential plant nutrient elements, in smaller quantities.

6.0 TUTOR-MARKED ASSIGNMENT

1. List all the different types of soil organisms.
2. list and explain any three of the important activities of soil organisms.
3. In a table form list all the simple decomposition products.

7.0 REFERENCE/FURTHER READING

<https://www.sare.org/Learning-Center/Books/Building-Soils-for-Better-Crops-3rd-Edition/Text-Version/The-Living-Soil/Soil-Microorganisms>

UNIT 4 SHIFTING CULTIVATION

CONTENTS

- 1.0 Introduction
- 2.0 Objectives
- 3.0 Main Content
 - 3.1 Shifting Cultivation
 - 3.2 Advantages of Shifting Cultivation
 - 3.3 Disadvantages of Shifting Agriculture
 - 3.4 Fallow System
- 4.0 Conclusion
- 5.0 Summary
- 6.0 Tutor- Marked Assignment
- 7.0 Reference/Further Reading

1.0 INTRODUCTION

Shifting cultivation is the most ancient system of agriculture in which soil fertility is restored by long periods of fallowing rather than by off-farm inputs of fertilizer and amendments, nutrients are recycled between natural vegetation and crops, and ecological balance is maintained by adopting diverse and complex cropping systems rather than monoculture.

Fallow system is a system of subsistence agriculture in which land is cultivated for a period of time and then left uncultivated for several years so that its fertility can be restored.

2.0 OBJECTIVES

By the end of this unit, you will be able to:

- discuss what is shifting cultivation
- list both advantages and the disadvantages of shifting cultivation
- explain the meaning of fallow land
- examine the features, advantages and disadvantages of bush fallowing.

3.0 MAIN CONTENT

3.1 Shifting Cultivation

Shifting cultivation is the most ancient system of agriculture in which soil fertility is restored by long periods of fallowing rather than by off-farm inputs of fertiliser and amendments, nutrients are recycled between

natural vegetation and crops, and ecological balance is maintained by adopting diverse and complex cropping systems rather than monoculture.

The process

First the farmer clears the land which he has to cultivate. He then removes all the plants and vegetation from the land. He burns the vegetation in a controlled way. Whenever the rain comes, it washes the ashes of plants into the soil and in this way the soil fertility enriches as potash is added to the soil. This cleared land which is now perfect for cultivation and for the next two to three years fit for the crop-production. Holes are dug in the soil in which seeds are sowed.

Females and children work together to raise the crops for their sustenance. After three years when the fertility of the soil starts decreasing and does not support the soil to nourish the crops. They abandon the lands and start cultivating the other land. They leave that old land to allow it to regain its fertility naturally so that they can cultivate it after fifteen to twenty years. Previously lands were left for 20 years for replenishment but now-a-days due to high population it is reduced to much lower 4-6 years.

Crops

Mostly cultivators who are practicing this method have used it for the purpose so that they can sustain their lives and eat two square meals. They produce large varieties of crops from the same field for themselves only. Crops like food grains, vegetables, paddy, beans, millets etc. are generally grown.

3.2 Advantages of Shifting Cultivation

- i. It is very useful for the people living in hilly areas. It is the easiest way to cultivate their crops. Small bushes and weeds can be easily removed with small manual instrument.
- ii. Within short period of time crops can be easily produced and harvested.
- iii. No danger of flood or drought as stream water in hills can easily irrigate this land regularly.
- iv. It helps the used land to get back all lost nutrients naturally without any help from the modern methods of replenishing the soil.
- v. It saves a wide range of resources as only a small plot is used for such cultivation.
- vi. This method is environmental friendly as it is organic.
- vii. It reduces the environmental degradation.

- viii. Soil borne diseases also decreases by using this method.
- ix. It reduces the use of pest control medicine.

3.3 Disadvantages of Shifting Cultivation

- i. Destruction of forest is the biggest disadvantage of this type of cultivation. Forest takes years to build but this method destroys them in a day for their own motives of livelihood.
- ii. Destruction of forest causes heavy soil erosion which in turn causes flood in rivers and low-lying areas.
- iii. Due to heavy population, the land provided for shifting agriculture is declining. The burden on existing land available for such kind of cultivation increases which results in the loss of more nutrients from the soil without replenishing it.
- iv. Large scale of deforestation increases global warming also.
- v. It is uneconomical.
- vi. It leads to the loss of bio-diversity.
- vii. This method is responsible for reducing the soil-fertility of crops as the land is abandoned when the soil is exhausted.
- viii. Shifting cultivation causes a high national waste as it converts the green land into a barren land. The land takes many years to replenish just at the cost of providing yield for 2 to 3 years.
- ix. It upsets the ecological balance as it disturbs many eco-systems of that region due to destruction of natural vegetation.
- x. One more disadvantage is that the cultivators do not have any private ownership of land. Therefore, they do not take any initiative for soil-conservation and improvement.

Solution

- i. Farmers should be given education so that they can understand the harmful effects of shifting cultivation.
- ii. Afforestation should be stated again to replenish the large abandoned are of land.
- iii. Close monitoring of the land should be done to keep a check on the land under such cultivation.
- iv. The practice of Agro forestry should be encouraged in which both the crops and the tall trees are cared by farmers simultaneously.

3.4 Fallow System

Fallow system is a system of subsistence agriculture in which land is cultivated for a period of time and then left uncultivated for several years so that its fertility can be restored.

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Meaning of bush fallowing

Bush fallowing is a system of subsistence agriculture in which land is cultivated for a period of time and then left uncultivated for several years so that its fertility can be restored.

Meaning of fallow land

Fallow land is a land that is left dormant between planting seasons in order to allow the soil to regenerate its mineral base. In general terms fallow means a field where nothing is growing.

Features of bush fallowing

- i. It is mainly practised by peasant farmers.
- ii. It involves the use of crude implements like cutlass and hoe
- iii. The family is the main sources of labour
- iv. It is common in rural areas with abundant farmlands
- v. Productivity per unit of land or per unit of labour is low
- vi. Food crops like yam, maize, cassava etc. are grown
- vii. Farmlands are left to fallow after one or two years of cultivation
- viii. It uses slash or burn method for land preparation
- ix. It is mainly for family consumption or subsistence
- x. It is practiced where population is very low
- xi. Pests and disease are not controlled
- xii. The "slash and burn" method of clearing the land is practised.

Advantages of bush fallowing

The main advantage of Bush Fallowing is the fact that it is low cost since the equipment used are simple. Other advantages include:

- i. It ensures continuous sustenance during periods of crop failure
- ii. It can be used to check soil erosion, leaching and weed growth
- iii. It is very easy to practice as low technology as required
- iv. It aids the natural restoration of soil nutrients during the fallow period
- v. It helps to control plant and diseases and pests

Disadvantages of bush fallowing

1. There is no provision for expansion in bush fallowing, as it does not need mechanisation.
2. It leads to wastage of land.
3. It leads to the destruction of valuable forest resources like timber.
4. It leads to land fragmentation due to increase in population.

Factors militating against bush fallowing in Nigeria

- i. The population is fast increasing, hence a reduction in the size of farmlands.
- ii. The land is also becoming scarce due to increase in population.
- iii. The practiced of mechanised agriculture now discourages bush fallowing.
- iv. There is no increase in the use of technology.
- v. There is no intensive use of fertilisers and manures.

4.0 CONCLUSION

The study of shifting cultivation and bush fallowing is essentially hinged on the knowledge of what is actually refer to shifting cultivation, bush fallowing and both their advantages and disadvantages respectively.

5.0 SUMMARY

In this unit we have learn that:

- a. Shifting cultivation is the most ancient system of agriculture in which soil fertility is restored by long periods of fallowing rather than by off-farm inputs of fertiliser and amendments, nutrients are recycled between natural vegetation and crops, and ecological balance is maintained by adopting diverse and complex cropping systems rather than monoculture.
- b. In the process of shifting cultivation, first the farmer clears the land which he has to cultivate. He then removes all the plants and vegetation from the land. He burns the vegetation in a controlled way. Whenever the rain comes, it washes the ashes of plants into the soil and in this way the soil fertility enriches as potash is added to the soil.
- c. Bush fallowing is a system of subsistence agriculture in which land is cultivated for a period of time and then left uncultivated for several years so that its fertility can be restored.
- d. The main advantage of bush fallowing is the fact that it is low cost since the equipment used are simple.

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6.0 TUTOR-MARKED ASSIGNMENT

Differentiate between shifting cultivation and bush fallowing.

7.0 REFERENCES/FURTHER READING

<https://www.studytoday.net/shifting-cultivation/>

<http://www.schoolmattazz.com/2016/10/bush-fallowing/>

Henry, D. F. & Boyd, G. Ellis (1988). *Soil Fertility*.