PRINCIPLES OF SOIL SCIENCE

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With Special Reference to Conditions
in the Indo-Pakistan Subcontinent

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To my former students at the
Punjab Agricultural College,
Lyallpur, West Pakistan, this
book is affectionately dedicated.

ERRATA

p. 76, line 4—Read Percolation for Perculation
p. 168, line 1—Read Al ions for Alions

PREFACE

To THE author's knowledge, a textbook on soil science tailored to fit the needs of agricultural students in the Indo-Pakistan subcontinent has not been available previously. Since so many people in the subcontinent are associated closely with agriculture, there can be little question as to the importance of this subject and the need for its being understood widely. It is hoped that this text will contribute to this end.

For the most part the material presented herein is of a fundamental nature and generally of universal application. Where possible, factual information of local origin has been used to illustrate and amplify basic points discussed, however. It is believed that this practical approach will make this book appreciably more understandable and valuable to the student.

The degree to which various phases of the text material have been emphasized derives principally from the author's personal evaluation of needs in the Asian subcontinent. It is probable that this evaluation is somewhat biased due to inadequate familiarity with agricultural and educational facilities throughout the subcontinent. The organization of the text material represents an effort to treat the various individual soil attributes separately and in an undetached manner. However, this approach can never be more than only partially successful, for most soil properties are highly interrelated. This fact becomes most obvious in the last two chapters on soil development and classification, where general characteristics of soils are shown to reflect the combined influence of various individual and identifiable properties.

A majority of the textbooks on soil science contain sections devoted specifically to the subject matter areas of soil conservation and soil management. Such treatments are largely of a practical nature in that they generally discuss the application of basic soil principles to the solution of problems in conservation and management. No such sections appear in this text.

Instead, reference to these fields is made frequently in connection with other related subjects. This is not done as an attempt to de-emphasize these two very important aspects of soil science, but to avoid repetition and to allow a more comprehensive treatment of fundamental concepts. Because of their practical nature, discussions of conservation and management could be treated most logically by one with broad experience in agriculture as it is practised in the subcontinent.

As is essential in the study of any science, the memorization of some terminology and factual information will be found necessary. The knowledge gained by this procedure will be useless, however, unless the significance and application of the facts are understood. This can be accomplished neither in a short span of time nor merely by committing subject matter to memory. With this thought in mind, discussions provided in the text have been developed in a manner considered adequate for a reasonable level of understanding without extensive supplemental tutoring. However, complete comprehension will still depend on the willingness of the instructor to assist and encourage the student in a sincere effort to learn. But the student's background is equally important, and an understanding of the material presented will not be possible unless he is already familiar with the more important principles of inorganic chemistry and with simple algebraic manipulations.

References to the literature occur periodically throughout the text, but they have been kept to a minimum intentionally. Those included have been selected to serve as a guide to students interested in pursuing further some of the points covered in the text in a somewhat cursory manner. The individual literature citations are by the author and the date of publication of the work to which reference is made. A separate listing of the specific papers will be found starting with page 412. Supplemental listings of references occur also at the end of the various chapters. These are intended solely to provide additional reading matter where a more thorough understanding of the subject material is desired.

Development and publication of this volume has spanned a six-year period. This is sufficient time in our rapidly changing world for many new discoveries to have been made and various concepts changed. Indeed, the author is keenly aware that some information imparted in the following pages is now no longer in vogue. The well-informed instructor will recognize such instances and address his classes accordingly.

Perhaps more time than is justifiable has been spent in an attempt to make the text material readable and editorially correct. Errors undoubtedly remain, however. Their presence can be largely attributed to impatience and weariness on the part of the author who accepts full responsibility for them.

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Pullman, Washington,
May, 1963.

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R. L. H.

LIST OF PLATES

CHAPTER ONE

THE GENERAL NATURE OF SOIL

IN A WORLD of rapidly increasing population, the importance of maintaining a supply of food equal to or in excess of the demand places greater and greater emphasis on the need for careful exploitation of soil resources. Modern civilization is awakening to the fact that the ancient methods of farming are no longer adequate for supplying even the minimum requirements of foodstuffs. Combating the problem requires efficient use of the land not only to acquire abundant yields through proper cultural management but, also, to preserve the soil against deterioration in place or loss by erosion. To accomplish this requires a thorough understanding of the properties of soil and their relation to its utilization and preservation. Thus a great responsibility is placed in the hands of the soil scientist. He must work continually not only to learn how to utilize the soil better but to conceive means of passing his knowledge on to the cultivator. No amount of scientific information will increase the productive capacity of soils if the farmer is not taught how to benefit from it.

The student of soil science needs first to realize that soil is a relatively complex substance displaying many interrelated properties. No single term or experimental value will describe it. A soil is considered to be good when each of its various components measures up to an ideal or optimum standard. Since the soil is complex, its complete characterization is difficult. Yet, if we are to understand the functioning of a soil as a medium for plant growth, we must endeavour to relate all of its many properties to its capacity for producing crops. As we learn of those conditions which favour abundant production, efforts then may be turned to developing them in less productive soils. This may mean merely the application of fertilizer to infertile land or the drainage of water from waterlogged areas, or improvement may come from several corrective procedures applied at the same time.

The term 'soil' has various concepts. A soil may be described in numerous ways—by its appearance, feel, position on the landscape, or by its physical and chemical properties as determined in the laboratory. To the cultivator, the soil provides mechanical support for plants, nutrients for their growth, and serves as a reservoir for the storage of water which must be supplied continuously in relative abundance throughout the life of the plant. To him a good soil is one that is productive and, perhaps, easily tilled. While the productivity of a soil is of utmost importance to the soil scientist also, his primary interest is in determining ways of measuring precisely and individually the numerous soil characteristics which, when operating together, contribute to this productivity. The appraisal needed for establishing the relative influence of the various characteristics and the interrelationships among them requires examination of the soil through chemical, physical, and biological means.

SOIL—A PRODUCT OF ENVIRONMENT

Soil usually is depicted as a layer of varying thickness covering a large part of the earth's surface and occurring in a generally unconsolidated or easily disrupted form. It is highly variable with respect to both its horizontal and vertical distribution. That is to say, soils may be noted to change in outward appearance from area to area and, at any one location, also will vary noticeably with depth.

Soils also change gradually with time; so gradually, in fact, that the differences affected may not be recognized during a man's lifetime. However, through the ages, the characteristics of the original, raw soil material are altered as the result of climatic, biological, chemical, and physical processes. The new characteristics which result will depend upon the nature of the imposed processes and their duration and intensity. The rate of change decreases with the passage of time and eventually becomes essentially static. As this point of apparent equilibrium is approached, we may consider that a soil in its true sense has evolved. Its characteristics are no longer those of the original material but now relate closely to the environmental conditions. The characteristics are rather specific for a given environment and provide for the soil an identity which serves as a basis for its systematic classification.

Normally in the study of cultivated soils, we do not observe conditions which reflect the state of near equilibrium referred to above. Ploughing the land alters the surface layer markedly. New types of plants are introduced and are harvested rather than being returned to the soil. Erosion may be accelerated and chemical fertilizers and manures added frequently. Owing to loosening of the surface soil through cultivation, the interchange of air between the soil and outer atmosphere becomes more rapid. The result of these and other alterations in soil conditions from the natural state is usually an abrupt and rapid change in soil characteristics. It may be stated generally that, because of man's apparent unwillingness to follow proper practices in soil management, changes which follow the introduction of cultivation are seldom of a beneficial nature.

The most obvious component of soils is the pulverized, mineral portion which generally constitutes the bulk of the soil body. With few exceptions, this material is derived as a consequence of the physical and chemical weathering of rocks and minerals. The finely divided weathering products may remain in place to accumulate as surface deposits of varying thickness, or they may be transported and re-deposited elsewhere. When these materials are subject to movement by wind and water, the identity of their origin may be lost. For those that remain in place, the extent to which weathering has progressed often is recognized by the degree of conversion of the original rock to finely divided soil material, especially clay.

In addition to the comminution of rock masses, weathering releases gradually the chemical components of the original minerals and, thus, tends to maintain in available form those elements necessary for plant nutrition and growth.[1] It may be seen, therefore, that an important part of weathering rests with its being a continuing process. Unfortunately, often in regions of high rainfall and warm temperatures, it proceeds at an excessive rate. This results in the rapid and almost complete removal of essential plant nutrients through leaching and produces soils of a very low nutrient-supplying power.

The accumulation of weathered mineral matter on the land surface merely initiates the formation of soil. At this stage of development, the soil is considered to be young or *immature*. Through the gradual development of a cover of plants and the accumulation of their decay products along with those of other soil-inhabiting organisms, the organic matter content of the soil will increase. The colour of the soil, especially at the surface, will darken to the extent allowed by the abundance of plant growth and other factors controlling the accumulation of organic matter in the soil. Not until the weathering processes are accompanied by noticeable biological changes, such as the accumulation of organic matter, does the weathered material assume characteristics of a *mature* soil. The mature soil is one that is at apparent equilibrium with its environment.

The changes that have taken place in a deposit of soil material may be assessed visually by observing a vertical exposure or *profile* of the soil. These observations may be made in specially dug pits, or along road cuts or other openings which bring to view both the surface and subsoil. The experienced technician is capable of evaluating many inherent properties of a soil through observations of this type. The diverse nature

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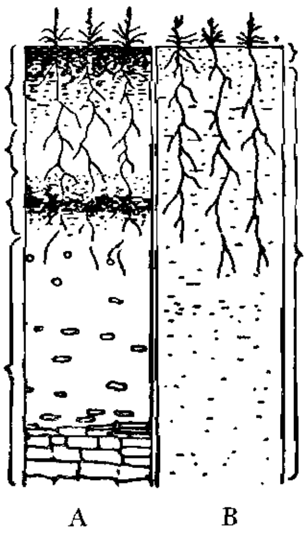
[1] There are 16 chemical elements known to be necessary for the growth and maturation of plants. Twelve of these *essential plant nutrients* are derived from weathering soil minerals. All are discussed in detail in later chapters.

of soils as products of widely differing developmental conditions is exemplified by the soil profiles diagrammed in Fig. 1-1.



Zone of accumulation of organic matter.

Zone of leaching.

Zone of accumulation of materials leached from surface soil.

Highly decomposed rock extending down to slightly weathered to unaltered parent rock.

Recent deposit.

Deep, weakly developed profile resulting from the continual burying of older soil material by fresh deposits. Soil properties are relatively uniform with depth although there may be layers of different textures due to variation in soil materials.

FIG. 1-1. Example soil profiles representing two widely diverse conditions of soil development. Profile A, developing from soil material produced by the weathering of parent rock in place, shows the effect of intense disintegration and decomposition processes operating continuously. The effect of percolating water, which dissolves materials and carries them downward in the profile, may be noted particularly. Profile B represents a soil developing from pulverized materials carried in and deposited by the action of wind or water. Weathering and soil-development processes operate continuously but their effects are covered up by the deposition of fresh materials on the surface. A distinct profile does not develop. The characteristics of the soil correspond more nearly to those of the original soil material. Properties of this soil show little relationship to the environment.

THE SOIL AS A THREE-PHASE SYSTEM

Many characteristics of soils important to their productive capacity may be judged simply on the basis of their general appearance and the way they feel when worked in the hand. Frequently, however, many soils have essentially identical appearance and feel and, yet, are markedly different with respect to their crop-producing powers. Thus, an examination of less obvious physical, chemical, and biological characteristics becomes necessary if the significant differences among them are to be distinguished. Ultimately, differences may be found to relate to the relative levels of chemical plant nutrients present, or possibly they may be due to the occurrence of some toxic substance in one soil but not in others. Many such factors may be involved. Generally, the evaluation requires the measurement of characteristics which permit comparison with other soils having known properties and behaviour. This approach is one in which the soil is considered as a *material* distinct from though related to the environment in which it has developed.

The properties of soil as a material are measured by the same techniques used for examining other materials having one or more corresponding properties. Thus, if we wish, we may compare the water-holding capacity of a soil with that of a sponge. If we were familiar with this property in sponges we would have a basis for evaluating the absorbing capacity of the soil. It would be more logical, however, to compare the absorption of two different soil materials and attempt to relate these absorption capacities to other factors; the rate at which these soils will supply water to growing plants, for instance.

Classified in the simplest of physical terms, the soil is recognized as a system containing three phases—solid, liquid, and gas. The first, or solid phase, consists of varying proportions of mineral and organic matter. The soil water comprises the liquid phase and contains varying concentrations of dissolved salts, gases, and organic compounds. The gaseous phase, or soil air, is a constantly changing mixture consisting primarily of nitrogen, carbon-dioxide, and oxygen. Whereas the solid portion of soil undergoes little apparent change with respect to time, the soil air and water are continually changing in both composition and amount. Their absolute quantities depend upon the total porosity of the soil. An increase in the amount of either one of these two phases is accompanied by a decrease in the other.

Our study in this text will be found to consist of a detailed consideration of the three phases comprising the soil. Each contributes its own part to the development of growing plants. Yet, they also function inter-dependently and so much so, in fact, that variation in one alters the influence the other phases have on plant growth.

The Solid Phase

The mineral and organic components of the solid phase of soils occur intimately mixed and in varying proportions to each other. The mineral portion persists in a wide range of particle sizes grading from coarse *gravels* and *sands*, through *silts* of intermediate size, and, finally, to *clays* which may extend in size into the colloidal realm. The relative quantities of sand, silt, and clay in a soil material determines its *texture*. Knowledge of the texture is important since it provides a good indication of the total pore space present and the relative proportion of large and small pores in a soil. This, in turn, can be related to the water-holding capacity and to the ease of movement of water and air through the soil.

Individual soil particles, especially those of small size and large total surface area, are strongly attracted to each other so that they form clumps or aggregates of various types referred to collectively as soil *structure*. The formation of these structures in a soil having a predominance of small pore spaces will result in the establishment of larger pore spaces which have a marked influence on water and air conduction. Good structure is considered essential in a soil material having a clayey textural configuration which tends to induce naturally a condition of low permeability to water and air.

A unique feature of the mineral particles is that the coarser fractions are comprised of fragments of original rocks or minerals. They acquired their size as a result of the physical breakdown of the original materials through weathering. Their properties correspond closely to the original minerals from which they form. Clays, on the other hand, usually are derived from chemical as well as physical weathering. The identity of the original materials has been lost through mineral degradation. Most particles of clay size result either from simple change in the chemical and physical structure of the original minerals or from the synthesis of new structures through the recombination of decomposition products of degradative weathering.

Although it is not always true, the coarser soil particles tend to remain unchanged because they are composed of minerals, usually quartz, which resist chemical alteration. Clays are relatively stable end products of chemical reactions typical of a particular weathering environment and also resist further change. We may conclude, therefore, that the bulk of the mineral matter in mature soils is static with respect to its basic properties. Any change that does occur will be at an imperceptible rate.

Organic matter, the second component of the solid phase of soils, has its primary source in the remains of a myriad of plants and animals which normally inhabit the soil. The residues, once incorporated into the soil, undergo extensive modification as they are attacked by living soil organisms, largely microscopic, seeking carbonaceous compounds as a source of energy. Some of the organic constituents are resistant to microbial decay, however. These substances tend to accumulate along with relatively stable, synthesized end products of the decay process to form an organic complex of variable composition known as *humus*. This complex is black in colour and occurs as a finely divided, colloidal substance. It darkens the soil but not necessarily in proportion to the quantity present. Therefore, the colour provides only a partially reliable basis for the estimation of the organic matter content of a soil. Being colloidal, humus presents a large total surface area and, consequently, has many of the properties of fine clay particles.

Soil organic matter, when considered *in toto*, is comprised of fresh and partially decayed plant and animal residues as well as accumulated humus. As such, it plays a vital role in soils. Various organic decomposition products act as binding agents and serve to hold soil particles together in the development of soil structure. In this way, organic matter affects the physical state of the soil. In addition, organic matter contributes to the fertility status of soils. Nutrients absorbed and retained in organic plant tissue are not subject to leaching loss. However, as the residues are decomposed, these nutrients are again released slowly and become available to growing plants. Moreover, nutrients absorbed from the subsoil by roots are carried to the surface where they are released during the decay process. As a consequence, the level of available nutrients in the surface soil may be increased materially over that in similar soil material where no plants are growing. This fact provides the basis for the frequent use of green manure crops which are grown and ploughed back solely for the improvement of physical conditions and the fertility status of the soil.

One very important characteristic of inorganic and organic soil solids is their ability to adsorb both water and ions to their surfaces. Since the adsorption is a surface phenomenon, it will be found that this property is most noticeable in the finer soil fractions which present the greatest total surface area per unit of mass.

Adsorption of cations by soils, and particularly by the clay they contain, is attributed to the accidental exclusion of cations from their normal positions within the molecular structure of the minerals. The omission results when these structures are formed. Because of the shortage of positively charged cations, the resulting soil particles normally have a residual negative charge. Electrical neutrality must persist in chemical systems, however. Therefore, a deficiency in structural cations is compensated for by the adsorption of other cations to the particle surfaces.

For the most part, adsorbed cations are held loosely and may be replaced by others occurring in the soil solution which bathes the particle surfaces. Thus, they are referred to as *exchangeable cations*, which distinguishes them from those bound securely within the structural units or those dissolved in the soil solution. Soil organic matter also has the capacity for retaining cations in exchangeable form, although the mechanism is through weak chemical bonding rather than the result of an electrical imbalance.

Several of the cations retained on soil particle surfaces are nutrients essential to plant growth. Thus, exchange reactions assume considerable importance in plant-nutrition considerations. Indeed, the ability of many soils to supply nutrients depends upon their total capacity to retain exchangeable cations. This capacity, in turn, is related closely to the quantities of clay and organic matter the soil contains.

The Liquid Phase

Water comprises the liquid phase of soils. The maximum amount of water which may be retained by a soil depends upon the total pore space present. The ability of the soil to transmit water, on the other hand, is determined largely by the relative proportions of the different sizes of pores occurring in the soil.

Water performs many functions in the soil. As a solvent, it carries essential plant nutrients to absorbing roots and brings about the solution of other ions or compounds from mineral sources. It assists in the renewal of the soil atmosphere by displacing air of low oxygen content from soil pores as wetting takes place. In reality, these are secondary effects, however, since the primary function of water in plant growth is as a constituent of plant tissue. Nonetheless, water must take part in all of its varied activities if normal plant growth is to be expected.

The water-absorbing and retaining characteristics of a soil are very important to plant growth. Water lost in runoff from the land surface is not only useless to plants but can cause severe damage through erosional processes. Proper management of surface soils can increase their ability to absorb rainfall and thereby minimize this defect. Cultural practices can also be applied which will improve the water-holding capacity of soils and thus raise the total quantity of water stored therein. That portion of stored moisture which can be utilized by plants is designated as *available water.* The ability to store water in an available form constitutes one of the principal physical characteristics of a soil and relates closely to its textural and structural configuration and to its organic matter content.

The Gaseous Phase

The volume of air in a soil varies as the reciprocal of the volume of water present. Thus, an increase in the moisture content will cause a reduction in the amount of air which serves as a source of oxygen so important to the functioning of plant roots. As water is removed by the roots, on the other hand, fresh air will enter the soil to fill the voids created thereby. This mechanism, along with gas diffusion, serves to provide a supply of fresh air to the roots at all times. The rate of gas interchange determines the state of *aeration* in a soil.

Soils of low porosity, such as those having a high proportion of microscopic pores through which air movement is slow, are characterized as having poor aeration. This emphasizes the need for the maintenance of good soil structure and a higher proportion of large soil pores within the soil mass. Further, since excessive amounts of water in the soil restrict air movement as effectively as will small pore size, conditions of waterlogging may be expected to limit soil aeration considerably.

Nitrogen is the major component of the soil atmosphere. As a gas, nitrogen is inert and has little effect on soil functions. On the other hand, nitrogen also is an essential plant nutrient by virtue of its place in protein molecules. To be absorbed and utilized by plants, however, nitrogen must be in an inorganic form, usually as either the ammonium or nitrate ion. Fortunately, a variety of soil microorganisms have the capacity for converting atmospheric nitrogen to inorganic form. This is a most important process because it accounts for the bulk of the nitrogen made available to plants in unmanured soils. Whether or not a soil is supplied with adequate available nitrogen has a more significant bearing on plant growth than does any other fertility factor.

Carbon dioxide, though present in small amount in the soil air, serves a very useful function as a result of its ability to form a weak acid in water. In the presence of an acid soil solution, minerals are made more soluble and greater amounts of certain essential plant nutrients may thereby be released for absorption by plant roots.

The principal source of carbon dioxide in the soil is that liberated by respiring plant roots and by soil organisms responsible for the decomposition of organic residues. The addition of fresh plant residues to the soil is followed almost immediately by an increase in the carbon dioxide content of the soil air and by an equivalent decrease in the soil oxygen supply.

When a sample of soil is taken to the laboratory for analysis, it is usually crushed and dried. Two of its three naturally occurring phases, air and water, are lost for further study. Laboratory investigations, therefore, are found to involve primarily the third or solid phase. An appreciation of this fact alone will lead the student to an understanding of the wide differences between the field condition of the soil and that as observed in the laboratory. Our studies of the soil in the laboratory through physical and chemical means can only approximate field conditions.

THE SOIL AS A BIOLOGICAL SYSTEM

Previous discussions picture the soil as a completely lifeless body functioning rigidly under a system of physical and chemical reactions. In reality, the soil may be considered as a living body if we take into account the microbiological, plant, insect, and animal life it contains.

Overall soil conditions are governed markedly by the presence of living organisms or *biota*, a grouping which includes plant roots. The availability of nitrogen and other plant nutrients is related closely to the activities of soil organisms. The indirect effect they have on soil structure through processes of organic matter decay reflects in the soil moisture and aeration conditions prevailing. Circumstances interfering with biological activity have a profound influence on many soil functions. We may consider a soil as 'sick' or 'healthy' depending upon its biologic behaviour.

Biological activity occurs primarily in the surface layer of soil. Here the major interchange of moisture and air takes place between the atmosphere and the soil, and between the soil and the plant. Plant roots are most concentrated in this zone, which accounts generally for the accumulation of organic residues near the soil surface.

The build-up of humus and other organic components in the soil is relatively rapid where conditions promote abundant plant growth and the return of copious quantities of carbonaceous residues which tend to blacken the soil. It is obvious, then, why the layman may consider a dark-coloured soil to have a high capacity for producing crops. Since, on the other hand, the soil scientist knows that soil colour can be a misleading index to the organic matter content, he will tend to consider various other factors which contribute to the crop-producing power of a soil. In the final analysis, the most suitable guide is the general appearance of plants which are growing on the soil in question.

EQUILIBRIA IN SOIL SYSTEMS

Many soil processes attain completion with the establishment of equilibrium conditions governed by physical and chemical laws. In chemical processes, the affinity between reacting ions is responsible for their uniting to form new compounds or substances. The union results in an arbitrarily defined, forward reaction which takes place at a rate dependent upon the specific affinity between reactants, their concentration in the reaction medium, temperature, and perhaps other factors. A reverse reaction normally opposes the forward reaction through reformation of the original reactants. When these two countering reactions proceed at an equal rate, an equilibrium will have been attained. An example of a chemical reaction of this type which occurs in soils is the exchange of ions from the soil solution for those held on particle surfaces. Physical reactions are involved in soil processes also, and usually are brought about by the need for an adjustment in energy levels associated with water, air, or heat in a soil.

Since most soil processes involve the establishment of an equilibrium, they may be expected to vary in rate with time. Normally they proceed most rapidly when initiated because the reactants will be in maximum concentration at that time. As the equilibrium state is approached and the reactants decrease in quantity, the reaction rate becomes progressively slower and will stop when the equilibrium point is reached. Reaction rates associated with the many soil processes are highly variable also. They range from the almost instantaneous chemical exchange reactions to those which may require centuries for completion, mineral weathering being an example of the latter type.

Soil processes seldom attain a true equilibrium state largely because environmental conditions are virtually always changing. Even reactions that proceed rapidly and, therefore, should have an opportunity to reach an end point, may not do so. For instance, rapid exchange reactions and the process of soil wetting, the latter tending toward the uniform distribution of water within the soil mass, must take place in an environment which is undergoing constant change as the result of the removal of both water and ions by plant roots. Mineral weathering, a much slower process, seldom can proceed to completion because of the gradual removal of surface material by erosion. Thus, residual products of weathering forming at the surface are carried away, and fresh, weatherable mineral matter is exposed from below to maintain the weathering rate at a fairly high level. Likewise, organic materials accumulate in the surface layer of the soil to a maximum level determined largely by imposed climatic conditions. However, this maximum level, which would correspond to an equilibrium state, cannot be reached if erosion continually removes small amounts of topsoil with its accumulated organic substances. Nonetheless, both organic matter accumulation and mineral weathering do proceed to a point where the change in soil properties they cause are just balanced by erosional losses. When this point is reached, soil characteristics remain relatively constant and would appear to be at equilibrium with the environment. Even so, the organic and mineral components are involved in continuous and rather profound reactions which have a decided effect upon soil behaviour. This state is described as a *dynamic equilibrium*, and is assumed to reflect rather precisely the influence of environmental factors controlling the development of a specific, mature soil. For all practical purposes, a soil at dynamic equilibrium with its environment will undergo but little apparent change with time, at least with respect to outward appearances.

Some soil processes are subject to rapid change and also to control by man, frequently to his personal benefit. Outstanding among these is the status of fertility and moisture in soils, both of which involve reactions taking place at or adjacent to soil-particle surfaces. The fertility status may be altered through the application of fertilizers and manures to the soil. Moisture levels may be controlled by conservation practices which reduce water loss resulting from runoff, through drainage of waterlogged lands, and through the addition of irrigation water to arid-region soils. Frequently where these processes cannot be controlled, the growth of economic crop plants becomes impossible.

In summary, the soil is an entity which may be described in terms of physical, chemical, and biological characteristics each of which is subject to some degree of change. If these characteristics are not conducive to abundant plant growth, the soil they describe will have but limited crop-producing capacity. The true value of such a soil will be determined, therefore, by whether or not the undesirable features are subject to ready and practical modification by man.

Virtually all soils have one or more features important to crop production which can be improved upon through proper management. A primary purpose of this text is to suggest means for obtaining such improvement. In pursuit of this goal, treatment of the text material in the following chapters is intended to provide a basis for understanding and evaluating the contribution each of the many soil attributes make toward plant growth, and to suggest means for improvement of any of them deemed to affect plant growth adversely. Whether or not improvement can be made in a specific instance will depend largely upon the availability of facilities and materials needed and upon the ingenuity of the individual in applying the corrective measures judiciously.

CHAPTER TWO

PHYSICAL PROPERTIES OF THE SOIL AND THEIR MEASUREMENT

A CONSIDERATION of the soil from the standpoint o its crop-producing ability usually involves two aspects: (1) the chemistry, as it affects the supply of plant nutrients, and (2) the physical condition, or *tilth.* Soil fertility has received by far the greater attention in efforts to improve the crop-producing capacity of soils. Undoubtedly this results from the fact that effects of adding supplemental plant nutrients are recognized easily. However, soil scientists realize that the physical state of the soil can be just as important a consideration in crop production as is fertility. Only if the physical condition of the soil is good can one expect an otherwise fertile soil to support abundant plant growth.

The importance of the physical state of soil is more apparent if we consider a number of the principal soil attributes it influences. Dependent upon the physical condition are:

1. The ease with which the soil may be cultivated.
2. Its capacity to hold water and some of the readily available plant nutrients.
3. Its state of aeration, or the rate of intake of O_2 which is needed by plant roots, and release of CO_2 which, though a product of root respiration, is toxic in excessive amounts.
4. The ease with which roots penetrate the soil, and its ability to support the plant mechanically.
5. Resistance of the soil to erosion by wind or water.
6. Soil temperature.

Physically the soil may be characterized in many ways. Each method attempts to evaluate soil properties as they relate to one or more of the six attributes listed above. Some of the major physical properties which, when considered together, define rather completely the physical state of the soil are (1) *consistence*, or the relative resistance of the soil mass to deformation, (2) *texture*, or the proportions of the different sized particles which make up the soil, and (3) *structure*, or the size and shape of aggregates formed as a result of the particles clinging together. Many specific measurements made on soils relate to and serve as a means of evaluating the effects of the above properties. Several of the measurements will be discussed in the sections which follow.

THE SOIL SEPARATES

The mineral portion of the soil may be divided into size groups which are referred to collectively as *soil separates*. The three major separates are sand, silt, and clay. However, since soil particles occur in an infinite number of sizes, it should be possible to divide or segregate them into an unlimited number of separates. In so doing, each separate would consist of particles of a very narrow range in size.

A soil sample may be described on the basis of the quantity of each of the soil separates it contains. Many of the physical properties of soil may be related to these quantities also. However, readily detectable changes in properties do not occur except through rather large changes in particle size. Therefore, the range in particle size within the various separates can be fairly large and, as a consequence, the separates into which soils are divided normally are limited in number.

Since the changes in properties related to particle size are gradual, it has been difficult to select meaningful points along the particle-size scale to represent where the different size-fractions should be separated. It should come as no surprise, therefore, that numerous systems for particle-size classification have been proposed. Yet, only two schemes are in common use by agriculturists today. These are the International system, as accepted by the International Society of Soil Science, and that used by the U.S. Department of Agriculture. Both systems consider the clay fraction to include those particles having an effective diameter of 0·002 mm. (2 microns) or less, and set the upper size limit of sand at

2·0 mm. The systems differ as to the point of division between silt and sand, and in the number of sub-divisions made within the sand range. These systems are compared in Table 2-1.

Table 2-1. The classification of soil separates according to the international system and that of the U.S. Department of agriculture (after soil survey staff, 1951, p. 207).

Separate	Diameter limits of separate	
	U.S.D.A. system	International system
	mm.	mm.
Gravel	Above 2·00	Above 2·00
Very coarse sand	2·00—1·00	
Coarse sand	1·00—0·50	2·00—0·20
Medium sand	0·50—0·25	
Fine sand	0·25—0·10	0·20—0·02
Very fine sand	0·10—0·05	
Silt	0·05—0·002	0·02—0·002
Clay	Below 0·002	Below 0·002

On occasion, further sub-division within the various size groups is made. Silt may be separated into two or more sub-groups. The clay fraction may be subdivided further into coarse and fine separates, with the fine particles often being referred to as colloidal clay.[1]

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[1] The term *colloid* carries numerous connotations. For our purpose, a 'oid may be considered as a particle so small that it cannot be seen in the ıry microscope. A diameter of 0·0005 mm. approximates the upper size soil colloids.

PHYSICAL PROPERTIES RELATED TO PARTICLE SIZE

Appearance and Feel

The designation of sands as a distinct size-group of particles originated because of man's ability to see and feel the individual grains. Being coarse they have a gritty, harsh feel. We may refer to sand as a 'coarse-textured' material. The finer the sand grains, of course, the more difficult it is to distinguish them. The perception by touch of sands which are sharp and angular will be somewhat easier than if the surfaces of the grains have been worn smooth through abrasion.

The individual particles of silt and clay can be neither seen nor isolated by feel. Silt in bulk is recognized by its smooth, powdery feel. Usually clays are characterized by their stickiness and pliability when wet, but the degree of exhibition of these properties is related to mineralogy as well as to particle size.

Cohesion and Adhesion

Cohesion is the result of attraction between like substances and is a surface phenomenon primarily. We note little coherence in either sands or silts because their total surface area is small. Clays, on the other hand, present a relatively large surface area per unit of mass. If a cube-shaped piece of gravel 1 cm. long on each edge is broken down into cubes having edges of 0·001 mm., which is within the range of clay-sized particles, the total surface area will have been increased 10,000 times. Because of their extremely high surface area, closely packed clay particles display outstanding cohesive qualities.

The coherence among particles will depend not only on their total surface area but also upon their shape and chemical make-up. If the particles have a high proportion of flat surfaces, more extensive contact between surfaces may be made. This results in greater cohesion than when rounded surfaces are dominant. This is particularly true if orientation of the particles can take place so that the flat surfaces come into close contact with one another (see Fig. 2-1).

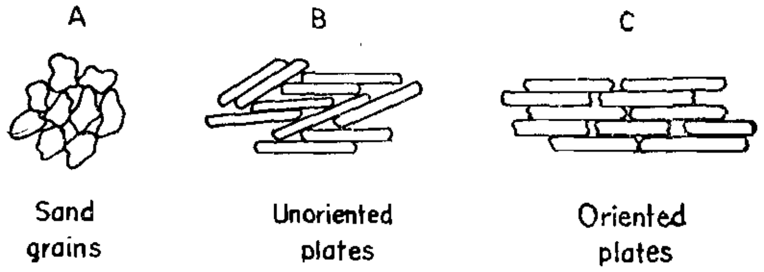

FIG. 2-1. The effect of particle shape and orientation upon cohesion. The point-to-point contacts, as in A and B, afford only a minimum of surface contact and cohesion. Strong forces of attraction will occur between particles in C, where flat, adjacent surfaces lie parallel to each other. The overlapping of particle surfaces in C also adds strength to the structural unit. A relatively large part of the silt and clay in the Indo-Gangetic basin soils should have this plate-like structure.

Clay wetted with sufficient water to form a paste behaves as a viscous fluid. As long as thick layers of water separate the particles, cohesive forces between them will be at a minimum. The clay will have a slick feel. The force of *adhesion*, or the attraction between unlike substances, plays an important role at this moisture level. This is evident from the manner in which the clay will stick to metal or glass, or to the fingers. As the moisture content is decreased, the fluid-like properties become less pronounced. At the same time, cohesiveness increases and, finally, a point is reached where this property overshadows the force of adhesion. The clay mass no longer sticks to the fingers. Instead it may be worked into various shapes which are retained readily. This pliability of clay is an outward expression of *plasticity*. It is the plastic quality of clay which makes it a prized material for use on the potter's wheel.

Many clay materials, when formed into a shape while moist and then dried, will all but defy further deformation. They can no longer be moulded or bent, only broken into fragments by shearing or crushing. During the drying process, surface tensional forces in the water films adhering to the clay operate to pull the particles together, while orientation of the particles occurs when in the moist state. As a result of intimate contact among the particles, cohesional forces attain their maximum intensity as the soil dries out. This explains why clayey soils, when ploughed wet and then allowed to dry, form clods which break down under further cultivation only *with extreme difficulty.*

The chemical nature of the clay will affect its cohesive properties. Clays formed during long periods of intense weathering, as are found more often in the humid, tropical areas, do not have the cohesive qualities of those which have formed as a result of a moderate weathering of primary minerals. As will be discussed in Chapter 4, the clays of the tropics tend to be of the kaolinitic type or occur as free oxides of Fe and Al. These substances have inherently low cohesiveness, since the highly weathered particles are more nearly spherically shaped and the total area of contact among them is limited.

Organic matter in the soil produces variable effects upon consistence properties. In bulk, well-decomposed organic materials are neither sticky nor plastic. As a general rule, increasing the organic matter content will cause the soil to crumble more readily and to present a more open or porous appearance. Clay soil can be made noticeably less sticky and plastic if a sufficient quantity of decomposed organic matter is added to it.

Certain gum-like substances, which occur both in fresh organic materials and as by-products of organic matter decay, have a binding effect upon the soil particles. Thus, groups of individual soil grains may become bound into relatively small, stable aggregates due to the cementing action of these substances. From all outward appearances, cohesive forces within these aggregates are strong. This process is important in clay soils because stabilization of aggregate forms provides permanence to the relatively large pore spaces which have been created. This normally results in an improvement in the rate of movement of air and water through the fine-textured soil material.

The binding effect of organic compounds can be of benefit to coarse-textured soils also. Sand grains, when bound together, are less subject to erosion by wind and water. However, organic matter contributes most to sandy soils by increasing their water-holding capacity and their nutrient-supplying power.

SOIL CONSISTENCE

Consistence has served as a basis for the general characterization of physical .properties of soil through the ages. Defined briefly, consistence is the manifestation of cohesion and adhesion as they affect the ability of the soil to resist deformation by externally applied pressure. By judging the consistence of a soil, the experienced field man can gain some idea as to the approximate proportions of the various size separates the soil contains. However, this is not always a completely reliable procedure. This latter fact may be understood more clearly after careful consideration of the previous discussion concerning the effect of organic matter and clay type on cohesive properties of the soil. A soil high in a clay formed as a result of only a limited degree of weathering may be highly plastic. A second soil of similar texture but containing clay that has been weathered very extensively may be essentially void of plastic properties.

Consistence measurements permit the qualitative evaluation of a soil with respect to a number of its physical properties. Some insight into the relative amount of pore space is possible. This will, in turn, relate to such features as the water-holding capacity, the freedom of water movement through the soil, and the state of aeration. Whether or not the soil is sticky or plastic will serve as an index to its ease of cultivation. Fluffiness of the surface layer of a dry soil is often used as an indication of a high level of salts in waterlogged areas.

Normally, evaluation of consistence properties requires examination of the soil at various levels of moisture. Plasticity in clays cannot be determined on the dry soil. Aggregates or weakly cemented granules of clayey soil may have the consistence of sand when dry. Upon moistening and rubbing between the fingers, however, these aggregates usually will break down and the true characteristics of clay then will become apparent.

Terminology in Consistence Measurements

Consistence properties in soils are customarily evaluated at three different moisture levels; wet, moist, and dry. These three levels are defined rather loosely. In the *wet* state, the soil should contain somewhat more moisture than would be desirable for ploughing. Further, if sufficient clay is present, plastic properties will usually be readily discernible. The *moist* state is considered to approximate as wet a condition as is suitable for ploughing. More often than not, this state is observable a few inches beneath the surface of the soil in the field and after a few days following a wetting rain or irrigation. At this moisture level the soil should be neither sticky nor plastic. Observations as to whether or not the soil is amenable to cultivation would be made at this moisture state. The *dry* state denotes a sufficiently low moisture content so that there is a noticeable difference in soil colour from that apparent in the moist state.

Consistence When Wet: The degree of stickiness and plasticity is determined on a soil in the wet state. The usual procedure for plasticity evaluation in the field is to determine the ease with which the soil may be rolled into a rod-shaped form and the resistance of this form to subsequent deformation. The test for stickiness must, of course, be carried out at a higher moisture content than is used for the test for plasticity. For sticky, plastic soils, greater care must be exercised in selecting the proper moisture condition for cultivation. These soils will be high in clay and may form very hard clods on drying if ploughed when too wet. Sandy and silty soils will not display extreme properties of stickiness and plasticity.

Consistence When Moist: In the moist state, observations are normally made as to the structural condition of the soil. Desirable consistence is in evidence when, under pressure, the soil crumbles into small, soft granules. This condition is denoted by the term *friable*. A soil with a discernible amount of clay usually will form a coherent cast when squeezed in the hand. Sandy soils, due to their lack of cohesion, will not form a cast and are considered to be *loose*. Moist soils which can be crushed only through the application of considerable pressure are described as being *firm*.

Consistence in the Dry State: The consistence of dry soils is described by the terms *loose*, *soft*, and *hard*. Generally, these terms denote the stepwise variation from sandy to clayey soils. Thus, hard soils usually contain a high proportion of clay. At the same time, however, a clay soil in a desirable physical condition can be somewhat loose or soft when dry. Usually this condition will be observed only when the organic matter level of the clayey soil is reasonably high.

SOIL TEXTURE

The texture of the soil is a property attributable to the combined effects of the various size fractions present. *Coarse*, *medium*, and *fine* are general terms used to denote texture. The size separates do not have equal influence on the textural qualities of the soil. For instance, the addition of a small quantity of fine-textured material will have a much more profound effect upon soil physical properties than will the addition of an equal quantity of coarse material such as sand. Perhaps this is the result of our expressing the amount of the separates occurring in the soil on a weight basis. Contributions to the physical properties are largely a function of surface area of the fractions present, not of their weight.

The origin of the term, soil texture, undoubtedly predates attempts to evaluate it quantitatively. Any cultivator will recognize textural qualities of the soil even though he may not be able to define them. Because of the close relationship between the proportionate amounts of the different size fractions and consistence, textural designations will relate to the workability, feel, and appearance of the soil. In many instances, it is possible for an experienced person to approximate soil texture in the field as a consequence of observing consistence properties.

Soil Textural Classes

Soils are seldom, if ever, composed of but a single size separate. Even those soils which contain a predominance of a single separate are relatively limited in number. All of the fractions present contribute to the physical properties of the soil.

The *textural class* into which a soil sample is placed depends upon the size of particles which dominate its mineral complex. The class names, on the other hand, allude not only to particle size but also to consistence properties as determined by feel and workability. Thus, we find three general class names in use today ; sandy soils, loamy soils, and clayey soils.

A consideration of the general class names would suggest that soils are comprised principally of sands or clays, or of a mixture of the two. Thus, we may assume the loam class to correspond to a combination of properties approximately intermediate between the extremes of those peculiar to sand and clay. This being the case, soils with a very high silt content will fall naturally among the loamy soils.

The entire range in properties within each of the three broad classes is quite large. It is possible to subdivide the major classes into less broad categories and still have each group retain a distinguishing set of physical characteristics. The names of the soil separates which contribute materially to the properties of a soil are included in the name of its textural class. Some of the class names in common use today are as follows :

Sandy soils	*Loamy soils*	*Clayey soils*
Sand Loamy sand	Sandy loam Loam Silt Silt loam Sandy clay loam Silty clay loam Clay loam	Sandy clay Silty clay Clay

Because of the noticeable differences among the various size fractions of sand, it is customary to express the textural class name on the basis of the dominant sand separate present. Thus, we will find frequent reference to such class names as fine sandy loam, coarse sandy clay, etc.

The Determination of the Textural Class of Soil

Before a soil can be placed in its proper textural class, the amount of each of the soil separates it contains, or the *particle-size distribution*, must be determined. The steps involved constitute the *particle-size analysis*[1] of the soil. The procedure includes (1) dispersion of the soil in water so that the individual particles will be separated from one another, and (2) grading of the particles into the various size groups. A more complete discussion of mechanical analysis will be found in later paragraphs.

Satisfactory agreement among soil scientists throughout the world has not been reached relative to the identification of soil textural classes. The problem arises from attempts to relate these classes, described in terms of percentages of clay, silt, and sand they contain, to physical properties formerly used to describe them. In the first place, textural designations of loam, etc., are abstract terms not readily defined in absolute values. Secondly, soil separates of different origin often contribute to textural characteristics differently. As discussed previously, two soils of identical textural composition may appear to belong to different soil classes because of dissimilarities in the physical properties of the clays they contain. Lastly, a universal system for expressing size limits of the soil separates is not used. For example, recall that the system used in the United States includes appreciably larger sized particles in the silt range than does the International system (see Table 2-1).

Procedures for converting mechanical analysis data to textural class names are variable. Less useable systems are comprised of tables listing the range, in percent content, of each separate within each designated textural class. A less complicated procedure involves the use of rectangular co-ordinates for expressing the percentages of the separates within each soil textural class. Such charts are shown in Figs. 2-2 and 2-3. Not only can the class of a soil be readily determined from these figures, but the relation to the other classes also may be seen at a glance. It should be noted, however, that use of these diagrams is possible only when all sand particles are determined or expressed as a single separate. If the individual sand

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[1] Commonly referred to as 'mechanical analysis'.

fractions are measured separately, tables must be used for an accurate textural classification.

It should be realized that the class designations in the figures are tentative and will probably continue to change from time to time, as they have in the past. It must be remembered that Fig. 2-2 is based on the measurement of silt as having a size range of from 0·002 to 0·02 mm., whereas the upper limit of 0·05 mm., as prescribed by the U.S.D.A. system, is to be used with Fig. 2-3. Note also that Fig. 2-3 shows two additional soil classes, silt and sand, which are not included in the International system.

The two schemes differ not only with respect to the particle size ranges for the sand and silt separates but also in that the textural classes defined by them do not represent the same proportion of each of the separates. As a consequence, comparison among soils is possible only if all have been classified by the same system. It is essential, therefore, that the presentation of textural analysis data include a notation as to which scheme has been used in the classification.

Particle-Size Analysis

The procedure for particle-size analysis consists of dispersing the soil in water and then measuring quantitatively the amount of each separate present. The dispersion is effected mechanically by stirring or shaking the soil-water suspension. Chemical dispersing agents are added to aid the process.

Dispersion may be hindered by cementing substances or binding agents which secure the individual particles in aggregate form. Among these agents are organic matter, lime, and oxides of Fe and Al. For accurate results, therefore, it is considered necessary to remove these substances by appropriate chemical treatment prior to mechanical dispersion. It would be pointless, of course, to remove them from soil samples which contain these substances as a dominant fraction of the soil since they would contribute materially to soil properties.

Dispersed, suspended mineral particles settle at a rate dependent upon their size. Therefore, with the passage of time, the quantity remaining in suspension decreases gradually as successively smaller particles settle to the bottom. We can

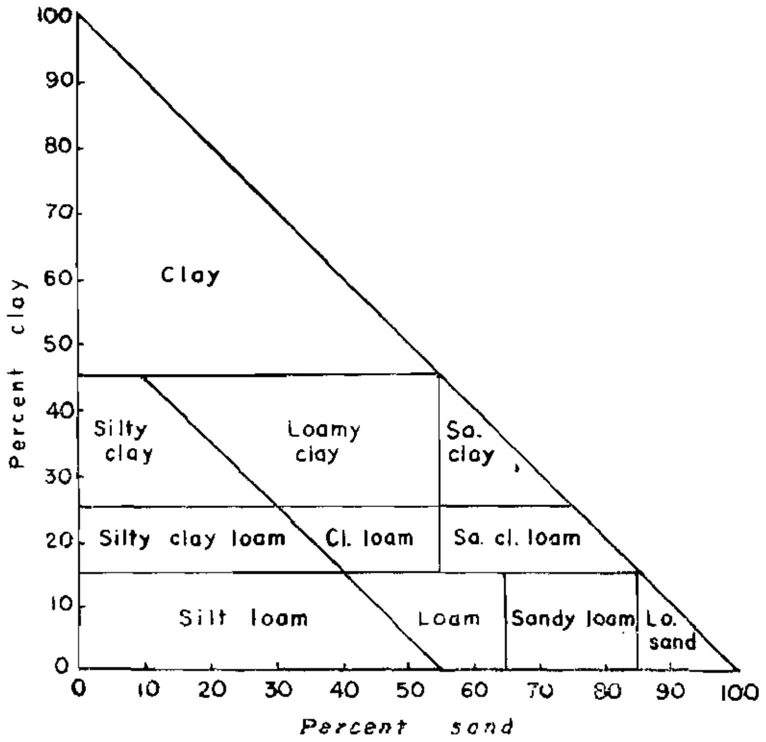

FIG. 2-2. Diagram for determining the textural class of a soil based on the International method of textural analysis in which particles in the silt fraction range from 0·002 to 0·02mm. in diameter. To use, draw a horizontal line corresponding to the percent clay and a vertical line corresponding to the percent sand in the sample. These two lines will intersect within the box representing the textural class of the sample. (Adapted from the *Potash Pocket Book* (1957), International Hondelmaabschappij voor Mestoffen, N. V. Keizersgracht 333, Amsterdam C, Holland.)

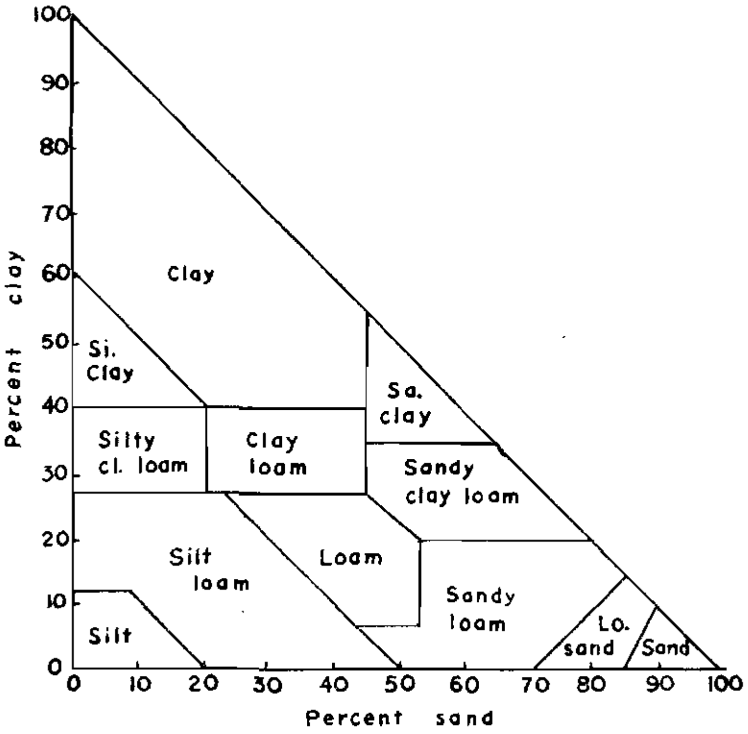

FIG. 2-3. Diagram for determining the textural class of a soil based on the U. S. Department of Agriculture system of textural analysis in which clay particles are less than 0·002 mm. in diameter, silt from 0·002 to 0·05 mm., and sand from 0·05 to 2·0 mm. For use, refer to Fig. 2-1. (Adapted from the *Soil Survey Manual*, Soil Survey Staff, 1951, p. 209).

calculate how long it takes for a particle of a certain size to settle out. Thus, by sampling the suspension after this specific time, we can measure the quantity of particles which are smaller than this size and, therefore, remain in suspension. By sampling at different times we can measure the quantities of any number of size fractions occurring in the soil. Relating these quantities to the weight of the original sample allows for the estimation of the percentage of each fraction the soil contains.

The equation from which settling velocities are calculated is based on Stokes' Law. It expresses the force acting on a sphere of a certain radius, r, in terms of the velocity, V, at which it moves through a liquid medium of viscosity, n. The velocity is proportional to the difference between the particle density, D_p, and the density of the liquid, D, and to the acceleration of gravity, g, as shown in the equation

$$V=\frac{2}{9}\,\frac{(D_p-D)r^2g}{n}. \qquad (2\text{-}1)$$

For practical use it is desirable to express this equation so that we may calculate more readily the time, T, required for a particle of specific size and density to fall through a specified distance, L. This may be done if we replace the velocity term by its equivalent, or the distance divided by time (L/T). Furthermore, we may replace the radius, r, by diameter, d, since this is the common method of measurement of particle size, and substitute the acceleration of gravity (980 cm./sec.^{2}) for g. Following this, equation 2-1 reduces to

$$T=\frac{0{\cdot}031nL}{(D_p-D)d^2}, \qquad (2\text{-}2)$$

where T is measured in minutes; n, in poise; L, in cm.; D_p and D, in g./cc.; and d is expressed in mm. The value for d is an *effective diameter* or the diameter of a sphere which would fall at the velocity of the particle in question.

In using the equation, the viscosity of the settling medium, which is temperature dependent, must be obtained from tables.[1] At normal laboratory temperatures, using a density value of

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[1] Such tables are found in reference handbooks for chemistry and physics.

one g./cc. for water is satisfactory. The particle density value of 2·65 g./cc. is suitable in most instances, except where the soil particles are made up largely of minerals high in Fe. In this latter case, particle-density values must be determined separately.

The Practical Value of Particle-Size Analysis

The complete physical characterization of a soil is not possible unless we know its textural classification. Conversely, knowledge of the textural class alone, though highly valuable when used as a basis for judging the general physical behaviour of a soil, is not entirely suitable for this purpose. Other information needed for a complete evaluation includes the organic matter content of the soil, the degree of aggregation of soil particles, and the stability of these aggregates. Since removing a sample of soil to the laboratory will alter its original physical state (i.e., its natural structural form), a part of the evaluation must be made in the field.

Information on the particle-size distribution is most useful in predicting the potential behaviour of a soil under a specified set of conditions. For instance, we may recognize in a sand the ability to conduct water rapidly, to erode readily, and to be drouthy. We would know that no practical treatment applied to a deep sandy soil would create in it the capacity to hold impounded water, as is done commonly in the culture of rice under irrigation. A clayey soil, on the other hand, would be expected to behave oppositely. Here, however, many other factors such as the organic matter content and the state of soil particle aggregation often are more important in determining overall soil properties than is texture alone.

A soil which is sticky and plastic by virtue of its high clay content may be difficult to plough. It may have a high water-holding capacity but, more than likely, will conduct water only slowly through the small pore spaces it contains. A second soil may have the same amount of clay, but because of a concurrent, high organic matter content and good physical state, it may be ploughed without difficulty and could conduct water with relative ease. Although these two soils may possess essentially the same capacity for storing water for plant use, their ability to supply the water and to support plant growth could differ markedly. This comparison emphasizes further the limits we must place on soil-texture values as indexes to soil behaviour, and it gives us further understanding as to why we must study thoroughly the various components or attributes of a soil if we are to characterize it adequately.

SOIL STRUCTURE

The arrangement or grouping of individual particles with relation to each other, or *soil structure*, greatly modifies the influence of texture on other soil properties. Without knowledge of structural configuration, application of data on textural composition is not practical.

The cause of structural development is not clearly understood. Since the soil structures assume various, rather distinct forms, it is probable that more than one mechanism is responsible. Factors contributing to the formation of structure include (1) the proportion and type of soil separates present, (2) the influence of actively growing roots and burrowing organisms, (3) the decay and accumulation of organic matter, (4) the expansion and contraction of the soil mass resulting from variation in the moisture content and in temperature, and (5) cultivation, particularly as it affects frequent alteration in the structural conditions within the surface layer of the soil. These factors, as well as others, will operate together in the creation of structural units which fit into a rather well defined system of classification.

In its original state, soil material is often without structure. The particles may persist in a *single-grained* state if there is no tendency for them to become bound together. This condition is exemplified by loose sand. If, however, there is fine-textured material or other binding substance present, a *massive* state may be assumed. In this condition the soil occurs as a weakly bound, continuous mass which, under the influence of crushing, breaks into irregular pieces having no particular shape or size.

Structural Classes

Although definite structural forms are not always evident in soils, they are observed very frequently where soil profile development is pronounced. Soil structural units may be classified on the basis of shape and size of individual fragments which may be pried or broken out of the soil profile from a freshly made exposure. The common structural types listed by the Soil Survey Staff (1951) are as follows:

1. *Platy*, or arrangement of particles along a plane. Normally, individual fragments tend to break out into thin sheets.
2. *Prismlike*, with the main structural unit being longer along its vertical axis and with external surfaces being relatively flat. This category is divided into two subgroups depending on the shape of the upper surface of the prisms, i.e.:
   (*a*) *Prismatic*, with the upper surface assuming a nearly level plane,
   (*b*) *Columnar*, with the upper ends distinctly rounded.
3. *Blocky*, in which the structural forms approximate cubes but may have irregular surfaces. Such units comprising the soil mass fit closely together. The blocky structure is also classified into two subgroups:
   (*a*) *Angular blocky*, in which the individual surfaces join as sharp angles, and
   (*b*) *Subangular blocky*, in which the sharp angles are replaced with rounded corners and edges.
4. *Spheroidal*, in which case the particles are arranged in a more or less spherical configuration. Two subtypes are recognized:
   (*a*) *Granular*, which consists of relative compact, nonporous aggregates, and
   (*b*) *Crumbs*, which are porous and easily broken down.

The various types of soil structure are illustrated in Fig. 2-4. Note that the prisms and columns are the largest structural forms, whereas the spheroids are often quite small. The relative sizes shown in Fig. 2-4 have little relationship to the actual sizes observed in the field.

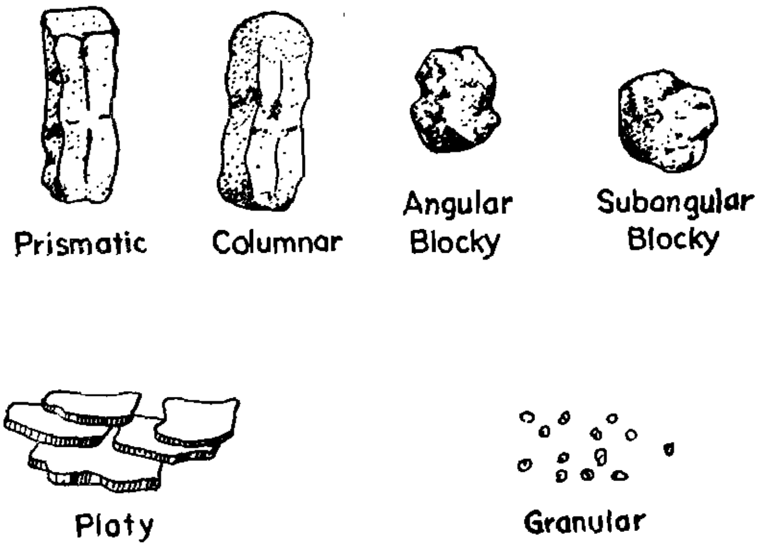

FIG. 2-4. Some of the major soil structural forms.

Formation of Soil Structure

The mechanisms of structure formation cannot be adequately explained. The forces responsible for establishing one type of structural unit do not necessarily apply to others.

Spheroidal crumbs and granules are most readily formed in the surface soil where chemical and biological activity is high. Factors contributing to their formation are (1) alternate wetting and drying of the soil, (2) the activity of plant roots and burrowing animals and insects, (3) decaying organic matter, particularly as affected by the formation of mucilagenous organic compounds capable of a cementing action, and (4) the presence of polyvalent cations capable of uniting adjacent particles because of their multiple positive charge.

Hydration and dehydration influence structural development by causing a physical shifting of the particles within the soil mass. Some types of colloidal clay tend to expand as they become hydrated and contract as the moisture disappears.

Furthermore, in the moist state, a certain degree of freedom of movement is afforded the particles allowing them to orient or group themselves and, under the influence of cohesion, they are capable of forming compact aggregates (Sideri, 1936). The particles are drawn into closer contact as the thin water films covering their surfaces disappear during dehydration. The fact that rehydration of the colloids will occur at a slower rate than dehydration lends a certain degree of stability to the aggregates.

The physical activity of roots and animals assists the above processes by mechanically disturbing the arrangement of the soil particles, Expansion of growing rootlets exerts a compressing effect on the loose aggregates surrounding them. Furthermore, the massive network of roots characterizing many plant species will create countless lines of weakness throughout a volume of soil which occurs in an otherwise nonaggregated or massive state. The roots permeate the soil with relative thoroughness and, by their wedge-like action, fragment the soil mass. By this means, granulation may be assisted or initiated.

Subjecting a 'puddled soil'[1] to repeated cycles of wetting and drying can create in it a decided granular structure. Freezing, which in itself is a form of dehydration, and thawing a moist soil is as effective in granule formation as is the actual wetting and drying process. These processes will ameliorate to some extent cloddy conditions in plouged soil.

The development of a crumb or granular structure does not in itself have a permanent effect on the soil. For this, stabilization of the structural forms is necessary. Stabilization is encouraged by orientation of the particles so they are in close contact, or through cementation by various substances, particularly by organic matter, lime, and iron and aluminium oxides. It is the purpose of this discussion to refer to these facts only and not dwell at length on the subject. For the

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[1] A puddled soil is one in which the particles are in a dispersed, nonaggregated form. The term usually is applied to the finer-textured soils. Puddling produces undesirable characteristics in that pore spaces are small and aeration and water movement retarded. On the other hand, in rice culture, puddling is a practice often followed prior to transplanting.

interested student, reference is made to the thorough consideration of granule stabilization as presented by Baver (1956).

Whereas granules and crumbs are observed more often in the surface layers of soil, the larger structual forms, such as prismatic, columnar, and blocky, occur principally in the subsoil. This undoubtedly relates to the lower level of physical and biological activity in the subsoil, a high level of activity apparently being necessary for the formation of more finely divided granules.

The principal direction of root extension in the subsoil is downward, and the development of lateral roots becomes less frequent there. Thus, the predominant effect of roots in splitting the soil mass will be along vertical planes. In addition, the formation of cracks due to expansion and contraction of the soil mass will occur in vertical rather than horizontal planes. Thus, one would expect natural cleavage lines, corresponding to the boundaries of structural units, to be more pronounced in a direction parallel to the vertical axis of the soil mass.

Aging or stabilization of the structural forms which develop in the subsoil usually consists of their becoming coated with a thin layer of nonplastic organic matter. Union at the contact surfaces between structural units is weakened by this, and subsequent cleavage takes place more readily. Such a condition characterizes the prismatic, blocky, columnar and, to a lesser extent, subangular blocky structural forms. Because of the presence of an organic coating, the surfaces of these units of structure are often much darker in colour than are their interiors.

The above-mentioned macrostructures may be broken out of the soil mass along the weakest planes of cleavage. In addition, however, they are normally permeated by various other horizontal and vertical cleavage planes along which separation is more difficult. Nonetheless, gentle crushing will normally result in the disintegration of the macrounits into successively smaller fragments of prismatic or blocky configuration.

OTHER PHYSICAL PROPERTIES RELATED TO STRUCTURE

Bulk Density

Whereas particle density expresses the mass per unit volume of individual soil particles, *bulk density* is the mass per unit volume of total soil, pore spaces included (see Fig. 2-5).

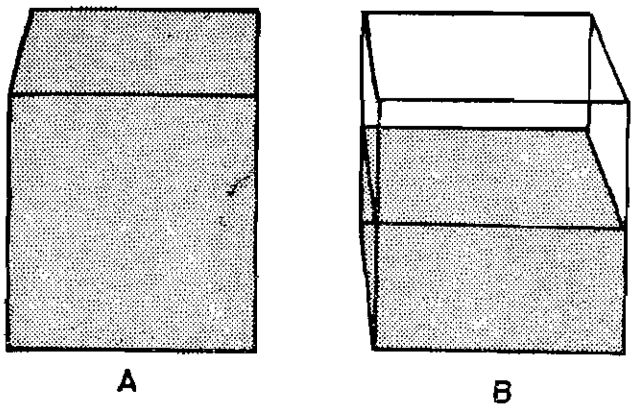

FIG. 2-5. Diagram illustrating the relationship between true density and bulk density of soil. Cube A represents a volume of soil as it might occur in the field. If it were 1 cc. in volume and weighed 1·3 g., its density, which would include the air in the pore spaces, would be 1·3 g. per cc. This would be the *bulk density*. If the cube were compressed so that all air was forced out, as in B, the weight of the soil would remain the same. Should the final volume now be 0·5 cc., the density of the compressed soil would be 2·6 g. per cc. This would correspond to the *particle density* of the soil particles.

It follows that bulk density is a highly variable quantity since it depends upon the degree of aggregation and porosity of the soil.

Bulk density is a highly useful measure because it relates so closely to the pore-space content of a soil. Compaction, or compression of soil, represents a reduction in pore space and an increase in bulk density. Bulk density is measured in soil profiles as a means of determining the occurrence of compact layers, particularly in the subsoil. The presence of such layers may indicate the downward movement of clay or clay-forming substances by percolating water, and the lodging of these particles in pore spaces within the subsoil. Compaction of the subsoil is considered to occur frequently as the result of the pressure exerted by overlying strata. Likewise of great importance is the very serious effects arising from the trampling of hoofed animals. The destruction of vegetation by penned or tethered animals is attributable to a large extent to the compacting effect they have on the soil underfoot. Layers having a high bulk density may result also from the chemical precipitation of cementing agents, either from percolating waters or at the surface of a fluctuating water table. From the standpoint of plant growth, compacted or cemented layers in the subsoil represent zones resistant to penetration by water and roots.

An interesting observation on the bulk density of soils near Rawalpindi has been made by Hamid.[1] The value increases with depth from about 1·3 g. per cc. near the surface to about 1·7 g. per cc. in the subsoil. There is a parallel increase in the lime content of the soil with depth. The increased density of the subsoil appears to be the result of the accumulation of $CaCO_3$ leached from the surface layers and re-deposited in the pore spaces of the subsoil. In other adjacent areas, where the surface soils have been removed and the calcareous subsoils exposed, the bulk density of the upper soil layers also approximates 1·7 g. per cc.

The bulk density of soils is usually lowest in the surface soil and particularly after ploughing. However, as settling of the cultivated land takes place, this value increases once again. The compacting effect of machinery and animals, especially when the soil is wet, will increase the bulk density appreciably. The surface layer of dispersed, puddled soils of low porosity may be expected to have a high bulk density. The effectiveness of cultural treatments designed to improve the physical condition of such soils may be determined by observing changes in bulk-density values.

[1] Abdul Hamid (1957), unpublished data from the A. C. Section, Punjab Agricultural College, Lyallpur, West Pakistan.

Pore Space

The volume of pore spaces in the soil is not constant but varies according to changes in soil structure. This is most important in the surface layer of soil where ploughing has a loosening effect and causes at least a temporary increase in the total porosity. Pore spaces are classified into two groups depending upon their relative size; the large or *macropores*, and the small or *micropores*. Differentiation between these two groups usually is on an arbitrary base.

A number of factors are known to affect the total porosity of the soil and the proportionate amounts of each of the two pore-size groups present. The total pore space can be shown to be related to the degree of aggregation. It is dependent upon texture and organic matter content of the soil, amount and type of cultivation and, in certain instances, the kind of plant grown. Pore space usually attains a maximum volume under continuous grass vegetation. This is attributable to the intense microbiological and root activity occurring beneath sod as well as the high level of soil organic matter produced by fibrous grass roots.

The determination of total pore space requires knowledge of the bulk density of the soil and the mean density of the particles comprising it. From these values the absolute volume occupied by the discreet soil particles may be calculated. The difference between this volume and the total soil volume represents pore space. It may be determined by the formula:

$$\text{Percent pore space} = \left(1 - \frac{\text{Db}}{\text{Dp}}\right) 100, \qquad (2\text{-}3)$$

where Db is the bulk density of the soil and Dp is the density of the mineral soil particles.

It will be found that from one-third to two-thirds of a volume of soil is occupied by pore spaces. Clayey soils generally have a higher total pore space than do sandy soils because of the greater aggregation and, consequently, lower bulk density in them. A major difference noted between these two textural types is the fact that, although lower in total porosity, there is an appreciably higher proportion of large pore spaces in sands. This is to say, the ratio of macropores to micropores in sands is higher than in clays. This is decidedly important in the matter of aeration, for the rate of air flow is greater through large pores. In clayey soils, therefore, it is often found desirable to attempt increasing the average size of pores through cultural practices which will increase aggregation. A large number of investigations have been conducted to determine the relationship between plant growth and the porosity of fine-textured soils.

Assessing Soil Structural Development

The degree of development and type of soil structure have a profound effect on the movement of water and air through the soil, and may determine to a large extent the vigour with which plants can grow. The degree of structural development is often observed readily in the field by the recognition of structural forms visually or by feel, or by observing the rate at which water moves into and through the soil. These methods are strictly qualitative, however. Frequently it is important to measure rather accurately differences in structural development and permanence in different soils. Discussions concerning such measurements are found in the text by Baver (1956).

Relationships between Soil Structure and Plant Growth

Cultural operations in crop production can influence the structure in the surface soil to a large extent but will have little effect on the subsoil. Generally, structure of the surface soil falls into a limited number of types and may be described by such terms as puddled, cloddy, granulated, etc.

The state of structural development in the surface soil serves as a measure of *soil tilth.* Tilth refers to the physical condition of the soil as it relates to plant growth. A soil having good tilth may be characterized as one persisting in a state of granulation which permits not only intimate contact between seeds and soil but also one in which pore spaces are large enough to allow for a ready exchange of gases essential for seed germination and root growth. In a cloddy soil, pore spaces where seeds come to rest will be large. Excessive movement of air within the pores results in a rapid loss of moisture to the atmosphere and prohibits, to a large extent, the absorption of adequate water by the seed pieces for germination. Thus, a soil properly prepared for a seed bed will have been worked to a point where a major proportion of the soil aggregates are small or crumblike.

An opposite effect is noted in puddled soils, or in weakly aggregated clayey soils. Pore spaces are of minimum size and very good contact occurs between the soil particles and seed. However, due to restricted exchange of gases, the level of O_2 available to the seed may be so low as to inhibit or completely prevent germination.

A good example of the effect of porosity on germination is presented by Baver and Farnsworth (1941) in which the stand of sugar beets sown to a clay soil was found to be closely related to the content of large, noncapillary pore spaces. As the noncapillary pores increased from a minimum of about 2% of the soil volume, the improved conditions of aeration ensuing brought about a marked increase in the number of seedlings emerging. The relationship is shown in Fig. 2-6.

The influence of soil structure on plant growth is generally of an indirect nature and relates primarily to soil moisture and soil air conditions. Direct influences occur, however, where structures of a compact, impervious nature prevent satisfactory root penetration. Furthermore, the downward movement of water may be so inhibited as to cause waterlogging of the soil above the impervious layer. Under this set of conditions, plant growth will be restricted both by limitation of the downward extension of the roots and the exclusion of air from the root zone saturated with water.

SOIL COLOUR

The colour of a soil will depend initially on the type of material from which the soil is forming. As weathering and soil developmental processes alter the original material, accumulations of weathering products and organic matter tend to bring about a change in the colour of the soil. Thus, this visible characteristic often will serve as an index to the

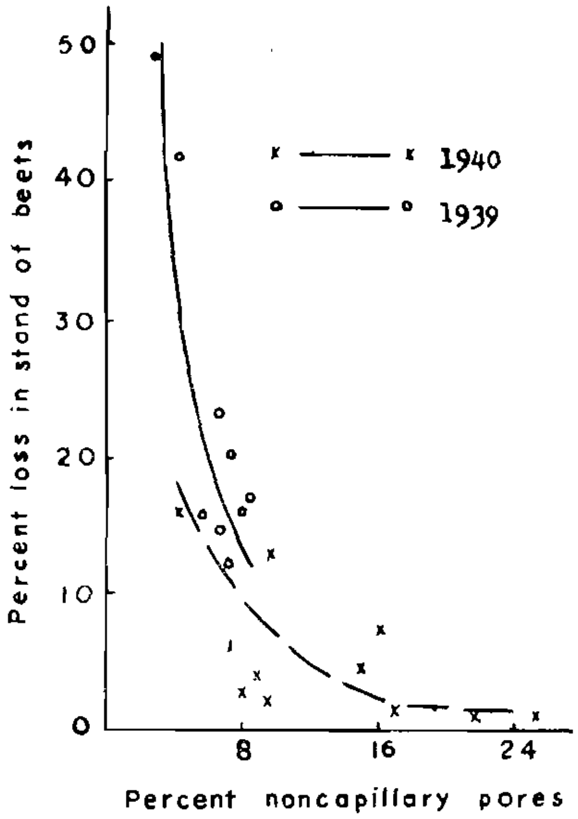

FIG. 2-6. The relationship between soil aeration and sugar beet stand. An increase in noncapillary pore spaces results in better aeration and more successful germination (from Baver and Farnsworth, 1941).

intensity and duration of the processes of mineral weathering and soil development.

A large part of the minerals in the surface crust of the earth are light in colour. Soils developed from them will be light in colour in their early stages of formation. As organic matter accumulates, the colour of the soil darkens, passing successively through various shades of brown and, finally, to black or brownish black. If climatic conditions encourage abundant vegetative growth, the return of relatively large amounts of organic residues results in a more rapid darkening of the soil. If, at the same time, conditions are favourable for the breakdown of soil minerals, other changes in colour will take place. This is usually the result of the formation of highly coloured oxides of Fe which may occur as coating over the original minerals, or, if degradation of the original minerals has been extensive, these oxides may comprise the bulk of the mineral skeleton of the soil. Iron oxides range in colour from yellow, through brown, to red. The more intense these particular colours in soils, generally, the more extensive has been the weathering of the original minerals from which they are derived.

Red iron oxides are either low in water of hydration, or they may occur as the completely dehydrated mineral. Those that are more highly hydrated generally have a yellow colour. Thus, reddish soils often are associated with good drainage. Yellow soils, on the other hand, are assumed to have been formed under conditions of inadequate drainage which permits hydration of the minerals as they are formed.

Red and yellow soils are observed in some locations where they obviously have not been subjected to extensive weathering. Outstanding in this respect are the widespread deposits of reddish materials occurring throughout the arid to semi-arid uplands in the northern and western reaches of the Indo-Pakistan subcontinent.[1] It appears that these materials acquired their characteristic colour in ages past and in areas where more intense weathering was possible. They occur today where they have been laid down by waters which carried them from their original point of formation.

Oxides and other types of Fe compounds produce characteristic colours which serve to indicate the presence of water-logged conditions in soils. Generally the bulk of the soil material subjected to long contact with water will have a gray to blue-gray cast. This is attributed to the occurrence

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[1] These deposits constitute the Siwalik and other formations as described by Wadia (1939). They skirt the southern edge of the Himalayan Uplands and occur as extensive beds in the Salt Range and in the mountainous regions of West Pakistan.

of Fe in a reduced state, the result of exclusion of O_2 from the water-filled soil pores. Where at least periodic access to O_2 is possible, some of the Fe will have been oxidized to rust-coloured, more highly hydrated forms. Iron is observed in this form as spots or streaks in the otherwise grayish soil material.

The texture of the soil will influence the intensity of colour, particularly where Fe compounds and organic matter occur as thin coatings over the mineral particles. This relates to total surface presented by the soil particles. The colour of sands may be affected more than is the colour of clay by a given quantity of Fe oxide or organic matter spread evenly over the surfaces of the individual grains. The same amount of colouring matter spread over a like volume of clay would occur as much thinner films over the greater total surface presented by the finer particles. Moisture influences soil colour, generally by causing the soil to appear darker than when it is dry.

Often it is assumed that a soil high in organic matter is fertile. As indicated previously, inclusion of organic matter in sandy soils usually will serve to improve them as media for plant growth. Since well-decomposed organic matter is black, or nearly so, a common tendency is to judge its level in the soil on the basis of colour. The relationship between colour and organic matter content is only qualitative, however. Highly dispersed organic substances may occur as films over a large part of the surface of soil particles and will impart a comparatively intense black colour to the soil mass. The same quantity of organic material occurring in aggregate form rather than as surface coatings may darken the soil only slightly. The soils of the Potwar Plateau in northern West Pakistan are examples which show a relatively intense black colour even though the organic matter content is low (1% or less).

Colour has a direct effect on behaviour of the soil only as it influences soil temperature. It is a well known fact that dark-coloured objects absorb heat readily and that light-coloured materials reflect it. Thus, dark soils warm up more quickly and attain higher temperatures when warmed from an external source than do light-coloured, reflective soils. At the same time, the dark-coloured soils radiate heat more readily and, therefore, tend to cool more rapidly when the source of heat is removed.

In summary, then, soil colour is a readily observed feature which may serve as an index to other soil characteristics. However, conclusions concerning the soil which are drawn as the result of observing soil colour usually must be considered as tentative until proven or disproven by other facts.

SOIL TEMPERATURE

The temperature of the soil is important because it influences various chemical, physical, and biological processes. Within practical limits, microbiological and root activity within the soil are observed to increase with increasing heat. The rate of water movement will vary measurably with the temperature. Evaporation from the surface of the soil is also a function of temperature, although other factors are more important in governing water loss by this process.

Factors Affecting Soil Temperature

The temperature of a soil is dependent upon several factors operating jointly to determine the rate at which heat is gained or lost. Major factors influencing soil temperature are the angle of incidence of radiation from the sun and altitude. These factors will be modified, sometimes to a large extent, by colour of the soil, the temperature of the air and its rate of movement above the soil surface, water content of the soil and rate of evaporation, and the amount of cover or shade.

Soils will receive maximum energy through solar radiation if they lie in a plane perpendicular to the rays of the sun. Thus, heating of the soil will be greatest if it occurs in equatorial regions. In the Northern Hemisphere, those soils occurring on south-facing slopes will be warmer than those on north slopes because of the more favourable angle of incidence of the sun's rays. At the same time, however, radiation from the sun must travel a greater distance through the atmosphere to reach land surfaces in the Northern Hemisphere. Thus, the farther north one goes, the less effective the sun becomes as a source of heat.

Aside from effects caused by variation in the angle of incidence of radiation, the temperature of a soil will be increased as the soil colour becomes darker and as the moisture content decreases. Dark-coloured soils absorb considerably more heat than do light-coloured, reflective soils. Water influences temperature in that it has not only a higher specific heat than the soil particles but, through its evaporation from the surface, will tend to maintain a cooler temperature.

Cover over the surface of the ground intercepts solar radiation and prevents its absorption by the soil. Cover is provided both by growing plants and their residues which accumulate on the surface of the ground. Clouds may be more effective in reducing soil temperature than plants, because they influence markedly the temperature of the air in contact with the soil. At higher elevations and in the more northern latitudes, snow affects soil temperature by limiting the depth to which frost will form. Furthermore, by reflecting the sun's rays, snow will prevent the early warm-up of the soil in the spring season.

Soil which is thoroughly protected by plant cover will acquire a temperature more nearly equal to that of the air above. However, if moisture at the surface is plentiful, evaporation may reduce the temperature to below that of the atmosphere. Exposed soil intercepting the direct rays of the sun may be as much as 30° to 40°F. above air temperature. The difference is appreciably greater with dark-coloured soils than it is with those of light colour.

Heat Conductance in Soils

The components which make up most soils do not vary markedly in their ability to conduct heat. Of greater importance is the texture and moisture content of the soil and, to a lesser extent, structure.

Heat transfer through a solid mineral is facilitated by the fact that the constituent atoms are in close contact with each other. If a solid block of the mineral is broken into small grains, its capacity for conducting heat will be decreased.

Transfer through the individual grains is as rapid as before. However, transfer from grain to grain must occur largely through the small contact points which are decidedly limited in their ability to conduct heat.

Air molecules within the pore spaces will conduct heat from grain to grain also but, because of the low density of air, this mode of transfer is much less than through the solid components. The finer the texture of the soil the greater is the resistance to heat flow. The greater the number of particles per unit volume of soil the greater the number of grain-to-grain contact points through which the heat is channelled.

Water will increase the ability of the soil to conduct heat. This arises largely from the fact that water is more dense than the air it replaces in soil pores when the soil is moistened. Structure will affect heat transfer through the influence it has on the solid-liquid-gas relations within the soil. In this respect, compaction of the soil will increase the rate of heat transfer by increasing the number and closeness of contact points between adjacent particles.

Heat conductance in soils influences the rate at which a soil body is warmed or cooled as the external temperature changes. A dark-coloured soil may absorb heat readily but, if it has a characteristically low conductivity, the soil beneath the surface will not warm up rapidly. This may influence the speed with which deep-planted seeds germinate, an important factor in crop production in areas of cool climate and short growing season.

REVIEW QUESTIONS

1. Define very briefly the terms texture, structure, and consistence as they apply to soils.
2. What are the size limits of sand, silt, and clay under the International and U.S.D.A. systems of particle-size classification ?
3. What is meant by a disperse system, and explain how dispersion of a soil is accomplished ? What is a suspension ?
4. What is meant by the particle-size analysis or mechanical analysis of a soil ? How does this relate to the textural class of a soil ?
5. Of what value is it to know the textural class of a soil ? Name some of the other physical and chemical soil properties that are related to texture ?
6. What is meant by true density of soil particles and bulk density of a soil ?
7. What is the relationship between particle density and rate of settling when soil is suspended in water ?
8. Examination of a soil profile shows it to be lime free and relatively high in organic matter in the surface soil. The subsoil has an appreciable amount of lime. If you were interested in knowing whether or not the texture was uniform throughout the profile, would you remove lime and organic matter prior to mechanical analysis or not ? Why ?
9. What is meant by massive structure, macrostructure and microstructure ?
10. Name the major structural classes.
11. Name four factors which contribute to the formation of structure.
12. How can you calculate the percent pore space if you know the bulk density and particle density of a soil ?
13. Does a high bulk density mean high or low total pore space ?
14. What is meant by aggregation ?
15. Explain the difference between macro- or noncapillary and micro- or capillary pore spaces.
16. What is the significance of the term aggregate stability ?
17. What is the meaning of soil tilth ?
18. What is aeration ?

19. In determining soil consistence, what properties are observed in the wet soil ? the moist soil ? the dry soil ?
20. Which soil separate affects consistence to the greatest extent ?
21. The particle-size analysis of 5 soils shows the following percentage contents of four particle-size ranges. What is the textural class of each soil as determined by the International and U.S.D.A. methods of textural analysis ?

Size range of particles in mm.

Soil No.	2·0—·05	·05—·02	·02—·002	·002
1	30	27	21	22
2	5	14	32	49
3	25	28	27	20
4	19	24	27	30
5	42	14	33	11

22. Using equation 2-2, determine the time required for particles having diameters of ·05, ·02, and ·002 mm. to settle to a 10-cm. depth in a water column. For convenience in calculation, consider the viscosity and density of water to both equal 1, and the particle density to equal 2·65.

SELECTED REFERENCES

1. Alderfer, R. B. and Merkle, F. G., 'The measurement of structural stability and permeability and the influence of soil treatments on these properties,' *Soil Sci.* 51 : 201-212 (1941).

2. Bouyoucos, G. J., 'Directions for making mechanical analyses of soils by the hydrometer method,' *Soil Sci.* 42 : 225-228 (1936).

3. Chu, T. Y. and Davidson, D. T., 'Simplified air-jet dispersion apparatus for mechanical analysis of soils', *Iowa State* (*USA*) *College Eng. Rept. 21.* (*1953*). N.B. Rept. 21 contains a number of papers dealing with various phases of mechanical analysis of soils.

4. Kilmer, V. J. and Alexander, L. T., 'Methods of making mechanical analysis of soils', *Soil Sci.* 68 : 15-24 (1949).

5. McCalla, T. M., 'Water-drop method of determining stability of soil structure', *Soil Sci.* 58 : 116-122 (1944).

6. Piper, C. S., *Soil and Plant Analysis,* University of Adelaide, Adelaide, Australia (1950).

7 Russell, M. B., 'Methods of measuring soil structure and aeration', *Soil Sci.* 68 : 25-35 (1949).

8. Tyner, Edward H., 'The use of sodium metaphosphate for dispersion of soils for mechanical analysis', *Proc. Soil Sci. Soc. Amer.* 4 : 106-113 (1939).

CHAPTER THREE

MOISTURE RELATIONSHIPS AND AERATION IN SOILS

Two VERY important functions of the soil are those of providing water and air to plant roots. Each of these functions is highly dependent upon the pore-space relationships in the soil. Not only do the pores provide a storage place for water, they serve also as the means whereby both water and air may be replenished from external sources. This requires, then, that the soil contain a suitably high total volume of pore spaces for water storage, and, at the same time, that the pores be of a size suitable for the ready movement of either water or air. Since soils with fine pores have the greater capacity for retaining relative large volumes of water and large pores offer the least resistance to water and air movement, the soil ideally will contain pore spaces of both sizes. An excess of either usually results in the display of undesirable moisture or aeration relationships.

Water plays an extremely important role in a wide variety of chemical, physical, and biological processes in the soil. Its presence is essential for the solution of soil minerals and the transport of nutrients to roots and within the plant, and it serves as a medium for the various chemical reactions which take place both in the soil and in the plant. Harmful effects can result as the consequence of excess water, for it may create conditions of poor aeration or, if it passes through the soil, will leach valuable plant nutrients from the root zone. Moisture excesses also contribute to a high water table which, in addition to promoting conditions of poor aeration, may encourage the accumulation of toxic salts in arid-region soils. Thus, while water is absolutely essential to soil and plant functions, it may serve also as major detriment to plant growth. Correct water management, and that affecting aeration as well, requires a thorough understanding of soil characteristics as they influence water and air relationships.

Capillary Concepts Important to Soil Moisture Discussions

Capillarity is a phenomenon familiar to most everyone. This process is responsible for the drying action exhibited by cloth or blotting paper and the rise of oil in a wick. A classical example of capillary action is the lifting of water in a vertically oriented capillary tube, this analogy being used frequently as a basis for explaining the retention and movement of water in soils.

The weight of water held in a capillary pore increases with the height of rise above a free water surface. This weight, which results from the downward pull of gravity, is supported initially by a strong force of *adhesion* which attracts water to the wall of the pore and thereby pulls it upward from the free water surface below. Since the rising water also fills the centre of the pore, this same weight must be supported by the *cohesive* forces which hold together all of the water molecules forming the column. Specifically, cohesion among water molecules at the surface is most important because it must support the weight of the entire column. Cohesion at this point in the column is a manifestation of *surface-tensional forces*. These forces are of constant magnitude for a given temperature and are of considerably lower value than are adhesive forces responsible for lift. Thus, the height to which water will rise in a capillary pore is limited by the magnitude of surface-tensional forces, for they are capable of supporting only a certain, limited weight of water per unit of surface at the air-water interface.

Whenever an object is subjected to the influence of forces pulling in opposite directions, a state of *tension* is created therein; that is, there is a tendency for the object to be pulled apart. However, no separation occurs so long as cohesional forces exceed the applied pulling forces. Tension is created in capillary-held water by the opposing forces of gravity and capillary lift. The tension is at a maximum at the surface of a capillary water column where its effects are just balanced by surface-tensional forces which hold the water molecules together.

As may be noted in Fig. 3-1, the surface of water held in a capillary pore is slightly higher along the wall than it is

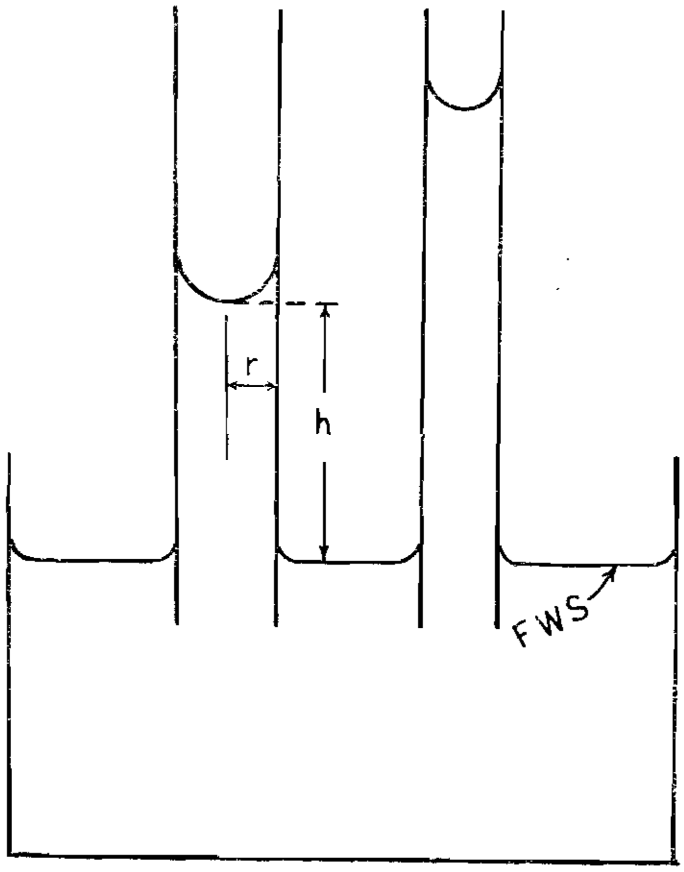

FIG. 3-1, The influence of pore radius upon the height of rise of water in a capillary tube. Shown are the radius (r) and height (h) factors which serve as a basis for calculating the weight of water forming a column above the free water surface (FWS). The weight, in turn, expresses the total downward pull supported by surface-tensional forces. Because of the concave shape of the surface film, total downward pull per unit of surface area is greater near the pore wall than it is in the centre of the pore.

in the centre of the pore. Thus, tension at the surface is somewhat greater around the circumference of the pore, and it is here that surface-tensional forces first become limiting to the height of rise. Since the magnitude of surface-tensional forces and the circumference of a pore can be determined,[1] one can compute the net lifting force in a given capillary system through use of the equation

$$\text{Total lift in dynes} = 2\pi rt, \tag{3-1}$$

where r is the radius of the pore in cm., and t is the surface tension in dynes per cm.

The downward pull exerted by a water column is proportional to its weight and can be expressed algebraically as

$$\text{Downward force in dynes} = \pi r^2 hdg, \tag{3-2}$$

where πr^2 is the cross sectional area of the column, h its height above a free water surface, d the density of water, and g the acceleration due to gravity. When the water in a pore stands at its maximum or equilibrium height, net upward and downward forces are equal as indicated by the equation

$$2\pi rt = \pi r^2 hdg. \tag{3-3}$$

Since the radius of a capillary tube can be determined, the height to which water will rise in it may be obtained by solving equation 3-3 explicitly for h ; thus,

$$h = \frac{2\pi rt}{\pi r^2 dg} = \frac{2t}{rdg}. \tag{3-4}$$

It can be seen from this equation that h and r are related inversely. Therefore, as is illustrated in Fig. 3-1, the smaller the radius of the tube the higher water will rise in it by capillarity.

Since all molecules in a capillary water column are subjected constantly to the opposing forces of lift and of gravity, they persist continuously in a state of tension. The magnitude

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[1] The surface tension of water at room temperature is approximately 72 dynes per cm., a dyne being defined as the force necessary to give a one-g. mass an acceleration of 1 cm. per sec. per sec.

of the tension at any point within the column will be proportional to the weight of water occurring below that point. Thus, the higher the level above a free water surface the greater will be the tension among water molecules. As suggested previously, maximum tension will be developed at the surface where the weight of the entire column must be supported by capillary forces.

One general and very important concept outlined in the foregoing discussion should be emphasized at this point. As shown in equation 3-2, the greater the height of a capillary column of water the greater is the downward pull which can be opposed successfully by limiting surface-tensional forces. This is to say, then, that the smaller the pore the greater is its capacity for retaining water against a force of removal. While gravity opposes the force of retention in the vertically oriented, uniform capillary tube, forces exerted by roots which result in the actual removal of water is the more common phenomenon observed in field soils. Thus, when filled pores in the soil are of large diameter, the force of retention for water is low and removal by plants will take place without particular difficulty. When the water is retained in small pores only, on the other hand, a much greater force must be applied to permit the transfer of water from these pores into the roots. If the force of retention by the soil becomes too great, roots may not obtain sufficient water to prevent wilting. This condition occurs when the moisture content of the soil is quite low. At this time, most of the water present forms only thin films over particle surfaces where it is held against removal by strong adhesional forces.

Tension in Soil Moisture

When the amount of water in a soil is insufficient to fill all pores, that present will fill those pores of relatively small diameter only and will occur also as films of varying thickness over particle surfaces which form the larger, unfilled pores. The thickness of these latter films determines the extent to which adhesional forces of attraction for water are satisfied.

Unsatisfied adhesional forces tend constantly to extract water from filled pores nearby. However, surface-tensional forces at the air-water interfaces prevent emptying of the pores, and, when unsatisfied adhesional forces and surface-tensional forces are the same, an equilibrium in moisture distribution between filled pores and films is obtained. Even so, these two forces oppose each other constantly and create a state of tension among the water molecules. As the moisture content of the soil decreases, pores of smaller diameter only remain filled and total capillary forces of retention at the air-water interface increases. At the same time, films in the unfilled pores become thinner and unsatisfied adhesional forces increase. Thus, with larger opposing forces at play as a result of the decreasing moisture content, tension among the water molecules increases. When the moisture content becomes so low that the residual water occurs only as thin surface films, unsatisfied adhesional forces are very large and the resultant moisture tension very high. Conversely, as the moisture content increases to where all pores become filled and the adsorptive forces for water are more nearly satisfied, tension drops to zero.

In poorly drained soils with a high water table, a substantial part of the system of interconnected pores above the free water surface is filled due to capillary rise. Here the development of tension corresponds to that in the uniform capillary system as illustrated in Fig. 3-1. Tensions under waterlogged conditions are generally low, particularly near the surface of the water table. Here, a minimum tension value of zero is observed at the free water surface.

Expressing Moisture Tension Values

The direct relationship between tension values and the height in a capillary pore above a free water surface allows use of the height measure as a simple, direct means of expressing moisture tension. Although any unit of linear measurement might be used for this purpose, designating tension values in terms of height expressed as cm. of water has been a very common practice. Applied to actual moisture conditions in the soil, tension values expressed in this manner normally range between zero, where the soil is completely wetted, to somewhere near 10 million cm. of water when the soil is in an air-dry state. Under the latter conditions, it is probable that only submicroscopic pore spaces within soil particles are filled ; the bulk of the small amount of water present will exist as thin films over particle surfaces.

Because expressing tension in dry soils in terms of cm. of water sometimes requires the manipulation of large, unwieldy numbers, Schofield (1935) proposed, as an alternative, the use of their logarithms to the base 10 and termed these derived values as the *pF* of soil moisture. This method of expression has been adopted by many soil scientists, particularly those in Europe. The relationship between tension values expressed as *pF* and as cm. of water is shown in Table 3-1.

Tension values may be expressed also in normal pressure terms. As noted in equation 3-2, the net downward force in a water column is equivalent to $\pi r^2 hdg$. This total force, exerted as downward pull against surface-tensional forces, is applied over the entire cross sectional area (πr^2) of the column. Thus, since pressure (P) is equal to force (F) per unit area (A), then,

$$\mathrm{P}=\frac{\mathrm{F}}{\mathrm{A}}=\frac{\pi r^2 hdg}{\pi r^2}=hdg. \qquad (3\text{-}5)$$

Therefore, pressure, like tension, increases in proportion to the height within a water column, and either pressure or height terms may be used to denote tension values.

The most popular pressure term used in soil moisture work has been the standard *atmosphere* (atm.) which is equal to 14·7 lbs. per sq. in. The bar, which has almost the same value as the standard atmosphere, has come into rather frequent use in recent years. In practical considerations, either an atm. or a bar of pressure is assumed to be essentially equivalent to 1000 cm. of water,[1] the relationship among these values being shown in Table 3-1.

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[1] More precisely, one atm. pressure is exerted by a water column 1033 cm. (33ft.) high.

Table 3-1. The relationship among various methods of expressing moisture tension values.

pF	Cm. of water or millibars	Atm. or bars
0	1	·001
1	10	·01
2	100	·1
3	1,000	1·0
4	10,000	10·0
5	100,000	100.0
6	1,000,000	1,000·0
7	10,000,000	10,000·0

The pressure[1] in any liquid system at rest, whether capillary or noncapillary, can be calculated through use of equation 3-5. If we assume *d* and *g* in the equation to be constants with a combined value of *k*, then the equation becomes simply $P = kh$. As noted above, *h refers to the height above a free water surface when dealing with a capillary system.* In a noncapillary or open body of water, on the other hand, *h* refers to the depth below the free water surface.

As illustrated in Fig. 3-2, the hydrostatic pressure at a free water surface is zero. The pressure below this surface is positive and increases with increasing depth. If the depth is measured in cm., then it is convenient to express the corresponding pressure as cm. of water. Atmospheres of pressure can be used also, of course. An identical relationship between values of *P* and *h* occurs in a capillary water column above the free water surface. However, since the water is lifted by capillary forces in opposition to gravity, the pressure developed

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[1] Pressure in a liquid at rest is referred to normally as *hydrostatic* pressure. It is sometimes called *hydraulic* pressure, although the latter term usually denotes pressure within or exerted by a liquid in motion.

is negative; that is, a state of tension is produced. The greater the distance above the free water surface the greater the tension or negative pressure as is noted also in Fig. 3-2.

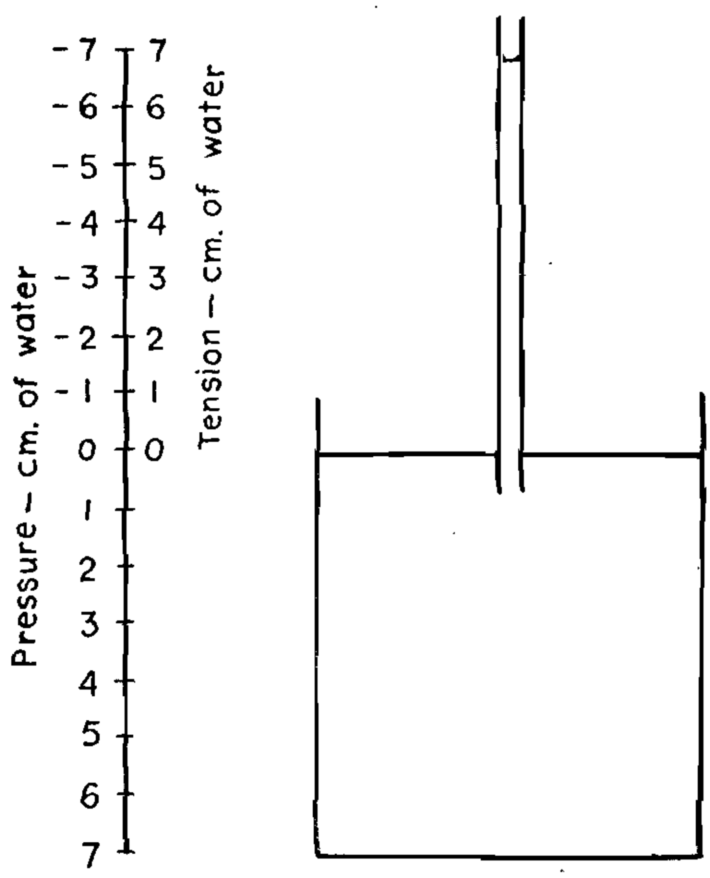

FIG. 3-2. Pressure and tension relationships in a continuous capillary-noncapillary system where pressure and tension are expressed as cm. of water. Pressure values increase positively below the free water surface, whereas they increase negatively above this reference level. Tension is equivalent to negative pressure although of opposite algebraic sign.

Moisture tension values in most agricultural soils supporting growing plants vary between zero and 15 atm. As mentioned previously, zero tension corresponds to the moisture level where all soil pores are filled with water. In well-drained soils, this condition will occur only rarely and then only during a brief interval when the soil is being wetted by rain or irrigation. The 15-atm. value corresponds to the average maximum tension normally developed as a consequence to the removal of water from the soil by plants. By the time this tension value is attained, most plants are being seriously deprived of water and are wilting or at least very close to this stage.

Expressing the Soil Moisture Content

Whereas the force with which water is held in a soil is expressed by tension values, the quantity of water present under a specified tension is denoted by a percent moisture content. Percentage values in moisture work are referred to the oven-dry weight of soil since this provides a constant point of reference. Thus, a soil sample which weighs 120 g. in the moist state and 100g. after oven-drying[1] will contain 20 per cent moisture. The range usually encountered is from 2-10 per cent in air-dry soils to 50-60 per cent in clayey soils which have all pore spaces filled. Highly organic soils, having both a low mean particle density and a low bulk density, may retain water in quantities equivalent to several times their oven-dry weight.

Characterizing Moisture-Retention Properties in Soils

Differences in moisture-storage and moisture-delivery capabilities among soils is related to textural variation for the most part. This fact is illustrated in Fig. 3-3 by the *characteristic moisture curves* for two soils of different texture. Construction of these curves involves the plotting of corresponding values of tension and moisture content, usually throughout

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[1] Drying is accomplished at 105-110° C. to an essentially constant weight. Normally, the process is assumed to be complete after 48 hrs. Once a drying cycle is well underway, fresh, wet samples should not be placed in the oven with those being dried. Otherwise the humidity of the oven is raised unduly and the required drying time prolonged.

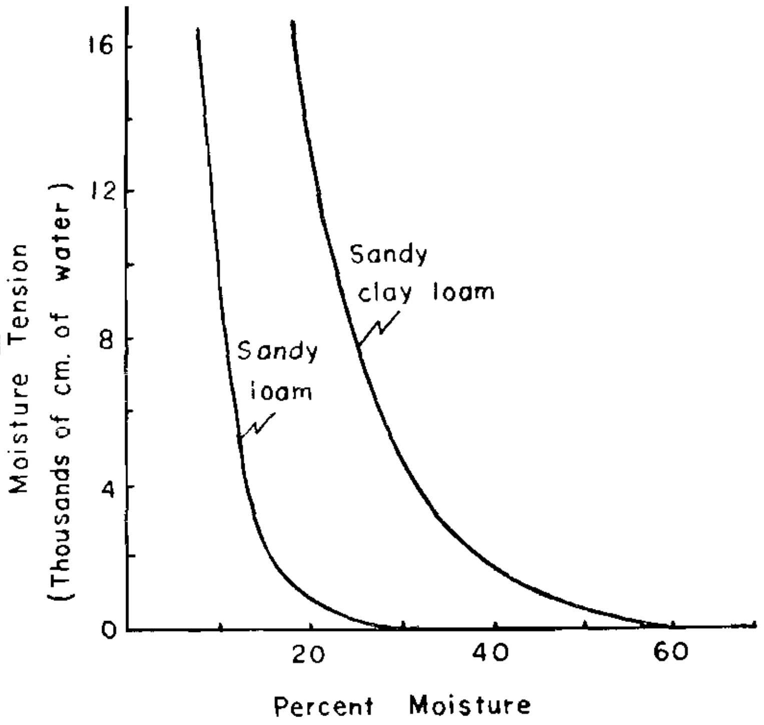

FIG. 3-3. Characteristic moisture-retention curves for soils of two textural classes.

the 0-15 atm. range. It can be ascertained at a glance from these curves that the finer-textured soil has the higher total capacity for holding water. Further, calculating the quantity of water lost as the tension is increased to 15 atm. indicates that this same soil will release a greater amount of water before the point is reached where wilting occurs. Thus, characteristic moisture curves provide a basis for the ready comparison among soils with respect to their moisture-supplying properties. Data for such curves are obtained virtually always on soil samples taken into the laboratory. The procedure involves the measurement of the amount of moisture held by the soil at different tensions developed by subjecting the samples to known forces of moisture removal.

The Field Measurement of Moisture Tension

One of the most useful instruments for measuring tension in field soils is the *tensiometer* (Richards and Gardner, 1936). The apparatus, which is illustrated in Fig. 3-4, consists of a porous ceramic cup that may be buried in the soil at any pre-selected depth and which is connected to a mercury manometer or vacuum gauge through a water column. When water from the interior of the cup passes through the pores and makes contact with the soil, it creates a continuous liquid system extending through the soil and into the tensiometer. Suction created in the system by the surrounding soil is reflected by a rise in the mercury column. The height of the mercury column, denoted by h in Fig. 3-4, provides a continuous measure of the soil moisture tension, therefore.

Ceramic materials used for tensiometer cups contain pores of an average size which permits the relatively rapid transfer of water but which also limits the measurement of tension to about 850 cm. of water only (Richards, 1949). Thus, tensiometers will give reliable readings if their use is restricted to soils kept fairly moist. If, as the soil dries, tension becomes so great that even the small pores of the cup are emptied by movement of water into the dry soil, air is drawn from the soil and into the interior of the tensiometer. Once air is trapped inside, the reliability of the tensiometer is destroyed. This is because air, unlike water, will expand or contract with changes in pressure and will not allow the true moisture tension to be shown by the manometer or vacuum gauge.

The moisture content of the soil may be measured indirectly by electrical means. Devices commonly employed are gypsum, nylon, or fiberglass blocks which are placed in the soil and which change in moisture content as the moisture level of the surrounding soil varies (Bouyoucos and Mick, 1940, 1947, 1948 ; Bouyoucos, 1954). The blocks contain two electrodes, and electricity will flow between them under the influence of an applied electrical potential in an amount dependent upon the moisture content of the blocks. Electrical characteristics vary from block to block. This requires that the blocks be calibrated individually by placing them in a

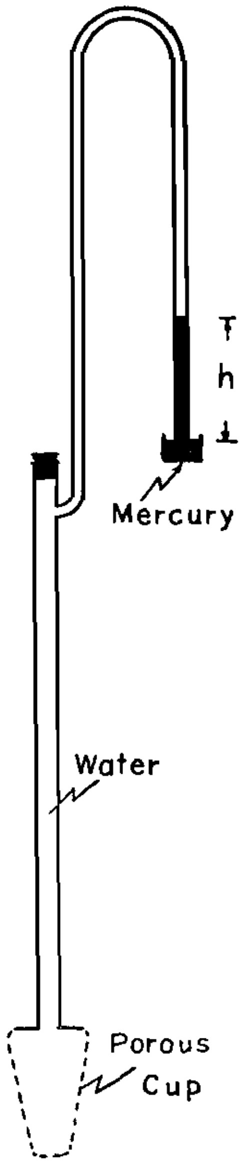

FIG. 3-4. Simplified diagram of a tensiometer.

soil and measuring their resistance at different, known moisture contents. Details on calibration are given by Kelley (1944).

Unlike the tensiometer, resistance blocks are most useful where tensions are high. Even then, changing salt concentration of the soil solution will affect the resistance of the blocks and thereby will limit their usefulness in many instances. Largely because of the technical problems involved in the use of tension- and moisture-measuring devices and because of their relatively high cost, they are employed primarily by the research scientist.

Moisture Stress

Moisture tension measurements take into consideration only the physical forces of attraction between soil solids and water. Where plants are involved, removal of moisture from the soil requires over-coming not only these physical forces but, also, those due to osmotic pressures caused by dissolved salts. The summation of all forces which operate to prevent removal of water by plants is called *moisture stress.* If the solute concentration is negligible, the moisture stress is equal in magnitude to the moisture tension.

THE MOVEMENT OF SOIL MOISTURE

For convenience in the discussion of water movement, the moisture level in the soil is normally described by one of the two generalized terms. These are *saturation*, where all pores are filled, and *unsaturation*, where, obviously, the volume of soil voids is shared by varying proportions of water and air. In a strict sense, saturation refers only to a single, maximum moisture content. Practically, however, this absolute moisture level is seldom observed in field soils and, therefore, the term is used also to denote near saturation, a condition arising from the presence of trapped gases in a limited number of pores within an otherwise saturated soil mass.

Forces Causing Moisture Flow

Qualitatively, the rate of transfer of water through soil pores is a function of the magnitude of the forces causing movement and the conductive capacity of the soil. Both factors vary with changes in the soil moisture level and its pattern of distribution within the soil body.

Forces responsible for moisture movement in soil have their origin in four primary sources. These include gravity, which provides a force of constant magnitude, and variation from one point to another within the soil body in either moisture tension, vapour pressure, or osmotic pressure. Since all may exert an influence at the same time, their separate effects are additive.

Differences in osmotic pressure and vapour pressure seldom cause effective moisture flow within a soil mass. Osmotic effects are of consequence only in soils seriously troubled with accumulating salts. While vapour-pressure differences can cause the diffusion of water vapour, such variations within the soil normally are too small to be of particular value, especially in contributing to plant growth. As a consequence, practical considerations of moisture flow may be concerned only with that caused by gravitational influences and moisture tension differentials.

In unsaturated soil, water tends to be distributed uniformly so that capillary forces everywhere are satisfied to the same extent. This state of balance, characterized by uniformity in tension throughout the soil, is disturbed if water is removed from or added to one point but not to the remainder of the soil mass. As a consequence, movement will take place and continue until an equilibrium state is established once again. Typical of any reaction which is completed when equilibrium is attained, moisture movement is most rapid when the degree of unbalance is large and slows gradually as equilibrium conditions are approached. Practically, however, since water is either added to or removed almost continuously from the soil, an equilibrium state is virtually never attained.

The degree of moisture unbalance in unsaturated soil is denoted by the *tension gradient.* This gradient is expressed as the difference in tension between two points divided by the distance between these two points. The steeper the gradient the greater is the motivating force influencing flow.

Gravity as a factor in moisture movement has little influence in unsaturated soils holding moisture at high tensions. As saturation is approached, however, the influence of gravity becomes more important and, at saturation, provides the only significant force for movement. The magnitude of the force causing saturated flow is expressed in terms of the *hydraulic head*, or the difference in elevation between points along a free water surface. The hydraulic head divided by the distance through which its influence is felt is designated as the *hydraulic gradient*. As the tension gradient affects unsaturated flow, movement of water between two points in a saturated soil increases with increasing hydraulic gradient.

Gravitational effects operate in a downward direction only. Lateral flow due to gravity occurs, of course, but only where such movement results also in the transfer of water from higher to lower elevations. As opposed to this fact, moisture flow along a tension gradient may occur in any direction and will be from points of low tension toward points of high tension. Thus, movement may be into drier soil zones around roots where active absorption is taking place ; or it may be in an upward direction toward the soil surface where a high tension is created by drying through evaporation. Downward movement in response to a difference in moisture tension occurs during the normal process of soil wetting by rain or irrigation water. In this instance, movement is from the surface, where tension is near zero due to the abundance of water, toward zones of higher tension in the drier soil beneath. Gravity contributes to this movement but its effects are significant only where a saturated condition is promoted temporarily during the application of water.

Permeability as a Factor in Moisture Flow

Tension gradients or hydraulic gradients do not function alone in determining the rate at which water will flow through a soil because different soils have different abilities for transferring water from one point to another under the influence of a given gradient. The ability to allow moisture transfer by a soil is referred to as its *permeability* or *conductive capacity* and may be characterized under both saturated and unsaturated conditions. It is a function of the total volume of conducting pore spaces, their mean size, and the degree to which these pores are filled with water at the time they are serving as the conducting medium.

The dependence of flow rate upon moisture content is related to changing total frictional losses with varying moisture content. The rate of flow of water through a channel is slowest along the wall where frictional loss is the greatest. The minute size of any soil pore thus imposes restriction on moisture flow in that all molecules must move in relatively close proximity to pore walls. Even so, whether or not water can pass through the centre of the pore, a condition possible only when the pore is filled, will have a marked influence on the conductive capacity of that pore under a given motivating force. Thus, as the moisture content of a pore is reduced and the proportion of water flowing near the wall increases, frictional losses per unit quantity of water transferred through the pore must increase. As a consequence, the unsaturated permeability of soil decreases with decreasing moisture content.

Pore spaces in coarse-textured soil material are predominantly large. When filled, movement of water through these pores suffers less frictional loss than will movement through the smaller pores dominant in fine-textured soils. Therefore, so long as saturation is maintained, the volume of water flowing in these coarser soils will be greater than it will be in finer-textured materials when the motivating force in each is the same. With increasing tension and degree of unsaturation, however, the large pores of coarse-textured soils are soon emptied. The pores which do remain filled with water are small and limited in number and may not provide an interconnected system through which the water may flow in an uninterrupted manner. Therefore, the conducting capacity of the soil will drop off sharply once the continuous system of water-filled pores is broken. This condition can occur at tensions of less than 0·1 to 0·2 atm. in sandy soils.

Fine-textured soils contain a large proportion of small pores which retain water under relatively high tension. As a consequence, the unsaturated permeability of these soils does not drop off as rapidly with increasing tension as it will in coarser materials. At the same time, however, the small pore size prevents rapid transfer of water even at saturation. Summarily, then, the conductivity of coarse-textured materials is greater than that of finer materials at saturation but the reverse is true under conditions of high soil moisture tension. The generalized relationship between permeability and increasing tension as affected by variation in soil texture is illustrated in Fig. 3-5.

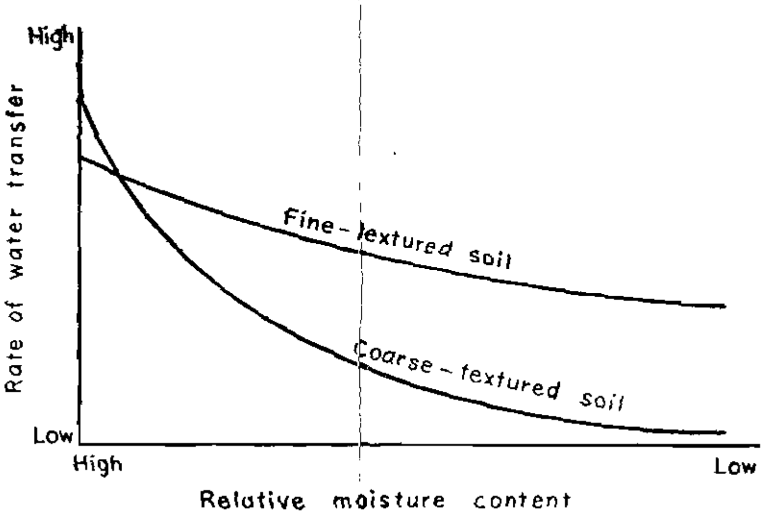

FIG. 3-5. The effect of soil texture on moisture conduction with variation in soil moisture content. At high moisture levels, the large pores remain filled and conduct water most rapidly. Finer-textured soils conduct the greatest amount of water at lower moisture levels because they contain a high proportion of small pores which remain filled for longer periods as the moisture tension increases.

SOIL WETTING

Because the adsorptive forces for water in a dry soil are largely unsatisfied, the application of free water under low tension at the surface results in the relatively rapid entry into the soil. The drier the soil the greater will be the difference in tension between the water held originally in the soil and that applied at the surface. Thus, the initial moisture content will be a determining factor in the rate at which the water moves into and through the soil pores. Obviously, the ability of the pores to conduct the water, a property dictated by pore size, is important also in controlling the rate of water movement from the surface to the unwetted portion below. Further, as the wetting process continues and the thickness of the wetted zone increases, water continually must move through a greater distance to reach the dry soil. Thus, as the depth of wetting increases, the tension gradient decreases and the net resistance to flow increases, both effects resulting in the gradual reduction in the rate of movement from surface to subsoil layers.

When the application of water at the surface ceases, water will continue to move toward the zone of higher tension below. However, the only source of water for movement into the dry soil below now is the filled or nearly filled pores near the surface. Consequently, as these pores drain, the tension there increases and tends to oppose continued moisture removal. Further, as the moisture content of the wetted zone decreases through loss of water to the subsoil, the rate of water conduction also is reduced. Either of these factors will result in a lowering of the rate of flow, and when their influence becomes sufficiently large, the drainage rate becomes negligibly small. Because of this, a moisture content is attained eventually which remains relatively unchanged in time, and therefore, corresponds to the maximum amount of water the soil will hold under the specified conditions of drainage. It is referred to as the *field capacity* of the soil and normally is considered to represent the upper limit of available water to plants. Each time the dry soil is wetted, this essentially constant moisture content will be re-established. Under usual conditions in the field, water removal by vigorously growing plants easily will exceed the loss by further drainage after the level of field capacity has been attained.

Soil Texture and Field Capacity Values

The concept of field capacity is not equally applicable to all soils. Those of a sandy nature, for instance, become nonconducting after a relatively short period of drainage and also at fairly definite moisture levels. Because drainage is inherently slow in fine-textured soils, the point in time where the drainage become negligible is not so precisely defined. This is illustrated in Fig. 3-6 where reduction in the moisture content and increases in moisture tension in the surface layers of thoroughly wetted fine- and coarse-textured soils are plotted against time. It will be noted that drainage in the sandy soil is complete for all practical purposes after about two days. The resultant moisture content of about 12 per cent represents the field capacity in this soil. No such point of relatively constant moisture content is observed for the silty clay loam. It may be noted also in the lower diagram of Fig. 3-6, that the initial increase in tension in the sandy soil is very rapid but does not increase above the relatively low value of about 0·1 atm. The tension in the silty clay loam, on the other hand, continues to increase due to continued water loss even though after 8 days the tension has risen to above 0·3 atm.

Soil Stratification and the Field Capacity

The above discussion concerning wetting and the establishment of the field capacity has application primarily to soil profiles of relatively uniform texture throughout. Many soils, on the other hand, are not uniform but consist of intermixed layers of different texture. Thus, in wetting, water must move at times between layers having different effective pore sizes and, therefore, different capabilities for attracting water by capillarity.

As water moves downward through a dry soil, the most advanced portion of the wetted zone, defined as the *wetting front*, has a comparatively low moisture content. Thus, if a wetting front advances through a fine-textured soil material to an underlying layer of coarse material containing predominantly large pores, transfer of water will not take place to any extent so long as the tension at the junction of the two soil layers remains high. This is because the large, empty pores below cannot extract water by capillarity from the smaller ones above until these small pores become essentially filled and the moisture tension therein is decreased to a low

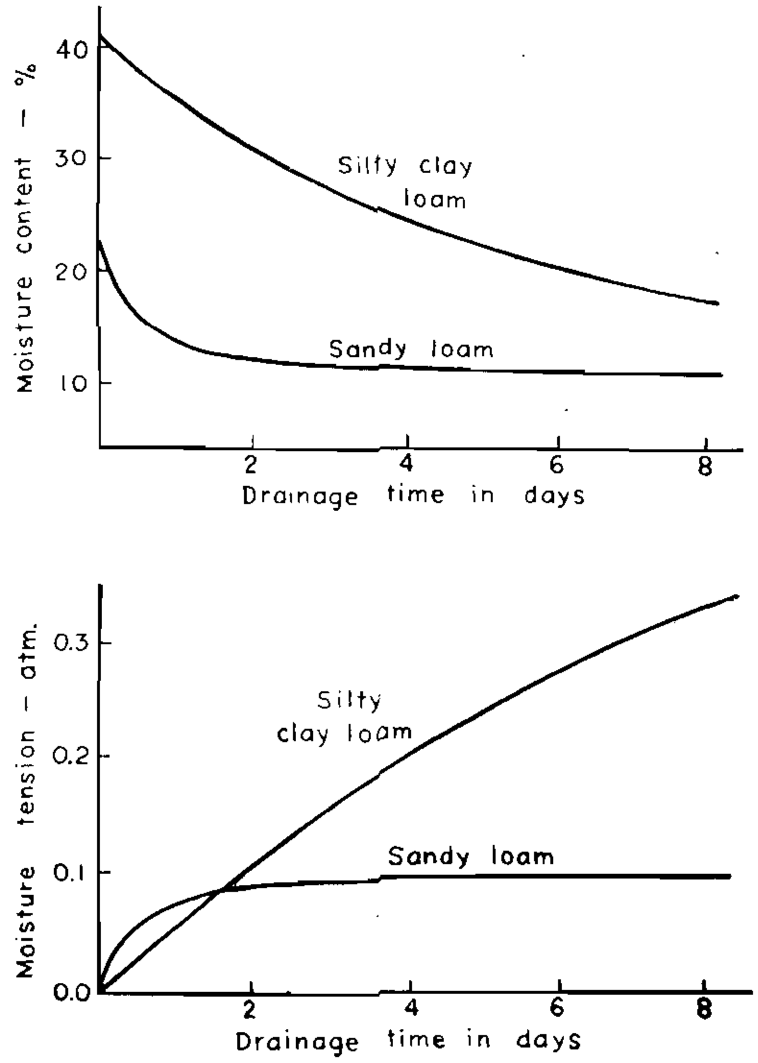

FIG. 3-6. The moisture content and moisture tension values for the surface layer of soils of widely different texture as affected by length of drainage time. The effect of moisture loss by evaporation or plant removal, though present under field conditions, is not considered in this idealized presentation.

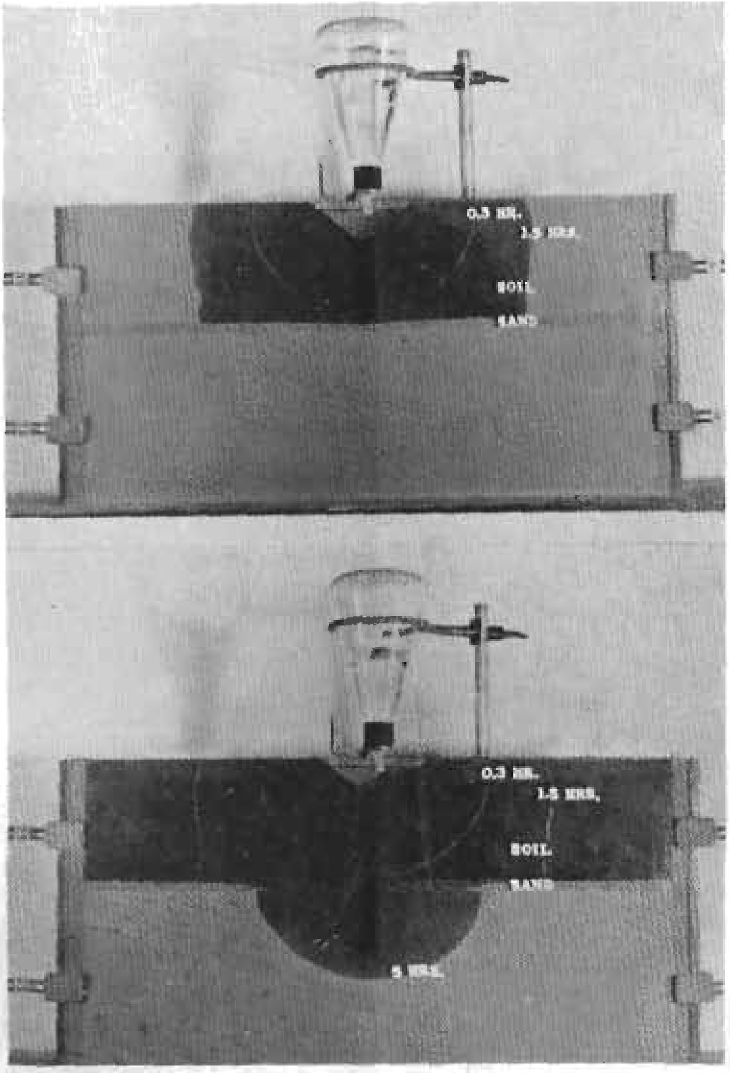

FIG. 3-7. The wetting of a fine-textured soil material containing a layer of sand. The line marked 0·3 hr. shows the semi-circular shape of the wetting front when it reached the sand layer. Only after about 1·5 hrs., when due to a gradual increase in the moisture content the tension at the wetting front became very low, was movement into and through the sand layer possible. (Photograph by courtesy of Dr. W. H. Gardner, Washington State University, Pullman, Washington.)

level. As a result of this, the wetting front will remain stationary at the boundary until more water moves in from behind to reduce the moisture tension there. This sequence of events is illustrated vividly in Fig. 3-7 where downward movement is impeded temporarily when the wetting front passing through a fine-textured soil material contacts a layer of sand. Once movement into the sand layer commences, it will continue without further interference.

A sand layer, because of its inability to extract water unless the tension is very low, will interfere also with the eventual drainage from the overlying layer. This is due to the fact that only a slight increase in the tension above the sand layer will again prevent movement into this coarser material. As a result, drainage will stop while the finer material is still very nearly saturated. A rather constant and relatively high moisture content corresponding to a field capacity value would thereby be established for the overlying, fine-textured material.

As opposed to the above example, a layer of fine-textured material in an otherwise coarse soil also impedes the downward movement of water during wetting and during subsequent drainage. However, the effect is the consequence of the relatively low permeability of the finer material even if the moisture level is near saturation. Water under high tension at the wetting front will move into the small pores of the fine material readily, but the transfer out into the coarse material below again will occur only if the tension is comparatively low. Thus, a clay layer in a normally well-drained sandy soil may, through impeded saturated flow, maintain the profile in a relatively moist state. If, because of this effect, excess water is retained in the soil above a slowly permeable subsoil layer for a prolonged period, a condition described as a *perched water table* results.

A Practical Consideration of Field Capacity Values

Realization that the field capacity denotes the maximum amount of water a soil will store against free drainage makes it a very important and useful concept. If negligible water is lost from the root zone once the field capacity is attained, then this value marks the upper limit of available water to plants. Thus, the field capacity serves as a means of characterizing partially the moisture-supplying ability of a soil. A limitation to this arises where one is unable to determine precisely the field capacity in slowly permeable, fine-textured soils.

Where land is cropped to vigorously growing plants, water removal by roots will take place while the water is being applied and during the drainage period which follows. Thus, even before drainage is completed, water will have been used by plants which cannot be related to a field capacity percentage. However, during the initial period following wetting, loss of water from the root zone is predominantly by the drainage process rather than through root absorption. It is not until the drainage rate becomes very slow that plant uptake becomes the principal agent for removal. Therefore, under these circumstances, drainage will proceed to the point where further water loss by this process is no greater than the loss caused by root absorption. The moisture percentage at this time is considered to correspond to a practical and useable field capacity value which approximates the upper limit of available water to plants.

According to Viehmeyer and Hendrickson (1949), the field capacity is obtained in a thoroughly wetted field soil after a drainage period of about 2 to 3 days. This arbitrary standard is used most frequently as the basis for the experimental determination of field capacity values. For soils of medium texture, a tension of about one-third atm. will have been created as a result of 2 to 3 days' drainage (Coleman, 1947). In coarse-textured soils, where the large pores become relatively nonconducting at low tensions, the value after this drainage period may be no greater than about 0·15 atm. (Richards, 1949).

INFILTRATION

The movement of water into a soil during an irrigation or rain is referred to as *infiltration*, and *infiltration rates* are commonly measured to determine this capacity in a soil. Initially, at least, the infiltration rate is affected by the original moisture content of the soil and its texture and physical state of the surface.

Infiltration can be controlled somewhat by tillage and soil management. As shown in Table 3-2, for instance, the rate of infiltration of water into well-granulated sod land is maintained at an appreciably higher level over a longer period of time than it is in cultivated maize land having a plow layer in a somewhat less stable state of aggregation. This may be of particular significance where the quantity of rain water absorbed by the soil is a function of the infiltration rate. Obviously, water that is not absorbed will drain away over the surface and will not become available for plant use.

Table 3-2. The effect of crop (cultural practice) on infiltration of water into an otherwise uniform soil.

Duration of Wetting (Hours)	Infiltration Rate (in./hr.)	
	Maize	Grass sod
0	1·10	1·10
1	·45	·80
2	·28	·50
3	·20	·37
5	·18	·30

As granulation affects infiltration rates, so will soil texture. In either case, the rate is largely a function of the conductive capacity of the soil where water is being applied continuously and, therefore, where transmission through the surface layer may be within pores that are essentially saturated. Subsoil layers which restrict the transfer of water also will, in time, affect infiltration rates when these layers are reached by the moving water.

Textural and layering effects are illustrated by the following tabulated comparison. Shown are infiltration rates obtained after the soil material in question has been subjected to an initial wetting of sufficient duration so that the wetting front will have reached any restrictive layer in the subsoil:

Soil Material Group	Infiltration Rate
	in./hr.
Coarse to medium texture	
Deep, well drained	·30—·40
Shallow over restrictive layer	·05—·15
Fine texture (Clayey)	
Well aggregated	·05—·10
Dense, dispersed	Negligible

The above values are, of course, subject to a great degree of variation depending upon both the textural and structural configuration not only of the main soil body but also of the subsoil layers present.

THE RISE OF WATER FROM A WATER TABLE

A classic example of unsaturated flow is the rise of water from a water table. This process can be observed in a simple demonstration consisting of columns of soil placed in contact with a free water surface. If several columns are used and contain soils of different textures, an observation as to the effect of the texture on rate and extent of rise can be made. It will be noted that, because of differences in pore size, the initial rate of upward flow in sand is more rapid than in fine-textured soils. However, significant movement is maintained for longer periods of time in the latter fine soil materials. These differences, which have certain similarities to the drainage patterns illustrated in Fig. 3-6, are shown in idealized graphic form in Fig. 3-8. Here soils of three textural grades are represented; sand, loam, and clay. The initial rapid rate of rise in the sandy soil results from the lower resistance to flow in the large pores. The same large pores, on the other hand, developing but limited capillary lift, also restrict the

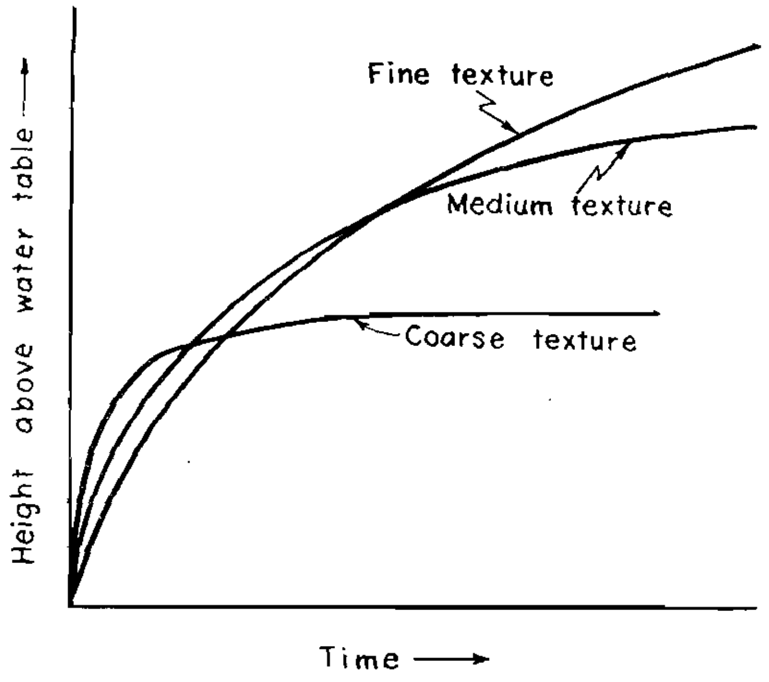

FIG. 3-8. The theoretical rise of water from a water table as a function of soil texture. The coarser the texture the more rapid the initial rate of rise. The finer the texture the higher the water will ascend before movement becomes negligibly slow.

height of rise. The maximum height to which water will rise is greater in the soil with the finer pore spaces, therefore, and this height is noted to vary inversely to the rate of flow during ascention.

The tension with which water is held in soil pores increases with the distance above a water table. Thus, water that has risen to appreciable distances by capillarity will occur only in the finer soil pores and its movement through these pores is relatively slow. For this reason, plants must approach reasonably close to the surface of a water table before the rate of water movement to them is adequate for normal plant functions. The rooting habits of plants is all important, therefore, in determining whether or not they can survive if the principal source of water is from a water table.

DEEP PERCULATION

When more water infiltrates into a soil than can be stored in the root zone, the excess will move into the substrata below. Here it may be added to a permanent water table or it may create a condition of temporary waterlogging until natural drainage removes the excess. The loss of water through deep percolation increases with increasing rainfall or with increasing application of irrigation water. Any treatment which raises the infiltration rate of a soil or reduces the amount of water lost as runoff will increase the percolation loss. Deep percolation is negligible in areas of low rainfall or where careful irrigation practices are followed.

Most soil profiles give visible evidence of the extent of percolation to which they have been subjected for long periods. For instance, soils of humid regions which have been influenced by considerable leaching are generally free of soluble salts. Many of their less resistant mineral components will have been dissolved and carried away in drainage water. Such soils, when placed under cultivation, suffer equally intense if not greater leaching loss unless removal of natural plant cover has resulted in an increase in the runoff of rain water.

Actively growing plants tend to reduce percolation loss through utilization of the water stored in the root zone. Thus, under comparable periods of precipitation, land well covered with plants may be expected to suffer less percolation loss than would occur were it under fallow. Countering this effect, however, is the fact that the cover provided by abundant vegetation may reduce the total runoff from sloping land and thereby increase total infiltration.

As a general rule in the subcontinent, most precipitation comes during the monsoon season and at a time when many plants are growing with relative vigour. The tendency would be, then, for these plants to lessen the intensity of percolation loss. Nonetheless, in Bengal, throughout the Himalayan submontane tracts, and elsewhere in regions of high monsoon rainfall, even the presence of abundant vegetation will have but an insignificant influence on total percolation loss. This is quite apparent from the highly leached appearance of the upland soils in these areas.

Nutrient ions absorbed by plant roots will not be subject to loss from the soil via the percolation stream. Therefore, a dense plant cover with a well-developed root system can contribute much toward the conservation of these nutrient elements. The effect involving nitrate-nitrogen is particularly noticeable since this ion is readily leached. Growing plants conserve nitrate because it is subject to rapid absorption by roots and in relatively large amounts.

While little control over percolation losses in humid regions is possible beyond the continual maintenance of a plant cover, this problem in irrigated lands can be held to a minimum by judicious irrigation practices. Often, where irrigation water is in abundant supply, the general tendency is to apply more water than is needed. This results not only in the loss of soluble plant nutrients but also in the intensification of waterlogging so common in irrigated areas. Problems of aeration, therefore, would increase through overirrigation. Further, the accumulation of salts in toxic quantities in the soil as a result of the capillary rise of salt-bearing water from the water table would be encouraged also.

Where losses due to percolation are extensive, it is imperative that careful management practices be followed to ensure maximum conservation of essential plant nutrients present in the soil. This is particularly true where these nutrients are supplied in expensive and scarce fertilizers or manure. Practices such as the use of forms of plant nutrients least subject to leaching, or their application to the soil in a way which will reduce leaching loss to a minimum should be followed wherever possible. These aspects of soil management are considered in greater detail in later chapters.

MOISTURE UTILIZATION BY PLANTS

Plants differ in their ability to utilize stored moisture particularly from the standpoint of depth of rooting. Further, when subjected to different environmental regimes, a single plant species may display a high degree of diversity in utilizing soil water effectively. It is important to recognize, therefore, that the soil alone will not determine the total quantity of water made available to plants during a particular period.

The Role of Water in the Plant

A single, continuous liquid system is formed by water which moves from storage points in the soil and into and through the roots, stems, and leaves of plants. A change in the moisture status at any one point will be reflected subsequently throughout the whole of the system. As in the soil alone, the effect results from the tendency for water to distribute itself uniformly in an effort to equalize differences in tension within the system. Therefore, loss of moisture from the plant increases moisture stress within the plants cells and encourages a more rapid rate of replenishment from the soil. Conversely, increases in the soil moisture tension as the moisture content of the soil decreases may be expected to reduce the rate of flow into the plant. Regardless of the cause of variation in the rate of flow, it must be maintained at an adequate level to supply the needs of the plant for water at all times.

Water enters into numerous processes within the plant. It serves as a solvent in which organic and inorganic components can react or can be transported to the various plant parts. It is used directly in photosynthesis where, with carbon, it becomes a component of many organic compounds such as sugars, starches, and cellulose. Water has an important role also in limiting variation in temperature of the plant, both by virtue of its high capacity for absorbing heat without marked temperature change or through the cooling effect accompanying evaporation. One of the main functions of water is to keep plant cells in a fully expanded or *turgid* condition. Loss of turgidity, recognized visually as wilting, interferes noticeably with physiological processes responsible for growth and reproduction.

In spite of its essential nature and frequent short supply in the soil, the use of water by plants is a relatively inefficient process. Normally, less than 0·5% of all the water taken up by a plant is retained in plant tissue. Most of the water is lost to the atmosphere through stomatal openings which occur in highest concentration on leaf surfaces. The process of water loss by this means is known as *transpiration.* Basically, it is an evaporation process and is influenced by such atmospheric factors as temperature, humidity, and wind, just as is evaporation.

Forces causing transpirational loss are often of very great magnitude. In terms of tension, the pull created by dry air on leaf-held water may exceed 1000 atm. Although plants normally have some morphological features which provide for reducing the rate of transpiration at times of great stress, the main defense rests with the continual flow of water at an adequate rate into roots. Obviously, then, where problems in supplying moisture to plants are common, it is of utmost importance to recognize those characteristics of soil which contribute materially to water storage and delivery to plants.

FACTORS AFFECTING THE SUPPLY OF AVAILABLE WATER IN SOILS

While physical characteristics (*i.e.*, texture) of a soil contribute materially to the storage and supply of water to plants, they are not the sole factors controlling these functions important to plant growth. The quantity of water infiltrating into a soil and efficiency of utilization of stored water by plants also play significant roles in determining the total delivery of water to plant roots. All of these factors must be considered in any evaluation of the potential moisture-supplying power of a given soil under a specific set of environmental and cropping conditions.

THE LOWER LIMIT OF AVAILABLE WATER—THE WILTING POINT

Water will be subject to absorption by plants if it can contact root surfaces while at a tension below that of the root-held water. Soil moisture tension, then, can serve as an index to the availability of water, for it relates to the force of retention by the soil. Further, since tension is related inversely to the unsaturated permeability of a soil, it will indicate also the relative difficulty encountered in moving water from points of storage up to the roots. The rate of delivery to the root appears to be as important a factor in determining moisture availability as is the force of retention exerted by the soil.

Water absorption continues even as a plant wilts. However, the rate of absorption at this time is inadequate to keep pace with loss through transpiration. The moisture content of a soil at which this phenomenon occurs defines the *wilting point,* and it may be as low as 2 to 3 per cent in sands and as high as 15 to 20 per cent in clayey soils.

Numerous studies have shown that most succulent plants, when grown on a single soil, will reduce the moisture content of that soil to the same point at wilting. Thus, a wide variety of plants might be used for determining the wilting percentage of soils. The standard procedure, however, involves growing sunflowers (*Helianthus annuus*) in a container filled with the soil and protected against loss of water through evaporation from the soil surface. The moisture content is determined when the plants wilt permanently. Logically, the moisture will not be distributed uniformly within the soil mass at wilting since the degree of removal will vary with the distance from the root surfaces. Greatest uniformity in moisture distribution within the soil and maximum removal will be attained if the wilting point is approached slowly during the test, therefore.

While the moisture percentage at wilting varies widely among soils of different textures, the moisture tension at the wilting point differs to but a limited extent from soil to soil. It may vary between 12 and 18 atm., but it is so often near 15 atm. that this value is considered as an almost universal expression of the wilting point in practical considerations.

It must be emphasized, however, that no single value expresses the tension precisely at all points within the soil mass at wilting. Indeed, tensions at the most distant points from root surfaces will be somewhat below the value for the whole soil, while, at root surfaces, they may be as high as 50 to 60 atm. at wilting.

Roots are actually capable of reducing the moisture content of the soil to very low levels. They do so by creating relatively steep tension gradients in the soil. This is possible because tensions developed at root surfaces are high and they must operate through relatively short distances in moving the stored water. Again, however, as the moisture content and the unsaturated permeability are reduced through absorption by roots, the rate of transfer under a given tension gradient decreases. Eventually the rate becomes inadequate to meet plant needs and wilting follows. This point is reached in sandy soils at lower tensions than in finer-textured soils because the large pores they contain are emptied and become nonconducting at lower tension values.

If, in a sandy soil, the instantaneous demand for water is great, wilting may occur even though the tension is no higher than about 12 atm. Plants can remove water from these same soils up to appreciably higher tensions if the wilting point is approached slowly and if the root system of the plants permeates the soil thoroughly so as to tap the maximum number of storage points within the soil mass.

Estimating the Available Water Supply in Soils

Any water held within the root zone at a tension less than that describing the wilting point may be considered as available to plants. Since the field capacity is deemed to denote the upper limit of moisture availability, *total available moisture*, then, may be computed as the mathematical difference between the field capacity and wilting point values. If the field capacity and wilting point values are expressed as percentages, the available water content will be a percent value also and will represent the mass of available water held per unit mass of soil. If bulk density values are included in the calculations, available water supply can be expressed on the basis of soil volume rather than soil mass. This, in turn, may be used in conjunction with soil or root-zone depth to estimate the total quantity of available water which may be stored in a field soil. The field capacity value used for these calculations will be that moisture percentage determined in the field soil 2 to 3 days after the conclusion of a thorough wetting either by rain or irrigation water.

Calculations used in determining the available water contained within the root zone of a soil may be adapted also to computing the amount of water necessary to raise its moisture content to the field capacity. By this means, overirrigation can be avoided to a large extent, an important factor where water is in short supply and where excessive additions will result in deep percolation or will raise the level of a subsurface water table. The need for the efficient use of irrigation water in many areas of the subcontinent emphasizes the desirability of understanding and using this procedure as a basis for the efficient use of irrigation water.

Textural Effects on Available Water Storage

Both the field capacity and wilting point values are dependent on total porosity and pore size distribution of soils. These features, in turn, are most closely related to soil texture. To a degree, structural configuration and organic matter content also will have an important bearing on the magnitude of these two values.

Significantly, both the field capacity and wilting percentages increase with increasing fineness in soil texture. However, soils of medium texture generally have the greatest capacity for storing available water. As shown in Fig. 3-9, the difference between field capacity and wilting point values is a minimum in soils of either very coarse or very fine texture.

The storage of but limited available water in sands results primarily from their low retentivity at the field capacity. The limitation in fine-textured soils comes from a more rapid elevation in wilting point values with increasing fineness of texture without a corresponding increase in the field capacity values. This would appear to be due to the fact that changes in the wilting point are related specifically to pore size throughout the entire range of soil texture, whereas field capacity

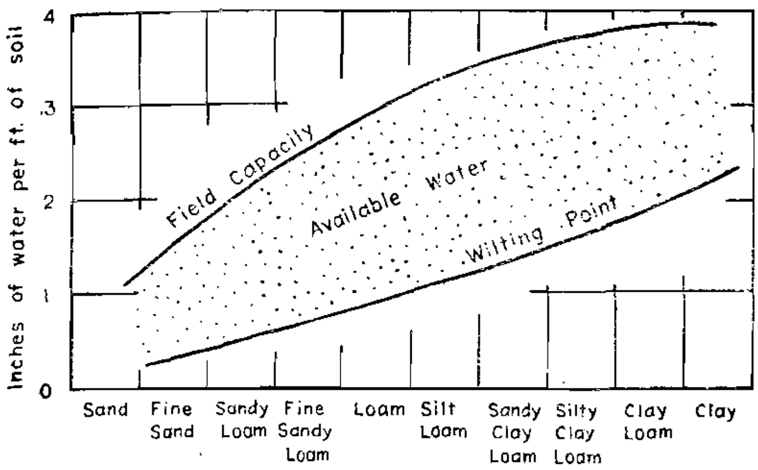

FIG. 3-9. The generalized relationship between soil texture and ability to store available water. Water held between the limits of the field capacity and wilting point percentages is considered to be available. (Adapted from the U. S. Dept. Agr. Yearbook for 1955, p. 120)

values increase only slowly after the average pore size attains a certain minimum. This would relate to the fact that the rate of drainage immediately following the wetting of fine-textured soils becomes so very important in determining field capacity values. Thus, since drainage rates are very low, even in silty or sandy clay loams, further reduction in average pore size as a consequence to an increase in fineness of texture would not be expected to alter either drainage rates or field capacity values materially.

Other Factors Controlling the Total Supply of Available Water

Total annual precipitation and its distribution throughout the year are important factors in determining the quantity of water which will be taken up by the soil to later become available to plants. In arid regions, the mere fact that the total amount of rainfall is low results in only a very limited wetting of the soil profile. Often, the rain that does fall is of a torrential nature and much of it is lost as runoff to limit further the available water supply.

Where the total precipitation infiltrating into the soil is inadequate for maximum plant growth because significant amounts of that received are lost as runoff, the initiation of moisture-conserving practices may be advantageous. Some of the procedures used are designed to increase infiltration by slowing the rate of runoff so that contact between the applied water and the soil is increased. Some of the programmes may be carried out by quite simple means. They may include maintaining the soil surface in a roughened or cloddy condition. Similarly, a natural barrier to runoff is provided by grassland vegetation. Thus, overgrazing, particularly by destructive goats and sheep, can result in the serious loss in the productive capacity of native rangeland of drier regions. Not only will the removal of plant cover reduce the amount of water retained for infiltration, but it also usually results in the irreparable erosion of valuable top soil. Obviously, self-imposed grazing restrictions which prevent the excessive removal of native plant cover at any one time are highly essential in the subcontinent.

Not all plants utilize soil moisture with the same efficiency. Differences in rooting habit, especially with respect to rooting depth, are responsible for this primarily. Deeply rooted plants such as lucerne are capable of utilizing moisture from depths of 12 to 15 ft. or more. Shallow rooted plants, potatoes and onions being examples, may, on the other hand, satisfy virtually their entire requirement for moisture from the surface 2 to 3 ft. of soil. They generally will do no better on deeper soils which have a larger total volume for moisture storage. Such crops do poorly when grown under conditions where the shallow soil reservoir cannot be replenished with water frequently, however.

Environmental conditions including the actual soil moisture status during the growing season may have a marked influence on plant-root development and ability to utilize effectively the total quantity of stored water. For instance, Bennett and Doss (1960) observed that frequent irrigations cause roots of forage plants to be distributed generally at shallower depths than when irrigation, though adequate, was performed at less frequent intervals. Limiting the distribution of roots by this process conceivably could subject the plants to later injury should an extended period of drought subsequently occur. Kmoch *et al.* (1957) observed that wheat grown under dryland (*barani*) conditions develops a more extensive root system and may grow to depths of up to 13 ft. when adequate moisture is present during the early stages of growth. When moisture is suboptimal in the early season, of course, root growth may be less extensive. An adequate level of plants nutrients will affect root development in much the same way as does the supply of moisture.

The average moisture-supplying characteristics of soils determines to a large extent the type of vegetation which occurs naturally in a specific climatic zone. In arid regions, plants are limited to desert species which are of short-life span or which have roots capable of exploring large volumes of soils for available moisture. They may also have leaf forms which minimize the loss of water through transpiration. At higher levels of total available moisture, desert species give way to grasses and, eventually, forest vegetation becomes dominant in areas of abundant precipitation. In transition zones between major vegetation groups, abrupt changes in soil characteristics are often observed to cause equally abrupt variation in vegetation type. Such effects are very closely related to the moisture-retaining characteristics of the specific soils involved.

Soil Moisture Level and Plant Growth

Roots can extract water held under relatively high tensions if the water can move to them rapidly. This is demonstrated by the fact that, where tensions are of sufficient magnitude so that the conductivity rate is somewhat limited, extraction from soils of fine texture is greater than from those of coarser texture. This relates to the greater capability of the small pores to remain more nearly filled and to conduct water at the higher levels of tension. This fact has been used previously to explain why the tension at the wilting point often may be comparatively low in sandy soils.

In some of the earlier work on soil moisture-plant use relationships, Martin (1940) and Kramer (1941) showed that a reduction in the amount of available water in the soil reduced the rate of movement of moisture through the plant. Such an effect, of course, does not necessarily indicate a corresponding change in plant-growth functions. In relating more directly plant growth to soil moisture levels, Wadliegh (1955) observed a gradual decrease in the rate of leaf elongation of pot-grown cotton as the moisture level was decreased to the wilting point. The growth of leaves was resumed, however, when the soil moisture content was increased through irrigation (see Fig. 3-10). Allmendinger *et al.* (1943) demonstrated that the more

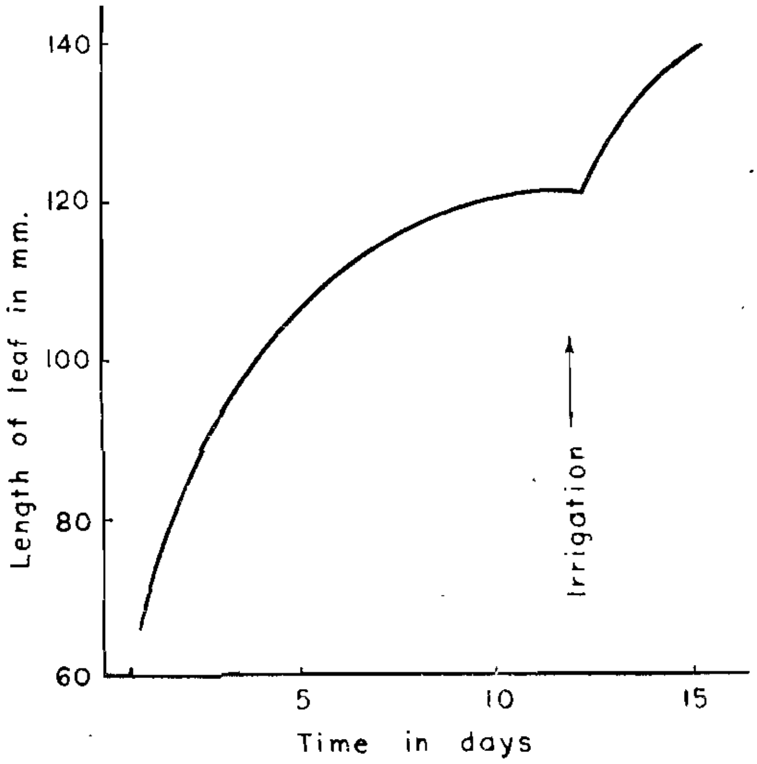

FIG. 3-10. The effect of change in level of available soil moisture on the rate of growth of cotton leaves (after Wadliegh, 1955).

complete the depletion of the available water supply prior to its replenishment through irrigation the greater the reduction in shoot elongation of young apple trees. It was noted also in this trial that the rate of photosynthesis, as measured by consumption of CO_2 by the apple leaves, was not affected unless 90 per cent or more of the available water was used between irrigations.

A consideration of the characteristic moisture curves for soils allows for a semi-quantitative inference as to the effect of tension level on moisture availability. The slope of these curves normally is relatively gentle at the lowest range of tension. This suggests, of course, that the percent moisture content of a soil at its field capacity or above can be reduced materially without a marked, initial increase in tension. For example, the moisture held in available form by the sandy clay loam considered earlier in Fig. 3-3 is approximately the difference between 20 per cent (15 atm.) and 60 per cent (0 atm.)[1], or 40 per cent of the oven-dry weight. If, through plant removal, the moisture content of this soil were reduced to the 40 per cent level, approximately one-half of the available water would have been lost. At the same time, however, the soil moisture tension would have risen to only 2 atm. Obviously, the removal of the remaining half of the available water would elevate the tension to near 15 atm.

So long as the soil moisture tension remains low, neither the force required to remove the water nor the inability of the water to move rapidly up to the roots should limit plant functioning materially. The transpiration rate and turgidity factors should be the first to reflect the effect of increasing soil moisture tension. Photosynthesis, which is largely a chemical process, logically would not be directly affected by even a moderately high stress within the plant, on the other hand. This process very likely would suffer only from such indirect influences as a reduced supply of CO_2 following the closure of stomata as the stress within the plant increased.

[1] For simplicity, the upper limit of available moisture is assumed here to be equal to that held at a tension of 0 atm. rather than that held at some slightly higher value which would correspond to the field capacity. The moisture content of this soil at its field capacity must be determined experimentally in the field and could easily be as low as 55 per cent.

Apparently, a degree of risk is involved in crop production if the soil moisture is not maintained at an adequate level at all times. Indeed, vegetative growth in the field has been observed in some instances to decrease slightly when tension throughout the root zone becomes greater than 1 atm. Certain plants are affected more seriously than are others, particularly if they are at a critical stage of growth. Maize, for instance, may never set seed if subjected to excessive stress at the time of pollination. On the other hand, high stress during the period of crop maturation is of little consequence, a fact that applies equally well to most crop plants.

A mild moisture stress is sometimes desirable in the production of non-vegetative growth. It is known that certain species of grasses and legumes, when grown in soils kept continuously moist, tend to reproduce vegetatively by rhizomes rather than through the formation of seeds. Obviously, where a plant of this type is grown specifically for seed production, a limited moisture shortage may be essential for satisfactory seed yields. The stress should not occur, of course, before the plants have made substantial vegetative growth.

The Pattern of Moisture Removal by Plants

Plants growing on a soil of uniform moisture content remove water from surface layers more rapidly than from the subsoil. Taylor and Haddock (1956) explain this by noting that the more rapid removal of water occurs where root density is the greatest and where the distance through which water must pass within the roots and up to the plant stem is the least. As the water content of the surface soil is decreased, the resultant increase in tension requires that more work be done there by plant roots to extract the remaining water. As a result, it becomes just as easy for the plant to remove water from greater depths even though more work is required to transport it through the longer roots. It is possible for crops to reduce the moisture content of the surface soil to the wilting point while active absorption continues in the subsoil. Although the rate of growth may be reduced at this time, the plant does not wilt until the moisture content of the entire root zone approaches the wilting percentage.

The pattern of removal of water by sugar beets from a soil that had been thoroughly wetted by an irrigation is shown by Taylor (1957). His data are presented in graphic form in Fig. 3-11 where the length of the horizontal bars shows the

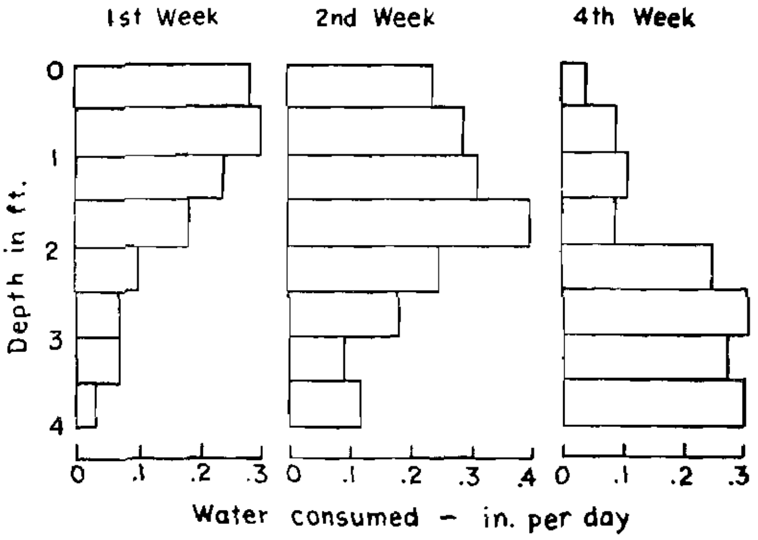

FIG. 3-11. The pattern of utilization of soil moisture by sugar beets during a 4-week period following an irrigation. The length of the bars represents the average water removal from successive 6-in. layers of soil expressed in in. of water per day for each of the 3 week-long periods shown (after Taylor, 1957).

rate of water removal in inches per day for each six-inch depth of soil up to 4 ft. During the first week, maximum water loss occurred in the surface 12 in. of soil. During the second week after much of the surface soil moisture had been used, water was taken up at the fastest rate from the second foot of soil. Finally, during the fourth week, moisture absorption proceeded at the fastest rate at the third-and fourth-foot depths. By this time, the rate of delivery of water had decreased to the point where some reduction in growth of the sugar beets was occurring.

Taylor (1957) considers moisture removal by plants to be governed by the total surface of actively absorbing roots, the rate at which these surfaces are increased through root growth, and the rate at which water moves through the soil to the roots. He points out that if the rate of water movement were constant at all levels of tension, the uptake of water would then depend strictly upon the ability of the root to adsorb it against the given tension level.

The Water Requirement of Plants

The total amount of water that must become available during a growing season depends on the type of plant and the conditions under which it is grown, particularly with respect to the climate. Generally, the larger the plant and the greater its total leaf surface the greater will be its need for water. However, a large, broad-leaved plant grown in a continuously cool, humid atmosphere may use appreciably less water than a much smaller plant grown in a hot, arid environment. Climatic factors are all important in determining the amount of water a plant will absorb and transpire into the atmosphere.

In the past, the water requirement of plants has been expressed as the ratio between the quantity of water used by plants and the amount of dry matter they produce. Since the amount of water used corresponds closely to that transpired, the value is referred to specifically as the *transpiration ratio.* This value provides a basis for comparison among different plant species when grown under the same set of environmental conditions. However, beyond this type of comparison, evaluation of the transpiration ratio is of little practical significance because of its high degree of variation from one climatic zone to another. For instance, small grains grown in the drier parts of the subcontinent will have transpiration ratios varying between 500 and 600 (500 to 600 lbs. of water transpired for each lb. of dry plant material produced). The same plants grown under the more humid climate of England have transpiration ratios of approximately one-half of these values.

Contemporary assessment of the water requirement of plants usually involves *consumptive use* measurements. Such evaluation has its greatest application under irrigation where water input can be controlled so that water shortages do not limit growth. Consumptive use is equivalent to the total quantity of water either transpired by the plants or evaporated from the soil surface (evapo-transpirational loss) during the period of growth of the plant.

Neither evaporation nor transpiration can be controlled effectively in normal crop production. Under careful irrigation, however, other sources of water loss such as deep percolation and runoff can be effectively minimized. These latter losses, therefore, are not considered as contributing to consumptive use. They are not an essential part of crop production in other words.

Consumptive use values will vary for a particular crop grown under different conditions just as will the transpiration ratio. The climatic factor, particularly rainfall and average humidity, is responsible for a large part of the variation.

A major advantage of consumptive use measurements is that they provide an estimate of the total amount of water which must be supplied to the soil for a particular crop during a single growing season. They have their greatest application in irrigated areas where an assessment of the total water requirement for a piece of land is of great importance to the overall planning of farming operations. Consumptive use values are expressed conveniently as the number of inches or feet of water necessary for optimum plant growth during a normal growing season.

AERATION AND DRAINAGE IN SOILS

The process of aeration is closely related to the moisture conditions in soils. Good aeration provides for the replenishment of oxygen so essential to biological processes, and for the rapid removal of carbon dioxide released by these same processes.

An adequate supply of oxygen is necessary for both root and microbiological activity. Summarily, limited oxygen in the soil atmosphere may restrict plant growth either directly or indirectly in the following ways:

1. By inhibiting the uptake of plant nutrients by roots, since nutrient absorption involves the expenditure of energy and requires oxygen for its continuation.
2. By limiting the exploration of the soil mass by roots since it is known that root growth is suboptimal where oxygen is limited. Roots restricted to a small soil volume will deplete the supply of available plant nutrients more rapidly than if they permeated a larger volume of soil.
3. By reducing the rate of decay of organic matter which is a microbial process wherein oxygen is needed for utilization of organic carbon compounds. Plant nutrients retained in organic residues cannot be released if these residues do not undergo decay.
4. By promoting the chemical reduction of iron and manganese to the ferrous and manganous forms. The solubility of these two metals in the soil solution may thereby be increased to the point where they are toxic to plant roots.
5. By encouraging the reduction of sulfate to sulfide, the latter ion being toxic to plants.
6. By preventing the conversion of the ammonium ion to nitrate, since the presence of nitrogen as nitrate often determines its availability to plants.

Other factors could be added to this list, but its present length should indicate clearly the importance of aeration to plant growth.

Composition of the Soil Atmosphere

As in the outer atmosphere, inert nitrogen gas molecules make up continuously about 79 per cent of the soil air. Thus, variation in the composition of the soil atmosphere is caused primarily by concurrent changes in the carbon dioxide and oxygen contents. A change in the concentration of one of these gases causes essentially a proportionate though inverse change in the concentration of the other.

The normal levels of carbon dioxide and oxygen in the atmosphere are 0·03 per cent and 21 per cent, respectively. A ten-fold increase in carbon dioxide to 0·3 per cent will cause an almost negligible change in the oxygen level. This is not to say, however, that major changes in the oxygen concentration do not occur. At considerable depths in many soils, the oxygen content frequently may be as low as 5 per cent. At the same time, accumulating carbon dioxide may make up about 15 per cent of the soil air. This concentration of carbon dioxide probably could be tolerated by most plants *without injury if the oxygen content were maintained near* its normal level of 21 per cent. However, plant growth often will be impaired somewhat if the oxygen content falls to 15 per cent, and may come to a virtual standstill if the concentration drops to as low as 5 per cent.

One of the primary determinants of the state of aeration in soils is the level of moisture present. Soil moisture as a key to aeration has its major influence in filling pores which normally would contribute to the transfer of gases between the soil environment and the outer atmosphere. Clayey soils of low permeability to water often drain so slowly that aeration in them frequently presents a problem. Medium- to coarse-textured soils, on the other hand, drain more rapidly and normally do not retain water at a level above the field capacity for prolonged periods of time. Thus, restriction in air movement in them due to excessively wet conditions will be of relatively short duration. In fact, the rapid interchange of air in very sandy soils may promote drying to the point where moisture loss by evaporation becomes a serious problem. This is particularly true when plants are in the seedling stage and provide little ground cover as protection against evaporation. Obviously, then, extremes in the textural grade of soils are the least desirable as far as soil aeration is concerned.

Oxygen shortages develop within a few hours after conditions of saturation are established in a soil. While these shortages are of a perennial nature under conditions of permanent waterlogging, the effect may be quite noticeable even where only temporary flooding occurs. The oxygen present in a saturated soil is limited to the small quantity dissolved in the water. Not only does this disappear rapidly but, also, the concentration of dissolved carbon dioxide builds up with equal speed and may accumulate to where it constitutes one-half or more of the dissolved gases present. The adverse relationship between carbon dioxide and oxygen concentrations is related partially to the greater solubility of the carbon dioxide in water.

Aeration problems are associated with conditions other than those of complete saturation. Boicourt and Allen (1941), for instance, found that aerating soils once a day doubled the growth rate of rose plants when the moisture content of the soil was maintained constantly at levels somewhat above the field capacity but appreciably below saturation. This may be related to the fact that moisture films on root surfaces allowed for only very slow transfer of oxygen to the roots. Oxygen will diffuse at a much slower rate when dissolved in water than it will when it occurs as a gas. Its low solubility in water accounts for this to a large extent.

Inadequate aeration likewise may be associated with conditions of poor tilth, particularly in fine-textured soil materials. So-called puddled soils, which by virtue of their lack of aggregation have a preponderance of small pore spaces, restrict the movement of air just as they impede the flow of water. Periods of low oxygen supply also may occur at other times when large quantities of easily decomposable organic residues have been added to the soil. Microorganisms responsible for organic matter decay seem to have first call on the oxygen present in the soil and thereby offer serious competition to plant roots for this essential element. The fact that microbes usually are more widespread in the soil probably accounts for this to a large degree.

The Adaptability of Plants to Waterlogged Conditions

The ability of crop plants to withstand conditions of poor aeration varies considerably. Many woody species withstand waterlogging to but a very limited extent. Cotton and numerous vines and trees fall in this category. Certain more succulent plants may survive submersion for considerable periods and may even produce some growth under these circumstances. However, irrigated rice is the only crop plant of importance which appears to require continual submergence for satisfactory growth. Dryland rice, like most other crop plants, requires a favourable oxygen balance in the soil atmosphere for satisfactory growth.

Following flooding and the rapid depletion of the oxygen supply in the soil, roots of sensitive plants become inoperative and finally degenerate and die. Those plants which can adapt to these conditions are ones capable of developing new root systems having the morphological feature of large internal spaces for the transfer of gases. Thus, the new roots are supplied with oxygen acquired by the leaves and then passed downward through the large, internal passageways. Growth of the adaptable species appears to be hampered to the greatest extent during the period when the new roots are being formed. Where adaptation does not follow flooding but results only in the deterioration of the roots, death may be traced largely to the decreased uptake of nutrients and water even though the degenerating roots are bathed continuously in water.

Drainage Requirements and Plant Growth

In regions where rainfall or irrigation provide adequate moisture for plants, the presence of a water table too near the surface normally will induce conditions of poor aeration in the soil. The most obvious effect is the restriction placed on the depth of root penetration. The lower limit of the root zone will correspond rather closely to the upper boundary of the *capillary fringe*, or that zone immediately above the free water surface which is saturated due to capillary rise of water. The thickness of the fringe will depend upon soil texture, which controls the rate of capillary rise from the water table, and the rate at which water is removed from the upper boundary of the fringe by plant roots.

Data of Elliott (1924) illustrate the sensitivity of one crop, maize, to conditions of waterlogging. Observations were made where this crop was grown on land troubled with a high water table but where this condition was partially overcome through the installation of a system of tile drains. As is typical with such systems, the greatest reduction in the level of the water occurred directly above the tiles. Thus, as shown in Fig. 3-12, the depth to the water table varied measureably over relatively short distances within the field. The effect on root development and top growth of the maize was very evident. Roots maintained a distance of 18 in. above the surface of the water table regardless of the depth to the latter. Further, the deeper the rooting the greater was the growth of the above-ground portion of the maize.

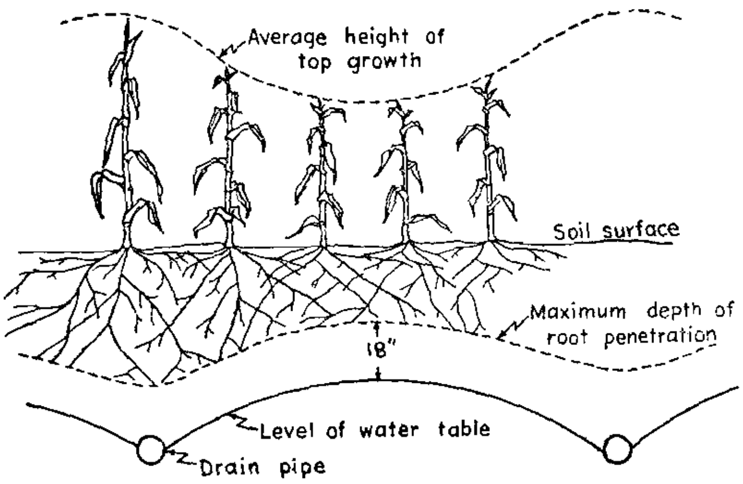

FIG. 3-12. The effect of depth to a water table on extent of rooting and top growth of maize (adapted from Elliott, 1924).

It is regarded generally that, where aeration is potentially a problem, a water table should be no nearer than 5 ft. from the land surface. However, the rooting habits and oxygen requirements of plants differ materially, as does the state of aeration over the water table in poorly drained soils of different texture. Thus, no single value can be assigned to indicate the most satisfactory minimum depth to the water table which will apply equally well to all conditions where drainage is restricted.

The Water Table as a Source of Moisture for Plants

Under certain circumstances, it is necessary for crop plants to acquire needed water primarily from subsurface sources. Where this is true, the presence of a water table may be considered as essential. Here, the depth factor is even more critical since a suitable balance between moisture supply and aeration must be maintained. Again, because the proper depth is dependent upon various factors, only generalizations can be relied upon as a basis for determining this depth value.

In coarse-textured, naturally well-aerated soils, the water table need not be as deep as in more poorly aerated, clayey soils. For instance, soil of coarse texture supporting grass with a rather limited root system may supply inadequate moisture if the depth to the free water surface exceeds 15 to 30 in. In clay soils with a thick capillary fringe, on the other hand, growth may be hindered as a result of poor aeration if the distance to the water table is less than about 35 to 40 in. On either soil type, even greater depths may be desirable for the production of horticultural or other crops grown in rows. Under these conditions, only a part of the land surface is covered with the crop plant and, therefore, an excellent opportunity is offered for weed growth because of the continually moist conditions of the soil.

A water table serving as a source of moisture to plants will be of little value if it occurs at depths in excess of 5 ft. This relates to the inability of the soil to conduct water upward at a sufficiently rapid rate to meet plant demands. Lucerne, even with its deep rooting habit, frequently will respond to supplemental irrigation if it has as a primary source of water a water table occuring at a depth of 5 ft. or more. Wheat and small grains, having a shallower root system, may suffer from drought if the water table is at a depth of more than 3 ft. This is particularly true during the early stages of growth before the root system is fully developed.

The Improvement of Aeration Through Drainage

Many lands in the subcontinent are affected by a water table. The broad flood-plains associated with the major rivers and their tributaries are particularly troubled. In the more arid regions to the north and west and in the central Indian peninsular area, the introduction of extensive irrigation systems has intensified the waterlogging problem. This arises from both the greater opportunity for seepage due to the establishment of irrigation canal systems and from the increase in the total amount of water applied to the land as the result of irrigation. Unfortunately, these induced high water tables commonly result from the build-up of subsurface waters of high salt concentration. Rise of the water from the water table and its subsequent evaporation has added, through the accumulation of salts, a problem of equal importance to that of poor aeration which also follows the creation of conditions of poor drainage.

Since the primary causes of these problems cannot be eliminated in their entirety, solution rests largely with the improvement of drainage. This, in turn, requires the rapid removal of excess water from the soil so that a reduction in the level of the water table is possible.

Drainage problems, because of their extensive nature, normally cannot be dealt with successfully by any single person or even a small group of individual land owners. Correction must come from cooperative action financed largely by the government ; for drainage systems, to be effective, generally must be capable of reducing the level of subsurface waters over wide areas.

Causes of Waterlogging

Whether or not a water table will be found under a given set of soil conditions depends upon a variety of factors. Basically, a water table is created where the input of water into a soil mass is at a rate in excess of the rate of removal. In general, the water may be derived primarily from one of two sources ; that is, seepage for permanent or semipermanent waterways, and that applied to the surface of the land by various means. Where both of these sources are functioning concurrently, the problems arising from waterlogging normally attain their most serious proportions.

The creation of a water table through seepage from a permanent waterway is illustrated in Fig. 3-13A. In this

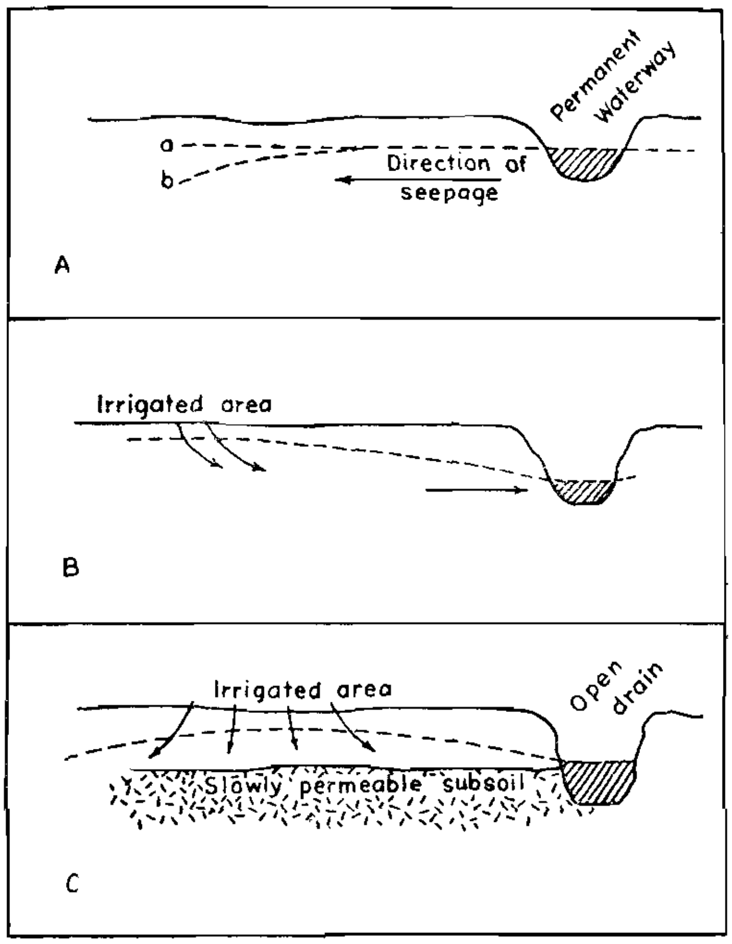

FIG. 3-13. (A) Diagram illustrating the maintenance of a permanent water table by seepage from a waterway. Dotted line **a** represents the free water surface in coarse soil of high conductivity for water. Dotted line **b** represents the water surface in fine-textured material where low conductivity limits the distance of seepage. (B) The influence of surface-applied water on the level of a water table derived initially through seepage from a permanent waterway. (C) The creation of a perched water table over a slowly permeable subsoil.

instance, the movement of water within the soil mass is away from its source. Under these conditions, the induced water table will occur over a greater distance from the source of seepage if the conducting soil material is open and porous. Where the water is derived entirely from seepage, the level of the water table will decrease with increasing distance from the waterway.

Surface-applied water also may contribute to the level of the water table as shown in Fig. 3-13B. When water from this source is added to a water table derived initially from seepage, the net movement will be toward the open waterway. Due to the increase this causes in the level of the water table, conditions of poor aeration may be more critical than where seepage alone is responsible for the waterlogging.

A temporary or perched water table is illustrated in Fig. 3-13C. As shown in the diagram, such a condition may occur over a relatively impermeable subsoil material. The degree and duration of waterlogging induced by this mechanism will depend upon the amount of water added and the rate at which it drains from the saturated zone by lateral movement. The ultimate disposition of the water may be into soil zones of better internal drainage or into open drains which border the waterlogged area.

Evaporation of water from the soil surface and loss by plant absorption and transpiration will affect the level of the water table. These causes of water loss may be of relatively large magnitude and of considerable importance where the input rate into the waterlogged zone is slow. Thus, during periods of high evapo-transpirational loss, the level of the water table may be observed to fall gradually. In fact, the variation in water level may be sufficiently great so that it can be measured readily on a daily basis. Where this is possible, a rather sudden and rapid increase in the subsurface water level normally follows the harvesting of the plants largely responsible for the removal of water. If, on the other hand, the waterlogged soil is coarse textured and exhibits high saturated permeability, seepage into it from a permanent waterway will be rapid. Thus, the rate of replenishment of water lost through plant absorption will normally be adequate to maintain an essentially constant water level.

Drainage Principles

A consideration of the factors causing waterlogging leads logically to the conclusion that the correction of this condition requires either a decrease in the rate of water input into an area or an increase in the rate of its removal. In the first instance, reduced input rates can be attained practically in but two ways ; through a reduction in the rate of application of irrigation water, or through the minimization of seepage from waterways. Should such control be impossible, then the only alternative is to attempt to increase the rate at which excess water is removed from the affected area.

A reduction in the level of the water table through drainage involves the transfer of water through the soil up to the point of discharge into a natural or man-made drainage system. Natural drainageways include, of course, streams and rivers which normally lead to the sea.

Artificial drains are of two general types: *open* systems as represented by ditches, and *closed* systems consisting of a network of buried pipes or similar channels which lead eventually into an open drain. Any artificial system is designed to intercept water below the surface of the water table and conduct it to natural drainage systems at a rate appreciably faster than would be possible by normal lateral movement through the soil alone. In essence, then, the primary purpose of drain installation is to hasten the rate of lateral flow of water in a waterlogged soil.

While the rate of discharge of water into a drain is a function of the conductive capacity of the soil material, it will depend also upon the characteristics of the drain. If the level of the water in an open drain is as high as that of the adjacent free water surface, no hydraulic head exists to promote the flow of water out of the soil body into the drain. Thus, basic to the success of the process is the rapid removal of the waste water and the associated requirement that the drain providing this removal has sufficient capacity and slope to carry away rapidly the water discharged into it. Restriction to flow as caused by accumulating silt or weeds must be carefully avoided.

Movement of water in the waterlogged soil itself is governed by the principles of saturated flow ; thus, textural effects are of paramount importance. Coarse gravelly and sandy materials, by virtue of their large pore spaces, are characterized by a high rate of flow when saturated. Water movement through them and into a drain will be sufficiently rapid so that the level of the water table will correspond closely to the level of the water in the drain. In other words, the depth of the drain becomes a primary controlling factor in determining the level of the water table in permeable soil materials. Further, the influence of the drain on the water table is observed over large distances in such soils. As a consequence, the drains themselves need not be spaced too closely. Often, half-mile intervals between them are adequate for suitable drainage control.

With increasing fineness of texture, the resistance to lateral water movement in the soil increases. Thus, to bring about a reasonable reduction in the water level, drains must occur at more frequent intervals. In clayey soils, drain spacings of 50 ft. or less may be necessary. Where such narrow spacing is required, a system of open drains becomes impractical because of the large area of land taken out of cultivation when the drainage system is installed. Although they are costly, cement or tile pipes constitute the only materials which provide for a suitably durable, closed drainage system. Thus, economics will determine whether or not their installation is possible.

REVIEW QUESTIONS

1. Why do we evaluate the force of attraction between soil and water in terms of moisture tension ? Why may we express tensions either as centimetres of water or atmospheres of pressure ?
2. Define the wilting percentage ; field capacity ; available moisture range or percentage.
3. As the moisture content of a soil decreases so does the rate of moisture movement under a given tension gradient. In which would you expect the rate of flow to decrease most rapidly, a sandy or clayey soil ? Why ?
4. Explain why *tension* and *negative pressure* have the same meaning.
5. What are the principles of operation of the tensiometer and moisture block used in measuring soil moisture or moisture tension ? What are the limitations of these two types of apparatus ?
6. Why does the tension gradient decrease as the depth of soil wetting increases ?
7. In which would salts probably accumulate more rapidly in the surface soil if a salty water table occurred 2 ft. beneath the soil surface — a clay loam or sandy loam soil ? Why ?
8. The field capacity of a uniform soil is 20 per cent and its wilting point is 4 per cent. If the bulk density of the soil is 100 lbs. per cu. ft. and it contains a uniform moisture content of 10 per cent, to what depth would the moisture content be raised to the field capacity by a 3-in. irrigation ? The available moisture in the upper 6 ft. of soil prior to irrigation corresponds to how many inches of water ? The available water in the surface foot after the irrigation corresponds to how many inches of water ? (*Note* : a cubic ft. of water weighs 62·4 lbs.)
9. Explain why moisture is absorbed by plants more rapidly from the surface soil than from the subsoil.

10. Assume the moisture content of the two soils referred to in Fig. 3-3 is 20 per cent. If they are placed in contact with each other, in which direction will the water flow and why ?
11. Why may a water table at 3 ft. have a more serious effect on plant growth in a clayey soil than in a sandy soil ?
12. Name four ways in which inadequate 0_2 in the root zone may influence plant growth.
13. Why may we say that the quantity of air in a soil is controlled by the moisture level ?

SELECTED REFERENCES

1. Baver, L. D., *Soil Physics,* John Wiley and Sons, New York.
2. Kramer, P. J. *Plant and Soil Water Relationships,* McGraw-Hill Pub. Co., New York.
3. 'Water', *The Yearbook of Agriculture for 1955,* Supt. of Documents, Washington 25, D. C.
4. Russell, M. B., (Coordinator) 'Water and its relation to soils and crops'. *Advances in Agron.* 11 : 1-131 (1959).

CHAPTER FOUR

THE MINERAL FRACTION OF SOILS

Most materials from which soils are formed are derived from weathered rocks and minerals. Weathering through physical breakdown of the original rock and mineral formations results merely in the formation of smaller fragments of the chemically unaltered materials. Where chemical weathering has been effective, characteristics of the original minerals other than size will have been changed. New products are formed as the result of chemical weathering. Since the comminution of mineral materials to clay-sized fragments usually involves at least a degree of chemical change, most clays are alteration products. Generally they are more resistant to weathering than are the unweathered minerals from which they form ; thus, they tend to persist in a soil body for longer periods than do most original mineral particles of like size.

Unweathered or moderately weathered minerals in soils serve as a reservoir of chemical elements necessary for plant growth (*essential plant nutrients*). The ability of a soil to supply nutrients in amounts adequate for plant growth depends upon the type and quantity of specific minerals present and the rate at which they succumb to weathering. Except where extensive weathering and leaching have occurred, soils generally contain the more common nutrient-supplying minerals in relative abundance. Even so, many of our agricultural soils are incapable of supplying all nutrients in adequate amounts, and, therefore, chemical fertilizers and manures are used frequently to supplement those derived from the soil minerals.

According to Clarke (1924), at least 95% of the outer 10-mile crust of the earth consists of essentially unweathered rocks. Although these rocks are comprised of many different types of minerals, they are made up largely of those few shown in the following list:

Feldspars	59%
Hornblende and augite	17%
Quartz	12%
Biotite	4%
Other minerals	8%
	100%

Except for quartz (SiO_2), the above minerals are aluminosilicates which serve as important sources of the Ca, Mg, Na, K, and Fe in soils. The feldspars and the micas[1] contain Na, K, and Ca as principal cationic components. Hornblende and augite, members of the ferromagnesium mineral group, are important sources of Fe and Mg. Quartz, which occurs in many crystalline forms, is not a source of nutrient elements to growing plants.

Rocks in the Earth's Crust

Rocks are classified primarily on the basis of their mode of formation. Three major groups occur: *igneous*, *sedimentary*, and *metamorphic*.

Igneous rocks are decidedly more abundant within the earth's crust than are the other two rock types. They are made up of mixtures of minerals which, for the most part, crystallized from molten magma during the early stages of formation of the earth.[2] They are found largely beneath the earth's surface where contact with the atmosphere, hydrosphere, and biosphere is limited. Major exposures of massive igneous rocks are observed in mountainous regions where weathered detritus is removed continuously by erosion.

Sedimentary rocks are formed as the consequence of the accumulation of weathered, fragmented mineral substances in layered deposits of varying thickness, very often beneath water. After deposition they become consolidated to varying degrees through cementation by precipitated calcium carbonate, silica,

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[1] The two most common micaceous minerals are *biotite* and *muscovite*.

[2] Minerals formed through crystallization of magma often are referred to as *primary* minerals.

or even iron and aluminium oxides. Sedimentary rocks range in texture from *conglomerates*, which include gravelly materials, through *sandstones*, and clayey *shales*. Those containing a large proportion of calcium and magnesium carbonates are *limestones*.

Metamorphic rocks consist largely of the same minerals found in igneous rocks. However, whereas igneous rocks have resulted from the initial solidification of mineral matter in the cooling crust of the earth, metamorphic rocks form through recrystallization of melts deep within sedimentary deposits. Due to the depth of these sediments, heat and pressure develop sufficiently to impart at least temporary fluidity to the mineral constituents. Eventual removal of the overburden through erosion again reduces the heat and pressure allowing the minerals to recrystallize into metamorphic rock formations. Normally, these rocks have a stratified or streaked appearance inherited from the layered sedimentary deposits in which they have formed.

There are various types of metamorphic rocks. Some are similar to specific igneous rocks and contain well-defined crystals of feldspars, quartz, mica, and ferromagnesium minerals. Sandstones high in quartz are converted to *quartzite* through metamorphosis, whereas shales are altered to fine-textured *slate*. Metamorphic rocks are more resistant to weathering than are their sedimentary counterparts.

The major occurrence of metamorphic rock, like igneous rock, is beneath the earth's surface. Therefore, only a small proportion of soils in the major agricultural areas are developing on materials derived principally from these rock types. Sedimentary deposits, including the unconsolidated formations laid down by river and stream action, are estimated to cover 75% of the total land surface. It is on these materials that a major part of the agricultural soils are found. Because of their previous contact with the atmosphere and hydrosphere, the minerals in sediments have been subjected to varying degrees of weathering prior to initiation of specific soil-forming processes.[1]

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[1] The formation of soils is the result of several processes including mineral weathering. These processes are discussed in detail in Chapter 13.

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WEATHERING PROCESSES

Rocks and minerals in the earth's crust begin to weather when subjected to environmental conditions under which they are no longer stable. Thus, igneous rock buried deeply within the crust of the earth persists in its original form until exposed to air, water, and biological action through removal of the surface layers above. The type and extent of change in mineral form which follows exposure will depend on the magnitude and type of change in environmental conditions.

Essentially, weathering is a process whereby minerals alter in a tendency to achieve an equilibrium with a new environment. Generally, weathering is viewed as a process which brings about the gradual destruction of original minerals and the loss of their components in water percolating through the weathering mass. However, weathering will occur even in the absence of water or, at the other extreme, it will take place in completely submerged sediments. Under the latter conditions, ions from the water may be taken up to replace those lost in a previous weathering cycle. Thus, where chemical change is involved, constructive as well as destructive alteration may occur. Nonetheless, the requirements of plants for a continual supply of nutrients necessitates their release from minerals as a consequence to destructive weathering reactions. The following discussions on weathering are concerned with the degradative aspects of the process.

Physical Weathering

Physical weathering occurs predominantly as the consequence to abrasion during periods of movement of rock materials. Weathering in place is attributable largely to chemical rather than physical processes. Perhaps the only significant physical weathering in place results from the wedging action of growing roots and forming ice. However, this will take place only in pre-existing cracks or fissures, and the effect will be one of extending or enlarging such breaks in rock masses.

Abrasive weathering, which occurs as two rock surfaces strike each other or are rubbed together, expedites chemical weathering markedly. It does so by increasing the proportion of fine-textured particles and total surface where chemical reactions may take place. The principal agencies of abrasive weathering are (1) moving ice (glaciers), (2) moving water, (3) wind, and (4) landslides. The first two means are considerably more important than the last and are responsible for a large part of the fine-textured soil materials distributed over the earth's surface.

Water, wind, and ice in themselves have little effect on the disintegration of rocks and minerals. As forces of transport of movable rock fragments, however, they bring about a grinding action among the fragments which they are transporting. During glaciation huge quantities of fragmented materials are moved *en masse*. Pressures within the moving mass are of great magnitude and cause rapid fragmentation and abrasion of the transported materials. Glacial *till*, the pulverized deposits from glacial activity, forms a very important soil material in the northern areas of the Northern Hemisphere and in mountainous regions where snow and ice accumulate. Gravelly soils and sediments in northern India and Pakistan are evidences of strong glacial activity in the subcontinent in times past.

Glaciation currently is a direct factor in physical weathering in the subcontinent only in the higher reaches of the Himalayan Mountains. The effects of this process are felt throughout the Indus, Ganges, and Brahmaputra River areas, however, because the pulverized debris is carried by stream action for subsequent deposition in the valleys of these great rivers. Further abrasion of rock fragments occurs during this latter period of transport.

Wind action undoubtedly has contributed much to the physical disintegration of soil materials in the subcontinent, particularly in the arid regions of the western parts. Not only does the wind bring about a reduction in the size of grains being moved, it also sorts them. Those particles of small diameter may be carried great distances, whereas the coarser grains are left behind. For this reason, much of the Thar and Thul Deserts are comprised of sands which are virtually free of clay or silt particles. The finer materials are deposited at varying distances from their point of origin depending upon the size of the particles and the velocity of the wind carrying them.

Chemical Weathering

The processes of chemical weathering differ from those of physical weathering in that they result in the change in composition of the original rock substances. The extent of degradative alteration depends upon the intensity of weathering reactions and their duration. Changes may be no greater than slight modification in the structure of the original minerals. However, extensive chemical weathering will result in virtually complete breakdown of mineral structures. Readily soluble products of such reactions normally will be carried away from the point of their formation by leaching water. Resistant alteration products, which are more stable in the weathering environment, remain and tend to replace the less resistant original minerals as the major constituents of the weathering mass.

Jackson and Sherman (1953) list five factors which control the rate and direction of reactions in chemical weathering. These are (1) temperature, (2) the quantity of water contacting the weathering surfaces, (3) the acid content of the solution, (4) biological forces, particularly as they affect the acid and cation status of the water, and (5) the oxidation state, or the aerobic-anaerobic relationships within the decomposing medium. All of these factors work simultaneously during mineral decomposition.

Solution Effects: Whereas most siliceous minerals dissolve to but a very limited extent in pure water, their solubility increases with an increase in the concentration of H ions, or the acidity of the hydrosphere. All water permeating the weathering mineral mass contains at least small quantities of CO_2 which, upon solution, forms carbonic acid (H_2CO_3). Although this is a weak acid, its continual presence in the weathering medium has a telling effect upon the permanence of the minerals present.

An important effect associated with reactions between H ions and minerals is the hydrolytic release of the metallic cations Na, Ca, K, and Mg into the surrounding solution.

Unless these cations are removed from the sphere of weathering, they tend to accumulate in the carbonate form. The increase in concentration of the cations in the solution phase reduces the tendency for the release of additional cations from the minerals. Furthermore, their occurrence as carbonates suppresses the ionization of H_2CO_3 and thereby minimizes the quantity of H ions present in the solution phase. For degradative weathering to be a rapid and continuing process, therefore, removal of the soluble reaction products, including Si and Al, is necessary. For this reason, the quantity of water passing through the weathering zone determines to a large extent the rate at which chemical decay will proceed.

Where plants have become established, respiring plant roots and micro-organisms involved in the decay of organic residues contribute considerably to the level of CO_2 in the hydrosphere. The acid this produces, as well as small quantities formed as microbes oxidize sulphur and ammonium compounds, all increase the capacity of the hydrosphere to bring about the chemical breakdown of minerals. Thus, biological activity, which is deemed essential to soil formation, hastens mineral decomposition.

During weathering, removal of alkali and alkaline-earth cations is more rapid than that of Si, Fe, and Al. This probably is due to the fact that these latter ions are capable of reforming new silicate structures which are relatively stable in the existing environment. Even these compounds are somewhat soluble and tend to disappear as leaching continues. Normally, silica is the more readily soluble and, therefore, disappears more rapidly than do either Fe or Al. As a consequence, the residual minerals occurring now primarily as clay-sized particles become progressively enriched in Al, or in Al and Fe. Because of this, the extent of weathering has been approximated frequently through determination of the ratio of Si to Fe+Al in the clay fraction of the soil. This ratio, known as the *silica : sesquioxide ratio*[1], varies from about 4 : 1 in slightly

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[1] The gravimetric determination of these ions involves measuring the weight of their oxides. Since Fe and Al commonly are precipitated and weighed together, they are often referred to jointly as *sesquioxides* and are represented symbolically as R_2O_3. Silica is SiO_2.

weathered mineral materials to less than 1 : 1 where weathering has brought about extensive mineral alteration.

Oxidation Reactions: Oxidation reactions important in mineral weathering are those involving the heavy metals Fe and Mn, and the conversion of C to CO_2. The latter reaction contributes CO_2 to the soil atmosphere and hydrosphere as discussed above.

Oxidation reactions involving heavy metals are exemplified by the change in Fe in the mineral crystal from the ferrous to the ferric[1] state. The conversion results in the gain of a single positive charge (*i.e.*, $Fe^{+2} - e \rightarrow Fe^{+3}$). This creates in the crystal an electrical imbalance which can be compensated for only by the loss of other cations. For this reason, the ferromagnesium minerals are among the first to decompose during weathering.

Temperature Effects: As with an increase in moisture, elevation of temperature also brings about more rapid mineral transformation. This relates not only to the speed of inorganic reactions, which are temperature responsive, but also to reactions within the biosphere, including the decomposition of organic substances and the biological oxidation of inorganic compounds. Since conditions within the humid tropics correspond to relative high temperatures and high levels of available moisture, weathering there proceeds at a comparatively rapid rate. Transformation of minerals to resistant products is swift as evidenced by the generally high proportion of sesquioxides in soils of these areas.

Effect of Time: In the strictest sense, degradative weathering is complete only when all components of the original materials have become solubilized and carried away in leaching water. Practically, however, we may assume an end-point to have been reached when original minerals, unstable in an imposed environment, have been converted to residual products which undergo further change at a very slow rate.

The initial conversion of minerals to more resistant products will occur at a rate depending upon the intensity factors of temperature and the quantity and composition of

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[1] Inorganic oxidation is the loss of electrons by an atom or ion which results in an increase in positive charge.

leaching water comprising the hydrosphere. Thus, if environmental conditions vary as a consequence to a reduction either in temperature or the degree of leaching, conversion will require a longer period of time for completion. Within a limited degree of variation in environmental conditions, time, then, may be thought of as compensatory for low weathering intensity.

WEATHERING RATES OF MINERALS

The rates of decomposition among the various minerals differ. Knowledge of this fact is important in understanding why specific minerals disappear or persist within certain size fractions of actively weathering soil materials. The rate of weathering serves only as a partial basis for predicting the quantity of a specific mineral which may be found under a certain climatic environment, however. Important also will be the proportions of individual minerals initially present, their particle size, the nature of the weathering environment, and the intensity and duration of weathering processes. A mineral matrix high in easily weathered minerals may be expected to change most rapidly under a given weathering environment.

Goldich (1938) has classified the more common minerals in soil-forming materials as to their relative rates of chemical breakdown. For particles of similar size, the ferromagnesium minerals and the calcium and sodium feldspars will disappear most rapidly. The micas and potassium feldspars assume an intermediate position, whereas quartz is relatively resistant. Since quartz does not contribute essential nutrients to plants, its slow rate of weathering is of little consequence to plant nutrition in agricultural soils.

Because of the differential in resistance to weathering among minerals, they may be expected to disappear from soil-forming materials at different rates. Since comminution to smaller particle size increases the rate of chemical weathering, usually the minerals in the clay fraction of soils which remain essentially unaltered from their original form will be of the more resistant types. The presence of easily weathered minerals as dominant constituents of clays indicates that physical rather than chemical weathering has been instrumental in bringing about the breakdown of minerals into the smaller-sized particles.

Mineral weathering in arid regions is dominated by physical processes which tend to produce coarse-textured soil materials and clays that result from but limited chemical alteration of the original minerals. In more humid regions, soils are characteristically finer textured and contain clays as products of extensive mineral alteration. Under either climatic condition, the length of the weathering period will determine partially the nature of the weathering products, of course.

Most agricultural soils are formed on transported mineral matter. Because of their being moved, these materials often are introduced into a climatic and weathering environment entirely different from that occurring at the point of their origin. Thus, prior weathering may determine the characteristics of the transported minerals to a greater extent than do weathering effects following transport. For instance, the nature of water-laid materials in the Indus, Ganges, and Brahmaputra River Valleys often relate more closely to the weathering conditions at the site of their origin in the Himalayan Uplands than to the environment in their present location. In the Indus and Upper Ganges Basins where aridity and low weathering intensity persist, river-deposited materials show evidence of greater alteration than do those formed as the result of the influence of local climate on rock native to these same areas. In the eastern reaches where weathering intensity is greater by virtue of more abundant rainfall, soil materials formed on local, upland rock are weathered to a much greater extent than are the river deposits. Obviously, the time factor in weathering has been important to the development of these observed differences. For example, the flood-plain deposits along the rivers are of relative recent origin, whereas the upland soils have been subjected to local weathering influences for a sufficient time so that they exhibit properties which may be related closely to the existing environment.

As indicated above, minerals occurring in the soil as alteration products of degradative weathering acquire a degree of stability not recognized in their original, unweathered predecessors. Generally, those that have undergone greatest change have done so under the influence of intense weathering imposed for relatively long periods of time. Since these reconstituted minerals occur principally in the clay fraction of soils, they have been referred to commonly as the "clay minerals". While this is a most convenient connotation, it should be realized that the minerals referred to actually may be of silt or even sand size. Thus, the term clay mineral generally suggests a mode of origin of the minerals or refers to properties not particularly related to size.

The previous discussions on mineral degradation and resynthesis have been limited to processes wherein some of the weathering products are lost through leaching and do not recombine as new mineral compounds within the weathering mass. The most readily soluble components are lost primarily and the residue consists of compounds relatively more stable than the original minerals. However, readily soluble ions can be returned to positions within the crystalline structure of alumino-silicate minerals to form stable products also. For example, partially weathered mineral particles which accumulate in marine sediments often recombine with soluble K and Mg present in abundance in the sea water. These new products are similar to some of the original minerals commonly present in igneous rocks. While stable in the saline water, they succumb rapidly to weathering in a leaching environment. Thus, elevation of the sediments to above sea level will, once again, initiate their breakdown. Minerals of this type are important in soils developing on sediments of marine origin.

PROPERTIES OF SOIL MINERALS

Mineral particles occur in the soil in a wide range of sizes. A decrease in the average size of these particles usually is accepted as denoting two conditions in that the smaller the particle and the greater the total surface area (1) the more reactive the mineral material is, both chemically and physically, and (2) the greater is the probability that the particles occur as resynthesized mineral matter produced in the course of mineral weathering. The larger soil particles, then, are viewed generally as consisting of relatively unaltered mineral fragments derived from the physical breakdown of larger masses of like chemical compounds. This concept relegates to a subordinate position large mineral particles derived through precipitation and cementation processes (*i.e.*, granules of precipitated lime, silica, and the sesquioxides). Such coarse-textured particles are common to many soils, however, and contribute materially to soil characteristics where present.

Mineral particles of clay size are observed to consist predominantly of resynthesized alumino-silicates and oxides and hydroxides of Fe and Al. As products of weathering under inumerable environments, one would expect these derived mineral compounds to exist in an almost unlimited number of structural forms, particularly those falling within the alumino-silicate group. While this is true to a degree, the manner in which Al and Si ions associate in molecular structures is relatively invariable. As a consequence, there are basic similarities among many of the compounds of this type in soils. Mineral types that are synthesized during weathering, then, are sufficiently uniform to permit their classification into a small number of groups which display a minimum of variation among the members contained within any of these groups. The system of classification, as will be discussed in more detail shortly, is based primarily on differences in molecular structure of these crystalline compounds. Intrinsic properties of the minerals are relatable to their structural forms. As a matter of fact, these properties served as the principal means of classification of microcrystalline and amorphous, colloidal mineral particles for many years.

Means of Characterization of Soil Minerals

Coarse mineral particles in soils are identified with relative ease. Often examination under the ordinary microscope discloses their identity. Usually, however, measurement of their optical properties with a special *petrographic* microscope and polarized light is helpful in distinguishing between species that appear to be similar from all outward appearances. Separation into specific-gravity groups by flotation in heavy *liquids is resorted to very frequently as a preliminary step in* identification by microscopic procedures.

Mineral particles of clay size are not identified easily by visual means. It is for this reason that early recognition of differences in clays depended upon total chemical analysis and measurement of other readily discernible physical and chemical characteristics.

Formerly, essentially all minerals comprising clays were thought to be amorphous. It was not until the development of methods for X-ray analysis that this opinion was shown to be false. Distinct X-ray *diffraction patterns*[1] are obtained for many of the minerals in clays just as they are for other substances known to be crystalline. Deductions have been made from such studies as to the molecular structure of the individual submicroscopic, mineral particles. In recent years, physical dimensions of the various groups of clay-sized mineral particles have been measured accurately through use of the *electron microscope* (see Figs. 4-3 and 4-4).

Structure in Silicate Minerals

It has been learned through X-ray analysis that silicate mineral particles consist of small structural units composed essentially of oxygen ions (or hydroxyls) enclosing one of several cations. The enclosed cations are most often Si and Al, but other cations are present frequently also. The ability of oxygens to enclose a cation depends, among other factors, upon the charge of the cation and its size or radius. Size of the cation is important because it determines the feasibility of satisfactory enclosure within the spaces between closely packed oxygen ions. If the oxygens are held too far apart, the crystal structure they form will not be stable.

The occurrence of Si in silicate minerals is in association with four O ions which are packed closely to form a four-sided *tetrahedron* (see Fig. 4-1B). By sharing O ions, numerous tetrahedra may be linked together to form a continuous structural system of relatively large size. The growth pattern may be organized in a manner characteristic of the particular

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[1] Identification of crystalline structures through use of X-ray diffraction analysis is discussed in most modern textbooks on Physical Chemistry.

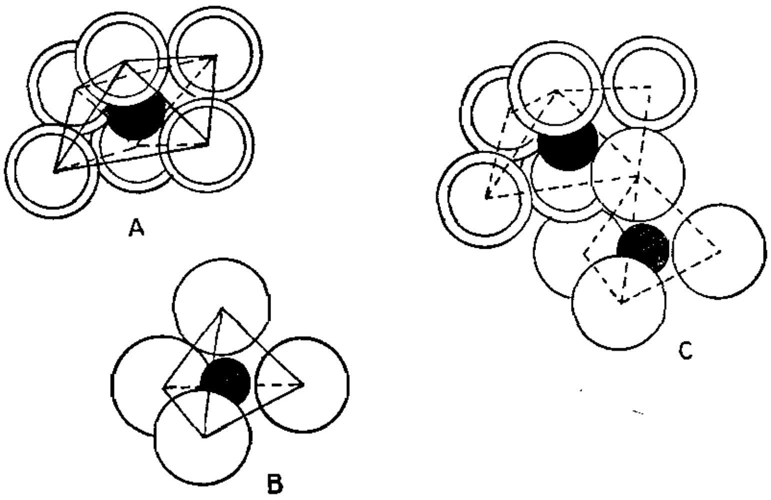


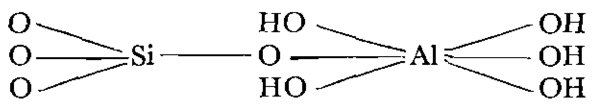
FIG. 4-1. Basic structural units of alumino-silicate minerals. Diagram A represents an octahedron formed by the enclosure of an Al ion by six hydroxyl ions. Diagram B illustrates the method of enclosure of a Si ion by four oxygen ions to from a tetrahedron. The octahedral and tetrahedral configurations are denoted by lines drawn between centres of the outer ions. Diagram C denotes the method of union between the two structural units wherein an oxygen ion from the tetrahedron takes the place of one of the hydroxyl ions in the octahedron. These diagrams are expanded and distorted somewhat to expose the enclosed cations.

mineral species of which the silica tetrahedra are a part. Where only these tetrahedra comprise the mineral, one of the many forms of silica (SiO_2) will result, quartz and amethyst being specific examples. If the pattern of growth is irregular, the resulting siliceous material will be amorphous.

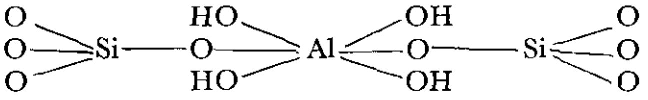
Cations other than Si occurring in silicate minerals also forms mall structural units with O or OH ions. Aluminium ions, for instance, occur most frequently in minerals in association with six oxygen or hydroxyl ions which form eight-sided *octahedra* (see Fig. 4-1A). Unless the alumina octahedra are linked directly to silica tetrahedra, the corner positions normally are occupied by hydroxyl ions (*i.e.*, as in the mineral *gibbsite*,

$Al(OH)_3$). Again, sharing of corner ions between adjacent Al octahedra allows for the development of a continuous octahedral system necessary for crystal growth, either in aluminium hydroxide crystals or those of the various alumino-silicate minerals.

Alumino-silicate minerals which develop within a weathering mineral mass are formed essentially through co-precipitation of Si and Al in their respective tetrahedral and octahedral configurations and in a layered arrangement. These layers are not separated, however, but are joined as the result of utilization of one O ion from a Si tetrahedron also as a corner ion for an adjacent Al octahedron. The bonding between a single Si tetrahedron and an Al octahedron may be pictured diagrammatically as:



The spacial relationship between the tetrahedron and the octahedron is illustrated in Fig. 4-1C. It is possible for two tetrahedra to be associated with a single octahedron. In this case, the octahedral portion is enclosed between the two tetrahedra as follows:



Either of these two configurations results in a stable chemical union. Where the ratio of tetrahedra to octahedra is 1 : 1, as in the first diagram above, the resulting mineral is said to be of the 1 : 1 type. Where two tetrahedra occur for each octahedron, the mineral is of a 2 : 1 type.

In the diagrams above, concurrent vertical extension of the Si and Al portions may be visualized as corresponding to crystal growth. The crystals formed are composed, therefore, of two separate parts, the *silica sheet* (or sheets) and the *alumina sheet.* The mixed layer of O and OH ions is considered as the boundary of these sheets, but, in reality, the mixed layer is shared by both. The completed structure formed by the tightly bound silica and alumina sheets is referred to commonly as a *crystal unit.*

The spacial arrangement of ions within a crystal is designated as the *crystal lattice.* Ions within a specific crystal species occur repeatedly in the same spacial relationship to each other and, thereby, form a characteristic lattice structure for that mineral. Those composed of sheets of tetrahedra and octahedra are known as *layer-lattice* minerals. Such structural form is characteristic of the bulk of alumino-silicate minerals which form in soils during weathering. Except for physically comminuted particles and precipitates of silica, lime, and the sesquioxides, layer-lattice minerals constitute a major part of the clay fraction of most soils.

A discreet layer-lattice mineral particle usually consists of more than a single crystal unit. Indeed, a particle within the range of clay sizes may have a thickness equalling several hundred times that of one of the crystal units comprising it. As might be expected, stacking of the platelets provides a means of crystal growth among the layer-lattice minerals along with lateral extension of the individual silica and alumina sheets. The physical stability of the stacked particles of layer-lattice minerals is a function of the force of bonding between adjacent crystal units.

Layer-lattice silicates are not limited solely to combined Al and Si sheets. Minerals formed particularly in environments rich in Mg may contain sheets composed of oxygen or hydroxyl octahedra with enclosed Mg ions rather than Al ions. Such layers are classed as *brucite sheets.*

Partial substitution for the Si and Al normally held in tetrahedral and octahedral positions occurs commonly during layer-lattice mineral formation. These substitutions are by ions of essentially the same size but often are of lower positive charge than are the Al and Si ions they replace. The resultant crystalline product, then, may be electrically unbalanced and, if so, will possess a residual negative charge. This charge deficit is compensated for partially by adsorption of positively charged cations to the surfaces of the crystalline particles.

CLASSIFICATION OF LAYER-LATTICE MINERALS

Consideration of previous discussions should indicate to the student that major differences in layered silicate minerals occur between those of 1 : 1 and 2 : 1 types. Within these major divisions, variation due to type and extent of cation substitution in the tetrahedral and octahedral layers provides a basis for further distinction among different minerals. Lastly, a separate and rather complex group of *mixed layer* minerals are recognized also. Basically, mixed layer minerals consist of 2 : 1 crystal units interlayered with other structural forms (*i.e.*, brucite sheets).

The Kaolinite Group

Members of this group of silicate minerals conform to the 1 : 1 type of lattice structure. The presence of but a single silica sheet for an alumina sheet in each crystal unit indicates a narrow $SiO_2 : R_2O_3$ ratio for these minerals, and suggests that they are the products of a relatively high degree of weathering. It is not surprising, therefore, that the principal member of this group, kaolinite, is a mineral prevalent in many highly weathered soils. It occurs in at least trace amounts in virtually all soils.

The structure of kaolinite, represented diagrammatically in Fig. 4-2, approaches an idealized 1 : 1 form in which negligible cation substitution for Al and Si has occurred. The lattice structure tends to be electrically balanced, therefore. Even so, this mineral does have some capacity for adsorbing cations, and the mechanism appears to involve a negative charge derived from broken oxygen bonds[1] at the edges of the crystal platelets. In acid soils, these sites are filled normally with H or Al ions which may be displaced by other cations.

The cystalline nature of submicroscopic particles of layer-lattice minerals is exemplified particularly by kaolinite. Electron photomicrographs, as presented in Fig. 4-3, show these mineral particles to be hexagonally shaped crystals for

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[1] These bonds, if occurring in the interior of the crystalline structure, are those responsible for the union of adjacent octahedra or adjacent tetrahedra.

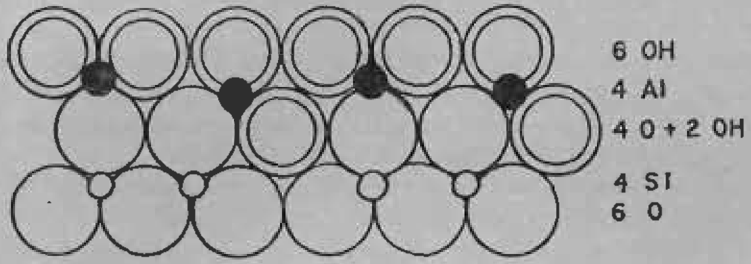

FIG. 4-2. Schematic diagram showing the approximate structure of an idealized kaolinite crystal lattice. Note that the union of a silica terahedron with an aluminium octahedron results in the replacement of hydroxyl ions by oxygens as part of the octahedral structures. Whereas the upper surface of the kaolinite crystal unit is made up of hydroxyl ions, the other is of oxygen ions only.

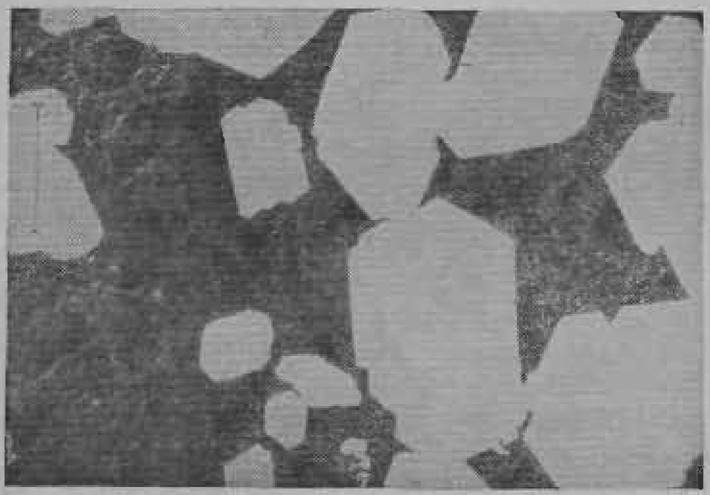

FIG. 4-3. Well-crystallized kaolinite particles from Georgia (U.S.A.) as photographed under an electron microscope. Some indication of the height of these crystalline stacks is afforded by the length of the shadows cast. The line shown represents a length of one micron (after Woodward, 1955).

the most part. The well-defined edges and corners of the particles suggest strongly that they are products of crystal growth, and that they have not acquired their particular size through physical comminution. One would expect clay-sized particles which have formed as the result of abrasion or grinding to display irregular edges and rounded corners.

The view of the kaolinite particles shown in Fig. 4-3 is perpendicular to the silica and alumina sheets. A side view of the crystals would provide some indication of their layered or stacked nature.

Discreet particles of kaolinite minerals generally are relatively large. This is due to the fact that the bonding between adjacent platelets is strong. The bonding force is derived largely from the sharing of H ions from OH groups in the alumina sheet in one crystal unit by the outer layer of tetrahedral oxygen ions in a neighbouring crystal unit. Thus, the stacked platelets orient themselves so that the silica and alumina sheets occur as alternating layers throughout the entire thickness of the particle.

The platelets making up the kaolinite crystal are bound together so tenaciously that ions and water cannot move in between them. If water could migrate between the platelets, the individual particles would swell on hydration and shrink on drying. However, the dimensions of kaolinite particles remain relatively constant throughout a wide range in moisture content, a fact which makes kaolinitic materials highly useful in the manufacture of pottery and other ceramic products. Occurring as relatively large particles with inherently low total surface area, kaolinitic materials display limited plasticity and stickiness. Because of a low degree of coherence among particles, soils containing a predominance of kaolinite show only a minimum capacity to form hard clods upon drying.

The intrinsic characteristics of kaolinitic crystals, then, tend to impart to soils rather desirable physical properties. However, because of their limited ability to hold cations important in the nutrition of plants, they do not contribute materially to the maintenance of soil fertility. Furthermore, occurring largely under conditions where weathering has been intense, soils high in kaolinite normally contain but limited reserves of other nutrient-bearing minerals.

Several layer-lattice minerals representing the 1 : 1 type of structure are classified within the kaolinite group. Among them, *halloysite*, *anauxite*, and *dickite* are referred to most frequently in the literature. Of these three, halloysite probably is the only one of importance in soils.

The Montmorillonite Group

A number of layered silicates go to make up this mineral group which acquires its name from its most prominent member, *montmorillonite*. All minerals in this group have a characteristic 2 : 1 structural configuration. However, the central octahedral layer does not always consist of an alumina sheet. Where a normal alumina sheet does occur and cation substitution within

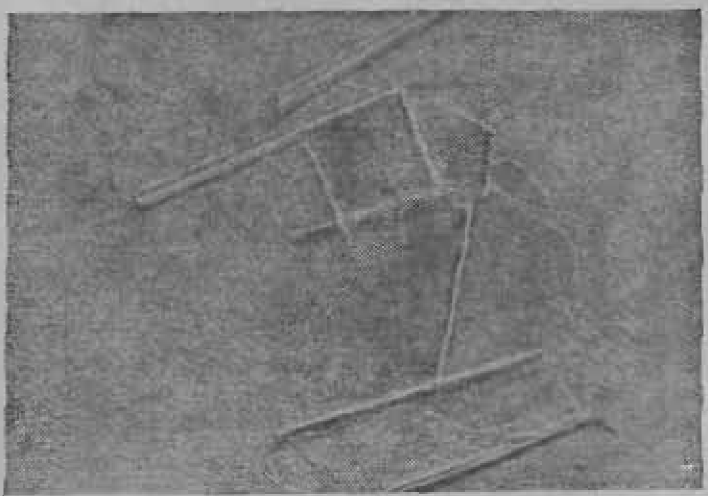

FIG. 4-4. Electron photomicrographs of highly dispersed montmorillonite. Note that the highly flexible nature of the thin laminæ permits folding and even conformation to the irregular surface of the carbon substrate on which the mineral flakes are deposited. Shadow measurement indicates these flakes to be about 0·001 micron thick. The line represents a length of one micron (after Nixon and Weir, Rothamstead Experimental Station, 1957).

the lattice is negligible, an electrically balanced structure is obtained. Such a structure is shown in Fig. 4-5 and corcorresponds to that of the layered mineral *pyrophyllite*. This mineral does not appear to be an important component of soils, however.

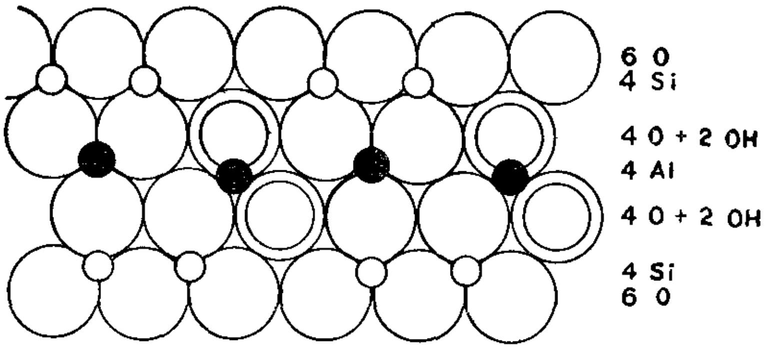

FIG. 4-5. Schematic diagram showing the approximate structure of the pyrophyllite crystal lattice. The replacement of hydroxyls normally occurring in association with aluminium by tetrahedral oxygens is very extensive. Both upper and lower sufaces of the crystal unit are composed only of oxygen ions, a condition typical of the montmorillitic group of layered silicates.

The mineral montmorillonite, which is very widespread in soils, has a structure very similar to pyrophyllite except that about one-sixth of the positions normally held by Al^{+3} are filled with Mg^{+2}. This substitution results in a charge deficiency which imparts to montmorillonite the capacity to adsorb cations to the outer surfaces of the crystal units. An essentially insignificant negative charge may result also from the presence of broken oxygen bonds occurring at the edges of the cystal units.

Nontronite is a second pyrophyllite-like mineral in the montmorillonite group. It is characterized as having complete or nearly complete replacement of the four central Al^{+3} ions by an equal number of Fe^{+3} ions. This type of substitution, of course, does not create a charge deficiency within the lattice. However, nontronite does have about the same capacity for adsorbing cations as does montmorillonite, and this is the result of the replacement of Si^{+4} by Al^{+3} in the tetrahedral layer.

A second series of minerals occurs within the montmorillonite group also. Its members differ from the pyrophyllite-like minerals in that six divalent rather than four trivalent cations occur in the central octahedral layer for each eight silicon ions. Magnesium and divalent iron are common cationic components of the middle sheet. When only Mg ions are present, the central layer is a typical brucite sheet. The 2 : 1 mineral formed through the union of a brucite sheet with two silica sheets is *talc*. Its structure is shown diagrammatically in Fig. 4-6. Like pyrophyllite, the lattice of talc is electrically balanced and is without a residual negative charge.

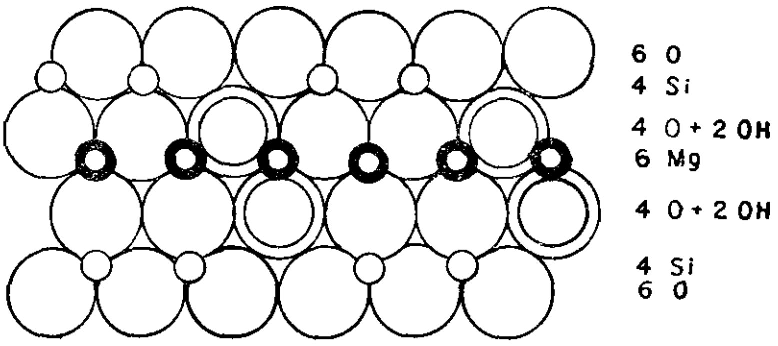

FIG. 4-6. Schematic representation of the crystal lattice of talc. Six divalent Mg ions are necessary to maintain electrical neutrality whereas only four trivalent Al ions are required for this in pyrophyllite and similar minerals. Talc-like minerals in soils have physical and chemical properties very much like those of montmorillonite.

Probably the most common talc-like mineral in soils is *saponite*. Its typical structure appears also to have positions within the central octahedral layer filled completely with Mg. However, as in nontronite, a negative charge essentially equal to that typical of montmorillonite results from the substitution by Al^{+3} for Si^{+4} in the tetrahedral layers. Other members of this mineral series are *hectorite,* which results from the partial substitution of Mg^{+2} by Li^{+}, and *sauconite,* in which Zn^{+2} occurs in the place of Mg^{+2} in the octahedral layer.

Aside from montmorillonite, none of the above-named minerals appear to occur in soils in significant quantities. Another mineral is referred to frequently which has the properties of montmorillonite and seems to be reasonably widespread in soil materials, however. This is *beidellite.* Structurally, beidellite is not markedly different from montmorillonite, although it has some iron in the octahedral layer and is thought to acquire some of its negative charge from substitutions for Si^{+4} in the tetrahedral layer.

As may be noted in Fig. 4-5, crystal units of 2 : 1 minerals have outer layers of oxygen ions only. Thus, in the absence of H bonding as occurs in the kaolinite crystal, attraction between adjacent platelets in montmorillonitic particles is weak. As a consequence, water and ions can move readily in between individual platelets making up the crystalline particles of the montmorillonitic minerals. Hydration and movement of water into the interlayer spaces causes swelling of the individual particles and of the soil mass containing them. This gives rise to the descriptive classification of members of the montmorillonite group as *expanding lattice* minerals.

Because their constituent platelets are split apart easily, montmorillonitic minerals persist in a higher state of dispersion than do kaolinitic minerals. As may be noted from electron microphotographs (see Fig. 4-4), montmorillonitic particles consisting of but a very few platelets are not rigid ; instead, they are capable of flexing and conforming even to irregular surfaces. The close contact this allows results in a relatively high degree of coherence among particles. As might be expected, soils high in clays composed largely of easily dispersed montmorillonitic minerals tend to be sticky and plastic when wet and to form hard clods when dry. They have a relative high capacity for adsorbing cations and water, the latter capacity contributing to the high degree of swelling they display. According to Wear and White (1951), those members of the montmorillonite group which are characterized as having more extensive cation substitutions within the lattice, particularly within the silica sheet, are noted also to have a high capacity for adsorbing cations.

Those montmorillonitic minerals containing Fe or Al as the principal octahedral cation have a higher $SiO_2 : R_2O_3$ ratio than do the members of the kaolinite group. This relates to the fact that the first-named minerals occur in soils which have undergone a lower degree of weathering. However, since some of the expanding lattice minerals contain normally very limited amounts of the sesquioxides, it is not safe to characterize the degree of weathering in all soils through evaluation of the $SiO_2 : R_2O_3$ ratio of the clay fraction only.

The Illite Group

As with montmorillonitic minerals, members of the illite[1] group are composed of stacked 2 : 1 crystal platelets. These two general groups differ markedly, however, in that illitic minerals are *non-expanding.* This results from the fact that an appreciable number of K ions are retained in the interlayer spaces and serve to bind platelets together as do H ions in kaolinitic crystal packets. Illite formed as the result of normal weathering appears to result from slight modification of original layered minerals high in K also (*i.e.*, micaceous minerals). Thus, the presence of a high proportion of illite in the finer soil fractions is assumed usually to indicate that weathering of the mineral matrix containing it has not been extensive. On the other hand, many illite-like minerals are assumed to have formed in sediments through the adsorption of K ions by expanding lattice 2 : 1 minerals while in contact with K-rich sea water.

The rather unusual ability of the K ion to bind platelets together results from the fact that its size permits it to fit more snugly in depressions occurring in outer layer of oxygen ions in the 2 : 1 mineral lattice. These outer layers are not filled completely with oxygen ions but, instead, consist of a network of hexagonally shaped holes bounded by tetrahedral oxygen

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[1] As yet, illitic minerals are not classified completely. They are referred to frequently as *hydrous micas.*

ions as illustrated in Fig. 4-7. The size of the hexagonal holes is such that adsorbed K ions will move into them to a

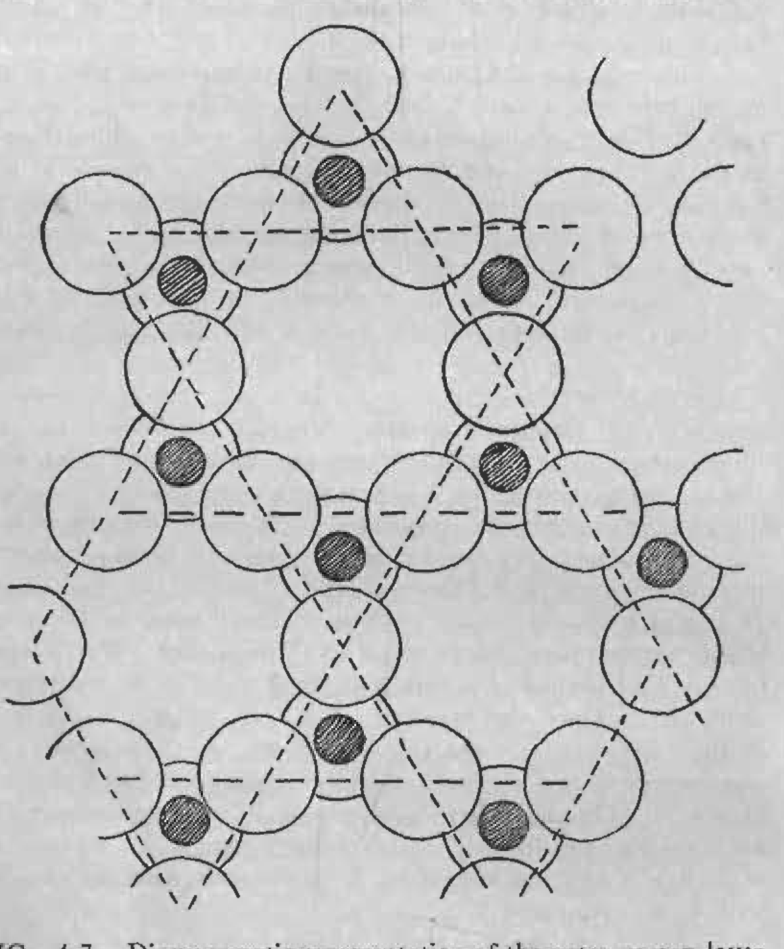

FIG. 4-7. Diagrammatic representation of the outer oxygen layer of a silica sheet showing hexagonally shaped depressions within which adsorbed ions may be retained. Each surface-layer oxygen occurs as a component of two contiguous tetrahedra. The structures shown are expanded somewhat to expose the silicon ion (crosshatched circle) normally enclosed within the tetrahedra, and the fourth oxygen ion (interrupted circle) which occurs in a plane above the silicon ion.

distance equal to about one-half their diameter. Thus, the remaining half of the adsorbed ion protrudes above the surface and will fit equally well into a hexagonal hole in the oxygen layer of an adjoining platelet. Sharing the charge of the K ions tends to pull the platelets together. Because of the precise fit of the K ion, neighbouring platelets may come into intimate contact and are bound together tightly as a result. Other ions of essentially equal size to K, particularly NH_4, are capable of binding together platelets of 2 : 1 minerals. Potassium, however, occurs naturally in much larger quantities than does NH_4 and, therefore, serves as the major bonding ion in illitic minerals.

The ability of a 2 : 1 mineral to display properties characteristic of the illite group depends partially upon the nature of the negative charge within the basic lattice structure. The ability to adsorb K with sufficient tenacity to bond adjacent platelets securely increases apparently with *major substitution* in the silica sheets. According to Wear and White (1951), those minerals in which principal substitution takes place in the alumina sheet exhibit a limited force of attraction for the adsorbed ion. This is because of the distance from the site of the negative charge to the adsorbing surface would be greater than if the substitutions were predominantly in the silica sheet. Substitution for Si^{+4} in the tetrahedral layer by Al^{+3} occurs extensively in illite.

Physical and chemical properties of illitic minerals are intermediate between those of the kaolinitic and montmorillonitic groups. Since adsorption of ions occurs mainly on external surfaces of the non-expanding crystal packets, this capacity in illite will be less than that for montmorillonite. The loss of K ions from interlayer positions leaves sites for adsorption of other ions, however.

Vermiculites

Vermiculites of various types are widespread in soils. Structurally, they are 2 : 1 minerals with a limited capacity for expansion upon hydration. Since they do expand, cations can migrate in between layers of the crystal units where they may become adsorbed. Vermiculites, like illite, have a characteristically high degree of ionic substitution within the silica sheets and a high cation-adsorbing capacity. The high affinity for cations contributes to their limited ability to expand. As a matter of fact, when placed in K-rich media, the adsorption of K within oxygen-layer holes results in a detectable contraction of the micelle because of the strong attraction between adjacent platelets and the K ions. The K ions held within the holes are not easily displaced and are said to be *fixed.* The fixation of K converts vermiculite into a non-expanded micelle with properties much like those of illite. The ability of vermiculite to adsorb other cations in interlayer positions is reduced measurably as a consequence to K fixation.

Montmorillonite, like vermiculite, will adsorb K in positions of fixation also. However, actual fixation and contraction of the micelle does not occur unless the mineral is dehydrated thoroughly, as by heating. Vermiculite, on the other hand, contracts and fixes K while in the moist state. The difference between these two minerals is attributed to the fact that the negative charge arises largely from the central alumina sheet in the montmorillonite crystal, whereas a more effective charge in vermiculite is derived from substitutions in the silica sheets. This allows for a much stronger force of attraction between the vermiculite platelets and the adsorbed K ions.

Chlorites

Members of this group of silicate minerals are representative of the mixed-layer types. The idealized chlorite structure is depicted as consisting of alternating layers of brucite and a 2 : 1 crystal unit (*i.e.*, talc). The development of the chlorite crystal appears to result from the deposition of the brucite layer within a pre-existing 2 : 1 lattice mineral in a Mg-rich medium ; for instance, chlorite is a common mineral in soils derived from ocean sediments. Aluminum may substitute for Mg in the brucite layer of chlorite. Being a non-expanding 2 : 1 mineral, chlorite possesses many characteristics which are similar to those of illite. The configuration of chlorite does not give rise to retention of K ions in oxygen layer holes, however.

Iron and Aluminum Minerals in Soils

Various Fe and Al compounds occur in soils. The dominant forms are the oxides, hydroxides, and oxyhydroxides. Since the two hydroxy forms lose water on heating and are converted to oxides, they sometimes are referred to as hydrated oxides. As a matter of convenience, it is customary to refer to this group of compounds as the *sesquioxides.*

The sesquioxides assume various crystalline forms in the soil, but all have one general feature in common ; they consist of many layers of O or OH ions enclosing the metal cations, Fe or Al. These layers may build up through crystal growth in any direction and, as a consequence, can cause the formation of relatively large particles. In fact, in highly weathered soils where large quantities of these compounds have accumulated, almost continuous layers of crystallized Fe and Al compounds may be found. Often gravel-sized concretions of these materials occur scattered throughout the profile of some moderately to intensely weathered soils.

Ideally, the structures of sesquioxide crystals are electrically balanced. There should be little tendency, therefore, for them to adsorb cations if interatomic spaces within the lattice are filled completely. Some sesquioxides have been observed to adsorb appreciable quantities of cations, however, a characteristic assumed to result from a shortage of Fe and Al ions within the lattice structure (Kanehiro and Sherman, 1956).

Like kaolinite, Fe and Al compounds accumulate as a result of extensive weathering whereby silica has been lost more rapidly than have the sesquioxides. Since the sesquioxides normally have a low capacity for adsorption of cations and water, the highly leached soils in which they commonly occur are noted for their inherently low level of fertility.

Where present in the soil in abundance, the sesquioxide minerals are characteristically well aggregated. Clays composed largely of these minerals, therefore, are not sticky, plastic, or clod-forming. The sesquioxides can occur in a finely divided, colloidal state, and often, when in this form, appear as thin films over less-weathered mineral particles. The more common forms of Fe and Al oxides are, respectively, *hematite* and *gibbsite*. These minerals impart vivid red and yellow colours to soils.

ENVIRONMENT AND SOIL MINERAL TYPE

Some very general relationships are known to exist between specific weathering environments and type of mineral clays persisting in soils. Environment affects the occurrence of organic colloids in the soil also. Both the production of plant residues and their accumulation or disappearance from the soil are closely related to soil environment, particularly as influenced by climate. The weathering environment cannot be expressed solely in terms of externally imposed factors of weathering, such as amount of leaching water and temperature. Involved as well are contributing factors such as the composition of the rock being weathered and the nature and amount of organic residues present. Consideration must be given also to the depth in the weathering profile when speaking of specific environments, for, as leaching waters are changed in composition while they pass through the profile, so is the weathering environment of which the leaching waters are a part.

The course of chemical weathering depends not only upon the quantity of leaching water passing through the weathering medium but also upon its content of acid- or base-forming ions. If the rate of water flow is slow, there is greater opportunity for solution effects to build up a substantial soluble cation concentration. This tendency will slow the rate of weathering. If, on the other hand, the quantity of water entering the weathering zone is large, percolation may be relatively rapid, and the concentration of soluble reaction products seldom reaches a significant level. This enhances weathering. Because of this behaviour, we may express the rate of weathering qualitatively in terms of the total amount of water which percolates through the weathering zone. The effect of variation in the amount of leaching water, then, is one of controlling the level of base-forming cations, silica, and other reaction products which may aceumulate in the hydrosphere.

Base-forming cations influence weathering reactions through the neutralization of acids. An increase in acidity not only encourages hydrolysis and dissolution of the weathering minerals, it also influences the solubility of reaction products and the rate of their removal from the zone of weathering through leaching. This fact appears to account partially for the differential removal of either silica or the sesquioxides from the weathering mineral mass.

Neither kaolinite nor the sesquioxides appear in quantity in the clay fraction of soils forming predominantly under conditions of limited leaching. Where K-rich minerals weather and the soluble products are not lost, illite-type minerals appear in considerable abundance. If chemical weathering is limited, the clay fraction of the soil may contain a relatively high proportion of physically disintegrated mineral particles. With an increase in weathering and leaching, but not to the point where soluble weathering products are removed rapidly from the profile, montmorillonite may become one of the more dominant minerals in the clay fraction. This is particularly true if there is an abundant supply of divalent cations, especially Mg, in the system. It is assumed often that illite is formed first under these conditions and, through eventual loss of interlayer K, is converted subsequently to montmorillonite or possibly to vermiculite.

Whereas the foregoing observed relations refer primarily to gross climatic effects, local variation in the environment, both external and internal, may alter these effects markedly. For instance, the weathering of limestone rock rich in Ca and Mg may result in the formation of montmorillonite even though soils in neighbouring areas, which are forming on igneous rock, may contain substantial quantities of kaolinite or gibbsite, $Al(OH)_3$.

Topographic conditions often contribute to pronounced differences in weathering. Soil-forming materials on steep slopes lose much of the precipitation they receive as runoff. Thus, in the light of limited percolation, only a minimum of mineral transformation may be expected. Furthermore, soils occurring on steep slopes are subject to excessive erosion.

As a consequence, new minerals resulting from weathering reactions may be lost about as rapidly as they are formed. Under these conditions, then, the prevailing climate would not serve as a useful index to the apparent rate of weathering.

An additional interesting example of the effect of topography is the formation of kaolinitic minerals in red soils of the Deccan Plateau and of montmorillonitic clays in adjacent black soils lying in lower positions. Whereas the red soils are well drained, the black soils are not. Drainage waters from the higher-lying areas tend to accumulate in the lower-lying positions and, because of their relatively high content of divalent cations carried in from the more elevated positions, encourage the formation of the montmorillonite mineral (Mukerji and Agarwal, 1943).

The generalized relationship between the effects of different weathering intensities and resultant soil mineral type has led to the characterization of the degree of weathering in terms of the dominant mineral type in the clay fraction of the soil. Simply, this means that easily weathered minerals persist in the clay fraction of the soil only if chemical weathering has been essentially negligible, and that the more stable minerals are dominant if the weathering has been intense or of long duration (Jackson and Sherman, 1953). It is obvious, of course, that if the materials on which a soil is forming have been transported from other areas, their observed form may be more closely related to their previous rather than their present environment.

It is concluded that when weathering is intense, mineral transformation will proceed in an orderly manner starting with the unaltered forms and will pass through the various intermediate forms until only the more resistant compounds, including the sesquioxides, remain. The sequence of stepwise transformations normally is depicted as occurring in accordance with the following scheme:

$$\begin{matrix}\text{Unweathered}\\ \text{minerals}\end{matrix} \rightarrow \text{Illite} \rightarrow \left\{\begin{matrix}\text{Chlorite}\\ \text{Vermiculite}\end{matrix}\right\} \rightarrow \begin{matrix}\text{Montmoril-}\\ \text{lonite}\end{matrix} \rightarrow$$

$$\text{Kaolinite} \rightarrow \text{Sesquioxides.}$$

Mineral weathering will not necessarily proceed through all intermediate stages until the stable sesquioxide form appears. Frequently, transformations to highly stable products occur which bypass various of the intermediate stages. For instance, Jackson (1957) observed vermiculite to form directly from mica without passing first through the illite stage. If illite were an intermediary mineral in this case, its transformation to vermiculite was too rapid to permit accumulation in significant quantities. Tanada (1944) observed that, in relatively uniform material of volcanic origin in Hawaii, either kaolinite or the sesquioxides formed directly from the original minerals. Kaolinite occurred if rainfall was sometimes limited, but gibbsite formed directly where rainfall was abundant and leaching essentially constant.

Sherman (1952) concludes that the exact nature of mineral transformation during weathering will depend upon the frequency and duration of alternating wet and dry periods. Under conditions where free leaching occurs during periods of moderate to high rainfall, the normal trend in weathering is toward the formation of gibbsite. Intermediate products, montmorillonite and kaolinite, will form if dry periods occur which permit at least temporary retention of silica in the weathering medium. The longer the periods of drouth the greater is the tendency toward montmorillonite formation.

It should not be concluded from these considerations that degradative weathering of soil minerals results always in eventual conversion to the sesquioxides. Should illite be stable in the weathering environment and should this environment persist indefinitely, there is no reason to assume that alteration to other forms would take place.

As has been noted in an earlier section, at least three-quarters of the soil-forming materials are of sedimentary origin. The history of the sediments will determine to a large extent the type of clay minerals present in soils forming from them. It is known that normal weathering processes are reversed during the period when the sediments are submerged in saline waters. Powers (1954) and Grim and Jones (1954) recognize the formation of chlorite and illite from montmorillonite and vermiculite in marine sediments. Weaver (1957) indicates that alterations of this type probably result from very simple chemical reactions within the submerged deposits. In his opinion, the changes may be no more extensive than the reabsorption of K and Mg ions by expanding lattice minerals to reform illite and chlorite. It appears probable that much of the illite-type of clay minerals present in soils have been inherited from sediments and occur less frequently as products of degradative weathering of other, less stable mineral compounds.

REVIEW QUESTIONS

1. Why is it important to the nutrition of plants for soil minerals to decompose slowly through weathering reactions ?

2. Differentiate briefly among the three major types of rocks occurring within the earth's crust. Which type serves as the principal source of soil-forming material ?

3. Explain the major differences between physical and chemical breakdown of rocks and minerals. Specify conditions which determine the predominating influence of either of these two processes in mineral weathering.

4. What is probably the most important consequence of physical breakdown of mineral matter ?

5. Define till, alluvium, and loess. Explain why wind-laid deposits frequently contain particles of limited variation in size.

6. Explain why mineral solution is largely a hydrolytic process. Why is chemical weathering most effective if soluble products are carried away in percolating water ?

7. What is the relationship between the time and intensity factors in mineral weathering ?

8. Explain the reason why layer lattice minerals possess a negative charge and how this charge is neutralized. Why does the negative charge vary from mineral to mineral ?

9. What are the similarities and differences between the methods of platelet bonding in kaolinitic and illitic minerals ? What is the relationship between the strength of bonding between platelets and (1) mean particle size of the layer silicate minerals, (2) the ease of dispersion, and (3) swelling and shrinkage ?

10. Explain the statement "Mineral weathering results in the formation of compounds which tend to be stable in the environment of their formation." How can this concept be applied to explain formation of illitic minerals either from degradative or constructive weathering processes ?

SELECTED REFERENCES

1. Grim, R. E., *Clay Mineralogy*, McGraw-Hill Book Co, Inc., New York, N. Y., 1953.

2. Marshall, C. E., *The Colloid Chemistry of the Silicate Minerals*, Academic Press, New York, N. Y., 1949.

3. Searle, A. B. and Grimshaw, R. W., *The Chemistry and Physics of Clays and Other Ceramic Materials*, Interscience Publishers, Inc., New York, N. Y., 1959.

CHAPTER FIVE

EXCHANGE REACTIONS AND pH CONTROL

CATION ADSORPTION AND EXCHANGE

CATIONS adsorbed to the surfaces of soil mineral particles by the negative charge within the crystal lattice are held much more loosely than are the ions occurring as a component part of the lattice structures. As a consequence, they may be displaced with relative ease by other ions which, in turn, serve then to neutralize the lattice negative charge. Since the surface-held ions may be replaced by other ions, they are said to be *exchangeable.* By this definition, they are distinguished from both nonreplaceable ionic components of the crystal lattice and from soluble ions occurring in the soil solution adjacent to soil particle surfaces.

The fact that ions occur in exchangeable form is of great importance to the nutrition of plants. These ions are subject to ready release to plant roots and, yet, they are not lost rapidly from the soil through leaching. If this were not true, it would be essentially impossible to maintain the fertility status in areas where leaching water passes through the soil in relatively large amounts.

The negative charge of minerals derived from ion substitution and from broken oxygen bonds does not serve as the only mechanism for cation adsorption in soils. Important are various organic materials which, by virtue of carboxyl (-COOH) and phenolic hydroxyl (-OH) groups they contain, are capable of exchanging cations also. Exchange reactions take place through the displacement of H ions from these organic groups. However, since these H ions are chemically bound in the organic groups, they are not easily replaced by metallic cations in acid media where the concentration of H ions is significantly high. Nonetheless, these exchange positions in organic matter are important in cation retention and, with those contributed by the mineral fraction, form the *exchange complex* of the soil.

The Nature of Exchange Reactions

The fact that cations held at soil particle surfaces and those in the soil solution are in constant motion accounts for their entering exchange reactions. The movement of adsorbed ions away from the particle surfaces, which is nothing more than ionization, leaves exchange sites vacant which then may be occupied by other cations derived from the adjacent solution. The retention of a soluble ion with the resultant exclusion of the one originally held constitutes a completed exchange reaction. Obviously, the probability of its ever taking place is increased with increasing numbers of ions in the solution phase. Thus, when soluble ions are added to a soil, exchange will take place to an extent depending upon the total number of ions added. The exchange reaction will proceed until an equilibrium between the adsorbed and soluble forms of all ions present is attained, and there will be a positive relationship between the final proportions of the various adsorbed ions and their respective concentrations in the surrounding solution. As an example, the exchange equilibrium which results when only two cations are involved is described qualitatively by the expression

$$\frac{A_{ad}}{B_{ad}} \rightleftharpoons \frac{A_{sol}}{B_{sol}}, \qquad (5\text{-}1)$$

where A_{ad} and B_{ad} are quantities of the two different cations adsorbed by the exchange complex, and A_{sol} and B_{sol} their respective concentrations in the associated solution which bathes the soil particle surfaces. We may see, then, that increasing either A or B in the solution phase will change the proportions of these two ions present in soluble form and, in turn, will result in a shift in their proportions on the exchange complex.

Equation 5-1 serves to describe a variety of exchange reactions. If B is the added cation and is accompanied by an anion which precipitates the original exchangeable cation, A, then the concentration of A in the soil solution is maintained at a low level. As a consequence, the added cation is adsorbed more completely. If an ion such as one in a fertilizer salt is added to the soil and then washed beneath the surface where it enters into an exchange reaction, the displaced ion originally on the exchange complex will be swept downward by the percolating water. Thus, at the point of exchange, the concentration of the displaced ion is kept at a minimum, and the ion added in the fertilizer salt is taken up readily on particle surfaces in the upper soil layer.

The ions from even a very dilute solution can be adsorbed completely in the upper soil layers as they percolate through the soil. Because of this, a weakly acid percolate, such as water charged with a small quantity of dissolved CO_2, is very effective in removing the original, adsorbed ions. This phenomenon accounts for the relatively complete removal of basic cations during weathering under high-leaching regimes.

Different cations are held by the soil with different tenacities. A tightly held ion will displace one loosely held with comparative ease. Considered in another way, if equal concentrations of two different ions are added to and mixed with the soil, the one for which the soil has the greatest affinity will be adsorbed in greater amounts. Cations have been rated as to their relative power of replacement and, generally, are found to conform to the following series:

$$\mathrm{Al} > \mathrm{Ca} = \mathrm{Mg} > \mathrm{K} = \mathrm{NH_4} > \mathrm{Na}.$$

The H ion will occur also in this series, but its position is open to question. It once was thought to be adsorbed more tightly than either Ca or Mg. The difficulty in determining the place for H arises from the fact that, when added to the soil, it enters into numerous reactions which are not exchange (*i.e.*, hydrolysis and solution of soil minerals).

The relative adsorbability of an ion is related to its *charge density*, which is a function of the valence and effective diameter of the ion. A polyvalent ion with small effective diameter will have a high charge density, therefore.

The principal ions held by the exchange complex in soils are those shown in the above replacement series. The cations Zn, Mn, and Fe, which also are essential plant nutrients, occur as exchangeable ions in but very small quantities only. Their scarcity both on the exchange complex and in the soil solution is due to the fact that they tend to form insoluble compounds in the soil.

Exchange Reactions and Acidity and Alkalinity in Soils

The degree of acidity or alkalinity in soil is a function of the H-ion and OH-ion concentrations in the solution phase. The concentration of these two ions is controlled largely by the type and amount of the various cations which are adsorbed in exchange positions. Being rather loosely held and subject to displacement in accordance with exchange equilibria, adsorbed ions can ionize, and, as the result of hydrolysis, their positions on the exchange complex are taken by H ions derived from the dissociation of water. The hydrolysis of exchangeable Na is very pronounced and the reaction involved is illustrated as follows[1]:

$$\boxed{\text{Clay}}\,\text{Na} + \text{H}^+ + \text{OH}^- = \boxed{\text{Clay}}\,\text{H} + \text{Na}^+ + \text{OH}^-. \tag{5-2}$$

The loss of H ions from the solution through the displacement of Na results in a relative increase in the OH-ion concentration which makes the solution more alkaline. If Ca and Mg were the exchangeable ions, the hydrolytic increase in alkalinity would be less than for Na. These divalent ions are more tightly held and, therefore, are replaced by H with greater difficulty. Further, the bases they form are not so highly ionized as is NaOH.

Aluminium is a common exchangeable ion in acid soils. Like Na, Ca, and Mg, it will hydrolyze also to form a base; in this instance, $Al(OH)_3$. This compound is insoluble even in moderately acid media. As a consequence, OH ions are precipitated from solution and it becomes more acid. Further, the H ions which replace Al ionize to a degree and contribute also to the acidity of the system.

If a neutral salt such as $CaSO_4$ is added to an acid soil, the resultant displacement of adsorbed H and Al will cause

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[1] The representation, $\boxed{\text{Clay}}$ Na, is intended to show symbolically a clay micelle holding adsorbed Na and corresponds to a Na-clay molecule of indefinite size.

an increase in acidity. The reaction may be described as follows:

$$\boxed{\text{Clay}}\begin{matrix}H_2\\Al_2\end{matrix} + 4CaSO_4 = \boxed{\text{Clay}}Ca_4 + H_2SO_4 + Al_2(SO_4)_3 \qquad (5\text{-}3)$$

The $Al_2(SO_4)_3$ hydrolyzes in turn to form sulfuric acid:

$$Al_2(SO_4)_3 + 6HOH = 2Al(OH)_3 + 3H_2SO_4 \qquad (5\text{-}4)$$

Since $Al(OH)_3$ is essentially insoluble in soils, OH ions are removed from solution and, therefore, will have no opportunity to counter the effect of the H_2SO_4 which lowers the pH.

The concept of acidity or alkalinity of a solution, as discussed heretofore, has been considered purely in the light of the concentration of H and OH ions. However, since the variation in concentrations of these two ions is of great magnitude, it is much more convenient to use pH values[1] which express concentrations in terms of logarithms. The pH or *reaction* of soils ranges from about 3·5 to over 10. Those of low pH have exchange surfaces filled largely with Al and H ions, whereas soils of extremely high pH contain essentially only Na in exchange positions. That is to say, they are Na *saturated* or nearly so. Often, appreciable quantities of exchangeable K occur in strongly alkaline soils also. Most agricultural soils considered suitable for moderately good to very good production have pH values lying between about 5 and 8·5. Soils having pH values below 5 are considered to be strongly acid, and those above 8·5, moderately to strongly alkaline.

The Cation-Exchange Capacity

The total quantity of exchangeable ions adsorbed by a soil may be determined if the ions are displaced into solution,

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[1] Specifically, a pH value is the log of the reciprocal of the H-ion concentration, or

$$pH = \log\frac{1}{(H^+)},$$

where (H^+) is the concentration of H ions in mols. per l. Use of the reciprocal expression results in numerically positive values. Usually the pH of a solution ranges from something less than one, as in a very acid solution, to about 14. The pH of pure water is 7 and corresponds to neutrality. At this point, H and OH ions occur in equal though very dilute concentrations (10^{-7} mols per l.).

washed from the soil, and measured quantitatively. This quantity, referred to as the *cation-exchange capacity*, or simply the *exchange capacity*, normally is expressed as me. per 100 g. of soil. It measures indirectly the total negative charge attributable to the mineral and organic fractions of the soil.

The salt used for displacement should not contain one of the ions being displaced, because the quantity of this ion derived through exchange then could not be measured. The concentration of the salt solution used should be sufficiently high to permit as complete displacement of original exchangeable ions as possible without using excessive quantities of the solution. In a leaching system such as one with which we are concerned, the total number of ions added in the displacing solution determines the completeness of displacement for the most part.

Mere leaching of the soil with a neutral or weakly acid, concentrated salt solution will not displace readily those H ions held by broken-bond oxygens and organic radicals because of the strong affinity of these exchange sites for the H. Removal of the H ions is more nearly complete if an anion is present in the leaching solution which also has a strong affinity for the H. Hydroxyl ions serve this purpose most readily since they combine with H to form slightly ionized water. Thus, increasing the pH of the leaching solution through the addition of OH ions will encourage more complete displacement of exchangeable H.

The determination of the cation-exchange capacity by separate analysis of each of the original cations displaced from the soil is a long and tedious procedure. It is considerably simpler to displace these ions by a single, easily-determined cation which also may be replaced subsequently and then measured in a single determination. The quantity of ion adsorbed from the single salt solution will, of course, depend upon the pH of the added solution, since the higher the pH the greater will be the displacement of the H ions adsorbed by the soil originally.

Conventional methods of exchange-capacity determination involve the use of such salts as ammonium or sodium acetate or sodium or barium chloride to provide the single displacing ion. The solutions of these salts are 1 N in concentration or greater and their pH usually is adjusted to either 7·0 or 8·2. Neither of these pH values is high enough to cause complete displacement of all replaceable H, however. Therefore, the determined value for cation-exchange capacity reflects merely the ability of the soil to adsorb cations other than H at the pH value of the solution used in the determination. In defining or expressing the cation-exchange capacity of a soil, therefore, it is necessary to specify the pH at which the value is determined.

The cation-exchange capacity denotes the ability of a soil to hold nutrient cations in readily available form. More important to growing plants, however, is the quantity of nutrient cations present in exchangeable form, not just the total capacity of the soil to hold them. For example, a soil with a high exchange capacity and a high proportion of exchangeable H and Al may be much less fertile than a soil with a low adsorptive capacity but a high proportion of exchange positions filled with essential nutrient cations. Yet, knowledge of the exchange capacity of acid soils is important for it gives some indication as to the amount of nutrient cations which should be added to displace the H and Al and thereby reestablish the nutrient status of the soil at a more nearly optimum level. The addition of Ca as gypsum, to correct for excessive exchangeable Na in strongly alkaline soils, and as lime to neutralize strongly acid soils is based largely on the capacity of the soil to adsorb the Ca.

Field treatment of acid soils seldom will result in elevating the pH above neutrality. As a consequence, evaluation of the cation-exchange capacity at pH 7·0 is probably as suitable as would be a value determined at 8·2. Alkaline soils, and particularly those that are calcareous, often occur at a pH near 8·0. Thus, it is advocated frequently that the cation-exchange capacity of these soils be determined at pH 8·2. This is the pH of a soil essentially saturated with exchangeable Ca and containing free lime in equilibrium with CO_2 in the air. Specific methods for determining the cation-exchange capacity of soils are discussed by Piper (1950), Bower *et al.* (1952), and Mehlich (1948).

Coleman *et al.* (1958) refer to the negative charge of the mineral lattice as the *permanent charge* in soils. The charge due to broken oxygen bonds and to acidic organic groups is designated as the *pH-dependent charge*. Soils high in kaolinitic minerals, which acquire a large part of their cation-adsorbing ability from broken oxygen bonds, hold an appreciably high proportion of exhangeable ions by the pH-dependent charge (Coleman and Mehlich 1957 ; Jenny, Ayres, and Hosking, 1945). Minerals of a 2 : 1 configuration, which have a relatively high permanent charge, show but a limited effect from variation in pH on the cation-exchange capacity. Among any of the major mineral types in soils, the effect of pH is most pronounced above neutrality.

The negative charge of organic materials appears to be entirely pH dependent. Organic matter has a negligible ability to hold cations other than H at a pH of 4·0 ; whereas, at pH of 7·0, it may retain these cations in greater quantities than can any of the soil minerals which enter into exchange reations.

One-normal ammonium acetate adjusted to pH 7·0 has been used widely as the exchange-saturating solution in cation-exchange-capacity determinations for many years. However, the diameter of the NH_4 ion is almost the same as that of K. Thus, when used on soils high in vermiculite, the NH_4 ion also is taken into oxygen-layer holes within the crystal lattice and can become fixed against subsequent extraction. This reduces the measured exchange capacity of the soil sample below that value which would be obtained if a smaller saturating ion were used. For this reason, Sawhney *et al.* (1959) recommend the use of either Na or Ca ions in both the saturating and displacing phases of the cation-exchange-capacity determination.

Cation-Exchange Capacities of Clays and Soils

There is considerable variation in the cation-exchange capacities among the several minerals common to clays. Generally, the values are high for the expanding-lattice types and low for the fixed-lattice minerals. Representative exchange-capacity values for the major mineral groups and for organic matter are shown in Table 5-1. The values listed are for determinations at a pH of 7·0.

Table 5-1. Representative cation-exchange capacities of the various common exchange materials in soils.

Colloid	Cation-Exchange Capacity
	me./100 g.
Organic matter	200-400
Vermiculite	100-150
Montmorillonite	60-100
Illite	20-40
Kaolinite	2-15
Fe and Al oxides	0-100+[1]

[1] Kanehiro and Sherman (1956) report exchange capacities for highly ferruginous (iron-rich) soils in Hawaii of slightly greater than 125 me. per 100 g. These capacities are attributed to the presence of hydrated iron oxides, possibly containing insufficient Fe ions to fill completely crystal lattices.

A comparison among soils containing the same types and proportions of exchange material will show that the exchange capacity increases with clay and organic matter contents. In the particle-size analysis of soils, removal of organic matter during the determination eliminates a fraction of the soil active in exchange reactions. Therefore, in soils containing appreciable organic matter, there may be no specific relationship between textural analysis and the exchange-capacity value determined on the whole soil. Furthermore, some mineral particles of silt size have the capacity to adsorb cations in appreciable quantity. Vermiculite, in particular, is found often to persist as particles with these dimensions.

Cation-exchange-capacity values for whole soil will range generally between 2 and 60 me./100 g. Some clayey soils in humid regions, such as East Pakistan and Southern India, often are high in kaolinite and sesquioxides. They may have exchange capacities that compare with much sandier soils in the more arid regions which have montmorillonite as an important type of clay present. Some representative exchange-capacity values (pH 7·0) for soils of different regions are listed in Table 5-2.

Table 5-2. Exchange-capacity values for whole soils from various locations in the Indo-Pakistan subcontinent.

Location and Description	Clay content	Exchange capacity	Reference
	%	me./100 g.	
East Pakistan,Joydebpur Forest, Upland	43	19·1	(1)
,, ,, Dacca, Upland	44	14·1	(1)
,, ,, Meghna Alluvium	9	15·0	(1)
,, ,, Brahmaputra Alluvium	24	18·4	(1)
West Pakistan, Punjab, Alluvium	20	11·6	(2)
,, ,, Punjab, Alluvium	12	5·2	(2)
South India, Coimbatore, Black Cotton Soil	38	39·4	(3)
,, ,, ,, Red Upland	16	7·6	(3)

(1) Karim and Islam (1957)
(2) Haque (1958)
(3) Raychaudhuri, Sulaiman, & Bhuiyan (1943)

It is important to realize that, because soil texture cannot be changed materially by ordinary treatment in the field, the characteristic cation-exchange capacity of a specific soil does not alter appreciably over short periods of time. For instance, changes in the organic matter content resulting from treatment of field soils with manure will have little effect on overall adsorptive capacities. If we assume the organic component of the soil to have an exchange capacity of 200 me./100 g., and this is a practical average value, then a soil with 1% organic matter will derive only 2 me./100 g. of exchange capacity from this component.[1] Assuming the weight of the surface-7 inches of soil to be 2 million lbs., an increase of 20,000 lbs. in the organic matter level is necessary

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[1] McGeorge (1930) working with a large number of soils having a wide variation in organic matter content found an average increase of 1·8 me./100 g. of soil for each percent increase in organic matter.

to raise its content by 1%. This magnitude of change is impractical to bring about under ordinary circumstances. On the other hand, improper management of land which results in the loss of native soil organic matter can result in a measureable decrease in the exchange capacity over the years.

pH RELATIONSHIPS IN SOILS

The measurement of pH in soils serves a number of useful purposes. Many soil functions vary with pH, the cation-exchange capacity included. The solubility of a large number of chemical compounds present in the soil is dependent upon pH. These compounds contain elements which are either essential or toxic to growing plants. Biological activity, such as that bringing about the decay of organic residues in the soil and even the growth of plants, is affected by soil reaction.

The H and OH concentrations of the soil solution and, therefore, the pH, are controlled to a large extent by the proportions of various cations present in exchange positions. Most commonly, the exchange complex of field soils is dominated either by Al and H ions or by Ca, or Ca and Mg ions. This fact relates to the strong attractive force between these ions and the soil particle surfaces. In humid regions, where leaching with acid-bearing waters is common, the loss of Ca, Mg, Na, and K ions causes the soils to tend toward H saturation. The slow release of soluble Al, Ca, and other metallic cations from soil minerals and decaying organic matter aids in preventing complete saturation of the exchange materials with H ions alone. In arid regions, where leaching occurs to only a limited extent, Ca is found normally as the dominant exchangeable ion. Depending upon conditions which govern the rate of removal of the metallic cations from the soil, we may consider the exchangeable cation status to represent a balance between Al and H on the one hand and the metallic cations, particularly Ca, on the other. Normally, none of these common ions are completely lacking from exchange surfaces.

THE MEASUREMENT OF pH IN SOILS

The bulk of acid-producing H and Al ions in soils occurs in exchangeable form. The determination of pH, however, requires the estimation of the quantity of hydrogen occurring in ionic form in the solution phase. Therefore, a pH measurement does not indicate the total acidity of the soil but, in reality, only a small portion of it. For convenience, the concentration of H ions in solution is referred to as *active* acidity and is represented by pH values. Adsorbed H and Al ions, which are capable of producing an acid reaction in the soil solution upon their dissociation from exchange positions, constitute the *potential* acidity of the soil.

The H-ion concentration of the soil generally is measured on a soil-water suspension or paste, either (1) with coloured indicator solutions or strips of indicator paper or (2) electrometrically with a pH meter. Various indicator solutions and electrometric devices are available. Comparisons among several of these have been made by Mason and Obenshain (1938). The present discussion deals only with the fundamentals of the two techniques.

Colourimetric Methods

Colourimetric procedures involve the use of either dye solutions or dye-impregnated papers. These dyes, or indicators, have the capacity of changing colour with variation in pH of the solution. When the H-ion concentration in the solution reaches a certain point, the indicator changes from the ionic form having one colour to the molecular form having a second colour. The pH at which this change occurs is different for different indicators. With a mixture of selected dyes it is possible to cover a wide range in pH with one indicator solution or impregnated test paper. The precision of determination is limited to about one-half unit of pH. However, for practical use, this is sufficiently accurate.

Electrometric Methods

The electrometric methods are much more precise than the colourimetric procedures. As a matter of fact, the accuracy of determination greatly exceeds the precision needed in most soils work. Its reliability and precision make it by far the most commonly used procedure.

Whereas the theory of electrometric pH determination is highly technical, the basic details can be described rather simply. The equipment centres around a pair of electrodes which are placed in the solution or suspension under examination. These electrodes and the intervening solution act as a small cell, the electrical potential of which is dependent upon the H-ion concentration of the solution. As the H-ion concentration changes, so does the minute voltage developed by the combined electrodes. The small voltages are measured on a meter calibrated in units of pH. Modern pH meters almost invariably employ a glass electrode as the H-ion sensitive component.

Titration Curves for Soils

If a naturally occurring, strongly acid, mineral soil is titrated with a base, a neutralization curve similar to A in Fig. 5-1 is obtained. If this same soil were leached rapidly with strong acid immediately before titration, a curve such as B would be obtained. According to Harward and Coleman (1954) and Low (1955), the distinction between these two types of curves rests with the fact that, in the first instance, the exchange complex is filled largely with Al ions whereas, for the second, the exchange materials are essentially H saturated. A comparison with similar titration curves for common inorganic acids would show the Al soil to behave like a weak acid, while the H soil would appear to have the properties normally ascribed to a strong acid.

A distinct difference between these two curves appears in that the Al soil displays the capacity for *buffering*, especially in the range of pH 5 to 7. The H soil, however, is not buffered to any extent in this range, as is indicated by the rapid rise in pH as neutrality is approached. The source of buffering is the exchangeable Al ion which forms an insoluble base as the titration proceeds and thereby prevents a rapid increase in the OH-ion concentration necessary for elevating the pH. The buffering capacity, which serves as an index of the potential

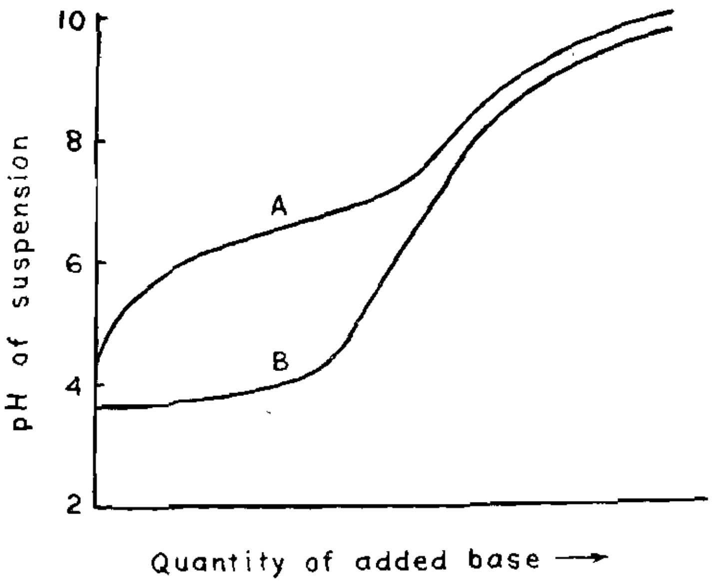

FIG. 5-1. Titration curves resulting from the treatment of acid soils with base. Curve A is typical of naturally occurring acid, mineral soils in which the exchange complex is filled largely with Al. The more nearly horizontal portion of this curve corresponds to the region of maximum buffering. Curve B is representative of a soil in which most exchange positions are filled with H ions initially.

acidity, is important in the treatment of soils considered too acid for good crop growth. A greater quantity of acid-neutralizing amendment is necessary than would be required were soils not so highly buffered.

The Relationship Between pH and Degree of Base Saturation[1]

Minimum pH values are observed in soils which approach H saturation. However, this condition is attained in field soils only if the exchange material is composed essentially only of organic colloids. These materials are stable in acid media.

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[1] *Base saturation* refers to the proportion of the exchange positions in soils filled by base-forming cations such as Ca, Mg, Na, K, and NH_4. The degree of base saturation is expressed as a percent of the cation-exchange capacity.

A mineral soil seldom will attain a pH much below 4·5, and then Al occurs as the dominant exchangeable cation. Saturation of the soil with Ca will result in a pH of about 8·2. Higher values, which often exceed 10·0, are observed where Na is the saturating ion.

The ready hydrolysis of loosely adsorbed Na accounts for the high pH of Na-saturated soils, the hydrolytic reaction resulting in the formation of NaOH. Often Na_2CO_3 occurs as a free salt in high-Na soils, and its hydrolysis also contributes to the elevated pH. Neutral Na salts, on the other hand, suppress dissociation of Na from particle surfaces or from basic salts because of the common ion effect. Therefore, accumulating neutral salts tend to reduce the pH of soils high in exchangeable Na. Dilution of the Na salts in such a soil, on the other hand, reduces their ability to suppress hydrolysis. The addition of water to a salty soil will cause a rise in pH, therefore.

Under average conditions, the adsorbed ions in soils of acid reaction are comprised of a mixture consisting predominantly of Ca, H, and Al. The more acid the soil the lower the proportion of Ca within this mixture and, consequently, the lower the quantity of this nutrient there will be for plant use. Because of this, strongly acid soils normally require a treatment which will raise the pH and also provide a higher level of exchangeable Ca. The desired effect customarily is obtained by adding pulverized limestone ($CaCO_3$) to the soil. This material will cause the displacement of exchangeable H and its ultimate conversion to slightly ionized water. Exchangeable Al is displaced also and precipitated as $Al(OH)_3$.

So long as only H, Al, and Ca ions dominate the exchange complex, a reasonably good relationship will exist between soil pH and the proportions of these three ions in exchangeable form. This is to say, the degree of base saturation with Ca can be predicted with fair accuracy from the pH of a soil. If the cation-exchange capacity is known, then a simple pH measurement is all that is necessary to determine the approximate quantity of limestone that is needed to raise exchangeable Ca to a desirable level.

Soils of very high pH and containing a predominance of exchangeable Na also are low in available Ca. Thus, the addition of an amendment containing Ca to such soils for displacement of Na is also advisable. However, because strongly alkaline soils generally contain variable quantities of salts which affect pH, no positive relationship exists between soil pH and the degree of saturation with exchangeable Na. Other means for determining the amount of Ca needed to displace exchangeable Na must be used, therefore. Such procedures are discussed in Chapter Twelve.

SOIL REACTION AND PLANT GROWTH

Plant growth is maintained over a rather wide range in pH. Arnon and Johnson (1942) found plants to tolerate the extreme values of 4 and 9 in solution cultures provided nutrients were kept in available form. Inhibition of growth between these extremes in pH usually is attributed to the decreased availability of the essential nutrients Ca or P in the nutrient solutions. Similar effects have been noted many times in soils.

Among other effects, extreme pH values represent conditions wherein availability of the essential plant nutrients Ca and Mg is impaired. Under strongly acid conditions, the shortage of available Ca and Mg results from their displacement by Al and H and eventual loss in drainage water. Potassium is lost from the soil by this process also. Where alkalinity is high, the divalent ions will again have been displaced, this time by Na. In the strongly alkaline medium, Ca in particular does not remain soluble but is precipitated as $CaCO_3$. The higher the pH the lower the solubility of $CaCO_3$ becomes.

Aside from effects involving competition among nutrient ions and Na or Al and H for exchange positions, the relationship between pH and nutrient availability depends largely upon mineral solubility at different pH levels. Although many compounds containing plant nutrients such as Fe, Mn, Zn, and Cu are readily soluble in strongly acid soils, elevation of the pH results in a decrease in their solubilities or the formation of less soluble compounds of these elements. Many of the nutrient ions will be found to behave similarly under varying conditions of acidity and alkalinity. These relationships are discussed in greater detail in Chapters Eight through Ten.

Certain toxicity effects can be related to pH of the soil. In strongly acid media, soluble Al and Mn have been observed to increase in the soil solution to the point of injuring plants. The harmful effects resulting from excessive quantities of boron are more intense under acid conditions and may be alleviated to a certain extent by liming the soil. The deleterious effect of strongly alkaline soils is attributable in part to their caustic reaction on plant tissue. From the several considerations above, we may conclude that the effect on plant growth arising from variation in the H-ion concentration is, more often than not, one of an indirect nature.

THE CONTROL OF SOIL REACTION

Normally, the pH of most soils lies within the range of 5·0 and 8·5. Some soils are observed to have pH values as low as 3·5 and as high as 10·0 or above. Most economic plants grow best within the narrower range listed above. If the reaction of the soil exceeds the limits of this range, adjustment of the pH to a more suitable level through use of chemical *amendments* is assumed to be desirable. A reduction in acidity is accomplished through the addition of a material such as ground limestone which consists largely of $CaCO_3$. Gypsum, $CaSO_4.2H_2O$, and agricultural sulfur are used to counteract strongly alkaline conditions. Treatments with these amendments customarily are designed to correct the adverse condition in the surface or 'plough layer' of soil only; that is, the upper 6 to 8 in. of soil.

Often, relatively large quantities of amendment must be added to be effective. The reason for this is understood readily when we consider the total acidity or amount of exchangeable Na in a layer of soil 6 to 8 in. deep and covering an area of one acre. If the soil in question has an exchange capacity of 10 me./100 g. and is saturated with H and Al, approximately 4·5 long tons (10,000 lbs.) of $CaCO_3$ would be required to neutralize the acidity completely. If the soil were Na saturated, about 7·5 long tons of high-purity gypsum obvious by comparing them in neutralization equations in which, for simplicity, the acid is assumed to derive from exchangeable H ions:

$$\boxed{\text{Clay}}\begin{matrix}\text{H}\\\text{H}\end{matrix} + \text{CaO} \rightarrow \boxed{\text{Clay}}\text{Ca} + \text{H}_2\text{O} \tag{5-6}$$

and

$$\boxed{\text{Clay}}\begin{matrix}\text{H}\\\text{H}\end{matrix} + \text{CaCO}_3 \rightarrow \boxed{\text{Clay}}\text{Ca} + \text{H}_2\text{CO}_3 \tag{5-7}$$

However, the weight of the CaO molecule is only 56% of that of $CaCO_3$. Since, in the reaction, 100 lbs. of $CaCO_3$ is equivalent to 56 lbs. of CaO, the neutralizing value of the CaO is $100/56 \times 100$, or 179. The neutralizing value of any liming material may be determined by dividing its moelcular weight into that of $CaCO_3$, or 100, and then multiplying by 100.

Rather than compare the neutralization values of various pure compounds, it is perhaps more important to take into consideration the purity of the many grades of limestone that are available. Seldom will one find a limestone without some degree of impurity. Purity is expressed on a percentage basis and, for a limestone material, will correspond to the neutralizing value. A purity of 80% indicates that the limestone contains 80% $CaCO_3$ and $MgCO_3$ and 20% of inactive ingredients so far as acid neutralization is concerned.

The purity of a liming material should be guaranteed by the supplier and the guarantee enforced by governmental law. If the purity is not known it may be determined by simple titration with acid (Piper, 1950, pp. 128-136).

Fineness of Grind

The relationship between the particle size of ground limestone and rate of neutralization of soil acid is shown in Figs. 5-2 and 5-3. From Fig. 5-2, it would appear that finely ground limestone (60 mesh[1] or smaller) is needed in smaller amounts to elevate the pH a specified amount than is required when the lime occurs as large particles. For instance, as shown in the figure, 10 T. of the coarse (4-8 mesh) material

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[1] Mesh refers to specific openings in sieves or screens as determined by the number of wires, or mesh, per linear inch. For the relationship between mesh number and the hole size in sieves, see Table 1 in the Appendix.

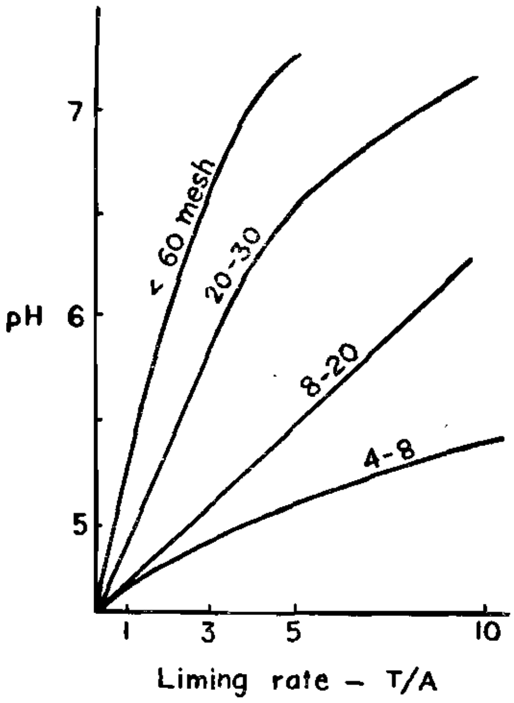

FIG. 5-2. Effect of rate and particle size of ground limestone on pH of the soil 18 months after application (Meyer and Volk, 1952).

raised the pH of a test soil to about 5·4 after 18 months, whereas only a little over 1 T. of the fine limestone (smaller than 60 mesh) was needed to bring about the same effect. Since full benefit from the coarse particles had not been realized after the 18-month period, it must be assumed that some of the added lime remained undissolved in the soil. Although complete solution of the coarse lime would be expected to occur eventually, the slow correction of acidity resulting from its use is not advantageous to the farmer. Obviously, it is not logical to apply a coarse material that will bring limited immediate gain when fine-textured materials can be obtained.

The data plotted in Fig. 5-3 show essentially the same results as have been cited above. In this latter instance,

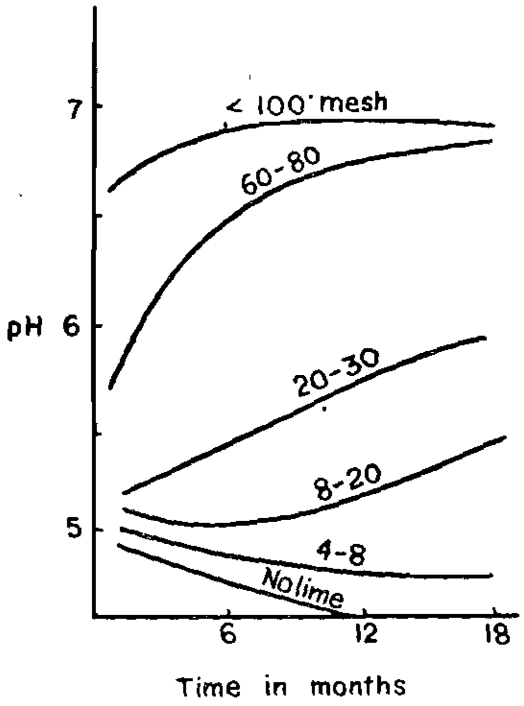

FIG. 5-3. Effect of limestone of different particle sizes on pH after contact with the soil for various lengths of time (Meyer and Volk, 1952).

comparisons were made wherein a single quantity of limestone separated into different particle-size groups was applied to an acid soil and the effects, as determined by increase in pH, were measured at frequent intervals during the following 18-month period. The limestone finer than 100 mesh caused very rapid correction of the acid condition. The 60- to 80-mesh material was as effective as the finer material but required the full 18 months to produce these results. Any grade coarser than 60 mesh fell short of producing maximum neutralization in the 18-month period. The 4- to 8-mesh material did not neutralize the soil acid at a rate sufficiently rapid to completely counterbalance the drop in pH occurring naturally in the soil during the period of the experiment.

DETERMINATION OF THE LIMING RATE

Various methods have been devised for assessing the need for lime by a soil. Probably the most precise method utilizes titration for measuring the amount of base needed to raise the pH of a soil sample to a suitable level. The rate of application of lime in the field on the acre basis is calculated by transposing the data obtained on the small soil sample to the equivalent amount of lime necessary for 2 million lbs. of soil. Unfortunately, the titration procedure, which must be carried out on soil suspensions, is more complex than are simple acid-base titrations where the neutralization reaction takes place instantaneously. It is appreciably too detailed to be applied practically to each and every field soil for which a lime recommendation must be made. However, within areas where climate and soil do not vary greatly, one may determine the general shape of titration curves on a limited number of soils and apply the findings to other soils in the same area. One such curve, as shown in Fig. 5-4, has been

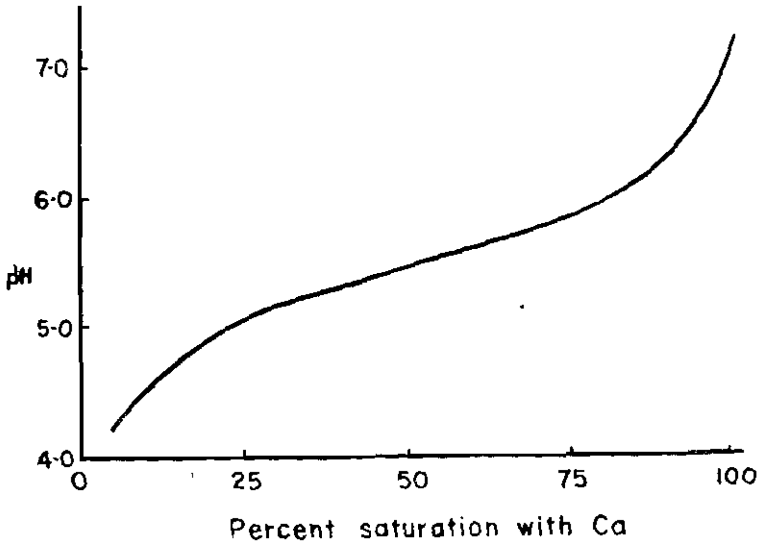

FIG. 5-4. Approximate average titration curve for a large group of acid sandy soils (after Peech, 1941).

developed by Peech (1941). It applies to a large number of sandy soils whose characteristics are similar because they are derived from like mineral matter under a rather uniform set of environmental conditions.

Use of a curve such as is shown in Fig. 5-4 requires knowledge of the initial pH of the applicable soil and its cation-exchange capacity. For example, a certain soil has a pH of 4·5 and it is considered desirable to elevate it to 6·5. The curve is then used to determine the change in percent Ca saturation (percent of the cation-exchange capacity filled by Ca ions) necessary to raise the pH to 6·5. This change in Ca level will, when multiplied by the cation-exchange capacity, convert the percentage value into me. Ca per 100 g. of soil. Since the milliequivalent weight of Ca is 0·020 g. and Ca makes up 40% of the $CaCO_3$ molecule, then, for each milliequivalent of Ca needed per 100 g. soil, 0·05 g. of $CaCO_3$ must be supplied. Conversion of these values for use on a field basis will show that 1 me. of Ca per 100 g. of soil corresponds to 1000 lbs. of $CaCO_3$ per 2 million lbs. of soil.

Since the above procedure can be applied to groups of soils which are relatively similar in all respects, it is unnecessary to determine separate cation-exchange capacity values for each of them. As a consequence, the determination of lime-requirement values necessitates only the measurement of soil pH.

Although less precise, a widely used procedure for predicting lime needs is one based on pH measurements and approximate knowledge of the buffering capacity of the soil to be treated. The latter characteristic is judged from an estimation of the type and quantity of exchange material present, for this determines not only the nature of the negative charge but also its magnitude. Thus, treatment of the soil takes into account whether or not the soil acid is associated with the permanent or the pH-dependent charge of the exchange complex. As shown by titration curves for many soils (see Fig. 5-1), more base is required to elevate soil pH by a specified amount as neutrality is approached. Consequently, lime rates are dependent upon the initial pH of the soil and the amount by which it must be elevated.

Allaway (1957) presents in tabularized form, as shown in Table 5-3, generalized recommendations for liming rates which are based largely on the principles outlined above. In this instance, climatic zone is used as a qualitative index to the type of mineral exchange material present in the soil.

Table 5-3. Approximate amounts, in long tons per acre, of finely ground limestone[1] needed to raise the pH of a 7-inch layer of soil as indicated (after *Allaway, 1957*).

Soil regions and textural classes	Limestone requirements		
	From pH 3·5 to pH 4·5	From pH 4·5 to pH 5·5	From pH 5·5 to pH 6·5
	T./A.	T./A.	T./A.
Soils of warm-temperate and tropical regions[2]			
Sand and loamy sand	0·3	0·3	0·4
Sandy loam	...	0·5	0·6
Loam	...	0·8	0·9
Silt loam	...	1·2	1·3
Clay loam	...	1·4	1·8
Muck[3]	2·3	3·0	3·4
Soils of cool-temperate and temperate regions[4]			
Sand and loamy sand	0·4	0·5	0·6
Sandy loam	...	0·7	1·2
Loam	...	1·1	1·5
Silt loam	...	1·4	1·8
Clay loam	...	1·7	2·1
Muck[3]	2·6	3·4	3·9

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[1] All limestone should go through a 3-mm. screen and at least one-half through a 0·15 mm. screen. With coarser materials, applications need to be greater. For burned lime, about one-half the amounts given are used; for hydrated lime, about three-fourths.

[2] Red-Yellow Podzol, Red latosol, etc. For a discussion of these soils and those under[4] below, see Chapter Thirteen.

[3] The suggestions for muck soils are for those essentially free of sand and clay. For those containing much sand or clay, the amounts should be reduced to values midway between those given for muck and the corresponding class of mineral soil. If the mineral soils are unusually low in organic matter, the recommendations should be reduced about 25 percent; if *unusually high*, increased by about 25 percent, or even more.

[4] Podzol, Gray-Brown Podzol, Brown Forest, Brown Podzol, etc.

As indicated in Chapter Four, extensively weathered tropical soils usually contain kaolinite and hydrous iron and aluminium oxides as dominant exchange minerals. Therefore, these soils normally will have a low cation-exchange capacity. Soils of the temperate regions are assumed to contain a predominance of montmorillonite, vermiculite, and illite in the clay fraction. Thus, soils of temperate regions should have a higher exchange capacity than those of similar textures occurring in the tropics. That these assumptions are subject to considerable error should be quite obvious to the reader by now. However, these errors may not be considered as serious even though they can be of frequent occurrence.

The recommendations shown in Table 5-3 are based largely on results observed in field and greenhouse trials. They may be derived also from titration curves obtained through treatment of acid soils with varying quantities of $Ca(OH)_2$.

THE FATE OF LIME ADDED TO ACID SOILS

So long as some of the $CaCO_3$ added to an acid soil remains undissolved, an equilibrium will exist between the exchangeable, soluble, and solid forms of Ca. The soluble form will be primarily as the bicarbonate. We may express the relationship between the three Ca forms by an equilibrium equation as follows:

$$\underset{\text{(Exchangeable)}}{\boxed{\text{Clay}}\,\text{Ca}} \rightleftharpoons \underset{\text{(Soluble)}}{\text{Ca(HCO}_3)_2} \rightleftharpoons \underset{\text{(Solid)}}{\text{CaCO}_3} \qquad (5\text{-}8)$$

Until the soil solution becomes saturated with $Ca(HCO_3)_2$, the reaction represented by this equation will move slowly to the left at a rate dictated by the rate of solution of the $CaCO_3$. Eventually, when all of the solid $CaCO_3$ has disappeared, a new equilibrium is established involving only the exchangeable and soluble forms. However, at this point, the direction of the reaction must change, for the removal of the soluble Ca in leaching water eventually will reduce its concentration in the soil solution to a negligible level. With the disappearance of the $Ca(HCO_3)_2$, the H-ion concentration of the soil solution again increases. Finally, through exchange with the H ions coming from acid leaching water, the level of adsorbed Ca decreases, and the soil may be expected to revert gradually to its original acid condition. We may surmize from this that, in an environment conducive to the formation of acid soils, a high level of exchangeable Ca cannot be maintained easily unless at least a small amount of free $CaCO_3$ remains in the soil. Liming should be looked upon as providing temporary correction of acid conditions and the procedure must be repeated on occasion if the Ca fertility and the pH of the soil are to be kept at optimum levels.

Certain management practices may be used to conserve the Ca applied by liming. It is advisable to keep a cover of growing plants for as large a proportion of the time as possible, and to return a sizeable portion of the plant residues to the soil. The plants sustain the level of Ca in the upper layers by extracting this element from less soluble forms in the subsoil, and by intercepting leached Ca as it passes through the soil in the percolating water. Eventually, when the plants are incorporated into the surface soil and decay, the Ca they contain is released. This Ca is available for adsorption by the exchange complex.

An additional advantage of the plant cover is the protection it affords against erosional loss. The removal of top soil by erosion can result in the disappearance of much of the lime that has been applied as an amendment. This represents a loss not only of the lime as such but also in the productive capacity of the soil.

One point which may not have become obvious from foregoing discussions is that the application of lime will not displace H and Al alone but also other cations adsorbed to the exchange complex. This fact is taken into account by the Staff of the East Bengal Soil Survey Scheme (1956) in their recommendations for liming, namely: 'liming may be done but one must take care, otherwise depletion of potash may occur as the percentage of potash is already very low.' It is a well known fact that liming the soil frequently increases the level of soluble K in the soil solution. The effect arises from the displacement of exchangeable K by the Ca added in the lime. However, it must be remembered also that Alions common in acid soils are much more capable of displacing K than is Ca. Therefore, it is to one's advantage to reduce the Al-ion concentration of an acid soil through liming, particularly if expensive potassic fertilizers are to be added. Certainly, if K is in low supply, plant requirements for this ion must be met from an external source, through use of either chemical fertilizers or manures.

The Control of Alkalinity

Soils with high pH values often are troubled with excessive quantities of exchangeable Na and limited amounts of available Ca. Alleviation of this condition is possible through use of amendments which increase the supply of soluble and exchangeable Ca in the soil. Since Chapter Twelve is devoted largely to this type of soil condition and its correction, further discussion at this point is considered unnecessary.

REVIEW QUESTIONS

1. Show the relationship between H-ion concentration and pH of a solution.

2. Define potential and active acidity. Which form do we measure when determining the pH of a soil ?

3. What is a buffer and why is buffering important in the neutralization of soil acidity ?

4. Why do soils cointain predominantly either exchangeable H and Al or Ca ?

5. Why does a Na-saturated soil have a higher pH than one saturated with Ca ?

6. Define 'neutralization value' as it applies to liming materials.

7. Why must one not expect a rapid and complete correction of acidity upon the addition of coarsely ground limestone ?

8. You have determined the pH on two soil samples. One is a non-plastic, red, clay loam with a pH of 4·5. The second is a black, organic soil with little mineral matter, and has a pH of 3·5. How much lime should be added to each soil in the field to raise the pH to 6·5 ?

CHAPTER SIX

SOIL ORGANIC MATTER

Soils which have been in the process of development for long periods of time tend toward a state of constancy with respect to their mineralogical make-up and organic matter content. Whereas previous mineralogical change has been largely the result of mineral breakdown and alteration, the history of change in the organic matter content has been one of accumulation. Thus, during the course of soil formation, inorganic materials initially void of organic matter are combined with varying amounts of organic substances derived for the most part from residues of green plants which have rooted, grown, and died in the altering soil material. Since the quantity of residual organic materials returned to the soil is governed by the abundance of plant growth, we may expect the level of soil organic matter to be associated closely with environmental conditions which control plant growth. Yet, these environmental factors influence not only the rate of organic matter production but also the speed with which organic residues are attacked by microbes and other soil-inhabiting organisms responsible for their decay. These organisms, in breaking down the residues, seek out carbon as a source of energy and as an essential component of cellular tissue. However, all carbon added to the soil, even that resynthesized into microbial tissue, is utilized eventually as a source of energy. As a consequence, the carbon is converted completely to CO_2 and lost from the soil. The decay process, therefore, tends to limit organic matter accumulation in the soil. The level finally acquired normally is considered to represent a state of balance between factors which control the accumulation and disappearance of organic carbon from the soil body.

The importance of organic matter as a part of the solid portion of the soil has been discussed in considerable detail in previous sections. Its presence was shown to influence materially the appearance of the soil profile (see Fig. 1-1). A darkened surface horizon is a normal, easily discernible characteristic of many soils. This is especially true where climatic conditions promote the growth of grass vegetation, or where soils have developed under conditions of imperfect drainage. Organic matter is of great importance in agricultural soils because it serves as a storehouse for essential plant nutrients, and especially nitrogen, which is required by plants in relative abundance for satisfactory growth. As shown earlier, organic matter also influences plant growth indirectly through the part it plays in maintaining desirable physical conditions within the soil mass.

THE ACCUMULATION OF ORGANIC MATTER IN SOIL

Among other factors which control plant growth and the quantity of organic residues added to the soil is the availability[1] of inorganic N. Nitrogen is an element essential to plant development because it is needed for the formation of cellular protein. However, significant quantities of this element do not occur in unweathered mineral substances. Thus, during early stages of soil formation, plant growth is limited by a dirth of available N even though other environmental factors, including climate, may be capable of supporting abundant vegetation.

Soil materials do not remain eternally void of N. Fortunately, there are means whereby inert gaseous N from the atmosphere is converted to inorganic form (*i.e.*, as NH_4 or NO_3) which may be utilized by growing plants. Special-purpose microorganisms which live in association with the roots of legume plants are responsible for a greater part of the conversion or *fixation* of atmospheric N. Lesser quantities are added to the soil following formation of NO_3 during lightning discharge in the upper atmosphere and transport to the earth in precipitation. Under some circumstances, fixation within the soil also occurs as the result of the activity of other specific microorganisms which do not live in association with legume

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[1] The available form of a nutrient is considered here to be that which plants can absorb and utilize. Available N, referred to in this and in succeeding paragraphs, is assumed to be the NH_4 and NO_3 ions.

plants. Nitrogen additions to the soils may come from various sources, therefore.

Nitrogen is needed in relative abundance not only by growing plants but also by microbial organisms which utilize organic carbon as a source of energy. A limiting supply of N, therefore, will affect both the rate of growth of plants and the rate at which their residues will disappear from the soil through decay processes. In the early stages of soil development, the N which occurs in small amounts is utilized completely by the growing plants. Decay of the residues returned to the soil is limited because any N released during decay is tied up immediately either by growing plants or within the tissues of microbial cells. As a consequence, disappearance of C does not keep pace with additions, and the level of organic matter in the soil increases. As shown in Fig. 6-1, this results in a relatively rapid initial rate of organic matter accumulation in the soil body.

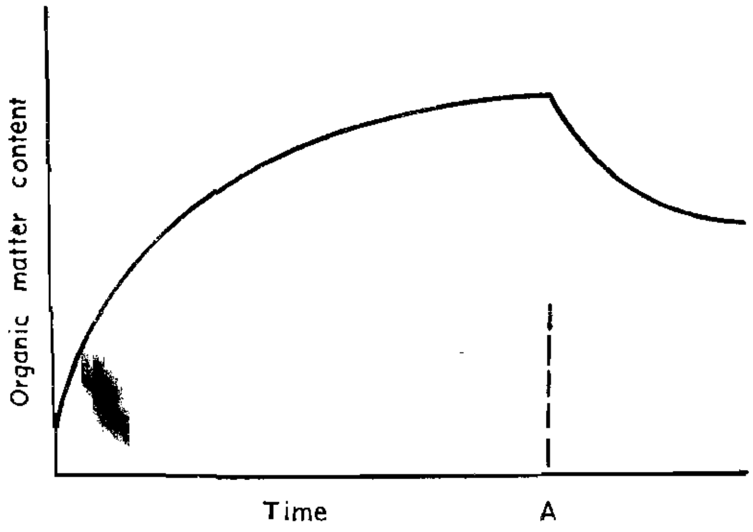

FIG. 6-1. The accumulation of organic matter in soil as a function of time, and the effect of cultivation, initiated at time A, upon the equilibrium level of organic matter.

The processes of decay result in the loss of C as CO_2. This causes a relative increase in the N content of the residual organic products of decay. The N remains largely in proteinaceous compounds which are either resistant to decay or which have been resynthesized as new microbial tissue. The N is thereby conserved against loss from the soil. Additions of N from the atmosphere continue, however, and there is a gradual increase in the level of N in the soil. A point is reached eventually where environmental conditions other than the level of available N (*i.e.*, total moisture supply) limit plant growth and the return of organic residues to the soil. At the same time, the level of N will no longer be limiting to the rate of decay of organic residues. Therefore, the rate of loss through decay will keep pace with the rate of addition of residues, and an equilibrium level of organic matter in the soil will have been attained.

As shown in Fig. 6-1, the initial rate of increase in organic matter in soil is rapid but decreases as the equilibrium point is approached. Once attained, the equilibrium level will tend to persist indefinitely. However, if the environment changes so will the organic matter level. For instance, the introduction of cultivation on virgin land, as illustrated by point A in Fig. 6-1, results in increased soil aeration and other changes which cause a more rapid rate of decay. As a consequence, the organic matter content of the soil decreases until a new equilibrium level is established. The extent of decrease reflects the magnitude of environmental change that has been induced by man.

THE NATURE OF SOIL ORGANIC MATTER

The accumulated organic matter in the soil is a mixture of substances in various states of decomposition. Although the bulk of these materials comes from plants, a small portion consists of living insects, worms, and microbes which are instrumental in creating the decomposed material. During the decay process, the more readily utilizable fractions of fresh residues disappear while the resistant compounds remain. Other complex products including cellular tissue are synthesized by the organisms responsible for decay. Thus, resistant materials accumulate to form the relatively stable *humus* fraction of soil solids. Subject only to slow additional change, humus contributes materially to the physical and chemical properties of the soil.

Properties of Humus

The humic portion of soil organic matter possesses various characteristics which contribute favourably to soil properties. It is an amorphous or noncrystalline substance, dark in colour, and occurs in a very finely divided or colloidal state. Many of its traits relate to its colloidal nature. Well-decomposed organic matter can contribute materially to the cation-exchange and water-holding capacities of a soil.

The intensity of dark colour of many mineral soils is associated quite closely with the humus content. Because of the importance placed on humus as a factor in crop productivity, dark soils usually are considered to be fertile. The dark colour is not always in proportion to the humus content, however. A striking example is the black cotton soils of the Deccan Plateau which have a lower humus content than many lighter-coloured soils.

Physically, humus has limited plasticity and cohesion. This point has been stressed previously as a factor in the maintenance of cleavage planes between soil structural units. Coatings of organic matter prevent the larger structural units from adhering to each other. Being light in weight, humus in large amounts can reduce measurably the bulk density of the soil. At the same time it increases the water-holding capacity.

The pH of organic colloids saturated with H ions will be lower than the pH of H-saturated silicate clays. This is because disintegration of the mineral clays on saturation with H prevents the lowering of the pH below a certain point, and because more H is ionized from the organic colloid.

Stability of the organic molecules decreases with an increase in pH. Advantage is taken of this fact in the separation of humus fractions from the soil during chemical analysis. Being soluble in alkaline media, organic matter dissolves to a certain extent in strongly alkaline soils and appears as a blackish film on the surface of the soil. This has given rise to the use of the term 'black alkali' to designate those soils whose high pH is derived from their high level of exchangeable Na.

Distribution of Organic Matter in the Soil Profile

The distribution of organic matter accumulated in soils is largely a function of the type of vegetation from which the residues are derived. Vegetation types native to soils may be separated into two widely divergent groups, namely grasses and trees. The contribution made by crop plants is negligible because, as indicated above, the introduction of cultivation to virgin land normally results in loss rather than accumulation of organic matter.

Vegetation consisting primarily of grasses contributes large quantities of organic matter from root tissue. Root systems of such plants are profuse and are characterized by a relatively short life and rapid replacement of the individual rootlets. The organic matter in soils under grass is, therefore, relatively abundant. While in greatest concentration in the upper layers, its presence is readily observed in the subsoil and to depths up to several feet. The level of organic substance is, of course, dependent upon environmental factors controlling the rate of accumulation and the period of time through which processes of accumulation have been operating.

As opposed to grass-type vegetation, the major contribution of forests to soil organic matter is as leaf litter. Characteristically, leaves of deciduous trees and needles of conifers do not succumb to decomposition as readily as do grass residues. As a consequence, organic matter is concentrated upon and within the surface layer of soil. Contribution to the organic matter level in the surface soil layer comes in part from decaying plant roots and partially from surface litter incorporated into the soil by burrowing animals, insects, and worms.

Under conditions where leaching occurs, dissolved organic matter may move down through the profile in association with dispersed clay. Accumulations of both the clay and organic colloids may occur as dark-coloured, compact subsoil layers referred to as *clay pans*. The presence of the organic matter, largely as coatings over cleavage surfaces, contributes to the separation of the soil within these layers into distinct structural units. Because of the clay present, these layers frequently are only slowly permeable to water.

Some protein molecules are adsorbed by clay and become enclosed within the interspaces of the expanding lattice crystals. In these positions, they are less subject to the attack of microorganisms which are too large to enter into the restricted interlayer spaces. This may explain partially why some proteins apparently are resistant to decay with the result that they tend to accumulate in the soil.

CHARACTERISTICS OF THE DECAY PROCESS

Any particular plant material contains C and N in a ratio that is characteristic for that material. Upon decay, however, since C is lost and N conserved, there is a gradual decrease in the ratio of C to N from the original value. When the organic components either have been broken down to simple end products or converted to substances resistant to further change, the decay process becomes less intense and the loss of CO_2 very slow. During this last phase of decay, the C content of the soil and the C : N ratio remain essentially constant until fresh residues are added once again.

Organisms involved in decay seek not only C but also other nutrient elements essential for their development. Therefore, unless these elements are supplied in adequate amounts by the soil, their availability to the organisms will depend largely upon amounts which can be liberated from the organic substances undergoing decomposition. The availability of N limits the growth and reproduction of soil microbes more often than does the availability of other nutrient elements.

If the ratio of C to N in organic residues is high and the C easily utilized, competition among organisms for the limited amounts of N available will be keen. If the C : N ratio is narrow, on the other hand, the N requirement of the microbes will be met more easily, and decay will proceed more rapidly. Thus, it is common to characterize organic substances as to their C : N ratio with the understanding that this ratio infers the relative rate at which decomposition will proceed. However, as will be discussed in greater detail later, the ratio between these two elements may not be so important as the form in which they occur. For instance, the C : N ratio of humus is narrow, but the C and N occur in it in forms which succumb to decay only very slowly.

The Organic Matter Cycle

The decay process is but one phase of an unending cycle in which the components involved return to the same state again and again. Thus, C will be found to change successively from plant to microbial tissue and humus and, finally, to CO_2. Eventually, however, the CO_2 is reabsorbed by growing plants and, through photosynthesis, is reconverted to plant tissue. By this means the cycle is reinitiated. In addition to C, other elements pass through a similar cyclic existence. Most similar to C are the constituents of protein, N, P, and S.

We may illustrate the cycle as in Fig. 6-2. The addition to the soil of raw organic matter commences the decay process. Nonresistant compounds are broken down very rapidly into simple end products. The more resistant products remain for longer periods but eventually are converted to the more simple compounds. During this period, however, these substances are an active part of the humus fraction of the soil.

Incorporation of organic residues into the soil will result in a rapid increase in the microbial population and a corresponding increase in the evolution of CO_2. These two factors are related so closely that measurement of the volume of CO_2 gas liberated serves as an indirect evaluation of the relative number of microscopic organisms involved in the decay process. The measurement of CO_2 evolved provides no information as to the specific groups of organisms present, but it does reflect the intensity of joint microbial activity.

Curves typical of the pattern of CO_2 evolution are shown in Fig. 6-3. Notable is the rapid increase in the rate of gas formation during the initial stages of decay. Two curves are shown, one for straw and the other for lucerne, the latter plant having the narrower C : N ratio. The differences shown are somewhat exaggerated for the purpose of illustration. They

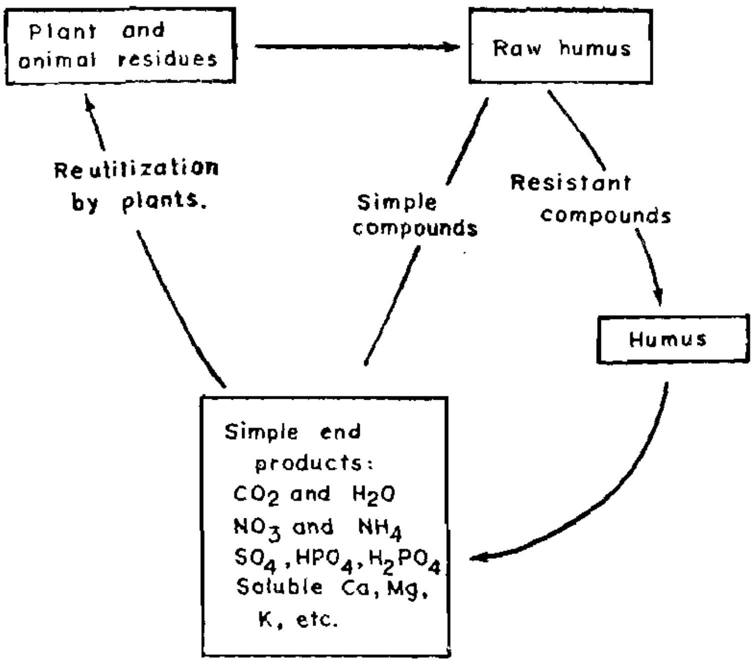

FIG. 6-2. Simplified diagram of the organic matter cycle. Humus is noted to be the accumulating organic compounds resistant to decay. The diagram suggests that all organic residues added to the soil are decomposed eventually.

indicate, however, that the rate of CO_2 liberation and, therefore, rate of decay proceed much more rapidly with the legume. A point is reached eventually where the curves for both materials attain about the same level and further decrease in activity (CO_2 evolution) is very gradual. During this final stage, the slow decomposition of the more resistant compounds is taking place.

The Mineralization Process

Whereas C is noted to disappear from the soil during the decomposition process, N[1], P, and S normally will be lost in

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[1] As will be seen later, under certain circumstances, N is lost from the soil through volatilization and leaching. Appreciable quantities of S may be lost also by these means.

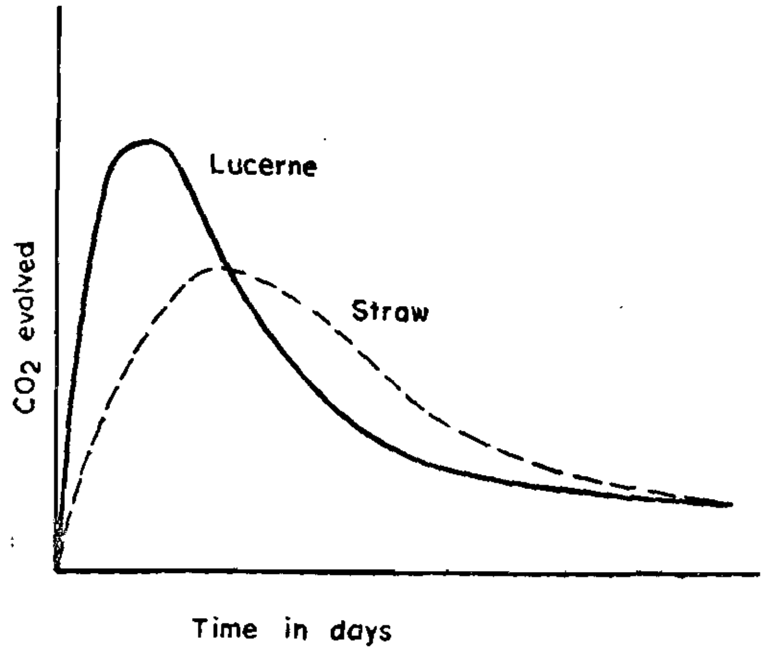

FIG. 6-3. The relative rates of decomposition of materials of widely different C : N ratios as approximated by the release of CO_2 with time.

limited quantities only. The extent to which these latter elements are involved in decay is evidenced to a large extent by their appearance as inorganic compounds in the soil solution. The conversion of these elements to inorganic form is referred to as *mineralization*.

The mineralization of organic N results in its transformation to NH_4 through *ammonification*. Where conditions are suitable, the NH_4 finally is oxidized to NO_3 by means of the *nitrification* process.[1] As a consequence to mineralization, we may expect, therefore, the appearance of NO_3-N in the soil solution. Ordinarily, however, this does not occur immediately where carbonaceous residues are added to the soil, because the N mineralized first usually is assimilated into

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[1] Ammonification and nitrification are discussed in greater detail in Chapter Eight.

microbial protein as soon as it is released from other organic forms. Thus, opposed to the appearance of free NO_3, we have a preliminary retention of N by microbes in protein form, the result of *immobilization* processes.

Kirkham and Bartholomew (1954, 1955) have observed that the two processes of mineralization and immobilization occur at the same time. As shown in Fig. 6-4, as long as there is a wide ratio of C to N[1] (30 or more) in the organic residues, immobilization proceeds more rapidly than mineralization. Nitrate does not appear in the soil solution. As shown further in the figure, when the C : N ratio approaches 15 : 1, conditions favour a more rapid rate of mineralization with the result that NO_3-N begins to appear in the soil solution. As the residual organic compounds resist further decomposition, a point is reached eventually where the decay process continues only slowly but persistently. The release of CO_2 is at a minimum and further change in the level of NO_3-N is detected with difficulty. Mineralization and immobilization proceed at about the same rate during this latter period.

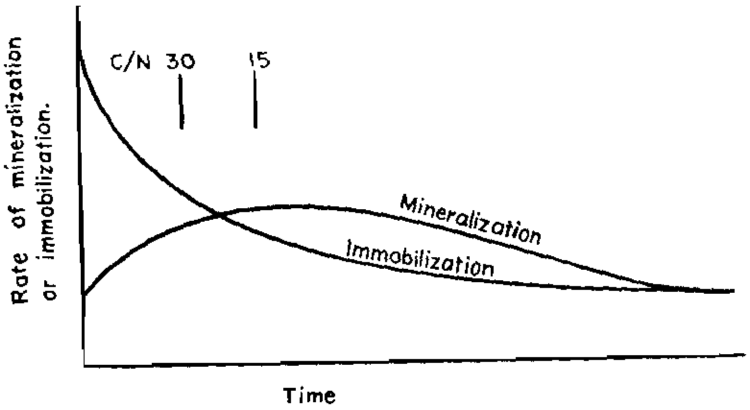

FIG. 6-4. The relationship between the C : N ratio of soil organic matter and the mineralization and immobilization of N. Mineralized N does not begin to appear until the rate of mineralization exceeds immobilization (after Kirkham and Bartholomew, 1955).

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[1] The C and N contents of organic matter are expressed as a percent of the oven-dry weights of the organic materials.

The release of P and S during decay follows very nearly the same pattern as that for N. Much less is known about the balance between mineralization and immobilization for these two elements, however. Mineralized phosphorus appears in the soil solution as HPO_4 and H_2PO_4 ions, the exact form depending upon the pH of the medium. Sulphur accumulates in the soil solution as the SO_4 ion.

The practical importance of decay and mineralization should now have become clear. These processes provide an important means for the maintenance of the fertility status of the soil. Nitrogen supplied in the form of organic manures will become available to plants when its release through mineralization exceeds its utilization by microbes. This will occur early if the manures have a narrow C : N ratio and explains the advantage to be gained from use of high-N legumes as green manure crops. It also explains why N deficiency will occur in many nonlegume plants if they are seeded too soon after the incorporation of highly carbonaceous residues into the soil. When this latter practice is followed, it is essential that sufficient N be added to allow for rapid organic matter decomposition without depletion of the supply of available N in the soil.

Factors Affecting the Decay Process

Factors which contribute to the rate of organic matter transformation are (1) temperature, (2) moisture supply, (3) soil reaction, (4) state of aeration, (5) supply of nutrients essential to the decay organisms, and (6) composition of the organic residues. These are discussed in turn.

Temperature: The activity of the organisms responsible for decomposition is a function of temperature. They are most active between about80 and 100°F. The upper limit of this range generally is exceeded only at the very surface layer of exposed and dry soil. Below 80°F. the rate of decay gradually decreases until the temperature offreezing is reached. Decompositional processes proceed at a negligible rate below this limit.

The effect of temperature on decomposition is apparent if we compare the organic matter content of soils developing throughout a wide range in temperatures but under otherwise similar environmental conditions. It will be noted that, under natural conditions where the native vegetation serves as the primary source of organic residues, the higher the temperature the lower the organic matter content of the soil. We should note also, however, that plant growth and the quantity of residues returned to the soil are greater in the warmer regions. Thus, even though an abundance of residues is available, the extremely high rate of decomposition results in a lower level of soil organic matter. Data of Jenny (1930), in which the amount of organic matter is assumed to parallel the total-N content of the soil, show the effect of temperature on the accumulation of organic matter (see Fig. 6-5).

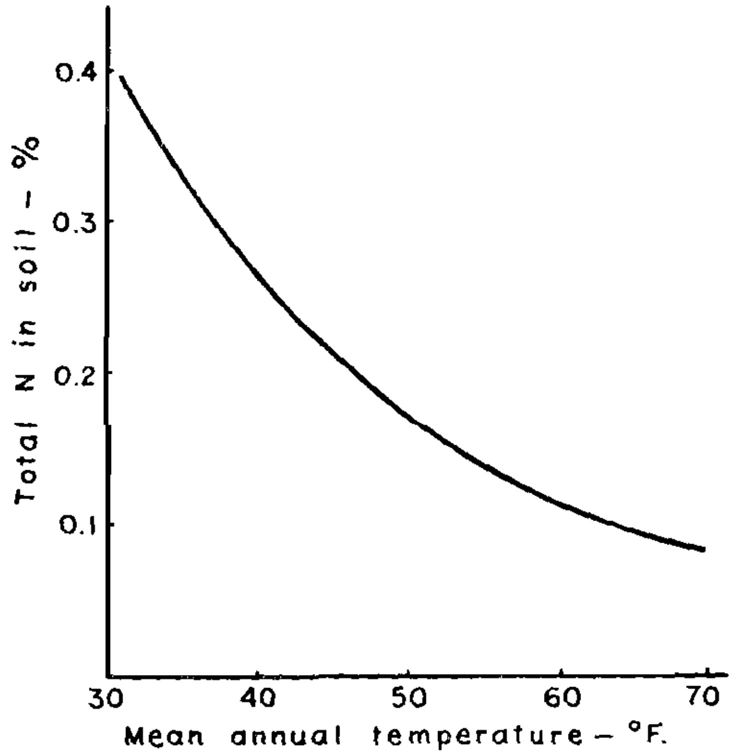

FIG. 6-5. The relationship between mean annual temperature and the accumulation of organic matter as reflected in the total-N content of the soil in subhumid grassland areas (after Jenny, 1930).

The high temperatures in the major agricultural areas contribute to the undesirably low organic matter content of soils of the Asian subcontinent. The low level of vegetative growth magnifies this deficiency in the arid regions. The high rate of decomposition in the more humid regions, on the other hand, aids in the maintenance of the supply of nutrients available to plants, especially where organic residues are returned to the soil. Unfortunately, however, throughout the subcontinent only a minimum of residues find their way back to the land. As a consequence, the small amount of humus remaining in these soils continues to decrease as the cultivators exploit this limited reservoir of plant nutrients through excessive cultivation and continuous cropping.

Moisture: Except under extreme conditions of drouth and waterlogging, the moisture conditions in the soil are usually suitable for the rapid decay of organic matter. If drouth is persistent, neither plant growth nor decay proceeds to any extent. Under conditions of excess moisture, on the other hand, certain aquatic plants grow well, but their residues do not decompose rapidly. This will result in appreciable accumulations of slightly decomposed organic materials which form *muck* or *peat* soils.

In regions where a shortage of moisture limits plant growth, an increase in the normal soil moisture supply will result in an increase in the organic matter content of the soil. The moisture effect is one which results from increased plant growth without a corresponding change in the rate of decay. The effect of moisture level on the native organic matter content of the soil is shown by data of Jenny (1930) reproduced in Fig. 6-6. Here, available moisture is represented by humidity measurements, and the organic matter content of the soil is assumed to parallel the level of total N. The data presented are for two regions of different, although uniform, mean temperatures. As indicated above, the lower the mean temperature the greater will be the quantity of organic matter accumulating in the soil.

Soil Reaction: In most agricultural soils, the reaction normally is such that other factors will have a greater influence on the rate of organic matter decomposition. Under some

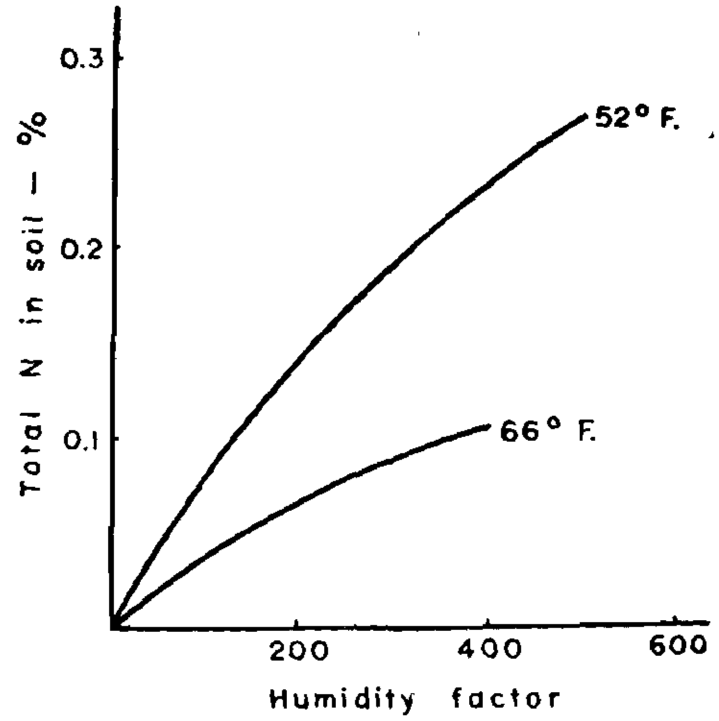

FIG. 6-6. The relationship between mean annual precipitation, as reflected in humidity, and the accumulation of organic N in the soil (after Jenny, 1930).

forest conditions in cool, temperate climates, however, leaf litter on the surface of the ground may develop an acid reaction which definitely retards the rate of decay. This results in the accumulation of a relatively deep deposit of organic refuse on the forest floor. A contributing factor to this is that the microbial population is restricted somewhat, consisting largely of acid-loving organisms. Other organisms, which normally share responsibility in the decay processes, are either absent or relatively inactive in the strongly acid medium. It should be pointed out, however, that additional factors such as a reduced mean temperature and a low level of nutrition may contribute to the slow rate of decay also under the forest conditions specified above.

Aeration: Microorganisms must have an adequate supply of O_2 to utilize organic C as a source of energy. We may expect, therefore, that clayey soil with poor structural development and inadequate aeration will not support the decay processes at the rate one might observe in well-aerated sandy soils. Any factor that will reduce soil aeration will slow the decay process, waterlogging being an example.

Nutrient Supply: Microorganisms responsible for decay require an adequate and reasonably well-balanced supply, of essential nutrients[1]. We may generalize by saying that if the nutrient status of the soil is suitable for abundant crop growth it will support a high level of microbial activity.

Composition of Organic Residues: The rate at which organic materials decompose depends to a large extent on their chemical composition. Conducive to rapid breakdown are a high proportion of compounds such as soluble sugars and simple protein and a high level of N and metallic cations, particularly Ca. Plant types vary markedly in their chemical makeup. Woody plants are high in complex organic substances such as lignin, waxes, and resins, and are low in N and the metallic cations. Therefore, they are slow to decompose. Because of their low content of Ca and Mg, they produce acidic residues as a general rule. Grasses and succulent plants, on the other hand, are higher in the simple organic compounds, in N, and in Ca and Mg. Their residues decompose more rapidly.

The range in amounts and the relative rates of decomposition of major components of plant materials are shown in Table 6-1. Listed also is the range in approximate composition of humic products formed under a wide variety of environmental conditions. Most noteworthy among these data is the change in proportion of compounds from plant tissue to humus. Humus is found to have the higher lignin content, evidence of the slow rate of decomposition of the latter substance. Protein is more abundant in humus also. This, however, is attributable to the fact that protein is resynthesized continually by the microorganisms involved in the decay

[1] Nutrients essential for plant growth are discussed in Chapter Seven. Nutrients required by soil organisms correspond closely to those used by plants.

process. Because of its composition, humus is referred to frequently as a *ligno-protein* complex.

Table 6-1. The approximate amounts of various groups of organic compounds in dry plant tissue and humus and their relative susceptibility to microbial decay (after Waksman, 1938, p. 95, and Waksman and Stevens, 1928, 1930).

Organic Group	Relative Rate of Decomposition	Approximate Content in	
		Plant Tissue	Humus
		%	%
Sugars and starches	Very rapid	1- 5	nil
Hemicelluloses	Rapid	10-28	5-12
Cellulose	Moderate	20-50	3- 5
Proteins	Very rapid[1]	1-15	30-35
Fats, Waxes, Resins	Very slow	1- 8	1- 5
Lignins	Very slow	10-30	35-50

[1] The decomposition of proteins proceeds at a variable rate but is generally rapid in fresh plant residues.

The C : N Ratio of Soil Organic Matter

Plant residues contain approximately 40% C and from 0·5 to 3% N on the dry basis, depending upon the nature of the plant. The C content of humus is considerably higher, approaching 60%. At the same time, the N content is found to have increased from 4 to 6%. Thus, in the first instance, we find the C : N ratio of the original residues to range between about 13 : 1 and 100 : 1. During decay, the ratios are narrowed appreciably, tending to stabilize at about 10 : 1 to 15 : 1 in many soils. The cause of the reduction in C : N ratio is the more rapid loss of C as compared to that of N.

The C : N ratio of the organic matter in an uncultivated soil remains relatively constant even though the total amount of each of these two elements may increase gradually through the annual addition of fresh residues. The constancy of the ratio is the result of the persistence of essentially the same microbial population and the same native plant species, both factors determining the nature of the humic product.

The C : N ratio of microbial tissue is quite narrow, usually ranging between 4 : 1 and 9 : 1. Thus, regardless of the C : N ratio of the added organic matter, if all of the C were utilized by the organisms, the resulting product remaining in the soil, microbial tissue, would have a narrow and relatively stable proportion of C to N. However, some of the carbonaceous material in plant residues does not succumb readily to microbial attack. We find among these materials, lignin, fats, and resins which are high in C and low in N and which effectively retard the complete conversion of added organic residues to a composition corresponding to that of the microorganisms.

Both the total-C content of the soil and the C : N ratio of soil organic matter represent a balance between the rate of decay and the rate of addition of fresh residues to the soil. The decay process tends to narrow the C : N ratio, whereas it is widened by the addition of undecomposed plant material. Any factor that will interfere with rapid decay and, at the same time, allow for continual accumulation of organic residues will tend to maintain a relatively high C : N ratio in the organic fraction of the soil. Normally the contributions to the organic matter level made by native plants and soil-inhabiting organisms are small compared with the residual humic substances remaining from times past. Therefore, the characteristics of soil organic matter are governed largely by the decomposition residues. Only where fresh residues are added in exceedingly large quantities can the organic C content and C : N ratio, at equilibrium with the environment, be altered appreciably.

The effect which climate-induced equilibria have on the C : N ratio has been studied in some detail by a number of investigators. Jenny (1930), for instance, observed that, under conditions of uniform rainfall, both the total-C content and the C : N ratio of the soil organic matter increase with decreasing temperatures (see Fig. 6-7). This trend is attributed largely to a reduced rate of decomposition and loss of C under cooler climates. Although plant growth and return of residues to the soil may be slowed also by a decrease in mean temperature, the effect probably is not so great as that observed for the decay

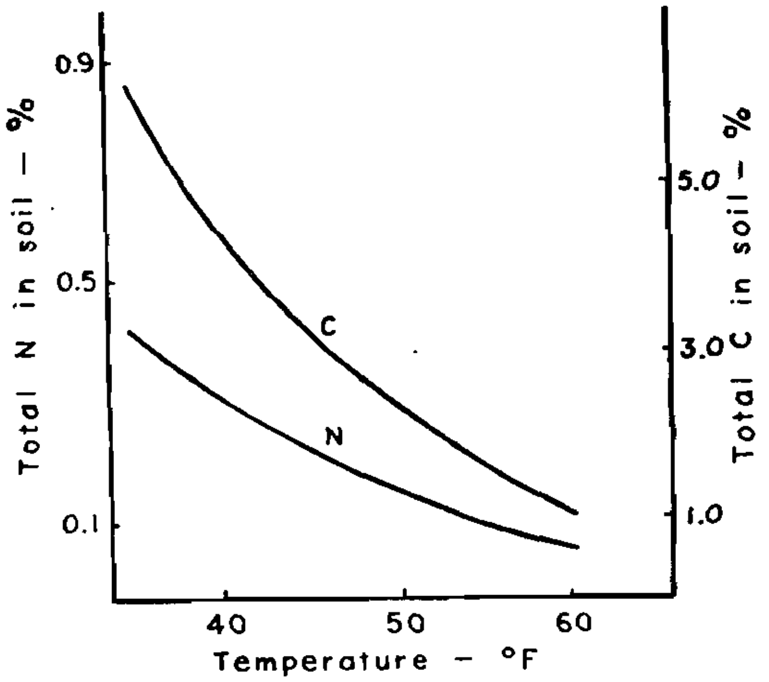

FIG. 6-7. The relationship between mean annual temperature and the C : N ratio of organic matter in virgin soils (Jenny, 1930).

process. Not only is the level of C in the soil increased at the lower temperatures, it remains for longer periods as incompletely decomposed organic compounds having a C : N ratio more nearly like that of the wide-ratio residues added.

One might conclude from this that the narrowest ratios would occur in soils of the tropics where biological activity responsible for decay is the greatest. This is not true, however, for C : N ratios of 14 : 1 and 16 : 1 are observed often in these regions. In explanation, this apparently results from the presence of an abundance of fresh, wide-ratio residues which tend to maintain high C : N ratios in the soil organic matter. Minimum ratios of 9 : 1 to 11 : 1 occur in arid regions, whereas the higher ratios are more characteristic of either cool, humid or hot, humid areas. This is borne out by the data of Anderson and Byers (1934) shown in Table 6-2.

Table 6-2. *The relationship between climate and the C content and C : N ratios of surface soils (Anderson & Byers, 1934).*

Great Soil Group[1]	Climate	Carbon content	C : N Ratio
		%	
Chernozem	Temperate to cool subhumid	1·45	9·0
Prairie	Temperate to cool humid	2·50	12·2
Podzol (Upland)	Cool to temperate humid	3·30	21·4
Podzol (Lowland)	Cool to temperate humid	5·40	22·6
Podzolic	Warm humid	1·25	12·6
Latosolic	Tropical, very humid	0·80	15·2

[1] For further dicussion of the great soil groups see Chapter Thirteen.

ORGANISMS RESPONSIBLE FOR DECAY

Microorganisms alone are not responsible for the decay of organic matter. Various animals and insects also play a very important role in the process, since they use vegetation for food and start its decomposition through digestion and return to the soil as manure. Frequently, organic substances undergoing decay will pass through a number of organisms before they become a part of humus or are converted to simple end products of decay. Last in the series of decomposing organisms will be the microbes which, as a composite group, are capable of breaking down almost any carbonaceous substance into stable or relatively stable products of decay.

Soil microorganisms may be classified into various groups based upon their physical appearance and behaviour. The major categories are *algae*, *bacteria*, *fungi*, and *actinomycetes* (see Fig. 6-8). Microbial organisms may also be grouped according to whether or not they can exist in the absence of O_2, on the basis of their source of energy, or the particular role they may play in soil processes.

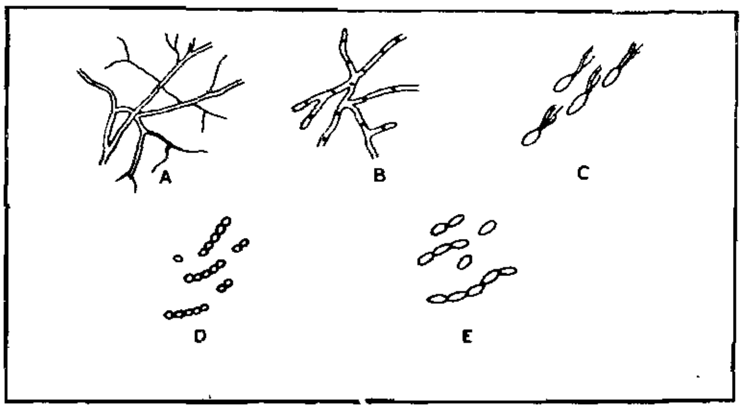

FIG. 6-8. Common forms of soil microorganisms: A, fungi; B, actinomycetes; C, flagellate bacteria; D, cocci bacteria; and E, bacilli.

Certain groups of bacteria thrive only in the presence of atmospheric O_2 and are designated as *aerobic* types. Others exist only in the event O_2 is absent. These are the *anaerobic* organisms which derive their O_2 from the reduction of various oxides in the soil. Still a third group, the *facultative bacteria*, appears to adjust well to either aerobic or anaerobic conditions.

Most of the microbes responsible for the decay of organic matter derive their energy from the oxidation of C from organic compounds. Those that derive their energy by this means are the *heterotrophic* organisms. While causing decomposition they also assimilate C and N into tissue and release unused N into the surrounding medium. Another group, the *autotrophic* organisms, receive their energy through the oxidation of various substances including NH_4, Fe, and S, and assimilate C directly from CO_2 present in the soil air.

Special-purpose microorganisms occur in both the heterotrophic and autotrophic groups. They play an important part in the maintenance of soil fertility but have little to do with the decay processes. Some are responsible for the oxidation of S and NH_4 to form, respectively, soluble sulphate and nitrate ions. Others convert atmospheric N from its inert, gaseous form into nitrate so that it becomes available to plants (N fixation). Whereas the special-purpose organisms are considered in detail in Chapter Eight, those groups affecting organic matter accumulation and decay are discussed briefly in the following paragraphs.

Algae

The algae are distinguished rather readily from other soil organisms because they are visibly pigmented. Groups of green, blue-green, brown, and red algae are recognized on the basis of their colour. Many algae occur as individual, microscopic cells. Some form giant seaweeds ranging in size up to hundreds of feet in length, on the other hand. Colonies of these organisms are observed commonly as a covering or scum over the surface of moist ground, or on the water in swamps and rice paddies.

The green pigment in algae is chlorophyll which, as in larger plants, permits these organisms to synthesize their own carbohydrate supply from atmospheric CO_2. Since they require light for the photosynthetic process, algae occur primarily where exposure to the sunlight is possible.

A number of algae appear to be capable of fixing atmospheric N as nitrate. De and Sulaiman (1950) found this to be true for paddy soils in East Pakistan, the process apparently being faster in the presence of the rice plant. This phenomenon may have contributed over the centuries to the maintenance of the N-fertility level in paddy soils. However, the rate of N fixation by algae is inadequate for optimum growth of rice or other crop plants.

It is interesting to note that the quantity of algae which can be grown in controlled nutrient-culture solutions is very large. Not only is the yield outstanding, but the protein content of the product is much above that of normal crop produce. It is possible that algae will serve as an important source of food in the future, but various problems related to their mass production remain unsolved.

The function of algae in organic matter decomposition is indirect. Since they are capable of increasing the supply of soluble N for use by decay organisms, they can encourage the more rapid destruction of highly carbonaceous plant residues in the soil. In many instances, the algae will in themselves contribute substantial quantities of high-protein substances which become a part of the humus fraction of soil.

Algae do not always assume a beneficial role in soil processes. We may predict from their behaviour, which is similar to that of higher plants, that competition between them and plants for nutrients will take place. Thus, the algae may, at times, actually interfere with the growth of more important crop plants.

Fungi

A second group of organisms, the fungi, are classed as plants along with the algae, but differ from the latter group primarily in that they are without chlorophyll. Many types of fungi occur. They range from single-celled yeasts, through fibrous or thread-like molds, to the more complex mushrooms. The common bread molds, and the smuts and rusts which cause much damage to grain crops, are other examples of fungi. Whereas the mushroom types have a role in the decay of organic matter, more important are the filamentous molds that occur as cotton-like masses in the soil or in organic deposits.

The fungi, being heterotrophic, cannot exist alone but must derive their energy from C supplied in organic combination. The fungi may be divided into three classes depending upon their mode of existence. The *parasitic* forms sustain themselves by infesting living animals and plants. They contribute nothing to their host. The *saprophytes* attack dead plant and animal tissue and, therefore, form a group important in the decay processes. The *symbiotic* fungi live in association with other organisms, and an interchange of food and nutrients between the two occurs to the advantage of each. Several fungi, for example, associate with algae to form special types of plants called *lichens*. These plants are found often on the surfaces of rocks and stones and contribute to the breakdown of mineral matter in them. Other associations, known as *mycorrhiza*, occur between fungi and plant roots. By permeating the root the fungus can have easy access to the carbohydrate supply of the plant. In turn, however, the fungus threads reach out into the surrounding soil and extract and transport nutrients into the conducting tissue of the plant root. This is beneficial to the plant because the absorbing power of its roots is increased many times over.

The mould fungi are characterized by their thread-like structural form. The threads, termed *mycelia,* are the vegetative parts of the plant. Spores, which develop on small projections from the mycelia, serve as a means for reproduction. The spores are pigmented. They are observed commonly as the blue or black colouration of bread moulds, or as the black, orange, or yellow smuts and rusts in seed grains. The spores are not the only means for reproduction, although, as microscopic 'seeds', they do serve this purpose. Mere extension of mycelia will maintain an abundance of fungus in the soil.

The fungi attack vigorously any organic residues in the soil. They are amenable to a wide variety of environmental conditions. Some thrive only where the pH is low, a fact which makes them especially important in the decay processes of leached, acid soils. On the other hand, their requirement for O_2 is relatively high, and they are not expected to function extensively under anaerobic conditions, such as in waterlogged soils or in the subsoil. The accumulation of deep deposits of undecomposed organic residues (*peat*) in swampy areas is attributable in part to the inability of the fungi to grow abundantly in the presence of the submerged plant residues.

Bacteria

These single-celled organisms are classified into three groups according to shape; (1) *bacilli,* which occur as thin rods, (2) *spirilla,* which are long and curved and which sometimes form distinct spirals, and (3) *cocci*, the spherical bacteria. Both the bacilli and cocci forms may occur as chains comprised of several individual cells linked together.

Each bacterial cell consists of a small mass of living protoplasm surrounded by a cell wall. Many forms possess distinct powers of motion provided by hair-like appendages or *flagellae*. This characteristic appears mainly among rod-shaped bacilli, but has been observed in other species as well. Bacteria obtain their nourishment by absorbing food substances from the surrounding media.

A basic requirement for bacteriological decay is an adequate supply of moisture. When adverse conditions develop, such as excessively high or low temperatures or drought, some bacteria assume a spore form which will tolerate the adversity much more readily than will the vegetative cell. Spores may persist in the soil for long periods and become active again when conditions are more favourable.

Anaerobic, aerobic, or facultative bacteria are found in the soil. There may be heterotrophic or autotrophic forms also. Because of their diverse characteristics, they are involved in wide variety of functions in decay. Many are special-purpose organisms, as has been mentioned previously. Because of their ability to survive under a wide range of conditions, bacteria are observed in all field soils. They appear to thrive best under conditions suitable to the growth of higher plants.

Actinomycetes

These organisms are similar to both fungi and bacteria. As for their cell structure and reproduction, they are more like the bacteria. On the other hand, they have an appearance quite similar to the fungi in that they are filamentous, or thread-like, and branched. Whereas actinomycetes are much larger and occur in smaller numbers than the bacteria, the total weight of these two organisms in the soil will be approximately the same. It is estimated that up to about 500 lbs. per acre of either of these groups of microbes will be found in the upper 7 in. of soil. This will correspond to approximately ·025% of the soil by weight.

Actinomycetes utilize the C in organic matter as a source of energy. They do not appear so active in the early stages of decomposition and increase in number more slowly than do either the bacteria or the fungi. This is associated with their ability to attack and breakdown the more resistant components of organic matter. These organisms play a very important role in releasing nutrients from organic combination, particularly the N from more resistant proteins.

Actinomycetes require an adequate supply of O_2 and cannot thrive where the soil is waterlogged. They are also sensitive to excess acidity or alkalinity, and occur in the largest numbers between pH values of 6·0 and 8·0. They may not be expected to exist in soil if the pH is much below 5·0. They are less sensitive to drought, however, and are often found to be active under dry conditions, even after the bacteria and fungi have become inactive. We may see that the actinomycetes form a very important group of soil microorganisms. They are able to extend or broaden the range of conditions under which active decomposition of soil organic matter will proceed.

MAINTENANCE OF ORGANIC MATTER IN CULTIVATED SOILS

Much has been said and written concerning the build-up of the organic matter level of agricultural soils. Many cultivators and agricultural experts are possessed with the belief that a sure corrective for all soil problems is the addition of large amounts of manures and organic residues. Whereas it is desirable to keep soil organic matter at an optimum level, a practical approach must be used in determining the level at which it should be maintained. Our fundamental knowledge of the behaviour of organic matter in the soil provides us with a basis for sound planning in this regard. For this it is necessary to take into consideration the factors of accumulation and decay.

It is estimated variously that the major part of the loss of organic matter present originally in virgin soil takes place during the first 10 to 30 years of cultivation. Following this initial rapid drop, there is a tendency toward a new equilibrium determined by the exact nature of the management applied to the land. The effect that differences in management can cause is illustrated by the data of Smith (1942) shown in Fig. 6-9. These data are for unirrigated maize land in the central part of the United States and cover the results of cropping for a period of 50 years. Oats, wheat, and maize, with and without farmyard manure, were grown continuously on the same land throughout this period, and the soils were sampled periodically to determine the effect of the different management systems on the level of N in the soil. The soil N and organic matter levels are assumed to parallel each other closely.

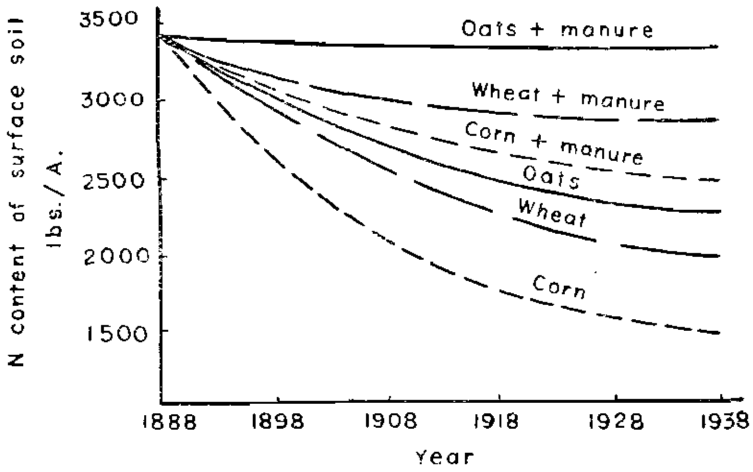

FIG. 6-9. Changes with time in the N content of surface soil from plots under continuous cropping with and without added manure (after Smith, 1942).

The data in Fig. 6-9 indicate that maize causes the most adverse effects on the organic matter level in the soil. This arises largely from the fact that small grains are not cultivated after sowing, whereas the soil on which the maize is grown is worked frequently for purposes of weed control. The increase in aeration this brings about hastens the mineralization process. Further, maize has an appreciably higher need for N. Both of these effects culminate in the greater loss of N from the soil as well as a more complete oxidation of C and loss of organic matter.

Figure 6-9 also shows that the addition of farmyard manure alleviates the adverse effects of continuous cropping. Treatment of oats with manure made it possible to maintain organic matter, as indicated by total N in the soil, at its initial level. Thus, whether or not one is able to maintain organic matter at its original level in the soil will depend largely on the cropping system followed and the method of management used.

From practical considerations, we may realize that it is very difficult to raise the organic matter content of a cultivated soil unless its present level is very low. Taking into consideration the texture and management, one should not attempt to raise the level much above the average for a particular area. This does not mean that practices of manuring should be stopped for those soils of relative high organic matter content. It must be kept in mind, however, that essentially all agricultural soils in the Asian subcontinent are low in native C and N, and it should not be assumed that these components can be increased easily to levels found in soils of the more temperate climates. The main benefit to be derived from manuring is the addition of mineralizable organic matter for the purposes of stimulating microbial activity and contributing to the level of plant nutrients in the soil. The high rate of mineralization caused by high temperatures will have resulted in an excessive reduction in the organic matter content of the soil after being brought under the plough. Thus, mineralizable N will be found almost always to be very low, and manure serves as a means of supplying it to meet plant needs.

The interrelation among soil texture, climate, and the organic C content[1] of some West Pakistan soils is shown in Table 6-3. The average C content, as determined by the dichromate oxidation procedure (see Walkley, 1947), is shown for 2 sets of samples from the Thal Development area and from the Potwar Plateau between Rawalpindi and

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[1] The percent of organic C times 1·78 equals the approximate organic matter content of the soil. This is based on the mean value of almost 57% C in soil organic matter.

Peshawar. It will be noted that the Thal sands have a much lower C content than do the clay soils from the same general area. The rainfall for the area of clay soils is slightly, though not significantly, higher than for the sands. The effective precipitation, on the other hand, may be appreciably higher in the clayey soils because of their greater water-holding capacity. The Potwar samples are included for comparison. These soils are intermediate in their clay content and are characterized as receiving about twice the amount of rainfall and occurring under slightly cooler temperatures than occur in the Thal. The highest percentage values shown for these 3 groups of soils approximate, for the respective areas, the organic C level one may expect to maintain by practical measures. These values may also be compared with those found in Table 6-2.

Table 6-3. Organic C and range in the clay content of soils from arid to semiarid regions in West Pakistan.[1]

Locality	Approximate clay content	Annual Precipitation	Organic-C Content	
			Range	Ave.
	%	in.	%	%
Central Thal (sands)	4-10	7-10	·10-·23	·16
Northern Thal (clays)	40-50	7-10	·26-·39	.33
Potwar Plateau (Silt loams)	15-25	18-·22	·42-·57	·50

[1] *Surface soils, 0-8", the mean of 5 randomly selected samples from* each area.

The effects of climate and poor management work hand in hand to maintain cultivated lands in the Asian subcontinent at almost the lowest possible organic matter level. In the first place, the exceptionally high mean annual temperatures over most of the region dictate that organic matter cannot be expected to accumulate to any great extent. Further, even in the regions of relatively high rainfall and humidity, the temperature effects essentially counteract the advantages expected from the increased moisture. Contributing much to the low level of organic matter in many areas, even where precipitation is plentiful on the annual basis, is the fact that maximum vegetative growth is not possible because of the long periods of drought.

In East Bengal, with rainfall varying from 60 to 70 in. per year and mean annual temperatures of approximately 70°F., the total N content of soils ranges between 0·05 and 0·10% (Staff, Soil Survey Scheme, East Bengal, 1952, 1956). Satyanarayana *et al.* (1946) found about the same range in total N for Indian soils where climatic conditions varied from arid through humid. Soils with a higher N content occur generally in the humid zones, as do those with higher total C and higher C : N ratios. Variations in the total N content of the soils in the northern plains region of West Pakistan cover about the same range as for the soils in India and East Bengal. The southern desert areas are lower in total N, however, having an average content of from 0·03 to 0·06%. The influence of combined temperature and moisture effects are noted in the submontane regions south of the Himalayan Mountains. The increased precipitation and lower temperatures there result in total N percentages of 0·15 or more.

If we consider the soils from throughout the major agricultural areas of the Indo-Pakistan region, it is probable that it will not be found to be feasible economically to maintain the level of organic C much in excess of 1%, or organic N in an amount greater than about 0·1%. Some exceptions to this may be expected, of course, particularly in fine-textured soils or in swampy areas where organic matter tends to accumulate, and in the cooler uplands.

Practical Considerations in Manuring

Manuring, as the term is used locally, refers to the application of various substances to the soil, including chemical fertilizers, for the purpose of enhancing the fertility of the land. If chemical compounds are used alone, they may increase plant growth and the amount of organic residues, particularly roots, returned to the soil. Organic manures, on the other hand, can bring about the same effect because of the nutrients they contain, and also contribute directly to the level of organic matter in the soil. Organic manures generally fall into two categories, common farmyard manure and green manure.

Farmyard manure usually is considered to be those materials collected in village waste heaps, including dung, straw, sweepings, etc. Its composition may be highly variable. According to Roberts and Singh (1951, p. 127), its content of major plant nutrients is approximately 0·40% N, 0·25% P_2O_5, and 0·15% K_2O.[1] An average composition usually listed in textbooks is 0·5% N, 0·25% P_2O_5, and 0·5% K_2O. Thus, local farmyard manure may be slightly low in N and very low in K. Its composition may be expected to vary markedly from area to area and will depend largely upon care of handling and storage.

Green manures are those crops planted especially for ploughing under when they are in a young and succulent state. At this stage of growth, the N content is relatively high, the C : N ratio narrow, and the rate of mineralization rapid. Legume plants, such as berseem or lucerne and which have a N content of from 3 to 4% on the dry-weight basis, are preferred as green manure plants. Nonlegumes, such as oats or rye, will have no more than about 1·5% N expressed on the same basis. Although the contribution made to the organic matter level by root residues of harvested crops is important, the total quantity usually is far from adequate to maintain the level present in the soil in its original, virgin state.

If we take into consideration the composition of fresh plant materials and of microbial tissue formed from them through decay, it is possible to estimate the relative effect of organic matter additions on the C content of the soil. The addition of 100 lbs. of a strawy material with 0·5% N should permit the retention of only 5 lbs. of the added carbon if the final C : N ratio of 10 : 1 is attained and no N other than that in the straw is available. The addition of 100 lbs. of a legume with 3% N would, under similar conditions, allow for the retention of about 30 lbs. of C. However, the initial loss of C as CO_2 will be much more rapid in the case of the legume (see Fig. 6-3), and the approach to the final equilibrium level of 30 lbs. of C will occur before the residual C content of the straw has been reduced to its equilibrium level. Thus, during

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[1] P_2O_5 and K_2O are standard means of expressing the P and K content of fertilizer materials.

the early stages of decay, the C content of the soil would be temporarily higher where straw had been added.

Two major factors controlling the maintenance of soil organic matter are erosion and the fertility status of the soil. Erosion can result in the loss of organic matter at a much greater rate than it can be replaced through manuring practices. The build-up of organic matter by natural means takes many decades. It is not uncommon in heavily eroded areas for from one-tenth to one-twentieth of a foot of soil to be carried away by eroding waters in one year's time. Even at the slower rate, one foot of soil can be lost in 20 years. There is no practical cultural practice that will allow for the replacement of organic matter in soil which disappears at such a rate.

The fertility factor is important because of the effect it has on plant growth and the amount of residues returned to the soil. The effect may be due primarily to an increased quantity of root material left in the ground. This is assuming that most of the above-ground portions are carried away in harvesting. Thus, increased fertility represents increased humus. This, in turn, results in an improved fertility status. Therefore, an increase in either the level of fertility or of humus will result automatically in an improvement in the other.

The use of legume crops as green manures is strongly encouraged. Two possibilities appear ; the ploughing in of the last residual growth of winter legumes, such as berseem, and the use of summer legumes, as guara, planted especially for this purpose. Since the major value obtained from such practices relates to the N added to the soil, further discussion on the use of legumes as green manures is reserved for Chapter Eight.

REVIEW QUESTIONS

1. What is meant by the term 'humus' ? Why does the level of humus tend to remain relatively constant in an undisturbed, virgin soil ?
2. Name the major groups of organic compounds in plant tissue and their relative resistance to microbial decomposition. Which groups tend to accumulate in the soil and why ?
3. What are the four major groups of microorganisms responsible for the decay of organic matter ?
4. Define the terms aerobic, anaerobic, and facultative ; heterotrophic and autotrophic.
5. Describe briefly the differences between fungi, bacteria, and actinomycetes.
6. Define the terms parasite, saphrophyte, and symbiosis.
7. What are the major factors affecting the rate of decay of organic matter, and how do they relate to the accumulation of humus ?
8. What differences would you expect in the amount of accumulated organic matter under the following sets of conditions :
   (*a*) Hot, humid climate, well-drained soil ?
   (*b*) Hot, humid climate, swamp ?
   (*c*) Hot, dry climate, well-drained soil ?
   (*d*) Cool, humid climate, well-drained soil ?
9. What change occurs in the C : N ratio of plant residues added to the soil ?
10. Is the C : N ratio for soil organic matter constant from region to region ? Explain your answer.
11. What bearing does the C : N ratio of plant residues have on rate of decay ?
12. Explain the cyclic nature of organic matter production and decay.
13. What is mineralization ? Immobilization ?
14. What effect does the introduction of cultivation have on the organic matter content of virgin soil ? Explain your answer.
15. What are the major reasons for adding organic residues to the land ?

SELECTED REFERENCES

1. Bartholomew, W. V. (1957) Maintaining organic matter. From *Soil,* The Yearbook of Agriculture for 1957, pp. 245-252, Supt. of Documents, Washington 25, D.C.

2. Broadbent, F. E. (1953) The soil organic fraction. Advances in Agron. 5 : 153-183.

3. Broadbent, F. E. (1957) Organic matter. From *Soil,* The Yearbook of Agriculture for 1957, pp. 151-157, Supt. of Documents, Washington 25, D.C.

4. Clark, F. E. (1957) Living organisms in the soil. *Ibid,* pp. 245-252.

5. Fraser, G. K. (1955) Soil organic matter. From *Chemistry of the Soil,* Firman E. Bear, (Editor), pp. 149-176. Reinhold Publishing Co., New York, N.Y.

6. Norman, A. G. (1942) Problems in the chemistry of soil organic matter. Proc. Soil Sci. Soc. Amer. 7 : 7-15.

7. Norman, A. G. (1946) Recent advances in soil microbiology. Proc. Soil Sci. Soc. Amer. 11 : 9-15.

8. Waksman, S. A. (1932) *Principles of Soil Microbiology.* The Williams and Wilkens Co., Baltimore, Maryland, U.S.A.

9. Waksman, S. A. and Starkey, R. L. (1931) *The Soil and the Microbe.* John Wiley and Sons, Inc., New York, N.Y.

CHAPTER SEVEN

CONCEPTS OF SOIL PRODUCTIVITY AND FERTILITY

PRODUCTIVITY

BY DEFINITION, *soil productivity* is the capacity of the soil to produce plants under a specified programme of management. The measure of productivity is yield, and it may be considered on a short- or long-time basis. Production is dependent upon the interrelationship among a number of *growth factors*, each of which must exist under optimum conditions ; otherwise, the growth of plants will be adversely affected. For maximum crop production, it is necessary to attempt to control these factors to the fullest extent possible.

We may list many of the factors of growth in the order of man's ability to control them successfully. Among these we find several generally uncontrollable items such as climate, physical composition of the soil, and topography. In certain cases, we are able to control, at least partially, such factors as soil moisture, disease and insect pests, waterlogging, and salt accumulation. For cultivated crops, relatively complete control is had over plant variety, plant-stand density, weeds, and various cultural practices including time of seeding, row planting, and fertility management.

The interdependence among growth factors is emphasized by a concept described by Tisdale and Nelson (1956, pp. 22-28). According to this concept, yield may be defined by the equation:

$$\text{Yield} = f(a \times b \times c \times d \ldots\ldots\ldots\ldots) \qquad (7\text{-}1)$$

where a, b, c, etc. are individual growth factors expressed as a percent of the total of that factor required for maximum growth. If two factors are limiting to the extent that either would result in a yield corresponding to 50% of the maximum, then, yield $= \cdot 5 \times \cdot 5 \times 1 \cdot 0 \times 1 \cdot 0 \ldots\ldots\ldots = \cdot 25$, or 25% of the maximum. Stated differently, the limiting effects of factor *a* will reduce yield to the 50% level. Imposing the second limiting condition, *b*, will again cut the yield in half, or to 25% of the maximum. It becomes clear that the effects of one deficiency will limit the benefit gained by the growing plant from all of the other factors of production.

Among all the factors important to crop production, our main interest lies presently in but one of them; that is, the soil. This factor, in turn, is comprised of various subfactors such as texture, structure, water-holding capacity, aeration, the presence or absence of injurious substances, and *fertility.* It cannot be overemphasized that maximum production of crop plants will be realized only if all the factors involved are in as nearly an ideal state as possible.

FERTILITY CONCEPTS

Fertility is defined as the potential of a soil for supplying nutrient elements in amounts, forms, and proportions required for plant growth. Direct measurement of the fertility status of a soil will be in terms of chemical elements or compounds. Generally, highly fertile soils should be productive, but this is not necessarily true in all instances. For example, a soil capable of supplying ample nutrients may be unproductive because of the limiting effects of waterlogging, insect damage, plant disease, and the like.

The Essential Plant Nutrients

The fundamental components of fertility are the essential plant nutrients (see Table 7-1) which are absorbed by plants and utilized for various growth and maturation processes. Summarily, nutrient ions occur in the soil in mineral or organic compounds, adsorbed to exchange surfaces, and in the soil solution. The soluble and adsorbed nutrients are extracted readily by plant roots or, in the laboratory, by chemical extractants. Because of their easy access, nutrients in exchangeable or soluble form are considered to provide *active* or *kinetic fertility.* Less easily extractable nutrient elements in primary and secondary minerals and in organic matter comprise the *potential fertility* of the soil. Often crop

Table 7-1. The essential plant nutrients[1]

Major Elements			Minor Elements	
Carbon	Nitrogen	Calcium	Iron	Copper
Hydrogen	Phosphorus	Magnesium	Manganese	Zinc
Oxygen	Potassium	Sulphur	Molybdenum	Chlorine
			Boron	

[1] Major elements are those required by plants in relatively large amounts; the minor elements, in small amounts. The latter nutrients also are called 'trace' elements. For detailed discussions of the nutrients, see Chapters Eight through Eleven.

production is highly dependent upon the rate of transfer from the potential to the active forms. Unfortunately the rate of this change cannot be characterized by the rapid laboratory tests commonly used to estimate the fertility status of the soil. Soil tests measure only the active fertility, and this fact represents a major limitation to their usefulness.

As shown in Table 7-1, there are 16 elements known to be essential to plants. Each has a specific role in the development of the plant, although all have not been characterized adequately in this respect as yet. The necessity of Cl as an essential plant nutrient has been only recently established, and it is quite possible that other elements will be found essential as techniques for identification of this fact become better developed.

The C, H, and O used by plants are derived from soil water and from CO_2 in the atmosphere. Whereas the water is absorbed by roots, the CO_2 is taken in through stomatal openings in leaves and stems and is utilized directly in the photosynthetic production of sugars and starches within the plant. These carbohydrates serve as energy materials in growth and reproduction and are converted to protein by combination with N, P, or S.

All other plant nutrients are absorbed from the soil through the root system. The abundance of their readily available form normally determines the rate of their absorption. Often inter-relationships among the nutrients themselves will determine to a large extent the ability of the plants to absorb and utilize them effectively. Such relationships determine the state of *nutrient balance* or *physiological balance* among the essential elements.

The plant nutrients customarily are grouped on the basis of the relative amount in which they are required by the plant (see Table 7-1). The major nutrients, C, H, O, N, P, and K, are needed in comparatively large quantities. Man has little control over the plant's ability to obtain the first three, C, H, or O. Therefore, they are seldom given consideration in fertility discussions beyond the point of indicating their essentialness. Such is not the case with N, P and K. These are the 'big three' of soil fertility. Investigations concerning the need for and use of these latter nutrients far exceeds the studies on all other phases of fertility.

Plants vary markedly in their requirements for the major plant nutrients. Maize, for instance, requires much more N for maximum growth than does wheat. For this reason we cannot specify a required level of availability of a nutrient which will apply universally to all crop plants. This is particularly true for soluble forms of soil N, all of which can be removed from the soil by plants during a single growing season. Total growth will be determined largely by the total amount of soluble N made available during the life of the plant.

While there are many reasons which explain why the major elements, N, P, and K, are studied in such detail, the primary causes for this are that (1) these nutrients are required by plants in large amounts, (2) their deficiency generally is readily recognized and corrected by field treatment, and (3) the cost of correction is much greater than for the minor elements and other major elements such as Ca, Mg, and S. It is this latter factor which probably is of greatest significance. The investment required for fertilizer materials, especially N, can be very large. The efficient use of fertilizers is necessary to bring suitable returns from the investment. Only if one understands the behaviour of the various plant nutrients in the soil can he be expected to make satisfactory use of fertilizer materials and manures.

The Availability of Nutrients

The term 'availability' is used commonly to express the amount of a nutrient which a soil can supply to growing plants. At other times, the quantity of an ion extracted from a soil sample by a chemical extracting solution also is referred to as the 'available' amount of that nutrient. Such a measurement may be expected to relate indirectly to the availability of a plant nutrient but will not measure the total amount that plants can extract from the soil during the growing season. This becomes obvious if one considers that chemical extractions normally are made on surface soil samples, whereas the plants feed throughout a considerable depth of the soil.

Availability, then, takes on aspects other than the ability of a chemical solution to remove a certain quantity of ion from the soil. It depends largely on the ability of the plant to seek out and extract the nutrients over a specified period of time. We find that the position of a nutrient within the soil, the chemical form in which it occurs, and other factors contribute to the specific availability of a nutrient to plants.

Positional Availability: A nutrient is positionally available when it can be contacted by plant roots. Nutrients at the very surface of the soil will not be absorbed if the soil is so dry that roots will not develop there. Similarly, a nutrient ion that has been washed down into the soil and beyond the reach of plant roots is positionally unavailable.

When a nutrient element is held in exchange positions by soil particles, it is not free to move about in the soil. Even though this element may be distributed uniformly throughout the root zone, only a small part of it will be positionally available because the roots do not contact all of the particle surfaces to which it is adsorbed. All NO_3-N dissolved in the soil solution within the root zone is, in effect, positionally available, on the other hand. Since this ion is not adsorbed by soil particles, it can move freely to plant root surfaces. The same may be said for the soluble, mobile forms of SO_4-S and BO_3-B.

Chemical Availability: A plant nutrient may be positionally available and yet unavailable because of its chemical form. An example is protein N which must be released through mineralization before it can be absorbed by plant roots. If the rate of mineralization is inadequate to supply plant needs, the N must be considered as occurring in a state of low chemical availability. The precipitation of ions, as often occurs with P, Fe, Ca, and Mn, also may result in their chemical unavailability. The great reservoir of plant nutrients, the minerals that make up the bulk of the soil, represents the major form in which nutrients commonly are available positionally but relatively unavailable chemically.

Physiological Availability: A nutrient may be both positionally and chemically available and yet may not be absorbed nor efficiently utilized by plants. This is often the result of an imbalance among ions, sometimes referred to as *antagonism*. For instance, in the nutrition of apple trees, it has been observed in some areas that the over-application of one of the elements, Mg or K, may result in the appearance of deficiency symptoms of the other. A similar relationship has been noted to exist between Mn and Fe.

THEORIES OF NUTRIENT SUPPLY

Solution Theory

This theory proposes that nutrients must first be in solution form to be absorbed by plants. Exchangeable ions will become soluble upon their displacement from particle surfaces by other ions, particularly H. The H ions, in turn, come primarily from H_2CO_3 formed by the solution of CO_2 liberated by plant roots. In other words, for this mechanism to be operative, H must become a part of the soil solution to displace the exchangeable nutrient cation. The displaced ion then would migrate to the root surface for absorption. Because of the slow diffusion of ions in the process depicted above, such solution effects would operate only over short distances from the plant roots.

The solution theory does apply well to ions that are not adsorbed onto colloidal surface. Such ions as NO_3, SO_4

and BO_3 are good examples and, in certain instances, the HPO_4 and H_2PO_4 ions.

Contact-Exchange Theory

Jenny and Overstreet (1939) have shown experimentally that exchangeable ions are absorbed by plants more readily when there is actual contact between the soil colloid and root surfaces. H ions produced at the root-colloid interface displace the exchangeable ions which then are absorbed by the root (see Fig. 7-1). Hydrogen ions from the root accumulate on the colloid surface and cause a reduction in the pH of the soil. The acid thus produced will be neutralized through the eventual solution of basic soil minerals. In this way, the

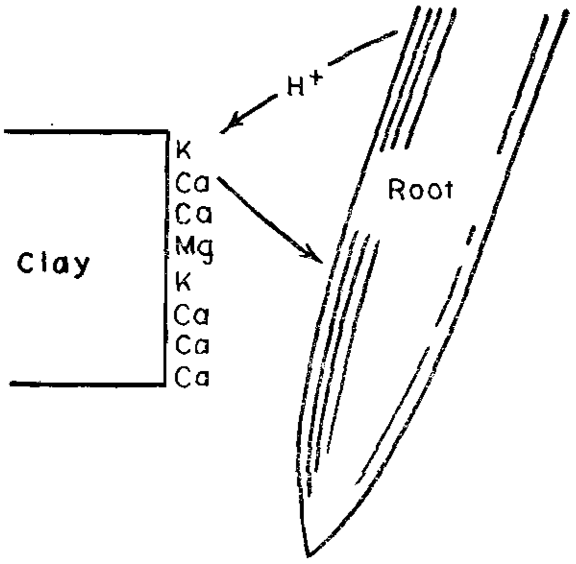

Fig. 7-1. The interchange of cations between plant roots and soil colloids. The nearer a root approaches an exchange surface the greater is the chance for a released H ion from the root to displace an exchangeable ion. In this example, H is displacing K, the latter being absorbed into the root (after Jenny and Overstreet, 1939).

supply of exchangeable nutrients is replenished. The contact-exchange theory will be operative for K, Ca, Mg, and other cations.

Efficiency of Nutrient Absorption

An exchangeable cation, such as the plant nutrient K, will be distributed throughout the mass of soil in the root zone. Since absorption of K by plants is subject to contact exchange reactions, uptake of this nutrient will depend upon how well roots permeate the soil and contact the individual exchange surfaces. Thus, we may see that only a small portion of exchangeable K may be absorbed during a growing season because of the inability of plant roots to contact all exchange surfaces. Not all absorption of K is the result of contact exchange. Some solution (hydrolysis) and migration to roots will take place. Nonetheless, the efficiency of absorption remains relatively low.

A similar relationship exists with the phosphate ion in calcareous soils. Phosphate derived from mineralized organic matter or dissolved soil minerals will remain in solution for only a short time before it is adsorbed onto the surfaces of $CaCO_3$ (precipitated, so to speak) and other soil particles. Absorption of the phosphate by plants depends largely on its being dissolved or displaced when the roots contact the phosphate-adsorbing surfaces. It has been observed in pot studies with radioactive P that a maximum of about one-third of the P applied as fertilizer is taken up by plants. Generally, the efficiency is much lower than this in the field. Thus, it appears that only a small portion of the total available phosphate will be absorbed by plants in a single season. In other words, the level of chemically available P must be sufficiently high so that the amount which becomes positionally available will be adequate for maximum plant growth.

Absorption of soluble, nonexchangeable nutrients from the soil is more efficient. Nitrate-N added to potted soils may be completely absorbed by plants, except where excessive amounts of N are used. The NO_3 ion moves freely with the soil moisture. Therefore, if most of the soil moisture is removed by plants, their roots will have an opportunity to contact most of all NO_3 in the soil solution. Complete absorption will depend largely on the need of the plant for the N, and the efficiency of absorption will decrease as the soluble N supply increases.

Bray (1954) considers that the quantity of a plant nutrient absorbed by plants during the growing season relates to the mobility of the element. Thus, by the end of the season the plant will have taken up essentially all of the NO_3-N that occurred in the soil solution during the season, except where large excesses prevailed. This quantity of N may determine maximum possible yield of the crop if all the available form is absorbed before growth is completed and no additional quantity is forthcoming.

An increase in plant growth is accompanied by an increased size of the root system. This allows for greater contact between plant roots and soil surfaces holding adsorbed, immobile elements. Thus, while the increased growth places a greater demand on the soil for the nutrient, such demand is met more readily through exploration by roots into previously untouched zones of soil. In effect, root extension increases the positional availability of a nutrient element and the total amount extracted by the plant.

PHILOSOPHY OF SOIL FERTILITY CONTROL

Soil fertility is only one factor involved in productivity. Many other factors may be responsible for poor production when it occurs in a particular soil. However, the fertility status of the soil very often can be evaluated properly and means of correcting nutrient deficiencies easily introduced. Complete advantage should be taken of this fact. The maximum productiou potential of a soil should be maintained constantly through the application of fertilizers as needed.

Only a very small fraction of the soils in the Asian subcontinent do not need fertilizer of some kind. Nitrogen is virtually always in short supply. It is only sound logic to maintain the fertility status continuously at the highest possible level, because any losses caused by low yields, even those of a transitory nature, can never be regained. In many instances, only a small investment in fertilizer can correct a deficiency which will otherwise reduce markedly the return the cultivator may expect for his time and for his cost of seed, water, and taxes. It takes neither more feed or larger animals, nor a better plough to cultivate land that will produce 20 maunds[1] of wheat than is required to prepare the land for a 10-maund yield.

If some other factor in production seriously interferes with crop growth, one must proceed with caution in the use of fertilizers. A waterlogged soil, which at best can produce only 10% to 25% of the maximum possible yield, cannot be made to respond well to fertilizers. Such a limiting factor must be corrected before the use of expensive fertilizers is undertaken.

A common belief concerning fertilizer use is that land producing poor plant growth will bring a better return from a fertilizer than will good land. The data of Wahhab and Muhammad (1955) on wheat yields in the Punjab disprove this belief exceptionally well. They have shown that largest increases from fertilizers are obtained on the soils that produce the highest yield without fertilizer. Results of their experiments have been summarized in Fig. 7-2. The results show that, as the indigenous productivity of the soil is doubled (wheat yield increased from 800 to 1600 lbs. per acre), the response to a 40-lb. addition of fertilizer N is almost tripled. The reasons for the initial differences in productivity among the experimental soils are not known, but they can be due to any one of a number of factors—available water or salt level in the soil, nutritional status of elements other than N, etc.

The following chapters will deal primarily with the essential plant nutrients, their forms and availability in the soil, and their supplementation with chemical fertilizers. Before proceeding it is important for the student to become familiar with the concept of crop response to fertilizer treatment. Furthermore, the objective in adding fertilizers must be clearly understood. These materials are costly and difficult to obtain in the Asian subcontinent. Therefore, it is absolutely essential that they be efficiently utilized. The great need for the application of fertilizer materials to most soils in the area

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[1] One maund equals 80 lbs., approximately.

should be appreciated to a greater extent than it is. The perennial shortage of foodstuffs demands this.

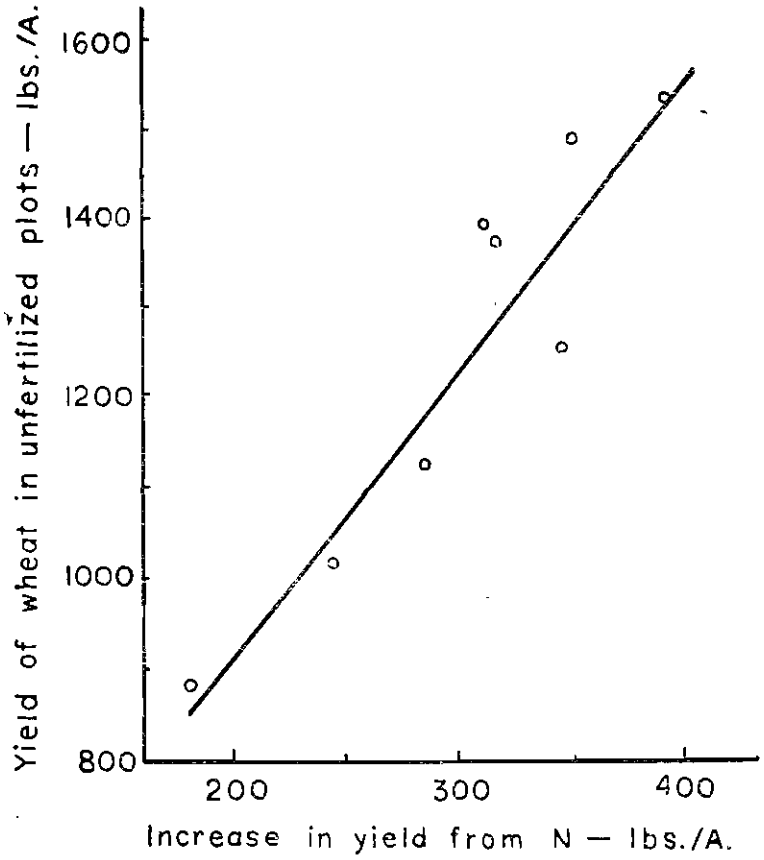

FIG. 7-2. Relationship between yield of wheat on unfertilized Punjab soils and yield increase from the application of 40 lbs. N per acre as $(NH_4)_2SO_4$ (after Wahhab and Muhammad, 1955).

An illustraiton of typical response curves is shown in Fig. 7-3. In this diagram, total yields of a crop, as produced by increasing increments of added fertilizer N, are shown on the

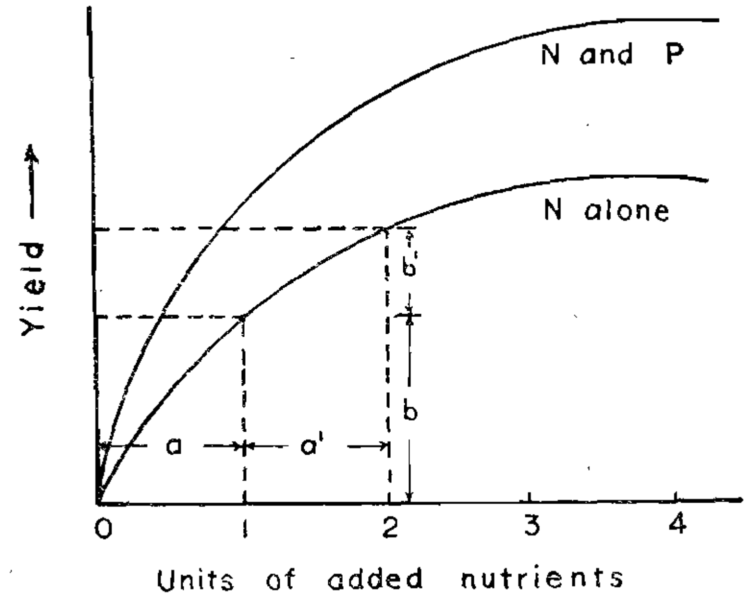

FIG. 7-3. A typical response curve showing the increase in yield due to an increase in quantity of added N fertilizer, with and without added P. Note that the increase in yield, b′, resulting from the addition of increment a′ of N, is decidedly less than the yield increase, b, caused by supplying the same amount of N as increment a. The addition of P increases the efficiency of utilization of the fertilizer N.

Y axis. Two curves are shown, one representing the response to the addition of N alone, and the other when this element is added along with a basal dose of fertilizer P. If these curves represent actual conditions, then they indicate that the treated soil is deficient in both N and P. That is, the plants will benefit from the addition of either or both.

The most striking result noted in Fig. 7-3 is that each increment increase in N applied causes successively smaller increment increases in yield. There comes a point where the addition of an increment of fertilizer causes no further increase in yield. This represents the maximum yield possible under the specific set of conditions involved. On the other hand, there may be a change which will alter this maximum yield value. An example is the addition of P along with the N. The combination results in a higher yield increase for each increment of added N than before. Furthermore, the maximum yield is higher than when the N was used alone. We may look upon the latter circumstance as an increase in the efficiency of use of the added N fertilizer.

Whether one would add P along with the N to the above soil would be determined by the economics involved. It may be seen by comparison between the two curves that, with a little added P, less N may be applied to obtain the same yields as before. If the increase in growth due to P more than compensates for the cost of this fertilizer, certainly it would be a paying proposition to use both fertilizers together. Not only must the total cost of the fertilizer be considered but, also, the value of the crop and the cost of cultivation, irrigation water, if used, and land taxes. None of these can be overlooked in evaluating the benefits derived from fertilizer application.

Another very important point illustrated by Fig. 7-3 is that the economic benefits obtained from the addition of the initial increments of fertilizer are relatively greater than when higher rates of fertilizer are used. This emphasizes the value that can be expected from the application of even small amounts of fertilizer on deficient soils. It also emphasizes the need for conserving the limited fertilizer resources available in the Asian subcontinent. If the available supply of a nutrient in the soil is initially high, on the other hand, further addition would result in only a small increase in yield. Failure to obtain a response to fertilizer may be associated with an initially high level in the soil of the nutrient element which has been added, or else some other growth factor may be limiting and thus prevent efficient utilization of the added nutrient by the plants. Exact conditions in the field must be known before failure for a fertilizer to produce increased yields can be explained. Some information can be gained on the status of various plant nutrients from reliable laboratory soil tests.

REVIEW QUESTIONS

1. Explain the difference between soil productivity and soil fertility.
2. Why does the total analysis of a soil mean little with respect to the fertility status of a soil ?
3. A soil with a high sand content might have both advantageous and detrimental characteristics as far as crop production is concerned. Why is this so ?
4. Name as many factors of growth as you can.
5. Define potential and active fertility. Would you expect to find a general relationship between the two ?
6. What are essential plant nutrients ? Name them. What is meant by the major and minor elements ?
7. Name three concepts of nutrient availability and explain why each may affect plant growth.
8. Describe the two theories of nutrient supply.
9. Why may soluble nutrients be more completely removed from the soil by plants than those adsorbed to soil particles ?
10. What two important roles does H play in the release of plant nutrients from the soil ?
11. Why should one attempt to maintain all plant nurtients at an adeqnate level ?
12. You have a field of well-drained soil with a moderately good fertility level and a correspondingly good productive capacity. You also have a waterlogged field which produces very poor crops. To which would you apply fertilizer and why ?

SELECTED REFERENCE

1. Bray, R. H. (1948) Correlation of soil tests with crop response to added fertilizers and with fertilizer requirement. From *Diagnostic Techniques for Soil and Crops,* pp. 53-86. The American Potash Institute, Washington 6, D.C.

CHAPTER EIGHT

SOIL NITROGEN

NITROGEN AND THE PLANT

NITROGEN OCCURS in the plant primarily as a component of protein, a building block of cellular tissue. As such, it is a basic constituent of the living part of plants, the protoplasm. Nitrogen also occurs in the chlorophyll molecule, the green pigment responsible for photosynthesis. When N is in short supply, chlorophyll formation is inhibited and, as a consequence, photosynthetic activity impaired. The resulting limited quantities of carbohydrates manufactured in the leaves are inadequate to meet the energy requirements of the plant. Further, because of a concurrent shortage of both N and carbohydrates, protein formation is restricted. It may be seen readily, then, why a limited supply of N can hinder markedly the growth of plants.

A reduction in the chlorophyll content of plant tissue results in a change in colour, first to yellowish-green and, eventually, to yellow. The degree of colour change depends upon the severity of the N deficiency. In extreme cases the leaves will die. Since N tends to accumulate in younger plant parts, older leaves are the first to show a yellow colour, or, with severe deficiency, are the first leaves to die.

Absorption of Nitrogen by Plants

The usual form in which N is absorbed by plants is either as the nitrate (NO_3) or the ammonium (NH_4) ion. The NH_4 form appears to be preferred by most plants during early stages of growth and by others throughout their entire period of growth, rice being a plant which fits into the latter category. For the most part, however, absorption of N by plants is predominantly as the NO_3 ion. This is due to the fact that the NH_4 ion is converted rapidly to the NO_3 form in most arable soils. As a consequence, the quantity of NO_3-N in the soil is, more often than not, greater than that in the

NH_4 form. Furthermore, the NO_3 ion is highly mobile and will migrate to the plant root. Ammonium ions, on the other hand, are adsorbed by soil colloids and must be sought out by the rootlets before they can be absorbed. Thus, not only does NO_3 occur normally in greater quantities than NH_4, it is more accessible to the plants.

Factors important in the absorption of the $NO_3^{\cdot}$ or NH_4 ions are summarized as follows:

(1) The uptake of the NH_4 ion may be limited in certain cases by an accumulation of toxic quantities of the anion accompanying it. This would apply primarily when the NH_4 ion is supplied as a supplemental fertilizer salt. The absorption of NH_4 ions is favoured by a neutral or alkaline soil reaction, a dilute concentration of NH_4 in the soil solution, and a nontoxic, associated anion, such as HCO_3 or CO_3[1]. Ammonium-N is absorbed more readily than NO_3-N during the early stages of plant growth.

(2) Nitrogen in the NH_4-ion form appears to be preferred over NO_3-N by those plants high in carbohydrates such as barley, maize, pumpkins, and potatoes. This is because NH_4 can be utilized in protein formation after simple conversion to NH_2 in the plant.

(3) The absorption of NO_3 is enhanced by its mobility or ready accessibility in the soil. Its absorption is at a minimum during the early stages of growth and a maximum at blossom time and thereafter. This probably is related to the period of maximum carbohydrate production by the plant and the consequent demand placed on the soil-N supply for protein formation.

The Utilization of Nitrogen by Plants

Following absorption into the conductive system of the plant, N is converted rapidly into the NH_2 form which then combines with free carbohydrate to form protein. A shortage of N, which, in effect, corresponds to an over-abundance of carbohydrate, results in stunted or impaired plant growth

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[1] These anions are nontoxic if they do not accumulate in the soil solution. They normally disappear rapidly through conversion to H_2CO_3 or through precipitation as $CaCO_3$.

because sufficient plant protein is not formed. The relative abundance of carbohydrates leads, then, to the formation of excessive structural tissue high in C and low in N (*i.e.*, cellulose). Nitrogen-deficient plants are stiff and woody, and form seeds with difficulty. When there is an abundance of N, on the other hand, the proportional amount of structural material such as cellulose is comparatively low. This, coupled with the high concentration of nonstructural protein results in a rapidly growing, succulent, and weak plant. Under these conditions, the lack of reserve carbohydrate for seed formation results in delayed maturity. An excess of N may result in plants which lodge easily and which remain vegetative for undesirably long periods.

The Source of Soil Nitrogen

The initial source of all N in the soil is the atmosphere which contains about 80% of this element in an inert form. To be usable by plants, gaseous N must be 'fixed', either biologically or chemically, into inorganic form. Normally this corresponds to conversion of the N to the NH_4 form. Various microorganisms occurring the soil are responsible for *biological fixation* of N. *Chemical fixation* may occur during lightning discharge, with the fixed N (NO_3, in this instance) being carried to the earth in precipitation ; or it may be transformed under controlled conditions, as during the preparation of commercial N fertilizers.

Because of the limited supply of commercially prepared fertilizers in the subcontinent, much reliance must be placed on other sources of N for soil treatment. Farmyard manure is suitable for this, but it is not available in sufficient quantities to meet even minimum crop needs for N. As an alternative, therefore, natural biological fixation of N must be utilized to the utmost in an effort to increase the N-fertility level of the soils of the region.

FIXATION OF ATMOSPHERIC NITROGEN

Nitrogen fixation is performed in the soil largely by two groups of bacteria. One such group, the *root-nodule* or *symbiotic* bacteria, lives in a mutually beneficial association with the roots of legume plants. Other *free-fixing* or *nonsymbiotic* bacteria occur in the soil and are equally capable of converting atmospheric N into inorganic form. Whereas the nodule bacteria acquire their food for energy from the plant with which they live, the nonsymbiotic organisms attack soil organic matter for this purpose and, therefore, aid in its decomposition. Symbiotic N fixation can be a considerably more important source of N than nonsymbiotic fixation in maintaining the nitrogen fertility status of agricultural soils.

Nonsymbiotic Fixation

Two heterotrophic bacteria are largely responsible for the nonsymbiotic fixation of N in the soil. One is an aerobic, oval-shaped bacterium, *Azotobacter.* It does not function well under strongly acid conditions, particularly if the supply of available Ca is low. Because a large part of free-fixed N results from the activities of the *Azotobacter,* nonsymbiotic fixation is referred to commonly as *azofication. Azotobacter indicum* is thought to be an organism which contributes fixed N especially in paddy soils (Starkey and De, 1939).

An organism of somewhat lesser importance, *Clostridium,* functions well under anaerobic conditions. Therefore, it is likely that this organism may play a role in fixation of N in subsoil, in waterlogged soils, or when the CO_2 concentration of the soil atmosphere is high.

Variable estimates are given as to the quantity of N fixed nonsymbiotically in soils. Amounts up to about 50 lbs. per acre per year are suggested. Because of the low supply of readily oxidizable organic matter in most soils of the Indo-Pakistan region, transformation of N by free-fixing organisms probably is of only minor importance. Thus, little dependence can be placed on this source of N for supplying the needs of crops. The most valuable contribution by these organisms has been in the part they have played in the build-up of organic N and humus in the soil over the centuries.

Some fungi are capable of fixing N nonsymbiotically, but the extent of their contribution is unknown. Algae, as mentioned previously, may aid materially by adding fixed N to paddy soils.

Symbiotic Fixation

The symbiotic fixation of N is a process that has been studied in great detail. Many questions relating to the exact nature of the process and its overall effect on crop production remain unanswered, nonetheless. The realization that all legumes have the power to acquire N through symbiotic fixation has led to the false belief that their N requirements are completely satisfied by this means. Some legume plants respond very favourably when treated with fertilizer N, however. Nitrogen fixation in field beans is particularly low, and these plants actually require fertilization with N for satisfactory yields in many instances.

A single group or *genus* of bacteria, the *Rhizobium*, is responsible for the fixation of N in association with legumes. Normally, these organisms are present in all soils and, therefore, a legume root will become inoculated with them soon after the growth of plants commence. It has been discovered, however, that there are several species of *Rhizobium* and each one is rather specific as to the host plant with which it will work most efficiently. For example, we find that the species *Rhizobium trifolii* is of greater benefit to the clovers than to other legumes, and *Rhizobium meliloti* is specific for lucerne. This knowledge has led to the common practice of inoculating legume seed with specific bacterial cultures prepared especially for each of the various important legume plants. Failure to use these specific bacterial cultures can cause serious loss through reduced yields of the legume. However, on older cultivated lands where a particular legume species has been grown previously, the proper strain of bacteria usually is present. In spite of this possibility, it is much safer to inoculate the seed with fresh, properly prepared *Rhizobium* cultures. These are available generally through the Department of Agriculture and from agricultural colleges.

Invasion of legume roots by nodule bacteria occurs initially by means of an *infection thread* which is produced by the infecting organisms and which penetrates and extends into the main body of the root. The bacteria enter into the root of the host plant by migrating through the infection thread. Once the organisms are within the root cells their number increases rapidly. Further, the plant root cells themselves no longer function normally but, in response to the presence of the microbes, swell and grow irregularly to form the characteristic gall or nodule on the root surface. The nodules are readily observed on the roots of inoculated legumes.

The association between the host plant and *Rhizobium* organisms is to the mutual benefit of each. By converting inert, atmospheric N to usable form, the organisms provide at least a part of the plant's need for this element. In turn, the organisms obtain from the plant soluble carbohydrates which are utilized for their growth and metabolic processes.

The N fixed by root-nodule organisms eventually may be utilized in several ways. Some of these are as follows:

(1) Small amounts are combined into protein by the *Rhizobium* itself.

(2) It is used by the host plant and eventually returned to the soil in plant residues, or it may be utilized when the plant is taken as feed by man or animals.

(3) Under certain circumstances, some of the fixed N may be excreted by the roots into the surrounding soil. Crops grown with legumes seem to benefit from the association, either through use of the excreted N or that released as dead roots from the legume slough off and decay.

MINERALIZATION OF ORGANIC NITROGEN

Indigenous soil N, or that occurring naturally in the soil, is a component primarily of the organic fraction of the soil. Generally it is considered to have accumulated during the normal processes of soil development. All added plant residues contribute to the available-N supply of the soil through mineralization as component proteinaceous compounds are broken down during decomposition.

The release of N from organic combination through mineralization is a step-wise process. The initial transformation results in the appearance of the NH_4 ion and is carried out by a wide variety of organisms. These organisms are incapable of causing further change in the N, however. Therefore, the transformation to NH_4 is considered a separate process and is termed *ammonification*.

Ammonium compounds may be transformed into nitrate salts or HNO_3 by a limited number of special-purpose microbes. The process is aerobic and is referred to as *nitrification.* During nitrification a small amount of nitrite may appear as an intermediate product. Since all NH_4 ions which may occur in the soil are not derived directly from organic decomposition, nitrification cannot be thought of as an essential phase of mineralization of organic N.

The respective rates at which the ammonification and nitrification processes proceed will determine the balance existing between the quantities of NH_4 and NO_3 in the soil. If the organic form of N present mineralizes at a slow rate, the production of NO_3 may exceed the formation of NH_4. On the other hand, where mineralization is rapid, the quantity of NH_4 ions accumulating in the soil can exceed the NO_3. Absorption of these two ions by plants and microorganisms can affect materially the relative proportions of each in the soil.

During mineralization, both ammonification and nitrification reactions go on simultaneously. As a consequence, assessment of the rate of mineralization requires the determination of the rate of accumulation of both the NH_4 and NO_3 ions. Because of this, the mineralization of N usually is considered to include nitrification and the entire mineralization process is expressed diagramatically as follows:

$$\underbrace{\text{Organic N} \rightarrow NH_4}_{\text{Ammonification}} \underbrace{\rightarrow NO_2 \rightarrow NO_3}_{\text{Nitrification}}$$

The Ammonification Process

Many types of the soil-inhabiting organisms are capable of decomposing organic N compounds and liberating NH_4 as a waste product. It is not surprising, therefore, that conditions under which ammonification proceeds need not be highly specific. Ammonification can be either an aerobic or an anaerobic process. Since plants are capable of utilizing NH_4 directly, this process can provide N in available form even under relatively adverse soil conditions.

Factors which control the rate of ammonification are those which influence the more general process of organic matter decomposition. Compounds of both C and N must be present and in forms susceptible to microbial attack. Where the ratio of C to N is wide, little or no NH_4 will be released. If NH_4 is liberated by one group of microbes under these conditions, it will be reabsorbed rapidly by other decay organisms. The addition of small amounts of carbonaceous compounds to the soil may cause an increase in NH_4 production as a result of stimulation of the existing microbial population. However, if these substances are supplied in large quantities and the demand for readily available N by the decay organisms becomes high, free NH_4 will not appear in the soil in measurable amounts.

The Nitrification Process

Nitrifying organisms are relatively specific in their environmental requirements. Like plants, they are capable of utilizing CO_2 directly from the air for assimilation into cell tissue. Both N and energy for growth of these organisms are derived from the oxidation of the NH_4 ion into NO_3. Only a small amount of the N involved will be utilized by the oxidizing organisms, the bulk being released to the soil solution as free NO_3 ions. This conversion is a two-step process expressed by the equations:

$$2NH_4^+ + 3O_2 \rightarrow 2NO_2^- + 2H_2O + 4H^+ + \text{Energy} \qquad (8\text{-}1)$$

$$2NO_2^- + O_2 \rightarrow 2NO_3^- + \text{Energy} \qquad (8\text{-}2)$$

The nitrite ion is seldom observed in the soil because, under normal soil conditions, its conversion to NO_3 is more rapid than its formation from NH_4.

Conditions for maximum production of NO_3-N are rather well defined. Factors which are considered to control the rate of conversion are (1) soil reaction and cation complement on the exchange complex, (2) the quantity of NH_4 present, (3) aeration, (4) temperature, (5) soil moisture, and (6) the relative amounts and forms of C and N in the soil organic matter which is being mineralized.

The presence of an abundant supply of O_2 is most important in the nitrification of NH_4. The above equations indicate that O_2 is required for both steps of the oxidation process. Seldom is there sufficient O_2 in the soil atmosphere to allow for a maximum rate of conversion of the NH_4 to the oxidized NO_3 form. Therefore, a basic requirement for the rapid formation of NO_3-N is good soil aeration. This fact explains partially why frequent and repeated ploughing appears to benefit food-grain production on soils of the Indo-Gangetic Basin. Aslam (1957) has shown that the NO_3 level of Punjab soil and the yield of wheat are increased as the number of preseeding ploughings is increased up to a total of 12. The increased aeration this has produced brings about not only a more rapid decomposition of the native or added organic matter in the soil but also the conversion of liberated NH_4 to NO_3.

Broadbent *et al.* (1957) studied the effect of high concentrations of ammonium fertilizer salts and various soil factors on the rate of nitrification. They observed several conditions which were thought to contribute to a slow rate of nitrification. Among these are:

(1) Excessively high pH, although nitrification was observed to proceed slowly above a pH of 9·0.

(2) Excessively low pH, which resulted from the conversion of NH_4 to HNO_3. Nitrification was much reduced at a pH of 5·0 although it continued even at pH 4·5.

(3) The presence of free NH_3, which in itself is a reducing substance, will interfere with the oxidation of NH_4 to NO_3. Free NH_3, which forms NH_4OH in the soil solution, will occur only under alkaline soil conditions. The above authors assume that this fact might account for the observed accumulation of NO_2-N in alkaline soils.

(4) Excessively high osmotic concentration in the soil solution where the amount of fertilizer salts is high.

Broadbent and his coworkers found the oxidation rate in acid, noncalcareous soils to vary in the order of $NH_4OH > (NH_4)_2SO_4 > NH_4NO_3$. In alkaline, calcareous soils, on the other hand, NH_4OH was converted at a slower rate than the other two, possibly because of an exceedingly high pH within the fertilizer zone in the soil.

The effect of the NH_4-ion concentration and of time on the rate of nitrification is illustrated clearly in Fig. 8-1. It will be noted in the data shown that, at low NH_4-salt concentrations, complete transformation may be expected in less than one week's time. It can be seen from the upper curve in the figure, which is for a low concentration of added NH_4, that more NO_3-N was recovered than was added initially to the soil. The excess N came from the organic matter present in the soil. At higher levels of NH_4, the relative rate of conversion is slower although the total amount of NO_3-N produced for a given period of time is greater. The lowest curve in the figure represents a treatment wherein the concentration of NH_4 salt is so high that it inhibits markedly the activity of the nitrifying organisms.

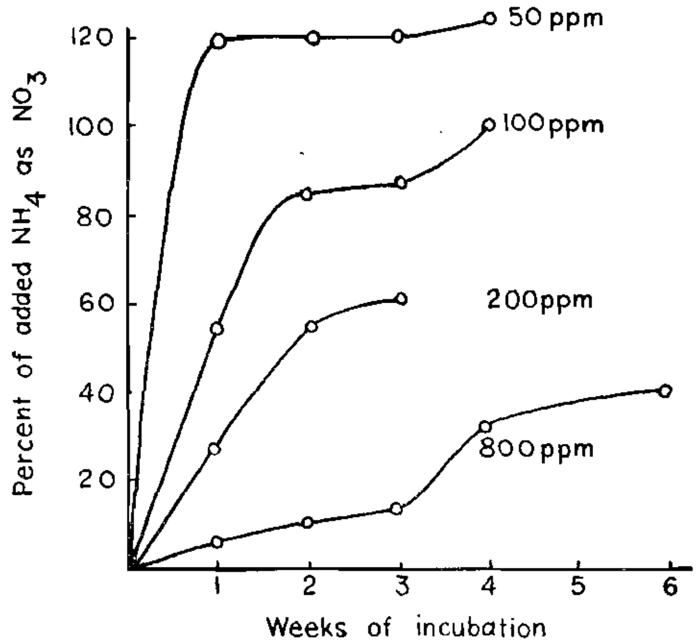

FIG. 8-1. Nitrate accumulation in Yolo loam as affected by time of incubation and initial concentration of NH_4 from $(NH_4)_2SO_4$ (after Broadbent, Tyler, and Hill, 1957).

The rate of accumulation of nitrate in the soil is used often as a measure of the rate of nitrification. This technique has certain limitations when applied in the field because other factors can influence the NO_3 level. For instance, at the time of lush plant growth or immediately following a period of pronounced leaching, the level of NO_3 may be low even though nitrification proceeds at a rapid rate. The presence of readily oxidizable organic matter having a wide C : N ratio also will affect the level of NO_3. Where the proportion of C to N is high, mineralized N is absorbed rapidly by the organisms active in decay processes and accumulation of NO_3-N in the soil solution is inhibited.

The population of soil microbes is greater in close proximity to root surfaces than it is at some distance from the root. This is attributable to the presence near the root of excreted or sloughed organic substances which serves as a source of energy material for the microorganisms. Therefore, where root density in the soil is high, as under grass sod, a continuously low level of NO_3 may be expected. In this instance, active absorption of N by both roots and microbes will assist in maintaining the low NO_3 level.

Although temperature has a direct effect upon the rate of oxidation of NH_4, the extremes normally experienced in the field, even in the hottest parts of Pakistan and India, would not be expected to interfere materially with the process. The maximum temperature at which nitrification will proceed at any appreciable rate is about 100°F. Whereas air temperatures often exceed this, soil temperature will not, except for the very surface layers of dry soil.

The threshold temperature at which nitrification begins is about 40 to 45°F, and the rate at which it proceeds approximately doubles for each 10 to 15° rise above this minimum. The rate is dependent also upon the level of soil moisture and the state of aeration. Important in this respect is the drying effect which elevated temperatures and low rainfall have on the moisture content of many subcontinent soils during late spring and early summer. The nitrifying organisms will have about the same requirements for moisture as will higher plants. Thus, if the surface soil is permitted to dry out, nitrification, and mineralization as well, may be reduced to a point where the early-sown *kharif* crops such as cotton and fodders suffer from a shortage of available N.

LOSSES OF NITROGEN FROM THE SOIL

A part of the N accumulated in or added to the soil is lost eventually. Various means of its disappearance are recognized. These include (1) use by plants, (2) loss by erosion, (3) leaching losses, and (4) volatilization into the atmosphere. The utilization by plants is, of course, desirable in most instances. At times, its consumption by soil organisms is desirable unless this results in competition with crop plants during a critical period of growth. Nitrogen absorbed by and held in the protoplasmic tissue of microorganisms eventually will be converted to usable form upon the death and decay of the organisms. Volatilization and erosional and leaching losses, on the other hand, constitute a serious wastage of vital soil N. Some discussion of each of these methods of loss is worthwhile.

Loss by Volatilization

Nitrogen is known to disappear from the soil in several volatile forms ; as NH_3 gas, elemental N, or as volatile oxides of N. Common heterotrophic organisms appear capable of causing the formation of these gases, primarily through the reduction of NO_3-N. The effect is essentially one opposite to the nitrification process and is aptly named *denitrification*. It occurs during the process of decomposition of organic matter if the O_2 supply is suboptimal. Denitrification will be more prevalent, then, under waterlogged conditions, in soils where the structural configuration interferes with good aeration, or where large amounts of highly carbonaceous organic materials are undergoing active decomposition in the soil. The latter condition causes a reduction in the O_2 content and an increase in the CO_2 content of the soil air. Denitrification losses are difficult to measure, but, at times, they appear to be appreciably large.

Wagner and Smith (1958) have noted losses of applied N through volatilization to be as high as 85%. They observed that losses by this means could not be related to any specific soil type or characteristic nor to type of N fertilizer salt. They were able to detect volatile forms of NH_3, nitric oxide (NO), nitrogen dioxide (NO_2), and nitrous oxide (N_2O). They felt that the latter gas was the form in which most N was lost from the slightly acid soils they studied.

The NH_4 ion, possessing a positive charge, is adsorbed to the cation-exchange complex. The NH_4 in a fertilizer applied to the land by broadcasting is taken up and held very near the surface of the soil. If the soil is alkaline, there is a strong tendency for the NH_4 ions to hydrolyze from the colloidal surfaces according to the reaction:

$$\boxed{\text{Clay}}\,NH_4 + HOH = \boxed{\text{Clay}}\,H + NH_4OH \qquad (8\text{-}3)$$

A solution of NH_4OH is, in reality, an equilibrium mixture of molecular NH_4OH, NH_4 and OH ions, water, and dissolved, molecular NH_3. This equilibrium is represented as:

$$2NH_4OH = NH_4^+ + OH^- + HOH + NH_3 \nearrow \qquad (8\text{-}4)$$

As indicated in the equation, the NH_3 form of N is subject to volatilization, a fact well known to anyone who is familiar with the odour of a solution of NH_4OH. Volatilization of NH_3 from the soil can and does occur and causes the reaction expressed in equation 8-4 to proceed to completion. Thus, NH_4 ions adsorbed at the surface layer of an alkaline soil can *convert to NH_3 and be lost. Volatilization will proceed most* rapidly as the soil dries. Therefore, losses of this nature will be retarded if evaporation of moisture is prevented. Volatilization will be lower in soils of high colloidal content and of no consequence if the pH of the soil is below 7·0.

Wahhab *et al.* (1957) predict that volatilization losses may be as high as 50% of the surface-held NH_4 in alkaline soils. It is obvious, therefore, that placement of NH_4 fertilizers below the soil surface constitutes one measure for its conservation. Broadcasting of an ammonium fertilizer just prior to ploughing provides one of the simplest procedures for placement and will minimize volatilization loss. Ploughing-down also places the N in a position in the soil where moisture is present for a longer period of time and where absorption by plant roots is more nearly assured. A furrow-turning plough which completely inverts the surface soil is most suitable for burying surface-applied materials.

Erosional and Leaching Losses

Loss of N and other plant nutrients due to the effects of excess moisture can be very large. This is particularly true if it is caused by the removal of topsoil through erosion. The fertility status of the topsoil normally is much superior to that of the subsoil. Thus, removal of a few inches of the surface layer can cause a disproportionate reduction in fertility and yields. This problem plagues those areas especially where precipitation is insufficient for adequate growth of a protective plant cover or where overgrazing prevents the maintenance of a good stand of plants, and where the rainfall is torrential and, therefore, erosive. Maintenance of a plant cover, particularly through use of close-growing types, is the best safeguard against erosional losses.

Leaching will occur in high-rainfall areas and in irrigated areas where over-application of water is practiced. Only N in the form of NO_3 will leach from the soil to any extent. Because NH_4-N normally converts rapidly to the NO_3 form in soils, even the N in ammonium fertilizer salts soon becomes subject to leaching loss. An advantage of maintaining N in organic form is realized, therefore, because slow mineralization and release of NO_3 during the growing season minimizes its accumulation and possible loss in percolating water.

The roots of green manure or cover crops, particularly nonlegume varieties, absorb free NO_3 from the soil about as rapidly as it is formed through the mineralization of organic matter. A plant cover on the ground is highly desirable, therefore, as protection against both leaching and erosional losses. In the subcontinent, this protection is more necessary during monsoon season than at any other time.

SUPPLEMENTATION OF SOIL NITROGEN

Providing plants with adequate N is a difficult task and a subject for much concern because of the expense of N fertilizers. Important to the proper use of supplemental N are (1) consideration of the exact needs of the plant being treated, (2) the availability of the N in the fertilizer material, (3) the availability of the native soil N, (4) the possibility of N loss from the soil by various means, and (5) the natural productivity of the soil as it relates to the efficient use of the added N. In southern Asia, nitrogen is in short supply and relatively expensive in terms of value of the crops obtained from the land. Efficient utilization of the limited quantities currently available is absolutely essential. Methods of use of N must be simple so that the cultivator can put them into practice without difficulty. Furthermore, he must be convinced that use of N fertilizers, and of other fertilizers as well, will represent a financial gain to him. Usually this comes about as a result of his observing successful use of supplemental plant nutrients by neighbouring farmers. Acceptance of new ideas is slow at best and then only by a very small proportion of the farming people. It is necessary to rely on these few to introduce new methods and their advantages to the agriculture of an area. This is true the world over.

Modern agriculture demands a continuing, increased use of chemical fertilizers to supplement those plant nutrients obtained from the soil. Nitrogen is one of the most needed as a supplemental plant nutrient. As use of this element is increased and, therefore, crop growth is stimulated, total requirements of the other nutrients in the soil will become greater. Thus, deficiencies of these other nutrients will begin to appear more frequently. Although facilities for the manufacture of synthetic manures are increasing in the Asian subcontinent, they will be inadequate for many years to come. In the meantime,every effort should be made to return to the soil the organic forms of N which are presently available, such as the manures and other plant wastes, and their use should be consistent with good management.

Inorganic Sources of N

Chemical manures containing N are derived principally from two sources ; from natural salt deposits or through the synthesis of NH_4 by chemical fixation of atmospheric N. Good discussions on the processes of manufacture of synthetic salts and the purification of natural materials are given by Collings (1955) and Tisdale and Nelson (1956).

When added to the soil, the NO_3 and NH_4 ions are markedly different, but rapid conversion of NH_4 to NO_3 erases this difference. In certain respects, therefore, the consideration of commercial-N fertilizers involves essentially only the NO_3 ion. Other effects of their basic differences will be noted shortly.

There are numerous compounds which serve as sources of fertilizer N. Only a few are given consideration here.

Ammonium Sulfate-$(NH_4)_2SO_4$: This salt contains from 20 to 21% N. Its preparation involves the reaction of gaseous N_2 and H_2 under high temperature and pressure in the presence of a catalyst to form NH_3. This product, in turn, may be reacted directly with H_2SO_4 to form $(NH_4)_2SO_4$, or it may be converted first to $(NH_4)_2CO_3$ and then to the sulfate by reaction with gypsum, *i.e.*,

$$(NH_4)_2CO_3 + CaSO_4 \rightarrow (NH_4)_2SO_4 + CaCO_3 \quad (8\text{-}4)$$

Ammonium sulfate has attained a position of great importance in the Asian subcontinent[1]. Its principal disadvantage is its relatively low N content which results in higher transportation costs. It is an acid-forming material in the soil. Oxidative nitrification of the NH_4 ion results in the formation of HNO_3 and excess H ions combine with SO_4 to form H_2SO_4. It is possible to reduce the pH of a noncalcareous, sandy soil by several pH units with heavy applications of $(NH_4)_2SO_4$.

Ammonium Nitrate-NH_4NO_3 : The N content of commercial NH_4NO_3 is about 33%. The salt absorbs water readily, and the granules or pellets into which it is formed usually are coated with a waterproof material to facilitate its handling. Ammonium nitrate is a very good source of N and, because of its higher N content, is less expensive to transport per unit of N than is $(NH_4)_2SO_4$. It is highly soluble in

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[1] Ammonium sulfate is particularly useful with rice. This plant appears to prefer N in the NH_4 form. Since it is held by the clay particles it will not leach. This is important in paddy soils which are under water for such long periods. The anerobic conditions in paddy soils preserve the NH_4 form and conversion to NO_3 will take place only to a slight extent.

water and provides both mobile NO_3-N as well as immobile NH_4 ions, the latter being adsorbed by the soil particles.

Ammonium nitrate is prepared by reaction of NH_3 gas with HNO_3. Because only one-half of the N is as NH_4, the residual acidity is not so great as with $(NH_4)_2SO_4$.

Sodium Nitrate-$NaNO_3$: Sodium nitrate is a salt obtained either from natural mineral deposits, or it may be produced synthetically. The largest natural deposits occur in Chile ; thus, the common name of Chilean sodium nitrate. It contains 16% N, all of which is in the NO_3 form. Because of the presence of the Na ion, it produces an alkaline reaction in the soil ; therefore, it is not recommended for application on soils which have a natural tendency to become strongly alkaline. Although sodium nitrate is a very good fertilizer material, its low analysis (percentage of N) is not in its favour.

Urea - $CO(NH_2)_2$: Urea is worthy of special consideration because of its relatively high N content (near 45%). It is an easily handled material, which should tend to make it popular with the average cultivator. Upon contacting moist soil, it is converted rapidly to $(NH_4)_2CO_3$ by a naturally occurring enzyme in soil, *urease*. After this change, it behaves similarly to $(NH_4)_2SO_4$, although the carbonate salt causes a much more alkaline reaction than does the sulfate.

The soluble carbonate resulting from transformation of urea is capable of precipitating exchangeable Ca. The NH_4 ion will take the place of Ca on the exchange complex. However, since the oxidation of NH_4 to HNO_3 leaves an acid residue, the $CaCO_3$ that has been formed will redissolve, and the soluble Ca may then reenter exchange positions. Through this series of reactions, urea is converted completely to NO_3 and CO_2. Both will disappear from the soil without leaving a residue.

Urea is formed by the reaction of NH_3 with CO_2 under high pressure.

Anhydrous or Aqueous Ammonia - NH_3 : A material that has become very popular as a source of N is NH_3, either in the gaseous or aqueous form. The gas contains 82% N, which is considerably more than any other popular fertilizer. This is not a great advantage, however, because it must be transported under pressure in heavy tanks. Gaseous NH_3 is usually convenient to use where the agriculture is highly mechanized. The tanks can be loaded onto tractors and the gas dispensed through tubes so that it may be injected directly into the soil. The gas is absorbed immediately in the soil water and is retained readily by the soil provided it is placed several inches below the surface. Injection near the surface of an alkaline soil can result in loss through volatilization of free NH_3 gas.

Aqueous ammonia is prepared by dissolving the gas in water. The maximum concentration obtained forms a solution containing 25 to 30% N, but it may be increased if salts such as NH_4NO_3 are added also. High-pressure tanks are not needed for the transportation of aqua ammonia.

Organic Sources of Nitrogen

The mineralization process whereby N is released from organic combination in a form available to plants is more complex than is the simple conversion of NH_4 to NO_3. In their attack on soil organic matter, the organisms involved seek out C both for energy and for growth. The use of C for energy results only in its liberation as CO_2. During growth, however, C is converted to protein and retention of a certain amount of N is necessary also. Thus, if the supply of C is disproportionately high, N derived from organic residues will be locked up in microbial tissue until the level of C has been lowered appreciably through CO_2 evolution. That is to say, immobilization of organic N initially exceeds its rate of mineralization and its release for plant use is inhibited. As a general rule, if an immediate increase in the level of available soil N is wanted, an added organic residue should contain no less than from 1·5 to 2·0% N at the time of its incorporation into the soil.

The use of organic materials for soil treatment may involve green manuring, which is the ploughing down of plants grown especially for this purpose, or the application of animal manures or other organic substances (*i.e.*, cottonseed cake, straw, etc.). Where a choice is possible, those materials high in total N normally are preferred. The total-N content of organic supplements is not the only important consideration in determining their benefit to growing plants, however. Of great importance is the rate at which the N is mineralized. This depends upon the form in which the N and C occur and the C : N ratio of the supplemental organic material.

Some forms of organic N are mineralized with great difficulty, particularly the more complex proteins, even though they have a narrow ratio of C to N. Organic substances high in these components seem never to succumb completely *to the mineralization process. When it occurs in a resistant* form of this type, the added N will behave like that in the more stable humus fraction of the soil and will become available to plants only very slowly.

The interrelationship among factors important to the rate of mineralization of N is illustrated by the data in Table 8-1. Generally, the higher the total N the greater is the total amount of NO_3 produced. However, this relationship does not hold true when a comparison is made between lucerne and raw bone meal. Although the N content of the bone meal is higher, it occurs largely in complex protein form which does not succumb readily to microbial attack. The nitrogenous compounds in lucerne, on the other hand, are broken down

Table 8-1. The total-N content and the rate of release of NO_3 from several organic materials (based on data of Rubins and Bear, 1942).

Organic material	Total N	Water-soluble N	% conversion to NO_3 in	
			20 days	40 days
	%	(% of total N)		
Cottonseed meal	7.2	7	49	54
Lucerne	2.8	48	24	32
Bone meal (raw)	4.2	0	7	10
Dried blood	13.8	2	60	66
Horse manure	1.5	20	-19	-16
Chicken manure	2.3	67	22	30
Tobacco stems	1.0	47	-14	5
Wheat straw	0.3	38	-16	-15

easily and allow for the relatively rapid accumulation of NO_3 ion in the soil.

The influence of the amount of oxidizable C in the organic material reflects in the rate of NO_3 production. Horse manure, tobacco stems, and wheat straw are high in carbonaceous compounds. As a consequence, the release of NO_3-N during the decomposition process is impeded. As indicated by the negative values for these materials in Table 8-1, some of the nitrate originally present in the soil was consumed by the organisms as they attacked these carbonaceous substances.

It will be noted in Table 8-1 that, in most instances, the rate of mineralization is reduced greatly after the first 20 days. This would suggest that a rapid change in the easily converted forms of N occurred during the early stages of mineralization. Later change, involving the more resistant protein compounds, proceeds more gradually.

Animal manure is one of the more readily obtainable organic supplements for use on the land in the subcontinent. The importance of this substance as a source of N for plants can hardly be overemphasized. Unfortunately, however, the practice of using manure as a fuel deprives the soil of a very valuable source of N. The reason for this is, of course, obvious, since manure generally is much less expensive than are other fuels. However, failure to apply manure to the land, especially where no other form of N is used, may be limiting crop yields seriously. It is possible that losses caused by this are of greater magnitude than are the savings gained through use of the manure as a fuel. There is a real need for experimentation to determine whether animal manure is more valuable as a source of greater crop productivity or more valuable when used as a fuel.

Manure should be used in a sustained or continuing programme. This arises primarily from the fact that the N in it is mineralized at a relatively slow rate. No more than about 25% of the N in manure is mineralized during the first year following its application. In some extreme cases, usually where the manure contains an abundance of low-N wastes, mineralized N may not appear in the soil at all during the first year. Under these conditions, the small amounts of mineral

N in the soil may be tied up temporarily by microbial organisms attacking the carbonaceous substances. As a consequence, before applying to the land, it is frequently desirable to allow partial decomposition of the manure to take place during the preliminary period of moist storage, or *composting*. Not only is the C : N ratio of the material narrowed during this period, but its physical state is improved also and handling at the time of spreading on the land is facilitated.

The relative importance of the rate of release of N from organic sources depends upon the type of crop being grown. Plants with a short growing period, wheat for example, will require an abundant supply of N for a relatively limited time only. To supply the N needs of this plant, organics which will mineralize rapidly are desirable. A slowly maturing crop, such as sugar cane, will make better use of organic supplements which have a slow rate of mineralization.

The use of organic sources of N can be encouraged only where less expensive materials are unavailable. Although various benefits may be gained from their use, the value of these organic materials must be determined by the N they contain. By comparing the cost for each pound of N in organic and inorganic sources, we find the relative cost almost always in favour of the commercially prepared salts. Furthermore, the composition of chemical manures may be relied upon to be fairly constant. Not so with organic materials, particularly manures. This fact must be kept in mind in determining the rates of material to be applied to the soil to meet a specific requirement for N as a plant nutrient.

An example of an advantageous use of organic wastes is provided by Karim (1948). He encourages the utilization of water hyacinth (*Eichhornia crassipes*) as a manure and source of N. The N it contains, which amounts to a little less than 2%, becomes largely available after a few months' decomposition in the soil. The hyacinth is a weed which forms a lush growth in the postmonsoon season throughout the eastern reaches of the subcontinent.

Organic substances as soil supplements may be used to advantage for reasons other than to supply N. Important is their content of plant nutrients in addition to N and, also, the effect these materials have on the physical properties of the soil. These advantages can best be evaluated by measuring the specific influence they have on crop growth.

RECOVERY OF ADDED NITROGEN BY PLANTS

Theoretically, if all of the N added to the soil were converted to the NO_3 form, complete recovery by the plant should be possible. However, such efficient utilization is observed very infrequently. In addition to the losses of N from the soil as discussed above, immobilization by microbial organisms and failure for roots to contact all of the available N present in the soil operate together to account for incomplete uptake by plants.

Additions of fertilizer N to the soil stimulate microbial activity, even where the soil organic matter is composed almost entirely of essentially stabilized humus. The reactivation of the organisms results in immobilization of a part of the added N, this effect being more pronounced with increasing rates of N application. The effect is temporary, however. When the organisms die and their bodies undergo decomposition, the immobilized N is released once again to become available to plants.

In a series of greenhouse studies, Pinck *et al.* (1948a, 1948*b*) observed that losses of N through volatilization, apparently a consequence of denitrification, also increased with the size of the fertilizer N application. These losses were of little consequence if the amount of N added was less than about 200 to 250 lbs. per acre. However, because of these losses, even if immobilization by microorganisms and leaching losses were avoided, one could not expect to regain in the plant all N added to the soil as fertilizer. The difference between the amount applied and the amount recovered is, of course, much greater for organic N than for inorganic fertilizer salts. Incomplete mineralization of the organic N is responsible for this.

Techniques for Conserving Fertilizer Nitrogen

Since the losses of N from the soil assume their greatest proportions when the quantity of this element in inorganic form is large, its conservation will consist logically of minimizing the concentration of NH_4 or NO_3 ions in the soil at any one time. This can be accomplished principally by applying fertilizer N in limited amounts only and at a time when its absorption by roots will take place rapidly. If leaching of the applied N is a problem, application to the soil far in advance of seeding should be avoided since plant roots would not be present to absorb it. Similarly, special care should be exercised in the application of fertilizer N during the monsoon season. Where serious loss of N either through leaching or volatilization is not experienced, methods and time of application and the amounts of fertilizer applied are not so critical and can be adjusted as is convenient to overall farming operations.

Where the total amount of N fertilizer needed to assure maximum yields of a crop is large, the concentration of inorganic N in the soil at any one time can be kept low through use of split fertilizer applications. In using this technique, only a part of the total quantity of N is applied initially, and the remainder is added after the root system of the plants has become well established. The length of time between the initial and subsequent applications will depend upon the growth habits of the plants being fertilized. Two to three weeks may constitute a suitable waiting period after germination of rapidly growing plants. A longer delay is permissible for a slowly growing crop such as sugarcane. In this latter instance, it may be desirable to add the N in three split applications of equal size.

Ideally, any fertilizer applied to a field where plants are growing should be placed below the surface so that immediate contact with the roots can be made. This approach is especially desirable in alkaline soils where the N is supplied in the ammonium form which may be lost partially through volatilization. Placement is not practical, of course, where the crop has been sown by broadcasting. If the plantings are in rows, on the other hand, a plough or cultivator equipped to place the fertilizer below the surface can be drawn alongside the plants without uprooting them. In addition, row cropping contributes materially to the control of weeds since such control can be carried out through use of animal-drawn cultivating equipment.

THE USE OF MANURE AS A SOURCE OF NITROGEN

Some points need yet to be drawn out with respect to the handling of manures and organic wastes as a source of N for plant use. All forms of supplemental N must be preserved carefully ; they are too valuable to be treated otherwise. Main considerations concern the faulty methods practised in the handling of farmyard manure and the almost complete lack of use of green manures.

A brief study of Fig. 7-3 on page 215 shows that the addition of N to a soil low in this plant nutrient may bring about spectacular yield increases. That is to say, where available N in the soil is very low, much greater gains from an application of N may be expected than if the level of the available form of this element in the soil were high. Most soils in the subcontinent are low in organic matter and available N, however. Therefore, even modest applications of N should cause substantial yield increases for most nonlegume crops. Each pound of N lost through poor handling of manure denies the farmer of these gains. Since increased N in the soil will result in more food, it can result also in cheaper food. Greater yields gained through common-sense practices of N conservation may be obtained without measurable additional cost to anyone.

Farmyard Manure

It is customary throughout the subcontinent to place farmyard manure in piles in the villages or on the field where, in the latter instance, it may be left untouched for periods up to several weeks in length. As long as the manure remains moist, mineralization can proceed at a rapid rate under the generally warm temperatures. Normally, however, the manure heaps remain in the field until they become thoroughly dry and, during the drying process, most of the NH_3-N that has formed will be volatilized. The amount of the original N that is lost cannot be estimated accurately, but it is quite likely that as much as 50% may disappear.

Allowing farmyard manure to stand for long periods in the field, as shown above, can result in the loss of valuable plant nutrients. Its immediate incorporation into the soil is essential to obtaining maximum benefits from an application to the land.

In the more humid regions, losses through volatilization from manure placed in the field may be replaced by those caused by leaching. Not only N but the soluble forms of any of the plant nutrients may be carried into the soil by excess moisture. In this latter case, the plant nutrients from manure stacked in the field would not be lost entirely to the plants. On the other hand, they would be concentrated in a single spot beneath the manure heap and would be of benefit to only relatively few of them. In some instances, nutrient salts leached from manure piles may accumulate in the soil to such a high concentration as possibly to cause damage to the plants growing in that very location.

When manure is carried to the field, it should be worked into the soil as soon as possible after application. This will prevent volatilization losses and the undesirable effects of leaching. However, it is not always practical to incorporate manure into the soil when it is carried to the field at frequent intervals and in small quantities. It is to the advantage of the farmer, therefore, to accumulate manure and other wastes in quantities suitable for treatment of an entire field, and to apply them to the land just prior to ploughing. During the storage or composting period, if suitable conditions are maintained within the storage heap, active decomposition of the organic wastes will take place. Substances of a wide C : N ratio will be reduced to products of a lower C content. Thus, at the time of spreading and incorporating into the soil, the C : N ratio will be nearer the level necessary for a more rapid release of mineralized N to the crop growing on the treated land.

Basic to proper storage is the maintenance of the manure in a moist and compacted state. Compaction reduces aeration and excessive moisture loss and, therefore, the volatilization of NH_3-N. Manure may be kept too moist, of course, which results in the leaching of soluble components. Protection against such loss must be afforded manure in humid regions if its strength for producing plant growth is to be retained.

Protection against excessive drying is achieved by storing the manure in pits. In this way it may be guarded on all sides by the insulating soil. The storage heap should be provided with a protective covering of soil, straw, or leaves.

The pit should be positioned so that it is not subject to inundation by overflow water nor by seepage from ground or drainage water. Where this may occur, above-ground storage is essential. Small enclosures may be constructed from mud and covered with straw or other substance to protect the manure.

Generally, the wider the C : N ratio of a material to be composted the longer should be the period of composting to allow for more complete breakdown. Organic substances which have a narrow ratio and which may be stored for long periods should be mixed with straw or other highly carbonaceous material to reduce the rate of release of mineralized N.

Green Manures

The seeding of crops especially for use as a green manure is a very neglected practice in the Asian subcontinent. For the most part, the average farmer does not understand that advantages may be gained from growing a crop for incorporation into the soil as opposed to harvesting it for use or sale. Yet, there are various ways to supply soil N in this manner which are adaptable and economically feasible. Proof of the need for green manuring is necessary, however, and to obtain this proof one must take fully into account the financial losses and gains resulting from the operation and the practicability of carrying out such practices. There are circumstances, of course, wherein green manuring would be difficult for the farmer to undertake.

Either legumes or nonlegumes may be utilized as green manure crops. However, the use of nonlegumes for this purpose is extremely limited. They are grown normally only to absorb N released through active mineralization or from applied fertilizer when there is considerable risk of losing this nutrient through leaching. The general procedure in green manuring with nonlegumes is to plough under the plants while they are still young and succulent. The young tissue has a narrow C : N ratio and the N is in an easily mineralizable form. Thus, the N absorbed by the green manure crop will be conserved against loss and still may be utilized readily by crops that follow.

It is a rather common belief that root residues from a legume crop will increase the level of C and N in the soil. This is true only infrequently, largely because most of the N in a legume plant is concentrated in the top parts. For some annual legumes such as peas and beans, no more than about 10% of the total N fixed during the life of the plant may be found in the roots when the plant is mature. For perennials such as clovers and lucerne, only about 30% of the total N remains in the roots. To obtain an appreciable gain in the level of soil N following legumes requires that some of the top growth be incorporated into the soil along with the residual root materials.

There is considerable variation among legume species in their capacity to supply fixed N to the soil. For instance, experimental work has shown that the contribution to the soil N level may be seven times higher for lucerne than for a crop such as field peas. The data from a 10-year study on such effects are summarized in Table 8-2. During the course of this particular experiment, various legumes were grown alternately with small grains in potted soils. All top growth was removed from the pots and, therefore, changes in soil N were the result of accumulations from roots only. Changes in total N in the soil were detected by analysis of samples taken at the beginning and conclusion of the experiment.

Table 8-2. Apparent N fixation by various legumes (after Lyon and Bizzell, 1934).

Crop	Apparent N fixation[1]		Gain or loss in *Soil* N
	Annual	Relative	
	Lbs./A.	%	Lbs./A.
Lucerne	241	100	607
Red Clover	146	60	595
Alsike Clover	136	56	532
Soy Beans	102	42	—42
Vetch	65	27	97
Field Beans	57	23	—100
Field Peas	46	19	—32

[1] Determined as the total N removed in plant tops plus the change in level of soil N for the 10-year period, all divided by 10 to place it on an annual basis. Lucerne, for example, fixed a total of 2411 lbs. of N during the 10-year period.

Table 8-2 indicates especially the value of lucerne and clover as contributors of N to the soil. Their contribution as N from root residues will be around 50 to 60 lbs. per year. This quantity will go far in providing the need for N by nonlegume crops such as cotton or wheat. It will fall short of the N requirement of crops with a high demand for N, however. Sugarcane, maize, and summer forages are in this latter category.

As with most plants, parts of legume roots are sloughed off continually during the lifetime of the plant. Such residues have a narrow C : N ratio and undergo mineralization readily. Upon conversion to NO_3 form, the N released from the root tissues is susceptible to leaching. Thus, where large quantities of water percolate through the soil, measurable amounts of the mineralized N may be lost.

Conservation of N released by legume roots may be accomplished either by the addition to the soil of organic materials with a wide C : N ratio, or by growing a nonlegume crop in association with the legume. In either event, mobile nitrate will not accumulate in the soil solution to a level where leaching losses will be serious.

It is realized generally that a legume plant will absorb NH_4 and NO_3 ions present in the soil. Further, when this happens, fixation of atmospheric N by the root-nodule organisms is reduced. Therefore, to obtain maximum fixation of N by legumes, it is necessary to maintain available forms of N in the soil at as low a level as possible. This is done at times by incorporating strawy residues into the soil when the land is being prepared for seeding to the legume. As mentioned above, nonlegume companion crops, particularly grasses, may be grown with the legume since they are very effective in removing essentially all mineralized N from the soil. Data of Lyon and Bizzell (1934) bear this out. In their 10-year greenhouse experiment, lucerne grown alternately with grains fixed a total of 2411 lbs. N per acre. However, since the legume was grown for only 5 of the 10 years, the quantity of N fixed each year it was grown was 482 lbs. per acre. Within the same experiment, lucerne grown continuously for 10 years fixed 2684 lbs. of N per acre. Thus, without a grain crop to remove excess

N every other year, lucerne fixed only 268 lbs. of N per acre per year.

Green manures cannot be effective if there is a shortage of soil water. Thus, the use of green manures may be restricted during dry periods. On the other hand, during the time of monsoons, land otherwise in fallow can be used effectively for green manuring. Not only will this result in the improvement of soil conditions, but, also, erosion by water can be reduced through the protection offered by the plant cover.

Berseem, because of its widespread popularity and heavy growth, is an excellent crop for green-manuring purposes. The use of berseem is advantageous because early cuttings can be harvested as a forage during the winter months. Therefore, costs of seeding and irrigation will be covered by the cash return from the crop. Utilization of berseem as a green manure would require the ploughing under of the final growth in late spring rather than cutting it as a fodder. Economic and other conditions at the time will determine whether or not advantages may be expected from this. Some of the necessary considerations are given in the following example:

Assume that the last cutting of berseem provides one long ton per acre of dry matter (120 mds. of fresh berseem) with a total N content of 3%. This would provide about 67 lbs. of N to the soil. Assume also that the cash value of this single cutting of berseem is about Rs. 32 per acre. To provide 67 lbs. of N, about 4 one-md. bags of $(NH_4)_2SO_4$ would be needed. At Rs. 8 per bag, the total cost of the fertilizer also would be Rs. 32. In addition, the variable cost of transportation of the fertilizer to the field must be determined and included in the calculation.

This comparison will indicate that the cost for N from either source will be about the same. However, any shift in the value of either material will change the comparative costs. If chemical manures are unobtainable to a village or area, it is pointless to draw any comparison in the first place. Furthermore, in those locations where fodders are very scarce and the value of the crop as an animal food very high, berseem as a green manure cannot be expected to compete with chemical manures as a source of N.

A major problem associated with ploughing in berseem is the fact that the last growth occurs at a time when the supply of animal forage is low. Therefore, the value of berseem as a fodder becomes disproportionately high. If it were found possible to dry earlier cuttings of berseem for use as hay during this period of scarcity, it would appear more practical to return the last growth as a green manure. More research is needed to determine the practicability and techniques for curing and storage of berseem as a hay with the objective of increasing the amount of the crop that can be used for green manuring. An alternative is to leave a part of one cutting in the field, that obtained during the lush-growth period, for incorporation into the soil later in the season.

Principal advantages in the use of summer green manure crops will occur on land which would normally be in fallow during the monsoon season and in areas where the rains alone will support the manure crop. The cost of the N obtained will include the expenditure for seed and land preparation. The water, coming as rainfall, will aid in increasing the level of fertility in the soil at no additional cost to the farmer. Green-manuring during the summer months should be encouraged provided water is available for this purpose. This is impractical, of course, where water from limited monsoon rains must be conserved in the soil for seeding *rabi* crops.

Drouth-resistant guara (*Cyamopsis tetragonoloba*) has been shown to be a very effective green manure for the *kharif* season. Nitrogen contents of up to 3% of the dry plant weight may be expected. Erdman (1948) found that, as is true for other legumes, specific strains of root nodule organisms are necessary for a maximum in both growth and N content of guara.

REVIEW QUESTIONS

1. Explain why N deficiency results in stunted, yellow plants. Why do symptoms of the deficiency appear first in the older leaves ?
2. Why does NO_3-N appear to be preferable over NH_4-N for absorption by plants ? Under what circumstances is the NH_4 form preferred ?
3. Explain the difference between symbiotic and nonsymbiotic N fixation. Which microorganisms are responsible for these processes ?
4. Explain the difference between nitrification and nitrogen fixation.
5. Can one always expect an increase in the level of soil N as the result of growing legume crops ? Explain.
6. What are the processes responsible for the loss of soil N, and how may each source of loss be guarded against ?
7. In an experiment on wheat grown under natural rainfall (*barani*) conditions, broadcast ammonium sulfate resulted in poorer yields than ammonium nitrate. After application, a rain fell which moistened the soil to a 6-inch depth. A dry period followed and the surface few inches of soil became dry. How do you explain the better results from the nitrate fertilizer ?
8. Why might an organic source of N prove superior to an inorganic source in an area of high rainfall ? Why might the reverse be true under conditions of adequate, though not excessive, rainfall where the total amount of N added in either form is the same ?
9. Explain why a greater yield increase for each pound of added N is obtained when 20 lbs. of N as compared to when 40 lbs. are applied to the field.
10. Describe the effects on plants of an overabundance of N and explain why this occurs.

SELECTED REFERENCES

1. Allison, F. E. (1955) The enigma of soil-nitrogen balance sheets. Advances in Agron. 7 : 213-250.

2. Allison, F. E. (1957) Nitrogen and soil fertility. From *Soil,* The Yearbook of Agriculture for 1957, pp. 85-94, Supt. of Documents, Washington 25, D.C.

3. Harmsen, G. W. and Van Schreven, D. A. (1955) Mineralization of organic nitrogen in soils. Advances in Agron. VII : 299-398.

4. Tisdale, S. L. and Nelson, W. L. (1956) *Soil Fertility and Fertilizers.* The Macmillan Company, New York.

CHAPTER NINE

SOIL PHOSPHORUS AND POTASSIUM

FREQUENTLY, both phosphorus and potassium are deficient in soils and, as a consequence, have an important bearing on the abundance of the world's food supply. Under most conditions, deficiencies of either can be corrected easily. Since these two elements are relatively immobile in the soil, their acquisition by plant roots is governed by different principles than is that for nitrogen in the mobile NO_3 form. This affects primarily the manner in which plants feed upon them, and the efficiency with which they are supplied by fertilizers and manures to absorbing plant roots.

SOIL PHOSPHORUS AND ITS USE BY PLANTS

Phosphorus is present in all plant tissue. It is found primarily in the cell nucleus where it enters into a number of physiological reactions. It occurs as a component of phytin, lecithin, and other phospholipids. Since it is so important in cell-growth processes, it is found to concentrate in the younger plant parts. Thus, it will occur largely in the meristematic tissue where cell division and growth are in greatest evidence. During the flowering stage, a relatively large proportion of P migrates into the expanding buds, and later its transfer to the seeds and fruit takes place.

Phosphorus is extremely important as an accessory substance in the oxidation of carbohydrates within cell nuclei. Energy released during the oxidation is held temporarily in phosphate compounds before final utilization in the growth processes of the plant. Phosphorus favours root development and tillering. It hastens maturity and tends to overcome the effect of excess N which has been applied to P-deficient soils. It also improves the quality of feeds, especially forages for animals. Cattle fed on low-P forage gain weight very slowly and never attain maximum size. Soils tend to be very deficient in both P and Ca in the more humid regions of the subcontinent, and this reflects in the bone structure of both animals and humans who have no other food but that grown within these areas.

Symptoms of phosphorus deficiency in plants are not distinct. A low level of P in soils results in reduced growth, slow ripening of the fruit, and the production of small and inferior seeds. If the deficiency is severe, leaves may become very dark green, sometimes with a reddish or purple colouration.

THE OCCURRENCE OF PHOSPHORUS IN SOILS

The bulk of P in soils is present in mineral and organic combination. Limited amounts are found in soluble form. The rate at which P is released from the solid phase into the soil solution determines the capability of the soil to supply this element in an available form to plants.

Mineral Phosphorus

The principal source of P in soils is the mineral apatite, $Ca_5(PO_4)_3 \cdot (Cl,F)$[1]. This mineral may be of igneous origin, or it may be formed as the result of the reaction of P-bearing waters on beds of limestone materials. The limestone, being more soluble than apatite, may be removed from a mixture of these two minerals when subjected to extensive leaching.

Apatite and certain other calcium phosphate compounds are relatively stable in alkaline media but become unstable as the acidity is increased. Compounds of P with Fe and Al are stable in strongly acid media, on the other hand. Therefore, since one or more of these three ions, Ca, Fe, and Al, are present in relative abundance in all mineral soils, it is possible for P to be retained in the soil in stable mineral form throughout a wide range in soil reaction. This explains, at least partially, why drainage water from weathering mineral masses often is very low in soluble P. Similarly, even though the total-P content of a soil may be high, seldom does the quantity of P in water-soluble form exceed 1 ppm. in the soil solution. Thus, while the stability of P compounds minimizes the loss

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[1] Various formulas are given in the literature for apatite and include $Ca_4(PO_4)_3 \cdot CaF_2$ and $[Ca_3(PO_4)_2]_3 \cdot CaF_2$.

of this important plant nutrient by leaching, it also limits the availability of P to plants.

The weathering medium in arid-region soils generally is characterized by an abundance of soluble ions in the soil solution, particularly Ca, and a pH above neutrality. Such a system is not conducive to the weathering of soil-forming minerals including apatite. It is not surprising, therefore, to find that the major part of mineral phosphate in arid-region soils is in slightly altered or unaltered apatite particles. Probably the most rapid release of P from mineral form occurs when CO_2-liberating microorganisms and plant roots reduce the pH of the soil temporarily. The rate of release will vary depending upon the size of the apatite particles present. Thus, if the minerals are finely divided and present a large total surface area, solution will take place at a relatively rapid rate. That P released during weathering but not taken up by plants may be expected eventually to revert to some stable, less soluble calcium phosphate compound.

Subjection of apatite mineral particles to weathering under high-leaching regimes results in their relatively rapid decomposition. Where large quantities of water move through weathering mineral matter, base-forming cations are lost rapidly. Iron and aluminium are released during weathering also, but they tend to form relatively insoluble compounds which prevents their loss in quantity from the weathering mass. Among the insoluble compounds formed by Fe and Al are those resulting from combination with P as it is released from weathering apatite. In general, then, with more extensive weathering of soil-forming materials, higher proportions of mineral P occur as Fe and Al compounds.

Precipitation of P by Fe and Al may be expected to occur extensively in the surface soil. However, Fe and Al also may form *complexes*[1] with substances derived from decomposing organic residues, and, in this form, migrate downward with the percolating water to accumulate in the subsoil. Being released eventually from the complexes, the Fe and Al then may

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[1] When complexed, the Fe and Al remain in soluble, nonionic form. They are free to move with the soil solution but, being nonionic, do not enter into chemical reaction with P.

intercept P which also is leached from the surface soil. Because of this, acid soils of the humid regions often are observed to increase in total P with increasing depth in the profile.

Soluble Phosphorus

Since P does not move to any extent within the solution phase of a soil, the availability of this element to plants often appears to depend upon the ability of plant roots to permeate the soil mass and contact more completely the P-bearing surfaces of soil minerals. However, Dean and Rubins (1945) have shown that plants absorb P equally well when their roots are not permitted to contact soil particles directly. It would appear, therefore, that the rate of release of P in soluble form determines its availability to plants.

The phosphate ion (PO_4^{-3}) has a strong affinity for H ions. Gradual acidification of an alkaline phosphate solution results in a change in form of the phosphate ion in accordance with the following series:

$$PO_4^{-3} \rightarrow HPO_4^{-2} \rightarrow H_2PO_4^{-} \rightarrow H_3PO_4$$

As shown in Fig. 9-1, only two of these forms, HPO_4^{-2} and $H_2PO_4^{-}$, occur in appreciable amounts within the normal range in pH of soils. At a pH of about 6·7, the P in solution is equally divided between these two forms. In acid soils, $H_2PO_4^{-}$ predominates; whereas, in alkaline soils, the HPO_4^{-2} ion is most prevalent in the solution phase.

The form of the phosphate ion and the type of cation with which it is associated will determine the level of soluble P in the soil. Where Na and K are the dominant cations in the solution phase, significant precipitation of P may not be expected to occur. Where the pH is such that a large part of the P present is as HPO_4^{-}, the Ca-ion concentration need not be very high to cause the formation of a precipitate of slightly soluble dicalcium phosphate, $CaHPO_4$. Monocalcium phosphate, $Ca(H_2PO_4)_2$, on the other hand, is a relatively soluble compound. However, where the pH of the soil is such that the most prevalent phosphate ion is $H_2PO_4^{-}$, its solubility usually will be limited as a result of precipitation with Fe and Al.

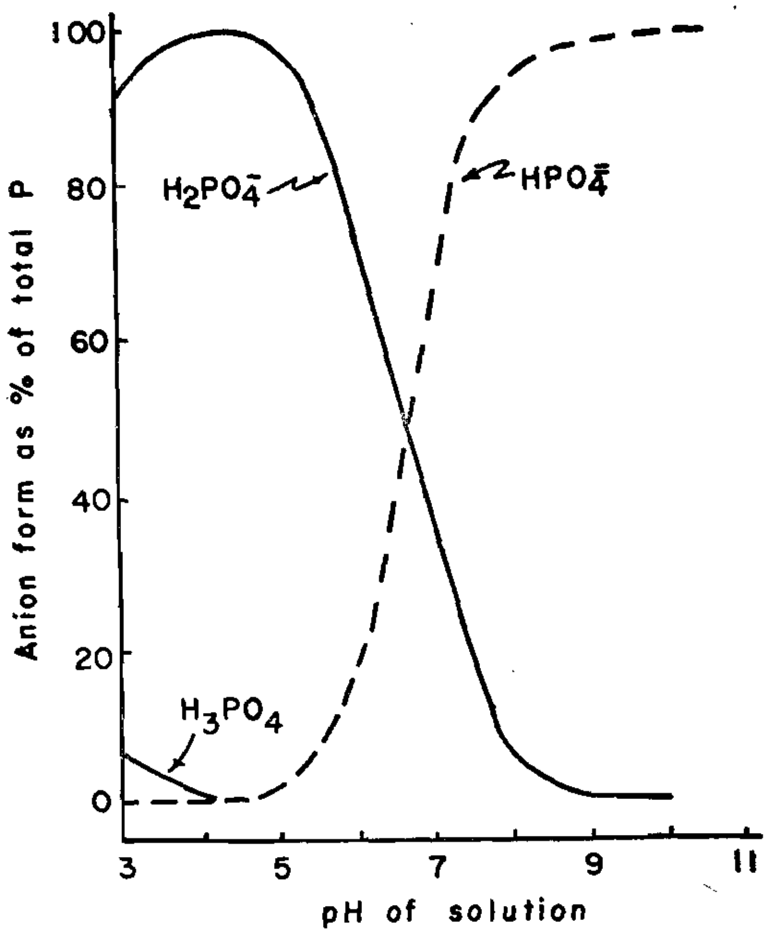

FIG. 9-1. The relationship between pH of the solution and the proportions of the various forms of the phosphate ions and of phosphoric acid present (after Buehrer, 1932).

Organic Phosphorus

Considering the 'big three' plant nutrients, phosphorus is next in abundance to N as a component of organic matter. It has been noted experimentally that the addition of organic matter to soils usually increases the level of soluble P. The increase has been found to be (1) in direct relation to the amount of P contained in the added organic matter, but (2) inversely related to the Ca or Fe and Al present in the soil solution.

Depending largely upon the climatic region, the amount of soil P occurring in organic form increases approximately in a constant ratio to the quantity of organic matter and organic nitrogen in the soil. In arid-region soils, where a large proportion of the soil P occurs in unweathered mineral form, very little will be found in organic combination, *i.e.*, usually less than 10% of the total. In humid regions, organic P may constitute as much as 50% or more of the total P. Because of the high degree of leaching in humid-region soils, however, the total combined inorganic and organic P will be quite low. For soils in the central part of the United States where leaching is only moderate, Pearson and Simonson (1939) observed that organic P constituted from 35 to 73% of the total P in the surface layer of the grassland soils they examined. They noted a very close relationship between total organic C and organic P in these soils.

The organic forms of P fall into 3 major groups: (1) phytin and its derivatives, (2) phospholipids such as lecithin, and (3) nucleic acid and nucleoproteins. According to Bower (1945) and Pierre (1948), a preponderance of the organic P occurs in the phytin and nucleic acid groups. Until released through mineralization, the P retained in organic combination is conserved against precipitating reactions which reduce its availability to plants. The ability of organic anions to complex Fe and Al will have some effect on the maintenance of P in available form when it is released through mineralization of organic matter in acid soils.

PHOSPHATE FERTILIZERS

Because soils are so frequently incapable of supplying adequate available phosphorus from indigenous phosphate minerals and organic matter, supplementation of the P supply through use of commercially prepared fertilizers or organic materials is a common practice. The limited supplies of organic wastes in the subcontinent places them in a subordinate position as a source of P, however.

All of the principal fertilizer compounds of P are prepared from the mineral apatite. Except where the preparation of fertilizer consists simply of grinding this phosphate rock, treatments usually are designed to increase the solubility of the Ca and P present either by converting them to more acidic compounds, or by removal of the fluoride ion. Fluoride may be dispelled by heating apatite to a very high temperature. It is lost also when apatite is acidified, the fluoride forming readily volatile HF acid. At the same time, this latter treatment converts the initial form of calcium phosphate to more readily soluble acid salts. The acidification reaction, where treatment of apatite is with H_2SO_4, is as follows:

$$\underset{\text{Insoluble}}{Ca_3(PO_4)_2{\cdot}CaF_2}+3H_2SO_4 \rightarrow \underset{\text{Soluble}}{Ca(H_2PO_4)_2}+3CaSO_4+2HF\rightarrow \qquad (9\text{-}1)$$

The soluble compound formed by this reaction is acidic monocalcium phosphate.

The Content and Availability of Phosphorus in Fertilizers

The amount of a fertilizer added to a soil should raise the level of available P until it is adequate for the crop being treated. When calculating the quantity of fertilizer to add, one must know the amount of P it actually contains and the extent to which it will become available once added to the soil. The phosphate content is, by custom, expressed relatively as percent P_2O_5 present rather than as elemental P[1]. However, application rates in the field normally are expressed on the basis of lbs. P_2O_5 needed per acre, and little confusion results from this method of expression, therefore.

Apatite and similar P compounds are essentially insoluble in water but will dissolve slowly in acid soils. The reaction

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[1] The P_2O_5 molecule is composed of 43% P and 57% O. Thus, the P_2O_5 content, expressed as a percent, when multiplied by 0·43 will give the actual percent P in the material.

bringing about solution is quite similar to that shown in equation 9-1 above. Some indication as to how rapidly P will be released from these insoluble forms once applied to acid soils can be obtained by measuring their solubility in ammonium citrate or citric acid solutions. Not only the chemical composition of the constituent phosphorus compounds but also the size of particles in which they occur determines their solubility rates. This is true both when the materials are mixed with the soil or when treated with citrate solutions. Because of this, the amount of P soluble in citrate solution is accepted rather widely as expressing the *available-*P_2O_5 content of phosphate fertilizer materials. Such measurements are made under uniform or standardized conditions in the laboratory (see Association of Official Agricultural Chemists, 1945, pp. 20-25, for prescribed methods).

More highly acidic calcium phosphate salts, such as $Ca(H_2PO_4)_2$ and $CaHPO_4$, are more soluble in the citrate solutions than is apatite. Correspondingly, they give up P to the soil solution more readily than does rock phosphate once they are mixed with the soil. Thus, their relative availability can be judged with reasonable confidence also by measuring their solutility in citrate solutions. True availability will, of course, be determined largely by subsequent reactions between the released phosphate ion and the soil.

A few of the more important types of phosphate fertilizers are discussed below. It should be noted that the order in which they are presented is also the order of their increasing solubility in water.

Raw Rock Phosphate

Rock phosphate, or apatite, contains from 24 to 45% total P_2O_5, the composition depending upon the source of the material. It must be finely ground to permit a suitably rapid rate of solution and availability. Of the total P_2O_5 present, only from 15 to about 60% may be expected to dissolve in ammonium citrate. Because of this it has the lowest solubility rating of all phosphate fertilizer materials. The solubility is closely related to the flourine content.

No benefit is derived from the use of rock phosphate on neutral to alkaline soils. This has been shown experimentally many times. Generally, the soils of arid to semiarid regions, which represent those with a neutral to alkaline reaction, may already contain apatite equivalent to several tons of P_2O_5 per acre from which plants roots may draw their needed P. A few hundred pounds of added apatite phosphate, which has about the same solubility as the indigenous soil minerals, will do little to affect the supply of available P in the soil. The use of ammonium citrate for evaluation of rock phosphate as a fertilizer for alkaline soils will overestimate considerably the availability of the P in this material, therefore.

Some benefits may be realized if rock phosphate is applied to acid soils. Even then the more soluble phosphates usually give superior results. Raw rock phosphate would be recommended only where it may be applied to acid soils in very liberal quantities and where other more soluble sources are not easily or cheaply obtainable. Since rock phosphate will dissolve slowly in an acid soil, extraction with ammonium citrate will estimate reasonably well the availability of the P it contains when it is destined for use on such soils.

Bone meal, produced by crushing animal bones, has about the same properties and value as rock phosphate.

Fused Tricalcium Phosphate (Fused Rock Phosphate)

Various processes are utilized for heating rock phosphate to volatilize the flourine it contains and thereby increase the solubility of the phosphate present. The P_2O_5 content of the fused material will vary depending upon the composition of the mineral from which it is prepared. Whereas the solubility is improved by this treatment, the finished product is still entirely unsuitable for use on alkaline or neutral soils. When finely ground it will produce reasonably good results in acid soils.

Dicalcium Phosphate

Dicalcium phosphate, $CaHPO_4$, is not prepared normally as a fertilizer salt. However, it does occur as an important component of several preparations which are becoming increasingly popular, namely, nitric phosphates and ammoniated superphosphates. Phosphate in the dicalcium form is sparingly soluble in water, but it is released fairly rapidly through solution if the fertilizer particles are of small size and if the medium in which it is incorporated is acidic. Dicalcium phosphate is essentially completely citrate soluble.

Superphosphate

Superphosphate is used more extensively than any other phosphate fertilizer. It is prepared by treating ground rock phosphate with sulfuric acid and aerating to remove hydrofluoric acid which has been formed during the acidification process. Superphosphate contains from 18-20% available P_2O_5. About one-half of the bulk of the fertilizer consists of gypsum formed by the reaction between H_2SO_4 and the Ca in the apatite. The amount of acid added in the manufacturing process is slightly less than that needed to react completely with the apatite mineral. The basic phosphate-carrying compound of superphosphate is monocalcium phosphate, $Ca(H_2PO_4)_2$, which is readily soluble in water.

Concentrated Superphosphate

In certain respects, an improvement over ordinary superphosphate is obtained by treating ground rock phosphate with H_3PO_4 instead of H_2SO_4. A fertilizer of higher P_2O_5 content, normally ranging between 40 and 50%, is obtained by means of this process. The material is referred to variously as *concentrated, double, treble,* or *triple* superphosphate. Since it is more concentrated, transportation costs per unit of P_2O_5 will be less than for ordinary superphosphate. Its behaviour in the soil is essentially identical to that of ordinary superphosphate because both contain monocalcium phosphate as the P source. The fact that it carries no sulfur may be a disadvantage in its use in areas where a deficiency of this element might occur also.

Phosphoric Acid

Commercial phosphoric acid is a concentrated, highly soluble form of P. It contains approximately 60% P_2O_5.

Even if it were currently available, transportation and handling problems would make it impractical for use under subcontinent conditions at the present time. Furthermore, its cost per pound of P_2O_5 is greater than for the other more common phosphate sources. If H_3PO_4 is used to treat phosphate rock, as in the manufacture of concentrated superphosphate, the total quantity of water-soluble P is doubled over that present in the original acid. It would appear wise to use it in fertilizer manufacture rather than to spend its solubilizing power on nonphosphatic soil minerals.

Organic Materials as a Source of Supplemental Phosphorus

The P content of most organic materials of vegetable origin is low, usually ranging between 0·15 and 0·35%. The P content will depend upon the type of plant from which the materials are obtained and upon the P-supplying power of the soil on which the plant has been grown. Young or immature plants have the highest total-P content, and may contain up to 2 to 3 times the total amount found at maturity. Further, the distribution of P within the plant is not uniform since this element tends to concentrate in meristematic tissue. Stems will be found generally to contain a smaller amount of P than will any of the other parts of a plant.

Relatively large amounts of organic residues must be returned to the soil to supply the P needed by a high-yielding crop. For instance, a legume may remove as much as 40 lbs. of P per acre in a single growing season. Ten T. of organic matter containing 0·20% P will be required to supply this. Since all of the P added would not be mineralized in one season, even greater quantities would be needed to provide a total of 40 lbs. in available form.

THE BEHAVIOUR OF FERTILIZER PHOSPHATE IN SOILS

The use of phosphatic fertilizers as a means of increasing the level of available P in the soil is confronted with a number of difficulties. As indicated in the preceding section, a large number of phosphate materials are available for use. Not only are there wide inherent differences among these various types of materials, but, also, reactions between them and the soils to which they are applied are highly variable.

Perhaps of greatest single significance in the application of phosphate materials to the soil is the fact that reactions between the phosphate ion and Ca, Fe, or Al almost always result in a reduction in the solubility of the applied P. Precipitation reactions may be so extensive that, in certain instances, very little if any benefit is derived from the applied material. A reduction in the level of soluble P, or P *fixation*, has its most profound effect in acid soils high in Fe and Al in soluble form. Often, either excessive rates of phosphate fertilizer materials must be added to these soils, or else the fertilizers must be placed in the soil in narrow bands to limit contact with soluble and exchangeable Fe and Al. This process of fertilizer application, known as *band placement*, permits the beneficial use of relatively small quantities of phosphate materials on soils of high fixing capacity.

The behaviour of P supplied in a fertilizer material depends upon its rate of release in the soil and the reactions it may undergo after becoming a part of the soil solution. The rate of release is governed largely by the solubility of the added material in the medium represented by the soil and its solution. Whereas an acid salt such as monocalcium phosphate is highly soluble under virtually all conditions encountered in the soil, the solubility of the more alkaline calcium phosphates depends largely upon the pH and the soluble-Ca content of the soil with which they are mixed.

Reactions of Monocalcium Phosphate with Soils

Most superphosphate and treble superphosphate materials, which contain P as readily soluble monocalcium phosphate, are prepared in granular form. This is desirable for it facilitates spreading in the field, either by machine or by hand. Further, as will be seen shortly, the P contained in relatively large granules is converted less rapidly to insoluble precipitates than is P in particles of the same material occurring in a fine state of subdivision.

Upon solution, monocalcium phosphate may be envisaged as yielding a mixture of equal quantities of dicalcium phosphate and phosphoric acid, *i.e.*:

$$\underset{\substack{\text{Monocalcium}\\ \text{phosphate}}}{\mathrm{Ca(H_2PO_4)_2}} \rightarrow \underset{\substack{\text{Dicalcium}\\ \text{phosphate}}}{\mathrm{CaHPO_4}} + \underset{\substack{\text{Phosphoric}\\ \text{acid}}}{\mathrm{H_3PO_4}} \qquad (9\text{-}2)$$

Lindsay and Stephanson (1959) have shown that a saturated solution of monocalcium phosphate separates into these two products in a manner very nearly as depicted above. The reaction takes place spontaneously in moist soil when a granule of the hygroscopic monocalcium phosphate absorbs sufficient moisture from the soil atmosphere to bring about its own solution. During the initial stages of the solution process, the concentration of Ca and P is so high that the solubility of slightly soluble dicalcium phosphate is exceeded, and part of the Ca and P present precipitate from solution in this form at the site of the original granule. The soluble Ca and P remaining form a solution of low pH consisting largely of phosphoric acid but containing some monocalcium phosphate.

Because of the increase in the moisture content at the site of the original granule, the residual monocalcium phosphate-phosphoric acid solution migrates outward into the neighbouring, drier soil. Movement of the P will continue until it encounters ions which bring about its precipitation. If the soil is initially high in soluble and exchangeable Ca, the migrating acid solution is neutralized gradually and the P converted slowly to the dicalcium phosphate form. The precipitate is probably distributed over the surfaces of other soil particles. Since precipitates of this type dissolve on contact with roots, they can provide adequate available P to plants. The availability decreases gradually over long periods of time, however, the result of growth of larger dicalcium phosphate crystals at the expense of the smaller ones, and probably the consequence of slow conversion to more stable, less soluble calcium phosphate compounds[1].

It is likely that the strongly acid solution flowing from the site of a monocalcium phosphate granule also will dissolve Fe and Al from soil minerals. Thus, even in high-Ca soils, precipitation of Fe and Al phosphates may be expected. However, when the soil at the site of these precipitates returns eventually to its normal pH, the P bound in this manner probably would be released and converted to a calcium phosphate compound.

The behaviour of a granule of monocalcium phosphate in an acid soil will be essentially identical to that in an alkaline soil high in Ca. However, the solution phase which separates from the precipitated dicalcium phosphate will react largely with solubilized Fe and Al in the surrounding soil. The availability of the P in the precipitates will, of course, depend primarily on the fineness of the phosphate crystals formed. It is probable, however, that a large part of the available P will come from the dicalcium phosphate residue remaining at the site of the original granule. This residue will dissolve as it is contacted by fresh increments of acid soil solution low in both Ca and P.

All of the P added to an acid soil as a calcium phosphate fertilizer will go into solution and, if not absorbed by plant roots or other soil-inhabiting organism, will be subject to entering a precipitation reaction with Fe and Al. For soils with a high P-fixing capacity, the P applied during one growing season is not expected to benefit plants the following year. This is not true normally for P applied to alkaline soils, however. In this latter instance, P applied in rather large dosages is observed to affect succeeding crops for a number of years. Usually, phosphorus shortages will occur again only after a

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[1] There is evidence that octacalcium phosphate, $Ca_4H(PO_4)_3 \cdot 3H_2O$, which is less soluble than dicalcium phosphate, will form in alkaline soils as the result of hydrolysis of dicalcium phosphate. Soluble phosphate is precipitated readily at the surface of lime particles in calcareous soils, probably also as dicalcium phosphate. It has been proposed that slow conversion of these precipitates to the more stable compounds of carbonato-apatite, $[Ca_3(PO_4)_2]_3 \cdot CaCO_3$, and hydroxy-apatite, $[Ca_3(PO_4)_2]_3 \cdot Ca(OH)_2$, occurs and results in a gradual reduction in the availability of the P (Buehrer, 1932).

large part of the nutrient applied has been removed from the soil by crops.

Reactions of Slightly Soluble Phosphates in Soils

Fertilizer materials of limited solubility in water are those which contain a high proportion of Ca in relation to the P present. Materials such as fused tricalcium phosphate and rock phosphate fit into this category. When mixed with high-Ca soils, these compounds are even less soluble than when mixed with pure water. For this reason, they are quite stable in most neutral to alkaline soils, and little benefit is derived from them when they are added to such soils as fertilizer.

Whereas the solubility of dicalcium phosphate in water is low, it is a much better source of P than are rock phosphate and fused tricalcium phosphate, particularly when applied to alkaline soils. Nonetheless, phosphate sources containing dicalcium phosphate should be in a very fine state of subdivision when added to soils, otherwise the total surface area of the particles will be inadequate to release P to the soil solution at a rate required by growing plants. According to Terman *et al.* (1958), phosphate fertilizers containing monocalcium phosphate release P in available form more readily in alkaline, high-Ca soils than do those containing principally dicalcium phosphate. The lower solubility of dicalcium phosphate can be compensated for, of course, by increasing the rate at which it is applied to the soil (Hausenbuiller and Weaver, 1954).

Calcium phosphates of low water solubility have their greatest usefulness in acid soils where they will dissolve more rapidly. Even so, very slowly soluble raw rock phosphate may not be able to supply adequate soluble P to plants if the soil is only moderately acid or if the particles of fertilizer are of relatively large size. The smaller the particles and the more widespread their distribution in the soil the greater is the opportunity for contact with roots. Direct transfer from particle to root provides the only real assurance against fixation of released P in acid, mineral soils.

Terman *et al.* (1958) report trials on acid soils in which relatively insoluble fused tricalcium phosphate appeared to supply P in available form as efficiently as did monocalcium phosphate. In the same trials, dicalcium phosphate sources were generally better than monocalcium phosphate. Much of the P from readily soluble monocalcium phosphate apparently was lost due to rapid fixation by Fe and Al. The dicalcium phosphate, being the more soluble, was superior to fused tricalcium phosphate in the acid soil.

Crop Type and Phosphate Fertilizer

Crop plants differ markedly with respect to their demand for plant nutrients. Influencing specific plant requirements are the total amount and rate of growth and the level of nutrients which must be maintained in the plant to sustain normal functions of growth and maturation. Short-lived, fast-growing plants must be supplied with nutrients in relative abundance even though it may be necessary for short periods only. Treatment of such crops with the more highly water soluble materials would appear to be desirable, therefore. Vegetable crops, particularly those grown during the winter months when the soil is cool, will respond much better to phosphates of high water solubility.

Perennial plants and the slow-growing annuals can be supplied with P adequately from materials which release P more gradually. This is especially true if the soil has a high capacity for fixing P and would reduce the level of available phosphate added to the soil long before plant growth was complete. Seatz *et al.* (1954), for instance, report results obtained on acid soils with fused tricalcium phosphate of low water solubility. They claim this material to be a relatively good source of P for long-season crops such as lucerne and pastures but significantly poorer when used with vegetables and cotton.

PHOSPHATE RETENTION IN SOILS

The phosphate ion is relatively immobile in soils. Normally, the addition of a soil material to a phosphate solution causes a rapid decrease in the concentration of the phosphate ion. Such reactions are indicative of the capacity of a soil to fix P and may relate, of course, to the availability of soil P to plants.

Retention of P by soils has been attributed to various processes. Several past studies have indicated the possibility of pronounced adsorption of phosphate ions by minerals in the soil through ion-displacement reactions at the particle surfaces (Stout, 1939 ; Murphy, 1939). This was thought to be true particularly for soils high in kaolinite. Such studies led to the conclusion that soils had characteristic *anion-exchange capacities* as well as cation-exchange capacities. A number of methods for measuring the anion-exchange capacity have been proposed. Among them are those of Dean and Rubins (1947) and Mehlich (1948). However, Low and Black (1948) more recently have concluded that retention of P by a material such as kaolinite is not due to surface adsorption as such, but, instead, results from a reaction between the P and exchangeable Al to form an insoluble precipitate. They suggest that exchangeable and soluble Al in a suspension of kaolinite occur in a state of near equilibrium with the solid-phase Al of the kaolinite crystal ; that is to say, the concentration of soluble and exchangeable Al is dependent upon the extent to which the Al in the crystal will dissolve in the solution phase. Therefore, removal of solution-phase Al through precipitation reactions would result in further solution of the kaolinite crystal. In keeping with this, Kittrick and Jackson (1954) have shown that kaolinite may be decomposed completely in a phosphate solution as the result of precipitation of an aluminium phosphate compound of greater stability (lower solubility) than kaolinite. Hemwall (1957), in an excellent review article on phosphate reactions in soils, points out that such findings essentially rule out anion-exchange as a possible cause of fixation of significant quantities of soluble P, therefore.

The importance of Fe and Al in P fixation has been shown also by Bass and Sieling (1950). They found that the amount of these two ions extracted from acid soils by citric acid[1] served as a good index of the ability of a soil to fix P against subsequent extraction. Soils high in hematite, Fe_2O_3, and gibbsite, $Al(OH)_3$, release large quantities of Fe and Al during citric acid extraction and also have a high P-fixing capacity.

The number of compounds which P may form with both Fe and Al in acid soils is considered to be large (Haseman *et al.*, 1950). It appears likely that one of the most prevalent is variscite, $Al(OH)_2H_2PO_4$, as proposed by Swenson *et al.* (1949). Cole and Jackson (1950) observed the solubility of this compound to attain a minimum at pH of 4·0.

Olsen and Watanabe (1957, 1960) and Cole and Olsen (1959) compare the retention of soluble P by soils to other reactions which result from adsorption to soil particle surfaces. In this instance, an equilibrium is visualized as existing between the adsorbed phosphorus and that in solution so that the greater the extent of saturation of adsorbing surfaces with P the higher its concentration in the solution phase. This concept does not eliminate precipitation as the responsible process for retention at the particle surfaces. However, should this type of reaction be involved, the precipitates which form are suspected to occur as extremely thin films (*i.e.*, monomolecular layers) over these surfaces. As a consequence, the precipitates in themselves present extensive surface so that a relatively rapid interchange of P between the soluble and precipitated phases can take place.

The validity of the above concept is strengthened by the fact that the level of soluble P in soils with high total surface area on which adsorption or precipitation may take place is not changed as radically by the addition of a given amount of phosphorus as it is in soils of coarser texture. This is illustrated in Fig. 9-2 by data derived from four alkaline, calcareous soils of widely different texture. Here, the level of soluble P is related to the quantity of this element added to these soils. It may be noted that more P must be supplied

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[1] As has been indicated previously for certain other organic anions, the citrate ion is capable of complexing Fe and Al, and thereby encourages their release from slightly soluble Fe and Al compounds. Thus, the effect of adding the citrate ion to an acid soil is not unlike that of P.

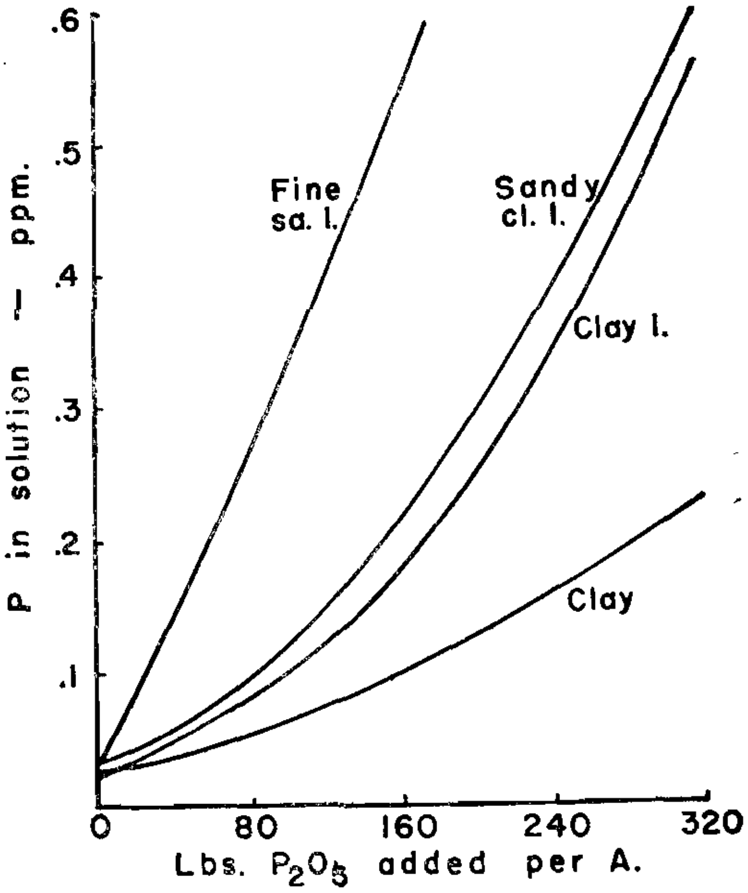

FIG. 9-2. The relationship between the amount of phosphate fertilizer applied to soils of different texture and the resultant level of soluble P (after Cole and Olsen, 1959, and Olsen and Watanabe, 1960).

to cause the same increase in level of soluble P as the texture becomes finer and, therefore, as the total adsorbing surface increases. For example, whereas about 75 lbs. of P_2O_5 applied to the fine sandy loam increases the level to 0·2 ppm., almost 300 lbs. are required to cause the same increase in the clay soil.

The importance of the level of soluble P to plant growth is exemplified in Fig. 9-3. In this figure, the quantity of soluble P measured in the four soils considered above is plotted

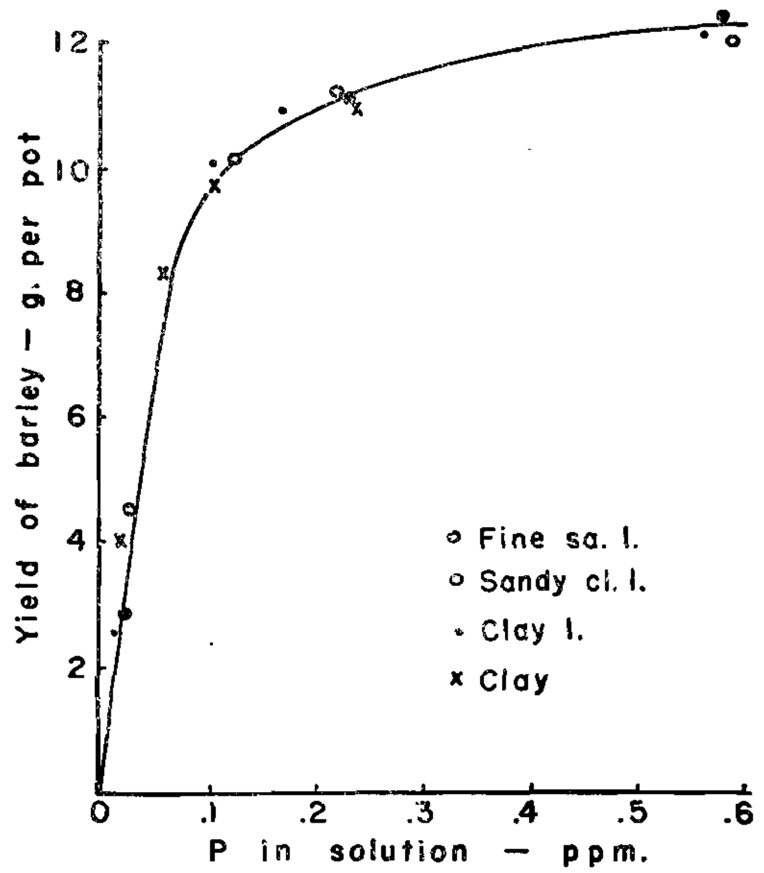

FIG. 9-3. The yield of pot-grown barley as a function of the level of soluble P established in four soils of different texture through the addition of phosphate fertilizer salts (after Olsen and Watanabe, 1960).

against the yield of barley grown on potted samples of these soils. It is apparent that yields are increased as the soluble P is increased through the addition of a fertilizer salt. By relating the results depicted in Figs. 9-2 and 9-3, however, it becomes immediately obvious that more P must be applied to fine-textured soils than to coarse-textured soils of low soluble-P content if corresponding yield increases in both are to be obtained. Once the retentive capacity for P is satisfied to the point where the soluble P in either type soil is adequate for maximum plant growth, on the other hand, the supply of available P will be maintained for a longer period by soils of finer texture.

While the trials discussed above were carried out only on alkaline, calcareous soils, there is little reason why similar results might not be expected from acid soils of humid regions. A major difference may be anticipated, however, in the qnantity of P which must be added to increase the level in soluble form by a given amount. Olsen and Watanabe (1957) have concluded from their work that, on the average, perhaps twice as much P must be added to acid soils per unit of adsorbing surface to yield an increase in soluble P which corresponds to that obtained in alkaline soils presenting the same total surface area per unit mass of soil. This, of course, conforms with the general recognition of the exceptional capacity of many acid soils to fix large amounts of P in a form which is unavailable to plants. As postulated for alkaline soils, the reaction in acid soils may result also from the formation of thin precipitates over particle surfaces. However, the chemical composition of these precipitates probably will be such that their solubility is low even though they present a large total surface area which is in contact with the soil solution.

SOIL POTASSIUM AND ITS USE BY PLANTS

Nearly all of the potassium absorbed by plants remains in a soluble, ionic form. It is not found in cell nuclei but occurs in the cytoplasm of all plant parts. The exact function of potassium in plants is not known, but it appears to be important in (1) the formation of protein and carbohydrates, (2) the synthesis of chlorophyll, (3) the transformation of NO_3 to NH_2, (4) the absorption of anions such as NO_3, H_2PO_4, and HPO_4, and (5) the storage of carbohydrates. Because of this last item, potassium has an important bearing upon the quality of plants and plant products. The storage and cooking quality of potatoes is impaired if grown on K-deficient soil. Wheat and other small grains produced under low potash fertility frequently have weak stems and, therefore, are subject to excessive lodging.

Soils low in available potassium produce plants with dull green leaves. If the deficiency becomes severe, edges of the leaves turn yellow and finally die, producing a characteristic deficiency symptom known as 'tip burn'.

Reitemeier (1957) lists the nature of K-deficiency symptoms for a number of crop plants. Generally, they follow the pattern listed above and occur first in the older leaves of the plant. A variation in observed symptoms is that their first incidence in some plants may be as small yellowish spots or mottlings along the edges of the leaves. Additional descriptions of deficiency symptoms for potassium, and the various other plant nutrients as well, are found in the publications edited by Hambidge (1941) and Kitchen (1948).

FORMS OF POTASSIUM IN THE SOIL

Potassium occurs in the soil in various forms and combinations. By far the majority of the K is contained in such minerals as biotite and muscovite micas and in feldspars. In unweathered soil materials, K may constitute as much as 2 to 3% of the total weight of the soil, and it will occur mostly in these mineral forms.

A second very important form of K is as an adsorbed ion on the exchange complex. As much as 10% of the exchangeable ions may be potash in normal, slightly to moderately weathered soils, although it is usually present in lesser amounts. Yet, there are some arid-region soils which have 50% or more of their exchange positions filled with this ion.

Potassium is present in the soil in soluble form also. The total amount of soluble K is usually small, and the exchangeable form will exceed it in total quantity by 20 or more times.

As was noted in Chapter 4, appreciable quantities of K may occur in 2 : 1 minerals where it is held in the oxygen layer holes in the silica sheets. This is particularly true for the mineral illite, and for vermiculite in soils which have been treated liberally with potassic fertilizers. While K held in these positions may be displaced by other ions, the reaction is quite slow. Therefore, for practical reasons, the K held in the oxygen layer holes is considered to be *nonexchangeable.* When K ions are added to the soil and migrate into the interlayer spaces to be retained in nonexchangeable form, they are said to be *fixed.* Even so, K thus held is available to plants although it may not be released to roots as rapidly as is exchangeable K. Consequently, fixation is not considered to be a particular hindrance to the K nutrition of plants.

The Potassium Equilibrium in Soils

The various forms of K in the soil occur in equilibrium with one another. Peech and Bradfield (1943) describe the equilibrium state through use of the equation:

$$\text{Nonexchangeable K} \rightleftharpoons \text{Exchangeable K} \rightleftharpoons \text{Soluble K} \quad (9\text{-}3)$$

The equilibrium concept is based on the frequent observation that the level of exchangeable K in the soil may remain relatively constant even though appreciable quantities of this element are removed by growing plants. Thus, it is not uncommon to find essentially the same amount of exchangeable K at the end of a growing season as was measured in the soil at the beginning of the season. This phenomenon is of most common occurrence in soils containing appreciable quantities of illite or other readily weathered minerals high in nonexchangeable K. Where the mineral complex of the soil is made up largely of kaolinite and the sesquioxides, on the other hand, the K equilibrium will centre largely around the exchangeable and soluble forms of potash only.

The addition of potassium salts to some illitic and vermiculitic soils will cause the reaction described by equation 9-3 to shift to the left. There is a limit to the change in this direction, however, for any soil will have a definite capacity for fixing K as is dictated by the number of oxygen layer holes available within mineral lattices. Normally, K fertilizer materials seldom are applied in quantities sufficient to saturate sites of fixation in the soil minerals.

The release of K from nonexchangeable form is of great importance to the maintenance of the levels of available K in the soil. A large amount of research has been conducted to evaluate this phenomenon. The results of one such study are summarized in Table 9-1 which shows the average amount of K released to plants grown in the greenhouse on 10 different soils. The harvested crops removed 119 lbs. of K per 2 million lbs. of soil, although the reduction in exchangeable K during the course of the experiment was only 43 lbs. Therefore, an average of 76 lbs. of the K absorbed by the plants is assumed to have come from nonexchangeable forms.

Table 9-1. The release to plants of nonexchangeable K by 10 soils. All values are in lbs. K per 2 million lbs. of soil (after Stewart and Volk, 1946).

Exchangeable K		Decrease in exchangeable K	K removed by crops	K obtained from nonexchangeable form
Initial	Final			
80	37	43	119	76

Plants differ in their ability to bring about the release of K from minerals or from positions of fixation. Evans and Attoe (1948) found that, when the supply of available K was low, oats encourage the release of nonexchangeable K to a greater extent than does clover. This is related to the ability of the oats to grow more thriftily on low-K soils than do many other crop plants. On the other hand, these authors point out that when the level of available K is high, release from nonexchangeable forms to clover is greater than is the release to oats. A greater depletion of exchangeable K results from its high absorption by clover. Therefore, maintenance of the K equilibrium in the soil cropped to clover requires a greater transfer of K from nonexchangeable to exchangeable or soluble form.

Fixation of K in the oxygen layer of the silica sheet represents a reversal of the weathering process whereby K is lost from mineral crystals. Volk (1934) found that soils treated for many years with K fertilizers contained clay high in muscovite, a K-rich mica mineral. Unfertilized soils from adjacent fields did not contain significant quantities of this mineral.

The extent to which fixation of K will take place is shown by the data of Evans and Attoe (1948). They found that almost 60% of a very heavy application of potash was transferred to nonexchangeable form. However, a high degree of fixation was invariably associated with a comparatively rapid release of the fixed K to growing plants. This suggests that migration of K into and out of the crystal lattice may take place with about the same relative ease.

The Potassium-Supplying Power of Soils

The ability of a soil to supply K to growing plants depends not only upon the total quantity of K-bearing minerals present but also upon the dominant particle size of these minerals. Clays and loams which contain a relatively large amount of easily weathered K minerals, particularly illite, may be expected to supply ample K for most crops. Sandy soils require the addition of potash fertilizers more often than do finer-textured soils, as a consequence. The large particles of the K-bearing minerals in the coarse-textured soils offer inadequate surface for the elements of weathering to bring about rapid release of the K they contain. Since the cation-exchange capacity of sandy soils is low, added K will not be adsorbed so extensively through exchange, and, therefore, is more subject to loss by leaching in areas of high rainfall. Frequent applications of small quantities of fertilizer K is advisable under these circumstances. In fine-textured soils, however, addition of K may be in such quantity as will supply the needs of plants for several years. Potassium that is not used during the season following application remains available for subsequent crops.

Losses of Potassium from the Soil

Available potassium is lost from the soil in various ways: (1) through leaching, (2) crop removal, and (3) by erosion. The prevention of leaching and erosional losses will correspond to those suggested for N.

Potassium frequently is absorbed by plants in amounts in excess of that needed to sustain satisfactory growth. Known as *luxury consumption*, this waste is of particular importance if an appreciable proportion of the K must be supplied in the form of fertilizer. It is possible to minimize the effect of luxury consumption if fertilizer applications are frequent and if the amounts supplied each time are not excessive. The return of K in crop residues and in manures is important in replacing K lost from the soil through over-exploitation by growing plants.

POTASSIUM FERTILIZER MATERIALS

All forms of K fertilizers are readily soluble in water. This means that the major sources will be about equal in their ability to supply available K to plants. The composition of potash fertilizers is expressed on the basis of percent content of K_2O. The most widely used fertilizer salts are as follows:

Muriate of potash, KCl, containing 47 to 61% K_2O.
Sulfate of potash, K_2SO_4, containing 47 to 52% K_2O.
Saltpeter, KNO_3, containing 43 to 45% K_2O.

The total amount of K_2O each fertilizer contains is dependent primarily upon the purity of the finished product.

The sulphate and nitrate forms of potash fertilizers may be preferable in those soils where either N or S is deficient also. However, the cost of the K-fertilizer material may be such that supplying these two nutrients in some other form is more economical.

POTASSIUM AND SODIUM RELATIONSHIPS IN SOILS

Recent research has disclosed that, in certain instances, the Na ion appears capable of assuming partially the role of K in the nutrition of some crop plants. For these particular plants, moderate K deficiency may be remedied by the addition of a Na salt. Complete lack of K cannot be compensated for by Na, however. In a limited number of trials, improved yields have been observed where Na was applied even in the presence of adequate K. Because of this, there is a tendency on the part of some to consider Na as an essential plant nutrient, at least for some plant types. Generally, however, most plants appear to thrive just as well without Na as when this ion is present.

Frequent papers on the role of Na in plant nutrition occur in the literature. Among them are those of Lehr (1951) and Truog *et al.* (1953).

REVIEW QUESTIONS

1. Describe the relationship between the form of the phosphate ion and pH of its solution. Why does the solubility of this ion differ at a pH of 8 depending upon whether it is accompanied by Ca or Na ?
2. Explain the basic differences in phosphate fixation under acid-soil and alkaline-soil conditions.
3. From which would you expect to find a higher level of soluble P, a soil at pH of 4 or the same soil which has been limed to a pH of 6·5 ? Why ?
4. Explain why placing P fertilizer in a band in a soil having a high phosphate-fixing capacity will increase its availability to plants. Why may liming the soil have the same effect ?
5. Why does raw rock phosphate produce poor results in neutral to alkaline soils ?
6. Describe the potassium equilibrium in soils.
7. What is the significance of the terms "fixed" and "nonexchangeable" potash ?
8. Why is the release of K from nonexchangeable form so important to the fertility of soils ?
9. Why are the NH_4 and K ions fixed by clay minerals in a similar manner (see p. 133) ?

SELECTED REFERENCES

1. Collings, Gilbreath H. (1955) *Commercial Fertilizers, Their Sources and Use.* McGraw-Hill Book Co., New York.

2. Dean, L. A. (1949) Fixation of Soil Phosphorus. Advances in Agron. 1 : 391-411.

3. Olsen, S. R. and Fried, M. (1957) Soil phosphorous and fertility. *From Soil,* The Yearbook of Agriculture for 1957, Supt. of Documents, Washington 25, D.C.

4. Olsen, S. R. (1953) Inorganic phosphorus in alkaline and calcareous soils. Amer. Soc. Agron. J. Monograph 4 : 89-122.

5. Reitemeier, R. F. (1951) Soil Potassium. Advances in Agron. 3 : 113-164.

6. Tisdale, S. L. and Nelson, W. L. (1956) *Soil Fertility and Fertilizers.* The Macmillan Co., New York.

CHAPTER TEN

SOIL CALCIUM, MAGNESIUM, SULFUR AND THE MINOR ELEMENTS

PLANT NUTRIENTS other than N, P, and K have received comparatively little attention in soil and plant research. Even so, investigations concerning these lesser elements have been very numerous. Many reasons may be listed for the secondary position given them. In the first place, they occur much less frequently in soils in amounts limiting to plant growth than do N, P, and K. For sulfur and the trace elements, at least, complexities in analytical procedures have discouraged their study also. However, recent advances in laboratory techniques have resulted in an increased interest in most of these plant nutrients.

CALCIUM

Calcium is unique in that it can occur in abundance, but not necessarily in excess, in calcareous soils; it can be present in harmful quantities in soils of high salt content; and essentially lacking in leached, acid soils. In most soils, the major part of the exchange positions is filled with Ca. This is true even under moderately acid conditions. Its preponderance on the exchange complex is derived from the fact that many soil minerals are rich in Ca, and, as an exchangeable ion, is held to soil particles more tightly than other common cations except H and Al. Most Ca minerals present in soil materials have a comparatively high weathering rate.

Calcium occurs largely in plants as a pectate, a constituent of the cell wall. Frequently it is found as insoluble precipitates such as calcium oxalate. Precipitation reactions with Ca are considered as inhibitory to the accumulation of certain toxic substances in plants. Only a small portion of the Ca in plant tissue is in a soluble form. It is virtually immobile in the plant itself and tends to accumulate in the older tissues. As a consequence, Ca-deficiency symptoms are peculiar to the younger plant parts, *i.e.*, buds and new leaves. Either the buds may fail to form, or the new leaves may have a characteristically irregular or uneven edge. Because of the limited transfer of Ca from old to new tissue within a plant, fresh supplies must be continuously available in the soil for absorption and transfer to the meristematic regions. Roots fail to develop properly in low-calcium soils, a fact which accounts for their frequent absence in subsoil layers of strongly acid, leached soils.

The availability of Ca to plants in lime-free soils is related to the proportion held on the exchange complex. Marshall (1948) has shown that the Ca-ion concentration of the soil solution of acid soils remains very low until the exchange complex becomes at least 70% saturated with this cation. This is particularly true for the montmorillonitic clay colloids, although Ca may not need to constitute so large a proportion of exchangeable ions where the exchange material has a high pH-dependent charge. Where the ion associated with Ca on the exchange complex is Na, even a higher percent saturation with Ca is necessary to maintain an adequate level for plant growth and good soil physical conditions.

As opposed to conditions in lime-free soils, the availability of Ca in calcareous soils is dependent upon the CO_2 content of the soil solution. Increasing the quantity of CO_2 in the solution reduces the pH and increases the solubility of $CaCO_3$ through its conversion to $Ca(HCO_3)_2$. The availability of Ca to plants in calcareous soils below pH 8·2 is considered generally always to be adequate. As the pH is increased above 8·2, the solubility of $CaCO_3$ is suppressed even to the extent that exchangeable Ca is precipitated. Above pH 9·0, Ca solubility may be so low that it will no longer be adequately available to plants.

The correction of Ca deficiency occurs almost invariably as a consequence to adjustment of the soil pH. This may be done either by liming an acid soil or by treatment of high Na soils with gypsum or other amendment. It cannot be said that major benefits from these operations are obtained through the addition of available Ca only, because a large part of the gains come from improved physical properties, changes in the availability of other plant nutrients with the change in pH, and the improvement in conditions of growth for beneficial soil microorganisms. That the addition of Ca ions alone may not always benefit supposedly Ca-deficient, acid soils is shown by Fried and Peech (1946). Whereas the simultaneous correction of soil acidity and low Ca supply with $CaCO_3$ brought about good recovery of deficient plants, similar responses were not obtained with gypsum. The latter salt did not increase the pH of the soil ; in fact, adding a salt such as $CaSO_4$ to an acid soil usually will reduce the pH. Upon adsorption of the added Ca, H and Al ions are displaced and result in the formation of H_2SO_4. In some instances, the further reduction in soil pH this causes may have detrimental consequences which far outweigh the benefits derived from increasing the available Ca supply.

An ample supply of available Ca is necessary for the proper functioning of soil organisms responsible for N fixation and organic matter decay. The author (Hausenbuiller, 1950) observed a marked increase in the decay rate of organic materials when lime was added to the soil even though the exchangeable Ca level was initially high. The exact role played by the Ca is not completely understood, but it does relate to the indirect influence it has through pH control and the availability of other plant nutrients. In addition, the requirement of the organisms for Ca is high.

The application of lime for correction of acidity is by far the most important use made of Ca-supplying materials. Occasionally gypsum is used for increasing the level of exchangeable Ca in high Na soils. Otherwise, Ca additions usually are incidental to the supplying of other fertilizer materials, particularly ordinary superphosphate. Major considerations in the use of lime are covered in Chapter Five.

Calcareous Soils

Calcareous soils, or those containing a measurable quantity of free lime, are common to a large part of arid mountainous regions of West Pakistan, the Indus and Ganges basins, and the Deccan Plateau. The most common occurrence of lime is as *kankar,* the lime concretions which are found at various depths throughout part or all of the soil profile. The origin of *kankar* appears to be from one of two sources, namely, precipitation at the surface of a water table, or the reprecipitation of lime dissolved in the surface soil and carried into the subsoil by rain or irrigation water. The *kankar* may occur as small nodules, often with a very irregular, knobby, or branched shape, or as subsoil deposits so extensive that they occur as continuous, compact, hardpan layers.

Formation of *kankar* layers at the surface of the water table probably results from a reaction of bicarbonate in the ground water with Ca carried down in the soil solution from above. Water rising by capillary action from the water table can lose CO_2 slowly which causes the conversion of the HCO_3 ion to the CO_3 ion and the precipitation of Ca, *i.e.*,

$$Ca(HCO_3)_2 \rightarrow H_2O + CO_2 + CaCO_3 \qquad (10\text{-}1)$$

Further, as water saturated with $CaCO_3$ rises from the water table and is utilized by plants, the $CaCO_3$ it contains will be precipitated within the soil. Similarly, water penetrating into the soil dissolves $CaCO_3$ in the surface layers and carries it downward to be precipitated in the subsoil by the same mechanism. The depth at which layers of precipitated lime occur in the soil will increase with increasing quantities of percolating water. Studies on the characteristics and occurrence of *kankar* in arid-region alluvium and in Deccan soils have been carried out by Singh and Lal (1946).

The presence of *kankar* in the soil should not be considered harmful. True, it is a source of precipitant for added P fertilizers, but, as discussed in Chapter Nine, phosphorus reverted in this manner retains a relatively high degree of availability.

In subcontinent soils of the drier regions, lime is virtually always present, frequently amounting to as much as 10 to 15% of the total weight of the soil. In locations where the soil surface is not changed by erosion or deposition of new materials, lime may be expected to occur up to the surface provided the rainfall amounts to no more than about 6 in. per year. With increasing annual precipitation, the depth to accumulating lime will increase. This depth is also a function of soil texture. It occurs at from 40 to 50 inches beneath the surface in the medium-textured soils along the approaches to the Himalayan Mountains with rainfall amounting to less than 25 inches per year. In clayey soils on the Deccan Plateau, lime may occur nearer the surface than this even though annual precipitation is somewhat higher. Any effort to show the relationship between the depth to lime and the climate of an area must involve studies on virgin soils where the effects of accelerated erosion and applied irrigation and flood waters have not confounded the effects of climate on soil properties.

Significance of Calcium Determinations in Soils

Frequent reference is made in scientific literature of the subcontinent to values for exchangeable calcium or acid-soluble calcium in the soil. Only under certain circumstances will these values have any significance.

Extraction of calcareous soils with either a salt or acid to determine exchangeable calcium usually releases appreciable quantities of this ion from $CaCO_3$. So much may be released, in fact, that the total Ca extracted by these procedures often exceeds the value for the cation-exchange capacity of the soil. If corrections could be applied for the $CaCO_3$ solubilized, meaningful exchangeable calcium values would be obtainable. However, there is no known method wherein Ca from this source can be evaluated accurately. Where free lime does not occur in the soil, extraction with neutral or slightly alkaline salt solutions will remove the exchangeable form of Ca without dissolving significant quantities of Ca-bearing minerals. Similarly, salt solutions or even weak acid solutions will extract essentially only exchangeable forms of Ca from normal, acid soils. Readily soluble calcium minerals will occur in minimum quantity in such soils.

The determination of exchangeable Ca in recently limed soils is determined with considerable difficulty at times. This is particularly true if some of the added lime granules remain undissolved in the soil. Calcium released from these granules in the field soil will maintain the adjacent clay particles in a state of near Ca saturation. Movement of the Ca ions away from the granules will be limited, however, and the distribution of Ca throughout the soil mass in the field will be highly variable. The availability of Ca to plants will be determined by the number of isolated zones high in Ca which can be contacted by foraging plant roots. If samples of these soils are examined in the laboratory through extraction procedures, a large part if not all of the lime particles will be dissolved and will appear to occur in exchangeable form. Thus, the laboratory analysis for Ca will lead to the conclusion that much more of this ion is available to plants than would be true under field conditions.

It may be seen, then, that caution must be exercised in selecting procedures for assessing the Ca status of soils. The use to be made of the specific analytical data must be ascertained before analyses are attempted, if they are to be tried at all. Methods for determining various forms of Ca in the soil are discussed by Piper (1950).

MAGNESIUM

Similarities between Ca and Mg in soils are many. Very frequently these two cations occur together in the same mineral, particularly in dolomitic limestone. Magnesium salts are generally more soluble than Ca, and, as a consequence, tend to disappear from the soil more rapidly as a result of leaching than do Ca salts. Magnesium is present in the soil in soluble and exchangeable form and as a constituent of slightly soluble minerals. One distinguishing characteristic of montmorillonite clay is that it contains a significant amount of Mg which occupies lattice positions normally filled by Al ions.

Magnesium occurs as the only metallic ion in the chlorophyll molecule and, for this reason, is highly essential for normal functioning of the plant. It is thought to aid in the translocation of starch within plants and in the formation of fats and oils. It appears also to affect the absorption and translocation of P.

A deficiency of Mg in plants reflects in stunted growth. As the deficiency becomes more acute, *chlorosis,* or loss of green colour due to the lack of chlorophyll, occurs. The chlorosis is not over the entire leaf surface, however, but is observed at tips or, more often, in the areas between the veins of the leaves (interveinal chlorosis). Unfortunately, deficiencies of several of the minor elements, especially Zn and Mn, also cause chlorosis in interveinal areas. Thus, the development of this visual symptom cannot be used alone as a positive identification of Mg deficiency.

Magnesium, unlike Ca, remains relatively mobile in the plant and is readily translocated to the young meristematic tissue. Symptoms of deficiency appear first in the older leaves, therefore. This is quite apparent in the cotton plant grown on low-Mg soils. The lower leaves may show a distinct brownish or reddish-purple colouration in the interveinal areas. In severe cases, the leaves eventually turn brown and die.

Supplementation of Mg in the soil usually occurs when lime is added for the control of an acid reaction. However, if materials of very high Ca analysis are used continuously, liming may result in a depletion of soil Mg or development of a physiological unbalance between this ion and Ca. Epsom salts ($MgSO_4.7H_2O$) or potassium magnesium sulfate ($K_2SO_4.MgSO_4$) are used most frequently for supplying Mg when the addition of a liming material is not desirable.

SULFUR

Sulfur in plants, like N and P, occurs largely as a constituent of proteins. Cystine, for example, is a S-bearing protein compound. In addition, S is contained in essential oils in some plants. The characteristic odours of garlic and mustard are attributable to sulfur-containing oils. Sulfur appears to be essential for formation of chlorophyll and protein in plants and affects the assimilation of carbohydrates.

A large part of the S in plants occurs as the SO_4 ion. In this form, it is mobile and tends to accumulate in younger plant parts. Upon maturation of the plant, S tends to concentrate in the seeds. Due to its mobility, visual symptoms of S deficiency appear in older leaves first, and are characterized by a general or overall yellowing of the leaves and stunted plant growth. This characteristic is particularly noticeable in lucerne.

Cotton, tobacco, and legumes require S in relatively large amounts, and generally are the plants most responsive to a

S application in S-deficient areas. Tea apparently has a high S requirement and has been noted to benefit from S treatment in tropical regions. Small grains and maize need lesser amounts. Usually the addition of from 20 to 30 lbs. S per acre to S-deficient soils will supply the needs of most crops.

A large part of the soil S is present in the organic fraction and is released to plants through mineralization. Deficiency may occur in sandy soils of low organic matter content where the leaching rate is high. The use of manures and ammonium sulfate and ordinary superphosphate fertilizers provide S to most soils in sufficiently large amount. Where these materials are used at least occasionally, there should be little question as to the presence of adequate S in the soil.

Since the S and organic matter contents of soils tend to parallel each other, we might expect soils of arid regions to contain the lowest amounts of this element. However, these soils frequently have high concentrations of soluble sulfate salts, including gypsum, and it is rare for S deficiencies to occur in them.

In the irrigated areas of the Indus and Ganges Plains, irrigation waters should supply ample S for plant growth. This is especially true for waters from wells, because ground waters in some areas are quite high in soluble sulfate. Though river waters are generally lower in S, as much as 50 or more pounds per year may be applied in irrigation water from this source.

Sulfur deficiency may be corrected by the addition of manures and S-containing fertilizers as mentioned above. Correction may also be brought about by the addition of gypsum or elemental S. In the latter instance, the insoluble S must first undergo conversion to the SO_4 ion. The oxidation of S usually is rapid because sulfur-oxidizing organisms, particularly *Thiobacillus*, are present in soils everywhere.

COPPER

The role copper plays in plant nutrition is somewhat obscure and highly variable. Apparently this metallic element is essential to several enzyme systems in plants and particularly those important in oxidation-reduction reactions. It may be necessary for the utilization of N in the ammonium form.

Plants grown in nutrient solutions with ammonium forms of N require higher level of Cu than when other forms of N are used.

Little is known about the forms of Cu in the soil. Seemingly, it is subject to strong fixation, but the seat of this reaction is not recognized.

According to Reuther (1957), only a small part of Cu added to soils is recoverable by exchange displacement. The fixation is greatest in fine-textured soils or those high in organic matter. Fixation is not so great if the soil is strongly acid.

Symptoms due to insufficient Cu in the soil vary so widely that no generalized description of them can be given. The interested student is referred to publications which provide specifically this type of information. Reference works edited by Hambidge (1941) and Kitchen (1948) list a large number of deficiency symptoms for Cu and most of the other plant nutrients.

Correction of a deficiency of Cu can be accomplished readily by soil application of Cu salts such as $CuSO_4$. Only very small amounts, 5 to 25 lbs. of the salt per acre, appear to be sufficient. Since Cu is highly toxic to plants, oversupply must be avoided.

ZINC

The behaviour of Zn in soils approximates closely that of Cu. Whereas many soils are characteristically deficient in both Cu and Zn, others are not. Zinc deficiency is much more widespread.

As with Cu, the function of Zn in the plant is associated with various enzyme systems. Of particular importance is *the role it plays in the formation of auxin,* the growth-promoting hormone in plants. Zinc-deficient plants are characteristically stunted due to limited shoot elongation. The distance between the leaf nodes is greatly reduced. As a consequence, leaves normally widely separated along a stem occur very close together. Lack of Zn also causes chlorosis in the interveinal areas of the leaves.

Zinc is mobile and accumulates in the actively growing parts of the plant. Thus it is common for deficient plants to show chlorosis in older leaves whereas the new leaves appear quite healthy. Viets *et al.* (1954) describe the Zn-deficiency symptoms of a large number of crop plants. They found the Zn content of plant tissue to serve as an index of the availability of the Zn in the soil.

Zinc deficiency may be corrected by spraying plants with Zn solutions or by applying soluble Zn salts to the soil. Excessive amounts are toxic. Some soils with a high fixing capacity[1] will not respond to light applications of the soluble salts. The use of chelated Zn compounds[2] are more effective in such instances. Zinc chelates tend to move freely through the soil, and, therefore, remain more highly available to plants than do adsorbed forms. Miller and Ohlrogge (1958a and b), however, have shown that certain organic chelating substances appear in manures and organic materials which, upon combination with Zn, reduces its availability to plants. This would suggest that Zn held in these organic combinations may not be absorbed readily even though the Zn is kept in soluble form. The occurrence of Zn deficiency in plants grown where manure has been stored previously has been recognized for many years. The presence of organic chelating agents in the soil would explain this behaviour.

BORON

The availability of boron to plants is dependent upon the quantity of its soluble form present. One to two ppm. of soluble B in the soil is ample to supply plant needs. Higher concentrations are toxic to some plants. The principal reserve form of B is as the relatively insoluble mineral tourmaline. In addition to the mineral and soluble forms, B is also held temporarily in organic combination in plant tissue and micro-

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[1] The affinity of most soils for Zn is so great that a solution passed through them is virtually Zn free when it drains from the bottom.

[2] *Chelates* are soluble, molecular compounds in which such elements as Zn and Cu are so tightly held that they will not enter into exchange or fixation reactions. Ethylene-diamine-tetraacetic acid (EDTA) is a chelating agent (or complexing or sequestering agent) for Ca. The Na form of EDTA will dissolve $CaCO_3$ and is used in the determination of Ca in salted soils.

bial cells, and is fixed by soil minerals. Although the mechanism of fixation is not known, it appears to be intensified by increasing the pH of acid soils, and has led to the occurrence of "over-liming injury" in some areas. An elevated pH alone does not cause the deficiency of B, however, for many alkaline soils are plentifully supplied and, sometimes, over-supplied with it.

The reduction in B availability with increasing application of lime is attributable partially to the increase in concentration of Ca in the soil. It can be related to the level of both available Ca and B in the soil. Either too high or too low Ca : B ratio proves detrimental to plant growth.

Crops differ in their B requirements and their ability to absorb B when grown under similar conditions. Berger (1949) reports on the B concentration in numerous plants, and states that it may range from 2 to 95 ppm. in different plants grown on the same soil. He rates a large number of crops as to their need for B, these requirements being based on the amount of hot-water-soluble B in the soil. A partial list of these plants is shown in Table 10-1.

Table 10-1. Boron requirement of some common field crops and vegetables expressed as the hot-water-soluble B content of the soil.

Plants with high requirement, greater than 0·5 ppm.	Plants with medium requirement, 0·1 to 0·5 ppm.	Plants with low requirement, less than 0·1 ppm.
Apple	Tobacco	Small grains
Lucerne	Tomato	Maize
Clovers	Lettuce	Beans
Red beets	Peach	Citrus
Sugar beets	Olive	Potato
Cabbage	Cotton	Grasses
Cauliflower	Peanut	Flax
Celery	Carrot	
	Onion	
	Pear	

The status of B in subcontinent soils is open to some question. River waters used for irrigation in the western areas appear to be relatively low in B (Eaton, 1953) and may give rise to B deficiency in some locations. However, the tendency for salts to accumulate in these areas may be sufficient to maintain adequate levels of B in most instances.

Boron-deficiency symptoms are very similar to those caused by a shortage of Ca. They appear first in the young, developing tissues and are characterized by growing tips which fail to elongate properly. As with Zn deficiency, leaves of B-deficient plants appear to originate from the same locus on the stem causing a "rosette", or closely grouped whorl of leaves. Often the younger leaves grow very slowly, and are yellow or yellow-red in colour. Boron-deficient plants may fail to set fruit, even though initial stages of flowering appear to be normal. The occurrence of B-deficiency symptoms in young tissue supports the belief that B is immobile within the plant and cannot migrate from older to newer parts. In this respect, it is also similar to Ca.

Boron appears to be involved in a large number of functions in plants. It has to do with cell division and growth, water use and movement, the translation of sugars through the plant, and the synthesis of organic compounds, particularly proteins. In the absence of B, simple carbohydrate compounds tend to accumulate within the plant.

Correction of B deficiency is possible through the addition of a soluble borate. Borax ($Na_2B_4O_7.10H_2O$) is normally used. Extreme caution must be exercised in the application of soluble borates to the soil. Usually from one to two lbs. of B, corresponding to 10 to 20 lbs. of borax, constitutes a suitable application for plants of a high boron requirement. Soil treatment with larger amounts of this fertilizer material may damage seriously those plants which are not tolerant to boron excesses.

IRON

The soil generally contains an abundance of iron. It is not uncommon for iron oxides to make up more than half of the soil mass where intense weathering and leaching have reduced the SiO_2 content to a low level. On the other hand, in strongly acid soils, the solubility of iron is such that it tends to disappear and may be reduced to as little as 10% of the soil or less. Even where the total quantity of Fe is high, that present in a form available to plants is frequently very low.

In young, slightly weathered soils, iron occurs largely as ferro-magnesium minerals or as pyrite (FeS). In older, highly weathered soils, its predominant form is as the oxide. Iron oxides will be found primarily in the clay fraction, the result of intense weathering.

In the oxidized (ferric) state, Fe is essentially insoluble and unavailable to plants. Under conditions of poor aeration, such as in waterlogged soils, conversion to the reduced state increases its solubility. However, this does not necessarily assure greater availability of the iron. Inundation of alkaline, calcareous soils by flood waters, for instance, often is followed very shortly by the appearance of Fe-deficiency symptoms in sensitive plants.

Iron is associated with enzyme systems in plants, particularly those having to do with respiration. It is also essential for chlorophyll formation although it is not a part of the chlorophyll molecule. Deficiency symptoms attributable to the lack of Fe result first in an interveinal chlorosis caused by the absence of chlorophyll. In severe cases, the entire plant leaf becomes yellow or almost white. Because Fe is immobile in the plant, chlorosis is apparent initially in the younger leaves. Woody plants show the symptoms most frequently. Grapes and peaches are highly sensitive to low levels of available Fe.

Iron is absorbed by the roots in ionic form. Either its absorption or utilization by plants is interfered with by a high concentration of other heavy metals such as Zn, Mn, and Cu. Iron deficiency may be accentuated by an application of a manganese salt, for instance.

The availability of Fe is governed chiefly by the pH and state of oxidation in the soil. Deficiencies occur most often in alkaline and calcareous soils where the iron is precipitated as $Fe(OH)_3$.

The most generally accepted method for increasing the availability of Fe in alkaline soils is through acidification (*i.e.*, with sulfur) or through application of an acid-forming iron salt. The acidifying agent or iron salt either should be banded in the soil or applied in coarse granular form. This will cause the acidification of a sufficient number of small, localized zones within the soil to supply adequate soluble Fe to foraging plant roots. Spraying plants with dilute solutions of Fe salts also gives a degree of correction of Fe-deficiency symptoms.

As with zinc, iron may be kept in solution by complexing or chelating agents. Certain organic compounds of Fe maintain it in a soluble state by this mechanism. Apparently these compounds are absorbed directly by the plant, a behaviour which is dissimilar to that of some organic Zn compounds.

Iron deficiency occurs often in vineyard and orchard plants. This appears to result from use of rootstocks which are incapable of extracting and transporting adequate amounts of Fe to the above-ground plant parts. The difficulty may be overcome in most instances through use of proper rootstock material.

MANGANESE

The behaviour of Mn closely resembles that of Fe, both in the plant and in the soil. These two elements are affected similarly by pH and state of oxidation, and are found to be closely associated within the plant.

Manganese-deficiency symptoms, like those of Fe, appear first in the young, actively growing plant parts. The symptoms for Mn deficiency are not so pronounced as for Fe, the tendency being for the chlorotic areas to remain relatively small except in severe cases. This fact has given rise to common descriptive names for Mn-deficiency symptoms such as "white spot" and "leaf spot". Sometimes deficiencies of Zn and Mn may be relatively indistinguishable, and the only positive means of identification rests with observing response to separate treatments with these two elements. Occasionally a plant that responds to Mn will benefit also from a simultaneous application of Fe, again attributable to their similar behaviours.

Manganese is considered to be involved in the control of the status of Fe in the plant. When the Mn level is suitable, iron appears to persist in the more mobile, ferrous state. A low level of Mn, on the other hand, results in excessive oxidation and immobilization of the Fe. At the other extreme, an excess of Mn interferes with the absorption of Fe by plant roots. The exact cause of this is not known.

Manganese deficiencies occur more frequently in strongly leached soils or in alkaline, calcareous soils. Liming an acid soil reduces the availability of Mn, apparently because oxidation to insoluble MnO_2 is encouraged. The low availability of Mn in alkaline, calcareous soils is attributable to this cause also.

Correction of Mn deficiency is accomplished most frequently by addition of a salt, usually manganese sulfate, directly to the soil. This proves to be a satisfactory procedure if the soil is somewhat acid. However, in neutral to alkaline soils, it is preferable to mix the salt with elemental sulfur and place the mixture in a band beneath the soil surface. The solubility of native Mn in alkaline soils may be increased by application of sulfur alone. This treatment, of course, results in acidification of the soil.

MOLYBDENUM

Molybdenum has been added relatively recently to the list of elements essential to plants, even though it has been known to be necessary for the development of microorganisms for many years. This element is required in extremely small amounts, and suitable rates of application for correcting a deficiency are sometimes less than one ounce per acre.

Molybdenum deficiencies are not widespread but, in certain instances, they may be very acute. The molybdate ion appears to be fixed in unavailable form in acid soils but becomes available upon liming. It is relatively soluble in neutral soils and may be leached from them readily.

One of the most notable functions of Mo is in the nutrition of nitrogen-fixing organisms. Both symbiotic and non-symbiotic bacteria fail to develop properly in its absence. Legumes grown in Mo-deficient soil may acquire a yellow colour typical of N deficiency, and the symptoms can be removed by the application of either Mo or N.

Since the rate of Mo applied to the soil is so low, it is advisable to mix the salt with some diluent (soil, sand) to allow its uniform distribution in the field. A common practice is to supply a soluble Mo compound mixed with another fertilizer such as superphosphate. Soils low in Mo are, more often than not, also low in available P.

The functions of Mo in plants have not been clearly defined. Their benefit is not only through stimulation of N-fixing organisms, however. Plants other than legumes have shown a need for Mo, particularly the *Brassica* family (*i.e.* cabbage, cauliflower, etc.).

REVIEW QUESTIONS

1. Why is the availability of Ca the highest near the neutral point and lowest at the two extremes in soil pH ?
2. Describe the difficulties associated with determining the level of exchangeable Ca in calcareous or limed soils. Why does the level of exchangeable Ca have little significance in such soils with respect to the availability of this element to plants ?
3. Explain how *kankar* (soil lime) is formed.
4. Why does a deficiency in magnesium result in chlorosis in the plant ?
5. Why do deficiency symptoms for Mg and Ca occur in different parts of the plant ?
6. What are the major sources of B and S in the subcontinent in (1) the arid, irrigated areas, (2) in the humid, tropical areas ?
7. Name the trace or minor elements and state their known functions in plants.
8. Explain what is meant by "chelating" agents, and tell why they are capable of increasing the availability of some cationic trace elements to plants.
9. Why are iron and manganese less available in alkaline, calcareous soils than in acid soils ?
10. Why has it been more difficult to prove the essential nature of the trace elements than for the major plant nutrients ?

SELECTED REFERENCES

1. Anderson, A. J. (1957) Molybdenum as a fertilizer. Advances in Agron. 8 : 164-202.
2. Gilbert, F. A. (1952) Copper in nutrition. Advances in Agron. 4 : 147-177.
3. Jordan, H. V. and Ensminger, L. E. (1958) The role of sulfur in soil fertility. Advances in Agron. 10 : 407-434.
4. Millar, C. E. (1955) Soil Fertility. John Wiley and Sons, New York, N.Y.
5. Mulder, E. G. and Gerretsen, F. C. (1952) Soil manganese in relation to plant growth. Advances in Agron. 4 : 222-277.
6. Nikitin, A. A. (1954) Technological aspects of trace element usage. Advances in Agron. 6 : 183-197.
7. *Soil,* The Yearbook of Agriculture for 1957, Supt. of Documents, Washington 25, D.C. Various Chapters on all of the plant nutrients.
8. Soil Sci. 57 : 1-65. Various papers on boron in the soil.
9. Soil Sci. 65 : 1-125. Various papers on calcium in the soil.
10. Thorne, W. (1957) Zinc deficiency and its control. Advances in Agron. 9 : 31-65.
11. Tisdale, S. L. and Nelson, W. L. (1956) *Soil Fertility and Fertilizers.* The Macmillan Co., New York.

CHAPTER ELEVEN

FERTILITY MANAGEMENT

Among factors controlling the production of agricultural crops, none are more important than the presence of an adequate and continuous supply of moisture and a suitable and well-balanced level of available plant nutrients in the soil. Of these two prime factors, the improvement in management of the fertility status is probably of greatest immediate concern to those interested in raising the average output of agricultural products in the subcontinent. Improvement of the moisture status includes increasing the supply of water for irrigated lands, removing excesses where waterlogging is a problem, and increasing the efficiency of utilization of moisture derived from natural rainfall in unirrigated (*barani*) areas. For the most part, changes necessary to improve the water status will come about slowly because the programmes involved are necessarily expensive and difficult to develop. Improvement in productivity through better fertility management can be realized almost immediately, on the other hand. For this, it is required that natural manures be utilized more extensively and in keeping with recommendations discussed in Chapters Six and Eight. Organic materials as a source of supplemental plant nutrients are inadequate to sustain the agriculture of the subcontinent, however, even if all were returned to land under ideal conditions. Therefore, the use of artificial manures must be increased, but not indiscriminately. Care is needed in selecting the supplemental plant nutrient to be used, for all lands do not require treatment with the same fertilizer materials. Their method and rate of application must also be along lines based on sound principles which permit efficient use of these scarce commodities. Summarily, then, not only must the need for fertilizers be recognized, but also proper techniques of fertility management must be followed to ensure maximum gains from their use.

TECHNIQUES IN FERTILITY MANAGEMENT

At least five factors must be considered in fertility management. They are as follows:

(1) Use of proper fertilizer materials.
(2) Use of proper methods of application.
(3) Use of fertilizers at suitable rates of application.
(4) Use of adapted and high-yielding plant varieties.
(5) Elimination of conditions, other than low nutritive status, which will interfere with plant growth.

These factors are discussed in the following paragraphs.

Determining the Proper Fertilizer Material

A fertilizer material is suitable if (1) the plant nutrient it contains is needed for increased crop production on the land in question, and (2) if the form in which it is applied is readily available to plants.

It is not a simple matter to determine which of the many plant nutrients may be limiting the yields obtained from a specific soil, but a number of means are available for this. Among them are (1) field experimentation with fertilizers, (2) greenhouse pot tests, (3) the visual identification of nutrient-deficiency symptoms in the plants, (4) plant analysis, and (5) soil analysis.

The most time-consuming and yet the surest procedure to determine nutrient deficiencies is the field trial in which a fertilizer containing the nutrient suspected of being limiting is applied to the land and resultant response by plants measured. Proper techniques of application must be used to minimize leaching or volatilization loss, excessive fixation of the nutrient by the soil, or its isolation in the dry surface soil where plant roots will not function. Check or control areas must be provided for by leaving a part of the land unfertilized. Only in this way can the true effect of the added plant nutrient be measured.

Pot tests, such as those described by Vandecaveye (1948), Colwell (1943), and Jenny *et al.* (1950), may be used successfully in determining which nutrient is in low supply in the soil. An advantage of this procedure is that the response to several plant nutrients applied to a large number of soils can be measured at one time and with a minimum of effort and expense. However, plants respond to some fertilizer treatments much more in pot trials than they do in field trials. As a result, soils which do not appear deficient in the field are deemed often to need fertilizer treatment as a result of observations from the pot trials. This is especially true for the nutrients N and P.

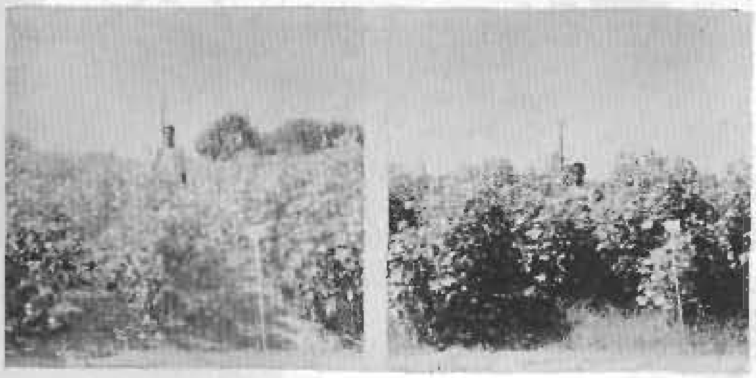
Deficiency symptoms, as discussed briefly in previous chapters, are often very helpful in diagnosing shortages of some of the plant nutrients. Since these symptoms are extremely variable, their use is practical only for one experienced in recognizing them. The same may be said for the use of plant and soil analysis in predicting the need for fertilizer materials. Interpretation of results from soil analyses is complex because numerous methods are used, some of which are specific for certain types of soil but meaningless when used on others. As will be emphasized shortly, a soil test is of value only when measured quantities of nutrient found in the soil by chemical means are shown to relate to the availability of that nutrient to crops grown in the field. Likewise, whereas the quantity of a specific nutrient found in plant tissue often is related to the level of available nutrient in the soil, this is not always true. Various factors influence nutrient uptake, and, unless these factors are understood and evaluated, plant composition may give an entirely false impression concerning the availability of a nutrient in the soil.

Probably the most useful method of diagnosing nutrient deficiencies is the rapid soil test. Because of the importance of this technique and the generally poor understanding of its value and limitations by most agriculturists, the soil test is discussed in detail in the closing sections of this Chapter.

Method of Fertilizer Application

Proper fertility management requires not only the correct identification of a nutrient deficiency, but also, when this nutrient is added to supplement that present in the soil, its method of application must permit efficient utilization by plants. Improper methods of application can, in some instances,

Maize showing the effects of *nonuniformity in soil* fertility. Correction is possible through the proper application of fertilizers and manure.



Cotton showing the effects of inadequate fertility in the soil (left) and the effect of treatment with manure (right). Note that the manured cotton stands as high as a man's head. (Experiment conducted by Prof. M. U. Din, Punjab Agricultural College, Lyallpur, West Pakistan)

Plate 2 *Facing page 300*

result in essentially complete impairment of the value of the added fertilizer. This may result from fixation reactions, loss through leaching or volatilization, or concentrating the nutrient outside the zone of greatest root activity, as in the dry surface layer of a soil.

The different methods of application and reasons for using or avoiding each of these methods have been discussed in Chapters Eight through Ten. Summarily, they fall under two headings, namely, broadcasting and placement. The most desirable procedure to follow depends largely upon the nature of reaction between the fertilizer and the soil. Placement affords protection against fixation and thereby increases the efficiency of the applied material if fixation is a problem. Broadcasting, on the other hand, allows for a wider distribution of the fertilizer particles in the soil mass. Thus, finely divided particles of materials of low solubility may be contacted by a larger number of roots if they are distributed throughout the root zone by mixing.

A large amount of experimental work has been done on proper methods of application for virtually all fertilizer materials and under essentially all soil and environmental conditions. The results of these studies can be readily adapted for use in the Asian subcontinent. Therefore, the need for research into this phase of fertility management takes a position of secondary importance. Of much greater value will be research into the kinds and amounts of fertilizers that can be profitably used under conditions prevailing in the area.

Determination of the Proper Rate of Fertilizer Application

Experimentation in the field normally provides the means for determining the rate and kind of fertilizer material to use. Since the availability of nutrients may vary from one fertilizer type to another, then so must the rates of application if similar levels of available nutrient are to be obtained from the different materials. This is particularly true for phosphates and for some of the compounds which provide the trace elements. It is not so true for the commonly used nitrogen and potash fertilizers.

The rate of application of a particular fertilizer salt is determined by adding it to the soil in increasing amounts until maximum growth is obtained. If such trials are carried out repeatedly at numerous locations within an area, the average level of nutrients in the soil may be evaluated and the average dosage of fertilizer necessary for maximum growth determined. Under normal circumstances, however, the nutrient status varies widely from field to field within any particular area. Therefore, an average recommended rate of fertilization may be of little value ; that is, a recommended rate based on average values may be too small for a field very low in the added nutrient and too large for one that contains an appreciable quantity of this nutrient in available form. In practical fertilizer management, it is necessary to take into account differences among locations. This often is possible through the use of those laboratory soil tests which have been correlated with yields under field conditions. A suitable soil test for available N is not available, however, and rates of application for this element must be based almost entirely on average result from field trials. Consideration must be given particularly to the previous history of the field and the N requirement of the crop to be grown in making recommendations for use of N.

The Use of Proper Plant Varieties

The tendency in crop-breeding work in the subcontinent is all too often directed toward producing varieties that will tolerate adverse conditions frequently confronting the cultivator. While it is important to develop varieties that are resistant to disease and adverse soil conditions, a greater proportion of the work should be directed towards introducing or developing varieties which will produce well under a high state of fertility. This relates primarily to the inability of present varieties of some plants (*i.e.*, wheat) to withstand heavy applications of nitrogen fertilizer without lodging.

It is a common practice in the Asian subcontinent to limit the rate of application of N fertilizers so that lodging will be minimized. Lodging definitely reduces yields and, for this reason, the lower application rates may be justifiable. However, the introduction of new plant varieties which will tolerate high dosages of N fertilizers appears essential in order to bring about maximum returns from the available land and to meet the ever-increasing demand for foodstuffs.

The Elimination of Adverse Soil Conditions

Soil conditions which interfere with crop production are varied. Some of the more important ones are (1) waterlogging and salt accumulation in irrigated areas, (2) erosion by wind and water, (3) poor soil physical conditions, largely the result of a universally low organic matter content in the soil, (4) inadequate moisture supply, and (5) soil-borne diseases.

The occurrence of any one of these problems will prevent the attainment of abundant yields. However, for the most part, their correction will require much time and great expense. An effort should be made to take land with such problems out of cultivation until it has been restored to a productive state. Fertilizers applied to them are essentially wasted and would be utilized to a much greater advantage on soil of high inherent productive capacity. This applies equally well to the frequent misuse of irrigation water. With water in such short supply, one is hardly justified in spreading it on poor land when production on good land is limited by a water shortage. Serious consideration should be given to laws which would prevent the use of scarce irrigation water for growing crops on land that does not produce above a certain practical minimum level.

SOIL TESTING

Soil tests are used for two reasons: (1) to aid in determining whether or not a particular nutrient is responsible for poor productivity, and (2) to determine the amount of a plant nutrient that must be supplied to raise its content in the soil to an optimum level. To be practical, a soil test must be rapid, and the analytical results it provides must relate to the amount of a nutrient made available to plants during a growing season.

The Nature of the Soil Test

Rapid soil tests involve the extraction of a small soil sample, usually with a salt or acid solution or with water, and measurement of the amount of nutrient element released during the extraction. The soil sample normally is shaken with the extracting solution until near equilibrium conditions are established. The quantity of the element in the extract is measured and the amount removed from the soil expressed on the basis of lbs. per acre or parts per million parts of soil. Since an acre furrow slice of soil is considered to weigh approximately 2 million lbs., one part of extracted element per million parts of sample is equivalent to 2 lbs. per acre furrow slice.

The results of laboratory tests obtained by extraction procedures, as described above, are meaningless unless they can be related to conditions in the field. The actual relationships are established through *soil-test correlations* which involve the use of fertilizer trials in the field. From these one may determine the response to supplemental plant nutrients and relate it to the level of extractable nutrient in soil samples taken from the experimental site.

Correlating Soil Tests

Even the most reliable laboratory soil test for a plant nutrient detects only a part of the total quantity of that nutrient present in the soil. Furthermore, the quantity extracted during the test and expressed as that occuring in an acre furrow slice of the soil under examination will not be the same as the quantity of that nutrient removed by plants grown on the soil under field conditions. However, for the test to be useful, the amounts extracted by the test solution and by the field-grown plants should be in a nearly constant ratio to each other.

The quantity of nutrient a plant removes will determine yield to a large extent. Therefore, the quick test predicts crop yields indirectly so long as that crop is grown under relatively ideal conditions. Unfortunately, the yield of a plant is controlled also by factors other than fertility. Thus, two soils showing the same quantity of extractable nutrient may still yield differently for reasons other than fertility.

Because of this, it is generally impossible to obtain a perfect relationship between the soil test results and crop yield. The use and interpretation of soil tests will predict the need for fertilizer accurately only a part of the time. It is important to understand this limitation in any soil test used and to search continuously for other more accurate tests.

Data from experiments on soils in the central part of the United States show that the level of exchangeable K can serve as a good index of the availability of this element to plants. This is true even though a large part of the K absorbed by plants is derived from nonexchangeable form. These results are logical because of the state of equilibrium normally existing between exchangeable and nonexchangeable K in the soil. Generally, the greater the amount of nonexchangeable K the higher will be the amount in exchangeable form.

Experiments with maize reported by Bray (1948) indicate that maximum yields cannot be expected unless the level of exchangeable K approximates 250 lbs. or more per acre furrow slice of soil.[1] This is shown graphically in Fig. 11-1 in which each plotted point represents results from a fertilizer experiment in the field. Of note is the variation in distance between these points and the average curve. In this case, the variation is not great, to be sure. It does show, however, that perfect correlation, where all points would fall exactly on the curve, cannot be expected. Outstanding is point A in the figure. It represents a soil which is relatively low in exchangeable K and yet did not respond appreciably to an application of fertilizer K.

As the soil test value is related to actual uptake of nutrients by plants by a proportionality factor only, so it is related to the amount of nutrient element needed in a soil to rectify completely a deficiency. This is illustrated also by data of Bray (1948) shown in Table 11-1. Listed in the table are suggested rates of potash, as lbs. K_2O per acre, to be applied to three groups of crops differing in their potash requirements.

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[1] The value stated here (250 lbs. per acre) should not be considered as universally applicable. Allaway and Pierre (1939) claim, for calcareous soils in the State of Iowa, that 175 lbs. exchangeable K per 2 million lbs. of soil is the dividing line between sufficient and insufficient levels. This value may be expected to vary from area to area.

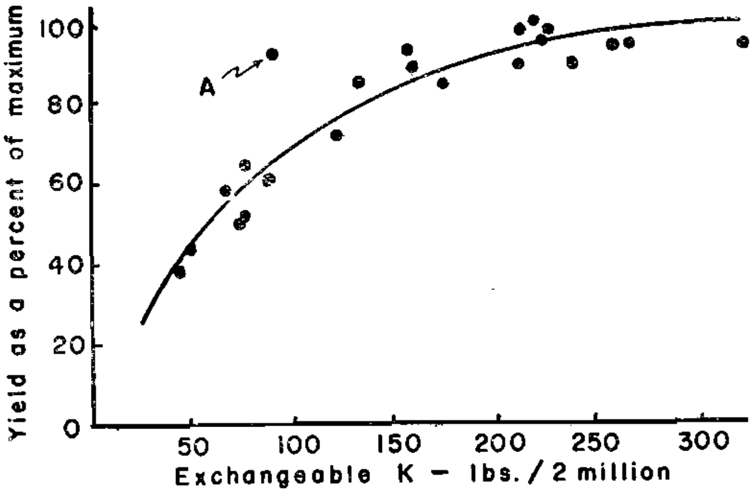

FIG. 11-1. The relationship between the level of exchangeable K in the soil and yield of maize. The plotted points show the yield obtained on unfertilized soil expressed as a percent of the yield obtained when ample K has been applied as a fertilizer. Thus, if the soil contains about 50 lbs. of exchangeable K per acre furrow slice, the yield will be approximately 40% of that obtained if the soil were fertilized adequately. Point A, which lies well above the average curve, represents a field trial where some factor other than K nutrition prevented the normal response to fertilization with K (after Bray, 1948).

Note that the low-requirement crops, wheat and oats, are not expected to require supplemental K if the soil test value is 130 lbs. of K per 2 million lbs. of soil (lbs. K per A.) or greater. As indicated above, this element is not considered adequate for maize unless the soil test shows a level of 250 lbs. K per A. or more.

Various extractants and extraction procedures for a number of plant nutrients have been found to correlate well with the availability of these nutrients in field soils. Measurement of exchangeable K, as discussed above, has been shown to be a satisfactory index of the availability of this element. It stands to reason, therefore, that an extraction procedure which removes a definite proportion of exchangeable K will

Table 11-1. The relationship between level of exchangeable K in the soil and amount of K_2O required to satisfy needs of several crops for maximum production (after Bray, 1948).

Exchangeable K	K_2O requirements for fertilization of		
	Maize or clover	Soyabeans	Wheat or oats
lbs./2 million	lbs./A	lbs./A	lbs./A
40	71	47	31
60	68	42	25
80	62	36	18
100	55	31	12
130	46	22	—
150	37	18	—
200	20	—	—

correlate equally well with field results. For this reason, equilibrium extraction with a variety of salts may be expected to measure the availability of soil K so long as the soil test is correlated through use of field trials.

Soil tests for ions which do not occur in exchangeable form present a different problem in extraction. This is particularly true if the ion in question forms compounds in the soil whose solubilities are affected by variation in pH. Phosphorus serves as an example of this type of ion. Its availability to plants is governed largely by the solubility of the compounds in which it occurs. Therefore, determination of the relative availability of P through use of laboratory tests requires that extractants used will remove P in proportion to the amounts which become soluble and available to plants during the growing season. It would be illogical to attempt a strong acid extractant with calcareous soil, for instance. Whereas the P from calcium phosphates is only slowly available to plants, these compounds are readily soluble in acid. Thus, extraction of an alkaline soil with acid normally would yield high extractable-P values not relatable to the actual availability of this element to plants.

The more successful tests for P are those utilizing extractants which do not encourage the solution of phosphate-bearing compounds. They may, on the other hand, suppress the solubility of the P-fixing ions, Ca, Fe, and Al, allowing for the release of a part of the P from compounds formed with these ions. This approach appears to provide the best correlations for P tests so far developed.

Among the more usable tests for P are those of Olsen *et al.* (1954), which uses 0·5 M $NaHCO_3$ extraction, and of Bray (1948), which uses a weakly acidified ammonium flouride solution. The latter test is designed for acid soils. The F ion added complexes Fe and Al and releases P during the extraction. The bicarbonate ion precipitates all three of the major P-fixing ions, Ca, Fe, and Al. The bicarbonate test appears to correlate well on a large variety of soils but has shown most promise on neutral to alkaline soils, with or without lime.

The Soil Sample

A frequently stated truism is that a soil test is no better than the sample on which it is run. A basic requirement for a suitable soil sample is that it must be representative of conditions in the field from which the sample has been obtained. Normally a number of individual samples are taken from a single field, mixed together, and a small portion of the composited soil used for the laboratory sample. If the field can be divided into areas of soils differing markedly in productive capacity, then composite samples from each such area should be taken for separate analysis. The individual samples may be taken as a uniform slice from the side of a freshly dug hole, or by means of a soil auger or sampling tube. Normally soil only from the surface 6 to 8 inches is used. In certain instances, samples may also be taken from the subsoil. Various good papers on sampling are available in the literature. Among them are those by Reed (1953), Reed and Rigney (1947), and Cline (1945).

THE FERTILITY STATUS OF SUBCONTINENT SOILS

Inference has been made in previous chapters as to the generally low fertility level of subcontinent soils. This knowledge is acquired largely from the observed response of crops to the field application of fertilizers. Virtually all areas are responsive to nitrogen fertilization. It is quite possible that, 9'23

where response has not been observed, other factors of a limiting nature have interfered with normal utilization of added N. It is felt that, in a large amount of experimentation, failure to use a phosphatic material with nitrogen has resulted in observing only a nominal response to the nitrogen.

It is probable that potential phosphate deficiencies are almost as widespread as are those for nitrogen. If it were not for the use of at least small amounts of farmyard manure by most cultivators, the status of phosphorus would be even more critical than it is today. Certainly, as the use of N fertilizers increases, phosphate deficiencies are going to become increasingly in evidence. If proper soil tests for P are developed, however, problems arising from low P supplies can be avoided to a large extent. A soil should not be allowed to become deficient before P is added as a fertilizer.

The need for P in subcontinent soils has been recognized for many years, but its use as a supplemental plant nutrient has not been stressed. Siddiqqi and Sardar (1950) and Sardar (1952) have noted the response to P in the upper Indus plains and hill areas. The author examined soils from approximately 75 widely spread areas representative of the Indus and Ganges plains and neighbouring uplands. Extraction with $NaHCO_3$ solution indicated that at least 50% of these soils were low in available P. The mineral materials from which the soils have been formed may be a primary cause for the low P status. For example, fresh river and canal sediments in the former Punjab are very low in this element, silts having only 2 ppm. and sands but 1 ppm. P extractable by $NaHCO_3$.[1]

Various factors point to the probability of low levels of available P in the more humid regions of the Indian Peninsula, Eastern India, and East Pakistan. Particularly important to this is the fact that many of the older soils in these areas have been depleted of reserve phosphate minerals through intense weathering. The relatively high level of sesquioxides in these

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[1] Correlation tests have not been made for 0·5 M $NaHCO_3$-extractable P on these soils. It is assumed, however, that a level of 8 to 10 ppm. (16 to 20 lbs. P per 2 million lbs. soil) or above is adequate for most crops of high-P requirement. This value has been found applicable to a number of the areas of arid-region soils in the United States.

soils indicates the potential of their retaining P in an unavailable, fixed form. Unless placement methods were used to ensure that added fertilizer P remained available, it is highly possible that a response from a soil treatment would not be obtained.

The above-mentioned low levels of available P prevail under agricultural conditions characterized by low crop production and, therefore, limited removal of this element in crop plants. Undoubtedly increased use of N fertilizers without P would rapidly deplete the available P supply in many soils now showing adequacy by chemical test. The effect of a low-P supply would be most noticeable on high-yielding crops as sugarcane and legumes, particularly Egyptian clover, gram, and lucerne. Response to P fertilizers by grasses and small grains occurs only if the supply of available P in the soil is very low.

Potash is probably in plentiful supply in most arid-region soils of the subcontinent. The more highly leached soils of the humid regions of India and East Pakistan have been shown frequently to be responsive to K fertilization. Little information is available on the need for trace elements.

A large part of the data obtained from fertilizer trials conducted in the past are of limited value. The limitations arise from two general inadequacies, (1) the application of fertilizer materials at rates too low to allow for maximum yield response, and (2) the failure to use combinations of fertilizer elements which, when used together, could cause much greater response than would any one of them when used alone. For example, the need for N has been evaluated under conditions which too often have been adversely affected by the shortage of some other plant nutrient, particularly P.

Stewart (1947) has evaluated numerous soil-fertilizer investigations carried out in the subcontinent. He favours the widespread use of test plots in farmer's fields to establish the average fertility status of an area. A more fruitful approach would be to use a limited number of these trials to correlate soil tests, and then apply this latter means rather than extensive field studies to determine the actual need for fertilizers in an area.

Population Pressure and Soil Fertility

Usual discussions on fertility management approach the subject from the standpoint of increased revenue to the cultivator. For the subcontinent, a much more basic consideration is the need for supplying adequate foodstuffs for the rapidly increasing population. The magnitude of this problem is brought into vivid focus by Anderson (1954).

If, through use of data presented by Anderson, we compare the trends in population on the one hand, and crop yields, as exemplified by those for rice on the other, some rather startling facts are brought to light. Such a comparison is made in Fig. 11-2. The change in the population of the area comprising present-day India is at an ever-increasing rate. The average yield of rice, however, has fallen steadily since 1910. Part of this decrease is probably the result of expansion of agriculture

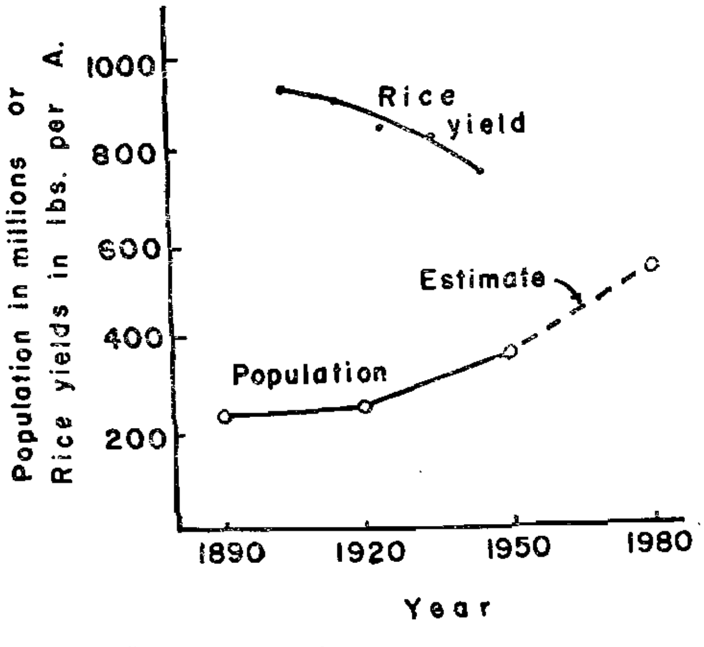

FIG. 11-2. Concurrent trends in the average yield of rice and in population growth in India.

into less productive areas (marginal and submarginal land). Undoubtedly it also is related closely to the decreasing level of fertility of the soil. These data compliment neither the agricultural research of the region nor the application of known basic facts of fertility management on the part of the farmer. If the trends in the curves of Fig. 11-2 are not halted or even reversed in the future, the consequence can be no less than catastrophic.

The solution to the problem of suboptimal crop yields rests first with improved research, improved extension, increased use of fertilizers, and introduction of better plant varieties. Fertility research must be placed on a sound footing with initial objectives being to determine which fertilizer materials are needed, their best methods of use, and suitable rates of application.

REVIEW QUESTIONS

1. Name the 5 methods outlined in the text which are used to diagnose plant nutrient deficiencies in soils, and explain in a general way how each method is used.
2. In what ways will the requirements for fertilization by lucerne and maize differ ?
3. Why would you apply a phosphate fertilizer differently to humid-region and arid-region soils ?
4. Yields of presently used wheat varieties cannot be heavily fertilized with nitrogen. Why is this, and how should this problem be overcome ?
5. Why might you expect extraction of an alkaline, calcareous soil with $NaHCO_3$ solution, on the one hand, and acidified sodium acetate, on the other, to yield markedly different amounts of extractable P but very similar quantities of extractable K ?
6. Maize grown on strongly alkaline soil was stunted and showed iron-deficiency symptoms. An acid extract of the field soil showed a large amount of iron present. Explain these apparently divergent observations.

SELECTED REFERENCES

1. Berger, K. C. and Truog, E. (1944) Boron tests and determination for soils and plants. Soil Sci. 57 : 25-36.

2. Mikkelson, D. S. (1955) Cotton phosphate fertilization. California Agri. 9 : 7.

3. Viets, F. G., Jr. and Hanway, J. J. (1957) How to determine nutrient needs. From *Soil*, The Yearbook of Agriculture for 1957, pp. 172-184. Supt. of Documents, Washington 25, D. C.

CHAPTER TWELVE

THE SALT PROBLEM IN IRRIGATED SOILS

SALINE AND ALKALI SOILS

SALINITY and alkali problems in the Indus and Ganges plains and the Deccan Plateau present one of the most serious threats to the soils of these areas. Good agricultural land is being continually forced out of production because of the spread of this 'disease'. Since there is such a high dependence on irrigated soils for food production, the need for control of further spreading and for the correction of waterlogging is of concern to everyone.

The basic cause of saline and alkali conditions in soil is waterlogging. Waterlogging and accumulation of salts go hand in hand; in arid regions, reference to one of these conditions usually indicates the presence of the other. The remedy for the problem is the improvement of drainage so that excess water, from whatever source, will move more rapidly to the sea rather than accumulate beneath the surface of the soil. When this is accomplished, the removal of salts is facilitated since salts may be moved out of the soil more rapidly by leaching.

The correction of waterlogging and removal of salts from subcontinent soils will be a tremendous undertaking. It cannot be accomplished by an individual or even a small group of persons. The programme involved must be considered on an area basis, and will require the attention of virtually all people within it whose livelihood depends upon sound argiculture.

It must be borne in mind that textbook procedures for reclamation cannot be applied indiscriminately to any and all fields. Each problem area must be examined individually, and the design of a reclamation programme adapted to soil properties and other conditions peculiar to the location in question. It is the purpose of the following discussions to acquaint the student with procedures for examination of salinized soils along logical and, yet, simple lines, and to provide some basis for developing reclamation programmes in accordance with the results of these examinations.

CHARACTERIZATION OF SALINE AND ALKALI SOILS

A *saline soil* is one which contains a high concentration of soluble salts ; an *alkali soil* is one in which a high percentage of the exchange positions are filled with sodium.[1] Generally, alkali soils contain damaging quantities of salts as well as a high level of exchangeable Na and are designated as *saline-alkali* soils. Some alkali soils are observed also to have very low levels of salinity because a large part of the soluble salts have been leached from them. These are known as *nonsaline-alkali* soils. The classification of soil into one of the above categories obviously is based upon information obtained through chemical analyses.

The Measurement of Salinity in Soils

Numerous methods have been devised for the characterization of the salinity problem in soils. One of the earliest and most widely used procedures has been the determination of *total dissolved solids.* After mixing the soil with a relatively large volume of water and filtering, the extract is evaporated and the salts determined gravimetrically. Results are expressed on a percentage or similar basis.

Two errors are introduced by this procedure. In the first place, if a slightly soluble salt, such as gypsum, is present in the soil, it will dissolve in greater amounts as compared to its solubility in the relatively small amount of water present in a soil at field capacity. This would cause the salt content of the soil to appear higher than it would be under field conditions. Expressing salt concentration as a percent of the soil weight does not take into account that like quantities of different salts yield varying quantities of ions. For instance,

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[1] The term *alkali* came into use to designate soils high in exchangeable Na because they often have an unusually high pH and an appreciable quantity of strongly alkaline, soluble Na_2CO_3. A current tendency is to replace this name with *sodic* to designate such soils. This permits use also of terms *potassic*, *magnesic*, and *calcic* to classify soils with properties affected by a predominance, respectively, of potassium, magnesium, or calcium on the exchange complex.

Na_2SO_4 will liberate almost 20% fewer ions into the soil solution than will the same weight of NaCl. Therefore, it is incorrect to assume that like percentages of different salts react similarly so far as salinity effects on plant growth are concerned.

The U. S. Regional Salinity Laboratory (Richards, 1954, pp. 7-17) recommends evaluation of salt level in soils by determining the electrical *conductance* of soil extracts. The conductivity of a solution provides a good measure of the effective quantity of salts present because the value of this measurement is dependent upon the number of ions dissolved in a given volume of water. It will be found that the ion concentration in the soil solution relates more closely to salt damage than does any other property. Furthermore, the osmotic pressure of a solution is a function of the concentration of ions also. Thus, ionic concentration (expressed as me. per l.), osmotic pressure, and conductivity are all directly proportional to each other.

Pure water has a high resistance to the flow of an electric current. As the ionic concentration increases, the resistance decreases and, therefore, under the influence of a constant voltage, the current flow increases. Whereas the unit of measurement for electrical resistance is the *ohm*, conductance (the reciprocal of resistance) is expressed as *mhos*, and the standard unit of measurement is *mhos per cm.*[1] The conductance of most soil solutions is so low that values as large as 1 mho per cm. are encountered only infrequently. As a consequence, it is more convenient to express conductivity as *millimhos* (0.001 mho) or *micromhos* (0.000001 mho) per cm. The conductivity values associated with soils not particularly affected by salts normally is below 1 millimho per cm., whereas values for seriously affected soils may range as high as 20 to 40 millimhos per cm. or more. Irrigation waters of low salt

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[1] According to Ohm's law, the current (I) which will flow in a conductor under a given applied voltage (V) is inversely proportional to the resistance (R) of the conductor, and is expressed algebraically as $R=V/I$. Since the conductivity (EC) is the reciprocal of R, then $EC=I/V$. Therefore, under a constant voltage, current flow is proportional to the conductivity. Standard cells used for measuring conductivity of solutions contain two electrodes placed 1 cm. apart. This provides, then, for the measurement of conductivity directly in mhos per cm.

content generally have conductivities which are measured in terms of micromhos per cm. For convenience in use, the units of measurement expressing conductivity values frequently are shown symbolically as:

EC = conductivity in mhos/cm.
$EC \times 10^3$ = conductivity in millimhos/cm.
$EC \times 10^6$ = conductivity in micromhos/cm.

The concentration of salts in the soil solution under field conditions varies with the moisture content and becomes higher as the moisture level tends toward the wilting point. As a consequence, the conductance of the soil solution is variable also. Therefore, to relate salinity effects among different soils, it is necessary to obtain extracts for conductivity measurements from them after they have been adjusted to comparable moisture levels. A commonly used procedure is to mix the soil with water until it almost acquires the properties of a fluid. A soil sample weighing from 500 to 1000 g., when moistened to this level, will give up sufficient extract under suction filtering to permit not only the measurement of conductivity, but also the quantitative determination of the soluble anions and cations which have an important bearing on properties of saline and alkali soils. The extract obtained according to this procedure is called the *saturation-paste extract.*

The use of the saturation-paste extract in assessing the level of salinity represents a compromise in moisture content which will provide an adequate volume of solution for chemical analysis but will not bring about the solution of excessive quantities of slightly soluble salts such as gypsum. Thus, one may expect the conductivity and ion concentration of the saturation-paste extract to relate very closely to corresponding values in the soil solution under actual field conditions. Readily soluble salts such as $NaCl$ and Na_2SO_4, on the other hand, usually dissolve completely in the soil solution under field conditions and, therefore, will be more dilute than normal when the soil moisture content is raised to that of the saturation paste.

For purposes of classification, a soil is considered to be saline if the conductivity of its saturation-paste extract is 4 millimhos per cm. or more. This point of division is strictly arbitrary, and has been selected on the basis of average plant response to saline soil conditions. Tolerant plants may grow well at conductivity values appreciably above 4 millimhos per cm. However, the growth of most common crop plants becomes noticeably hindered when salts are of sufficient concentration in the soil to produce this level of conductivity in the saturation-paste extract.

The Exchangeable Na Status of Alkali Soils

According to the most widely accepted standard of classification, an alkali soil is one in which the level of exchangeable Na is 15 percent or more of the cation-exchange capacity. Whereas high levels of exchangeable K and Mg are known to induce undesirable characteristics in soils, a limiting level has not been established for them as has been done for exchangeable Na. Neither ion is as effective as Na in creating adverse physical conditions in the soil. For this reason, reference to the term 'alkali', which normally is associated with soils of high pH and poor physical condition, relates to soil having an excess of exchangeable Na unless otherwise specified.

Properties of Saline and Alkali Soils

Aside from the possible toxic effect of certain ions in saline soils, the principal direct harm caused by accumulating salts is their osmotic influence which reduces the availability of water to plant roots. In soils classed as saline but nonalkali, the damaging salts normally will be NaCl, $CaCl_2$, or Na_2SO_4. Calcium sulfate may accumulate in the soil as a precipitate, but will not occur in harmful quantities in solution form. The maximum conductivity of a solution saturated with $CaSO_4$ is about 2·2 millimhos per cm. High concentrations of other Ca and SO_4 salts will reduce the solubility of $CaSO_4$ because of the common ion effect.

Two alkaline salts, $NaHCO_3$ and Na_2CO_3, may accumulate also in soils. When first introduced into a normal soil high in exchangeable Ca, these salts will disappear to a large extent. The Na from them enters exchange positions to displace Ca, and this latter ion is precipitated as very slightly soluble $CaCO_3$. So long as appreciable exchangeable Ca remains, the HCO_3 and CO_3 ions will not become important constituents of the soil solution. Soluble Na_2CO_3 may accumulate finally when most of the exchange positions become filled with Na ions, however.

Absorbed or exchangeable Na has two principal effects on soil properties. The first of these is that it has a dispersive influence on particles which might otherwise persist in aggregated or flocculated form. Because of its low charge density and relatively high degree of ionization from particle surfaces, Na is the least effective of all exchangeable ions in neutralizing the negative charge of soil colloids. As a consequence, even when Na saturated, the particles retain sufficient residual negativity to cause them to repel each other and thereby become dispersed. Alkali soils, therefore, are noted to puddle readily and to form dense, compact masses through which water and air move only very slowly. Not only does this restrict plant growth, but also reclamation of such soils is made extremely difficult because of the problem encountered in leaching out the accumulated salts. In fact, most alkali soils high in clay leach so slowly that their reclamation generally is quite impractical.

The second important effect of exchangeable Na is that of causing a relatively high soil pH. When an alkali soil sample is placed in water, as is necessary for the measurement of pH, the ready ionization of Na from particle surfaces allows for the adsorption of a few H ions derived from the concurrent ionization of the water. The removal of these H ions from solution leaves dissociated OH ions to cause a rise in pH. In essence, the reaction is caused by the *hydrolysis* of exchangeable Na to form a small quantity of highly ionized, strongly alkaline NaOH in the soil solution.

The readily soluble Na_2CO_3 which might be present in an alkali soil also can undergo hydrolysis when placed in water and effect a rise in pH. This reaction results in the formation of equal quantities of NaOH and H_2CO_3 which are, respectively, a strong base and a weak acid. Soils with very high pH values normally may be expected to contain at least modest quantities of Na_2CO_3. When present, it can be removed from the soil in a saturation-paste extract and its quantity determined by titration with standard acid.

The accumulation of neutral Na salts along with high amounts of exchangeable Na will, surprisingly enough, maintain the soil at a relatively low pH and will reduce the dispersive effect of the exchangeable Na. The depressing influence on pH arises from the reduced hydrolysis of exchangeable Na as the number of dissolved Na ions in the solution phase are increased. These latter ions compete favourably with H ions for exchange positions. Thus, the H ions are forced to remain in association with the OH ions and a significant rise in pH prevented. Soils without neutral salts but essentially saturated with exchangeable Na may have pH values of 10 or more. The same degree of Na saturation may result in pH values no higher than 8·0 to 8·5 in the presence of an appreciable quantity of neutral salts.

As the quantity of salts increases in the soil, more and more positively charged cations crowd against particle surfaces to reduce the residual negative charge. As a consequence, when the salt content is high, the particles may approach each other closely and flocculate. Indeed, saline-alkali soils frequently are well aggregated and appear to be in good physical condition. Water, when first applied, will move into them rapidly. As the salts are dissolved and carried downward, however, dispersion follows and the rate of water movement subsequently reduced.

Puddled alkali soils normally are difficult to cultivate or prepare properly for seeding. Due to the high pH, the essential plant nutrients, Ca, Mg, Fe, and Mn, are precipitated and rendered less available. Phosphorus availability, on the other hand, may increase as the result of the formation of relatively soluble sodium phosphate compounds.

Frequently, a dark film of organic matter occurs over the surface of strongly alkaline soils. This film develops as organic matter dissolved in the alkaline soil solution precipitates when the moisture rises to the surface of the soil and evaporates. The presence of this film has been used by many people in identifying high-Na soils, and the name 'black alkali' soil has been applied to them. Similarly, 'white alkali' has been used often to designate a saline soil, the name being derived from the white inflorescence associated with them. The use of either of these names is discouraged, however, for neither of the visual features they refer to serve as positive means of identifying a specific type of salted soil.

THE CAUSES OF SALINE AND ALKALI SOIL CONDITIONS

The introduction into or the redistribution and concentration of salts within a localized area of soil is necessary for the formation of saline and alkali soils. Initially the salts are derived from the weathering of minerals in the earth's crust. They may accumulate within the locale of their formation, or they may be transported by water to lower-lying soils wherein they subsequently may accumulate.

In arid zones, salts tend to accumulate in soils because the water lost from the soil surface by evaporation exceeds that entering the soil in the form of precipitation or irrigation. In other words, the net movement of water and salts dissolved in it is in an upward direction. Salts native to a soil may be redistributed or concentrated as the result of movement of seepage water from streams or ponds or by applied irrigation water.

Salts may be caused to accumulate or concentrate as the result of nonuniform wetting of soil where the surface of the ground is uneven. Precipitation falling upon high ground tends to run to lower areas where it moistens the soil more completely. Since moisture in the soil moves continuously from zones of higher water content, the net internal movement will be away from the low areas and toward the more elevated, drier knolls. Salts will be transported in the water to accumulate at the more elevated positions, whereas soil in the lower-lying positions will be kept more nearly free of salt. This effect is most apparent if the lower-lying positions serve as the beds for temporary or permanent waterways. As illustrated in Fig. 12-1, salts carried in by the water will accumulate with indigenous salts in the adjacent land as seepage takes place. This effect is noticeable also in land adjacent to some irrigation canals.

The effect of permanent streams and rivers on the salt content of adjacent lands is modified where frequent flooding occurs. Salts accumulated as the result of lateral movement of seepage water from the streams may be flushed from the soil partially when floodwater traverses the surrounding low-lying lands. The effect is more pronounced in depressions where the water residual from the floods cannot move away by natural surface drainage but disappears largely by percolating through the soil (see Fig. 12-1). In so doing, the percolating water leaches out salts which may have accumulated previously in the soil. Normally, this process of reclamation is inadequate to overcome completely the effects of Na-salt accumulation and alkali soil formation brought about the constant inundation of areas adjacent to the permanent waterways.

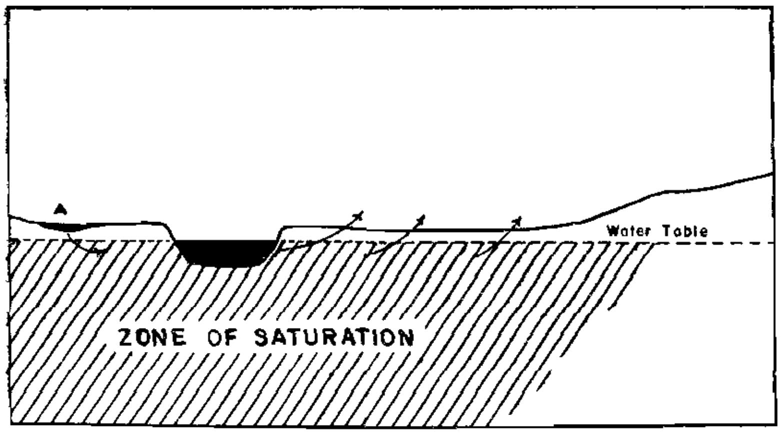

FIG. 12-1. The effect of permanent streams on the accumulation of salts in arid-region soils. Seepage water from the stream saturates the subsoil of adjacent land and serves as an external source of salt . Movement by capillarity from the zone of saturation and concentration of the solution at the surface by evaporation causes salinization of the soil. Entrapped flood water, as at A, tends to leach salts from depressional areas minimizing accumulation there.

The effects discussed above are important in explaining the origin of naturally-occurring saline and alkali soils. Contributing also to the formation of these problem soils in many of the irrigated areas of the subcontinent is the increase in water-logged soils apparently brought about largely by the initiation of the present, extensive irrigation systems. Frequently, the water table occurs at or above the land surface. Generally, these waters are high in salts. Their nearness to the surface provides a continual source from which water may rise, evaporate, and leave behind an ever-increasing level of salt in the soil. This problem, as it applies to conditions in the subcontinent, is discussed briefly in a later section.

The Influence of Ground Water Composition on Soil Properties

The chemical composition of ground water has a profound effect upon the type of soil which will develop under water-logged conditions. Of major significance in this respect is the ratio of Na to Ca + Mg in the water, or, perhaps of greater importance, the Na : Ca + Mg ratio of the soil solution resulting from the concentrating effects of evaporation on inundating waters. The level of exchangeable Na in a soil is governed by exchange equilibria dependent upon the ratio of Na to other cations in the solution phase. In other words, the greater the proportion of Na in the soil solution, the higher is the percentage of Na on the exchange complex.

The total quantity of salts in the ground water will, of course, influence the rate at which these salts will accumulate in a soil during a specific period of time. If the salts are largely chlorides, the Na : Ca + Mg ratio does not change as the total-salt concentration in the soil increases. If sulfate is present and accumulates to a level sufficient to exceed the solubility of $CaSO_4$, soluble Ca will be lost through precipitation and the Na : Ca + Mg ratio will increase with time. This will be reflected in a more rapid increase in the level of exchangeable Na than would occur where the accumulating salts essentially are only of the chloride form.

The HCO_3 ion has the most profound effect of the common anions on the Na : Ca + Mg ratio of accumulating salts. By virtue of the reaction,

Land made unfit for farming by seepage of water and salts from an irrigation canal.

Salt-infested land may be rendered useless for distances of up to one-half mile from unlined canals where soils exhibit moderate to high permeability to inundating water.

$$2HCO_3^- = CO_3^= + H_2CO_3 \qquad (12\text{-}1)$$

carbonate ions are formed which are capable of precipitating virtually all of the Ca and Mg present. Loss of CO_2 from the H_2CO_3 causes the above reaction to shift to the right increasing the CO_3-ion concentration and the effectiveness of this reaction in precipitating the divalent cations.

Eaton (1949) proposes that if the $HCO_3 + CO_3$ concentration of water exceeds the Ca + Mg concentration, essentially all of the soluble divalent cations will be precipitated eventually in the soil. Any $HCO_3 + CO_3$ remaining has the potential of forming Na_2CO_3 and can cause precipitation of exchangeable Ca in the soil. This, of course, increases the tendency for alkali soil formation. Eaton refers to the calculated excess of $HCO_3 + CO_3$ over Ca + Mg, when expressed as me. per l., as the concentration of 'residual Na_2CO_3' in the water.

Conditions at the surface of the soil in arid regions are ideal for Na_2CO_3 formation from bicarbonate waters. Here, loss of CO_2 from the system will proceed spontaneously.[1] Because of the loss of water through evaporation and use by plants, salts in the solution phase become more concentrated and precipitation of Ca and Mg as carbonates is encouraged. As with $CaSO_4$ precipitation, the resultant increase in the Na : Ca + Mg ratio of the solution phase causes an increase in the level of exchangeable Na in the soil even though the total quantity of Na ion in the soil may not be changed.

Eaton (1949) applies the concept of 'residual Na_2CO_3' to explain the formation of naturally-occurring saline and alkali soils. Accordingly, salt-bearing inundation waters containing 'residual Na_2CO_3' result in alkali soils, whereas those having no 'residual Na_2CO_3' produce saline soils. This concept appears to have wide application in the Asian subcontinent, particularly for lands in the Indus-Ganges watershed. The waters of the rivers in these basins, though low in total salt, do have 'residual Na_2CO_3', and soils that have been

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[1] A solution of $NaHCO_3$, standing open, changes gradually to a mixture of $NaHCO_3$ and Na_2CO_3 because of the loss of CO_2 to the air.

inundated by them frequently and over very long periods of time are strongly alkaline and high in exchangeable Na.

Water Table Depth and Salt Accumulation

The presence of waterlogged subsoil conditions does not result always in the accumulation of salts at the surface. The total distance through which rising water must move is important in determining the overall effect of this mechanism of salt accumulation. Of importance also is the counterbalancing effect produced by the percolation of surface-applied waters through the soil.

The effect of distance to the water table on the upward movement of water is illustrated by the data of Gardner and Fireman (1958) as shown in Fig. 12-2. In their study, evaporation from the surface of soil columns was determined as the depth to a free water surface at the lower end of the columns was varied. Two soils of widely different texture were used, a clay and a sandy loam. Their data show that, regardless of the texture, when the depth to the water table is less than 2 to 3 ft., evaporation is high. Within this range of depth, external factors (temperature, wind velocity, etc.) governed the rate of evaporation. Below a depth of 6 ft., a sharp decrease in the evaporation rate occurred with both soils and, beyond this depth, evaporation decreased more rapidly in the sandy soil than in the clay.

A very useful general conclusion may be drawn from these data. To prevent a significant rise of salt in the soil profile, the depth to the water table in the subsoil should be at least 4 and, preferably, 6 ft. There will still be some rise of moisture unless sufficient irrigation water is applied to counterbalance it. Under these conditions, the rate of ascent will be greater in soils of higher clay content.

RECLAMATION OF SALINE AND ALKALI SOILS

The reclamation of a soil characterized as having a salinity problem only is less complex than if an alkali condition prevails. In either event, leaching to remove excess salts will be necessary in most instances. However, where the level of exchangeable Na is high, soluble Ca must be added prior to

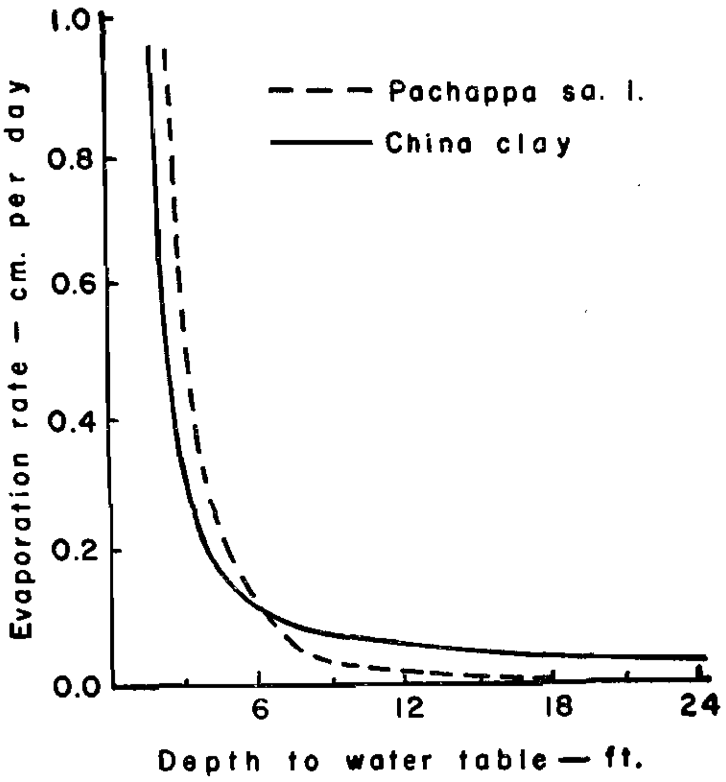

FIG. 12-2. Evaporation of water from Pachappa sandy loam and China clay as a function of depth to simulated water tables (after Gardner and Fireman, 1958).

leaching to encourage displacement of the Na. An amendment which will increase the level of soluble Ca in the soil is used for this purpose. Because of the expense involved, it is essential that the need for an amendment be ascertained positively prior to initiation of the reclamation programme. This is possible through use of relatively simple soil tests as outlined in the U. S. Department of Agriculture Handbook 60 (Richards, 1954).

Diagnosis and Reclamation Planning

The severity of a salinity problem is determined readily through use of conductivity measurements on saturation-paste extracts. If it is apparent that salts alone constitute the problem, reclamation requires only removal of these salts through leaching. Their total quantity is of little consequence in determining the total amount of water necessary to effect satisfactory reclamation.

Problems in salted soils are considered most serious when it is shown that an amendment is needed to return the soil to a normal, productive state. Usually this condition is associated with alkali soils having levels of exchangeable Na great enough to impair plant growth. Among other things, proper treatment requires the determination of the amount of Ca which must be supplied through use of an amendment to reduce exchangeable Na to an essentially harmless level. Preferably, no more than 10 to 15 percent of the exchange positions of a soil should be occupied by Na.

When classified on the basis of their Ca requirement, alkali soils may be placed into three categories, namely, (1) those containing soluble Na_2CO_3, (2) those containing precipitated gypsum (*gypsiferous* alkali soils), and (3) those in which exchangeable Na alone is the major source of alkalinity. In assessing the need for an amendment, the presence of either free Na_2CO_3 or gypsum must be taken into account. The quantity of Na_2CO_3 in a soil is determined by acid titration of an aliquot of the saturation-paste extract. The total milliequivalents of Na_2CO_3 plus exchangeable Na per unit quantity of soil indicates the amount of Ca needed.

Gypsiferous alkali soils are recognized by means of a separate determination for gypsum in the soil sample. Often the amount of Ca that may be dissolved from the naturally-occurring gypsum is adequate to displace completely the exchangeable Na present. When this is the case, the soil will respond to leaching essentially as though it were a saline soil. If neither Na_2CO_3 nor gypsum are present in an alkali soil in significant quantities, the Ca required for reclamation will be equivalent to the quantity of exchangeable Na which should be displaced.

A recently devised *gypsum requirement test* (McGeorge and Breazeale, 1951 ; Schoonover, 1952 ; Richards, 1954, p. 50) simplifies greatly the analytical procedures for determining amendment needs. The test is performed by mixing a small soil sample with a relatively large volume of saturated gypsum solution and measuring the Ca lost from solution after reaction with the soil. Sodium salts in an alkali soil are so diluted by this treatment that virtually complete displacement of exchangeable Na by Ca from the gypsum solution occurs. The Ca adsorbed, when calculated on the basis of tons of $CaSO_4 . 2H_2O$ per 2 million lbs. of soil, expresses the gypsum requirement. When an amendment other than gypsum is used, the amount necessary may be calculated by correcting the gypsum requirement value with an appropriate conversion factor (Richards, 1954, p. 49).

The gypsum requirement test works well if either Na_2CO_3 or gypsum are present in the soil. Sodium carbonate will precipitate Ca from the gypsum solution and will cause an equivalent increase in the gypsum requirement value. The same reaction would occur following treatment of the soil in the field, of course. If the soil under test contains indigenous gypsum, the test solution will remain saturated with respect to this salt even though some Ca may be utilized in displacing exchangeable Na. Calcium salts more soluble than gypsum, if present in the soil sample, will increase the amount of soluble Ca in the test solution and will result in a calculated gypsum requirement value which is negative. It is apparent from these considerations that a major advantage of the gypsum requirement test rests with its capability to account for all factors affecting the uptake of Ca by an alkali soil, and with but a single and relatively simple laboratory determination.

Saline Soil Reclamation

Steps deemed necessary for the satisfactory removal of excess neutral salts are as follows:

1. *Establish Adequate Drainage*—In diagnosing the problem of an area, it will have first been determined if a water table is present, if it lies sufficiently near the surface to be a current source of trouble, and if it can be lowered by improving

drainage. Should the ground water be within 3 to 4 feet of the surface of the soil, then it must be assumed that salts carried down by leaching eventually will return in damaging quantities. If a reclamation process is to have a permanent effect, the water table should be at least 4 to 6 feet below the surface. The optimum depth will be determined by the amount of leaching water available to counterbalance that rising by capillarity from the water table. The lower the quantity of irrigation water available the greater should be the depth to the water table.

2. *Level the Land*—Most of the agricultural land in the Indus-Ganges plains area is sufficiently level so that little additional land preparation is necessary prior to the initiation of leaching. Reclamation requires that the land be level so that the surface of the soil can be covered with water. The fact that much of the area is suitable for rice culture makes it also adaptable to leaching operations.

3. *Leach*—Sufficient water must be passed down through the soil to dissolve most of the soluble salts and carry them into the subsoil and away with the drainage water. No specific recommendation can be made as to the absolute amount of water that should be added. It is probable that less than 12 inches will be of little value. Generally speaking, crop production will continue to improve as the amount of leaching water is increased to at least a total of 3 feet. Water lost by evaporation must be taken into account when determining the total amount of water entering the soil.

Alkali Soil Reclamation

From the standpoint of reclamation, an alkali soil may be considered as one in which the level of soluble Ca must be increased before the satisfactory removal of exchangeable Na is possible. The most effective method of increasing the level of soluble Ca is through the addition of an amendment. Gypsum, though not readily obtainable in some areas, appears to be the most logical amendment for use in the subcontinent.

For best results, gypsum should be of fine particle size and should be mixed thoroughly with the surface soil so that it will dissolve and saturate the soil solution relatively quickly.

As the Ca is released in soluble form, it will enter exchange positions and displace exchangeable Na. The displacement reaction will cause an increase in soluble Na, however ; and when the concentration of this ion becomes so great that it can compete equally with Ca for exchange positions, an exchange equilibrium will have been attained. This equilibrium state will be attained prior to complete displacement of all exchangeable Na. Therefore, to effect a more thorough displacement, it is necessary to leach out the accumulating soluble Na. The exchange reaction then will continue as more of residual, undissolved gypsum comes into solution.

The percolating solution carrying soluble Na away from the reaction site will be saturated with gypsum. As a consequence, all of the added amendment will not be utilized in displacing exchangeable Na. In fact, up to as much as 10% of added gypsum may be lost from the surface layer of soil to which it has been applied. If the alkali problem were the result of more tightly held K ions, the efficiency of utilization of added gypsum would be reduced still further over that observed for Na-affected alkali soils.

Elemental S may be employed also as an amendment, and is oxidized to H_2SO_4 after incorporation into the soil. Most of the acid produced will react with $CaCO_3$, if present, to form $CaSO_4$. The rate of oxidation varies with the size of the S particles, and even material of very fine mesh requires a number of weeks before its conversion to acid is essentially complete. Therefore, S should be applied to the soil well in advance of the initiation of leaching operations. A material which contains a large proportion of S granules smaller than 40-60 mesh is preferable.

Chemical Reactions in Reclamation

The reaction of gypsum with the soil is one of simple exchange between soluble Ca and exchangeable Na:

$$\boxed{\text{Clay}}\begin{matrix}\text{Na}\\ \text{Na}\end{matrix} + \text{CaSO}_4 = \boxed{\text{Clay}}\,\text{Ca} + \text{Na}_2\text{SO}_4 \qquad (12\text{-}2)$$

The sulfuric acid produced from oxidized sulfur may react with free $CaCO_3$ in the soil, or directly with exchangeable Na:

$$2S + 3O_2 + 2H_2O \rightarrow 2H_2SO_4 \qquad (12\text{-}3)$$

$$H_2SO_4 + CaCO_3 \rightarrow CaSO_4 + H_2CO_3 \qquad (12\text{-}4)$$

$$\text{or} \quad H_2SO_4 + \boxed{\text{Clay}}\begin{matrix}\text{Na}\\\text{Na}\end{matrix} = \boxed{\text{Clay}}\begin{matrix}\text{H}\\\text{H}\end{matrix} + Na_2SO_4 \qquad (12\text{-}5)$$

The pH of a calcareous, alkali soil reacted with gypsum, as in equation 12-2 above, will not be depressed below about 8·2. This is the approximate pH of a soil which has all exchange positions filled with Ca. Where S is used, the pH may be reduced appreciably below 8·2 if there is no free lime present to neutralize the acid that is formed. Most alkali soils have an abundant supply of $CaCO_3$, however.

It is recommended frequently that organic residues be added to assist in the reclamation of an alkali soil. Fibrous organic materials tend to keep the soil open and, through improved permeability, permit more rapid movement of water and air through the soil. In addition, the oxidation of carbon to CO_2 results in the formation of H_2CO_3. Though weak, this acid will increase the solubility of $CaCO_3$ and the supply of soluble Ca necessary for displacement of exchangeable Na. The release of CO_2 through the oxidation of native soil organic matter may be as important as that derived from added organic residues, however.

MANAGING SALINE AND ALKALI SOILS

If reclamation procedures cannot be undertaken, certain practices may be followed which will help in obtaining improved yields on salted soils. These procedures may also improve soil conditions or at least prevent them from becoming worse. Recommended practices are as follows:

1. Use more irrigation water than is necessary for growing crops in order to encourage the downward movement of soluble salts. This practice is pointless if the water table stands very near the surface of the soil, however.

2. Light applications of irrigation water will tend to move salts from the surface of the soil down into the root zone ; heavy irrigations will tend to move the salts on past the roots. Therefore, it is preferable to irrigate after reasonably long intervals with relatively large amounts of water.

3. Grow only crops which are tolerant of saline and alkali conditions. Inclusion of rice in a rotation appears to be a very worthwhile practice since the required flooding will encourage leaching (Overstreet and Schulz, 1958). Salt removal, however, will not be extensive if the soil has a low permeability because of a high clay content.

4. If the surface layer of soil throughout an entire field is saline to the point of damaging germinating seeds or young seedlings, proper row-planting of the seeds on elevated ridges is beneficial. Greene (1953) describes such a system as used in Egypt for planting cotton. As shown in Fig. 12-3, cotton is seeded in ridged rows and irrigation water is applied between the ridges. When the field is irrigated, part of water moves laterally from the furrow toward the centre of the ridge

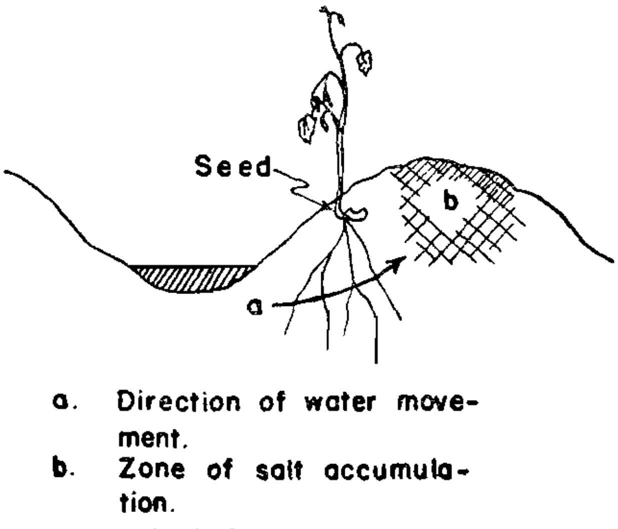

FIG. 12-3. Method of planting seeds on the sides of ridges in irrigated soils to minimize salt injury at the time of germination.

dissolving salts as it goes. Thus, after an irrigation, the soil between the furrow and ridge centre will have a reduced salt content. Seeds planted in this zone, and the tender seedlings they produce, will run less risk of being damaged by salts than if reduction in salt concentration were not possible. During the succeeding few weeks, when the land is hoed for weeds, the ridge soil is gradually moved over so that the plants occur finally on top of the ridge. By this time they are less susceptible to injury from salt.

Perhaps the most expedient method for obtaining some benefits from salted land is embodied in point 3, above, the growing of tolerant crops. A wide variety of plants have been classified on the basis of their salt tolerance. A listing of these is presented in Table 12-1. Many of the crops grown in the subcontinent will be found in this listing.

THE SALINITY AND ALKALI PROBLEM IN THE INDUS-GANGES BASIN

That waterlogging and salt accumulation constitute a major problem in the arid regions of the subcontinent does not need further emphasis here. Figures presented by Asghar (1952) indicate that, of the 12 million acres of land under irrigation in the former Punjab, about 230,000 acres are seriously affected by salts, 2·3 million acres moderately affected, and 4·6 million acres at least mildly affected. From 20,000 to 40,000 acres are said to be added to these categories each year.

Data concerning the specific properties of these soils is meagre. No indication can be given as to the relative proportions or total acreages of saline and saline-alkali or alkali soils that are present. Mehta (1940), in his classification of alluvial land in the Punjab, indicates that all types occur.

Most alkali soils are characterized as having the highest exchangeable Na percentage in the surface soil layer. That this is not always true in subcontinent soils is indicated by data of Hoon and Mehta (1937) and of Puri, Taylor, and Asghar (1935). Whereas most the soils they studied had a higher pH and percent exchangeable Na in the surface layers, there were profiles sampled in which the reverse was true. Many irregularities can be expected in soils of alluvial origin

Table 12-1. The relative tolerance of some crop plants to salts in the soil. For a more complete listing, see Richards, 1954, p. 67.

FRUIT CROPS	FIELD CROPS	FORAGES	VEGETABLES
Plants of high salt tolerance			
Date palm	*Barley* Sugar beets Rape Cotton	*Alkali sacaton* Saltgrass Nuttall alkaligrass Bermuda grass Canada wildrye Western wheatgrass Tall wheatgrass Birdsfoot trefoil	*Garden beets* Kale Asparagus Spinich
Plants of medium salt tolerance			
Pomegranate Fig Olive Grape Cantaloupe	Rye Wheat Oats Rice Sorghum Corn Flax Sunflower	Sweetclover Perennial ryegrass Strawberry clover Sudan grass Dallis grass Lucerne Tall fescue Orchard grass	Tomato Broccoli Cabbage Cauliflower Lettuce Sweet corn Potato Carrot Onion Peas Squash
Plants of low salt tolerance			
Pear Apple Citrus *Plum* Almond Apricot Peach	Field beans	White Dutch clover Meadow foxtail Alsike clover *Red clover* Ladino clover	Radish Celery Green beans

where normal development processes can be modified by frequent additions of fresh soil material during periods of flooding. Because of the varied sources of salts and methods of their accumulation, marked variation in saline and alkali soil properties may be expected.

Seepage from rivers and canals and the use of poor quality irrigation water contributes to the salted conditions in irrigated soils and the high water table beneath them. Roberts and

Singh (1951, pp. 179-182) point out the very important part also played by monsoon rains in raising the water table. According to their calculations, additions of moisture in the Punjab as rain may amount to $2\frac{1}{2}$ times that applied as irrigation water during the 3-month monsoon season (July-September). Although they recognize loss of water from canals through seepage they do not consider this in their calculations. That rainfall has a definite effect is seen readily in the rise of water in open wells during the late summer period.

The original source of seepage water contributing to the level of the water table will determine its chemical composition and, therefore, the effect it will have on the properties of the soils it affects. Because of the low salt content of river and canal water in the Indus-Ganges basin, land affected by it might not be expected to develop salted conditions rapidly. Yet, extremely high concentrations of salts sometimes are found in seepage areas adjacent to canals. Without suitable data, the reason for this cannot be given. Undoubtedly such conditions are partially the result of soil textural characteristics which permit rapid lateral movement of moisture from the canal. The effect may be accentuated by an initially high indigenous salt content of the affected soils and the presence of saline ground water, the level of which is raised by seepage water from the canal. Available data on the effect of canal seepage on salt accumulation in neighbouring land is virtually lacking. Preliminary observations made by the author indicate that seepage affects the land for relatively large distances from the canal.[1] This fact is recognized by the occurrence of deteriorated land parallel to canals in many locations. Nonetheless, one frequently observes only slightly affected land adjoining the canal even within an area which otherwise has become highly salinized apparently as a result of seepage.

Soils in the Indus-Ganges plains which lie high enough to avoid inundation by river water and monsoon floods are

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[1] Conductivities of extracts from surface soil samples as high as 60 millimhos per cm. and exchangeable Na percentages of up to 37 have been measured on soils from these seepage areas. The effect of seepage appeared to persist even up to one-quarter mile from the canal.

more often saline and only infrequently alkali. This is believed *due to the presence of a high proportion of Ca to Na in the* indigenous salts of the soil. Removal of the salts would return these soils to a normal state. In many instances they occur where the water table is relatively deep, 30 feet or more. Yet, extensive areas of soils are found containing essentially only neutral salts and a water table within 3 to 4 feet of the surface. Such areas must be considered as being isolated from an external source of salts (*i.e.* seepage water from rivers) and, because of the relatively recent rise in the ground-water level, conversion of these strongly saline soils to alkali soils has not taken place. The slow precipitation of the Ca by bicarbonate ions in the ground water may be expected to intensify the alkalinity problem with the passage of time, nevertheless.

A second condition is observed frequently in soils which have high levels of exchangeable Na but not necessarily in association with excessive salts. This is seen most often in low-lying areas plagued by at least occasional flooding. The latter circumstance suggests that direct river action or flood waters have been responsible for the accumulation of soil material and its subsequent inundation. Flood waters, by periodically covering the ground, serve to leach salts from the soil and minimize their accumulation. Yet, because of their low-lying position, these soils are or have been influenced by ground waters which contain sufficient bicarbonate to encourage the pronounced build-up of exchangeable Na.

A high proportion of the soils showing alkali characteristics are frequently high in clay also. The fine texture of these soils indicates that they have been laid down by slowly moving water, probably as a consequence of a decrease in velocity after spilling over river banks during periods of flood. Soil material deposited in this manner remains subject to frequent inundation. Its clayey, dispersed nature makes it difficult and often impractical to reclaim. This fact cannot be overemphasized. Soils of this type are classed as *bari* soils locally.

Much work needs to be done in the characterization of the alkali and saline soil problems in the subcontinent. A positive step is being made through the development of soil survey programmes. An important contribution of these programmes will be observations made on the properties of the deep subsoil, particularly as they relate to internal soil drainage.

The correction of existing salinity and alkalinity problems presents a major challenge to the agriculturist. At least three major obstacles must be overcome, namely, (1) the establishment of proper drainage and facilities for disposal of waste drainage waters, (2) the development of an adequate and economical source of gypsum, and (3) either the increase in the amount of irrigation water necessary for leaching operations, or development of more efficient methods of utilization of the present water supplies. If disposal of drainage waters into canals or streams were contemplated, a serious problem might arise in that the quality of the water for re-use may be seriously impaired. Eaton (1953) suggests that large areas be established where waste water can be allowed to accumulate. Excess water would be dissipated through evaporation. Such areas could be located on land already unsuitable for cultivation.

IRRIGATION WATER QUALITY

Poor quality irrigation water can have profound effects on soil properties. Saline waters may cause the rapid conversion of normal soils into ones with a high salt content. Prolonged use of many well waters may result in severely affected saline-alkali soils. However, waters of moderately poor quality can be used for long periods and without serious detrimental effects if proper methods of application are followed. The analysis of irrigation water provides a basis for making recommendations for its proper use.

DETERMINATION OF IRRIGATION WATER QUALITY

The quality of irrigation water is determined by its total salt content, the ratio of Na to the divalent cations Ca and Mg, and the quantity of HCO_3+CO_3 it contains. The standards of classification advocated by the U.S. Regional Salinity Laboratory (Richards, 1954) have rather wide acceptance and provide one of the better approaches to water quality evaluation.

As in salinity measurement of soil extracts, the salt content of irrigation water is estimated through use of conductivity measurements. The ion concentration of most waters is so low, however, that it is more convenient to express such measurements as micromhos per cm. ($EC \times 10^6$). Good quality water has a conductivity of 250 micromhos per cm. or less. A conductivity value above 750 micromhos per cm. is indicative of poor quality water and its use requires special irrigation management to prevent excessive salt accumulation in the soil. Ground waters in the subcontinent are sometimes observed to have conductivities of 2000 micromhos per cm. or more.

The higher the Na : Ca+Mg ratio of irrigation water the greater is the tendency for alkali soil formation. A frequently accepted limit is a soluble-Na percentage of 60. Since Na, Ca, and Mg usually constitute virtually all cations present, the two divalent cations will make up 40% or more of the total cation concentration in water of acceptable quality.

Acceptable limits of 'residual Na_2CO_3' in irrigation water, as calculated by the method of Eaton (1949), have been established tentatively by Wilcox *et al.* (1954). According to their observations, waters containing less than 1·25 me. per l. of 'residual Na_2CO_3' are considered acceptable, whereas values above 2·50 me. per l. are believed to be unsafe. Experiments by Haque (1958) indicate that, at medium to high salt contents (conductivities of 500 micromhos per cm. or more), any level of 'residual Na_2CO_3' may be harmful. Where there is a potential for complete precipitation of Ca and Mg in such waters, the large quantity of Na salts added can soon cause an appreciable reduction in the level of exchangeable Ca in the soil.

QUALITY OF IRRIGATION WATER IN THE SUBCONTINENT

The quality of irrigation water in the subcontinent has not been characterized very thoroughly. Some very useful information on water quality is provided by Eaton (1953) and Haque (1958). Rivers having their origin in the Himalayan uplands are low in total salts, seldom exceeding 300 micromhos in conductivity. Because they also contain some HCO_3 and CO_3, precipitation of Ca and Mg from the water after it has been applied to the soil may cause the effective Na concentration eventually to exceed two-thirds of the total soluble ions. At the same time, however, the loss of soluble Ca and Mg will reduce the conductivity of the water (Doneen, 1954) and, theoretically, resultant *effective-salinity* values will be as low as 100 micromhos or less. Proof that these waters are of relatively high quality rests with the fact that they have been used continuously for generations on well-drained land without obvious impairment of productivity.

Ground waters pumped from wells in the irrigated areas of the subcontinent are frequently very high in salts and in 'residual Na_2CO_3.' This requires that they be examined carefully prior to their use for irrigation of productive land.

USE OF INFERIOR IRRIGATION WATER

The application of poor quality irrigation water to soils can be recommended only with reservation. Circumstances do arise where the use of such water becomes desirable. However, if good land is to be spoiled by this procedure, it should be realized that it may be returned to a normal, productive state only at great expense.

We may draw conclusions as to how much salt a water can contain and still be used without disastrous effects to the soil. Water with a conductivity of 1000 micromhos per cm. (1 millimho), when applied to the soil, will create in that soil a salt level corresponding approximately to 1 millimho conductivity. This in itself is not recognized as a dangerous salt concentration. However, if additional irrigations are applied without removing residual salt by leaching, salts will accumulate roughly in proportion to the number of irrigations. Theoretically, four such irrigations would raise the salt content of the soil to a critical level.

Application of excess water each time the land is irrigated will displace residual salts downward and will minimize their accumulation. Thus, we may recognize the importance of proper management in the use of poor quality irrigation water. It may be concluded that waters having salts concentrations appreciably above the level just cited might be satisfactorily utilized where excess water is available for leaching and the Na to Ca+Mg balance in the water is suitable. The permeability of the soil must be taken into account, since this property will determine the feasibility of maintaining low salt concentrations through leaching.

The use of water with a high Na percentage or an appreciable 'residual-Na_2CO_3' content will cause soil deterioration regardless of management. Such water will increase the level of exchangeable Na in the soil. Whereas leaching minimizes the deleterious effects of these waters, they still may be expected to render the soil unfit for use, particularly if the conductivity of the water is high. Should the total salt content of the water be low, on the other hand, deleterious effects from use of the water may not develop even after long usage. As cited above, this is more or less proved by the continual use of river waters in the subcontinent which have a small amount of 'residual Na_2CO_3.'

If the salt content of water is not seriously high, it is possible to improve a poor Na : Ca+Mg balance through the addition of gypsum. The Ca to be added in gypsum should be equivalent to that necessary to precipitate any 'residual Na_2CO_3' plus an additional amount equal to two-thirds of the total Na present in the water. The resultant salt mixture will have a potential Na concentration of approximately 60 percent of the total soluble cations in the water.

If water is high in both Na and total salts, it is pointless to attempt to adjust the Na : Ca+Mg ratio with gypsum. The cost of the amendment will be excessive and only will add to the level of effective salinity of the water.

REVIEW QUESTIONS

1. Define saline, alkali, and saline-alkali in terms of the chemical analysis of soils. In general, what are the properties of soils falling into each of these three types ?
2. What errors are involved in expressing the salt content of soils on a percentage basis ? Why is use of conductance values preferable ?
3. Explain the advantage of using saturation-paste extracts over extracts made from more dilute soil-water suspensions in determining the salt content of soil.
4. Under what conditions may one expect a saline soil to develop ? An alkali or saline-alkali soil ? Why are arid climates normally more conducive to the formation of such soils ?
5. Why does the bicarbonate ion in ground water encourage the formation of an alkali soil ? What is meant by the term 'residual Na_2CO_3' ?
6. Recalling the influence of soil texture on the height of rise of water from a water table, explain why a salted condition may be expected in a fine-textured soil lying over a relatively deep water table (*i.e.*, 12-15 ft.) whereas none may occur in a sandy soil in which free water occurs at the same depth. From which soil would salts be removed most readily ?
7. In the conduct of the gypsum requirement test, Ca is not lost from the saturated gypsum test solution in proportion to the amount of exchangeable Na if the soil contains considerable exchangeable K or soluble Na_2CO_3. Why is this ?
8. Explain why a gypsiferous soil may have a relatively high level of exchangeable Na and still behave like a saline soil.
9. What is the difference in reclamation procedures applied to saline soils as compared to those used on alkali soils ?
10. A tubewell water is found to contain 3 me. Na and 20 me./l of Ca+Mg. Would you consider this suitable for irrigation ? Why ?
11. What determines if an irrigation water is a 'residual-Na_2CO_3' water ? Why is such water so effective in producing an alkali soil ?
12. Why will excessive irrigation prevent salts accumulating as rapidly as will occur where scanty irrigations are used ?

13. An irrigation water has a moderate level of salts and 'residual Na_2CO_3'. Explain how the heavy use of this water and the application of gypsum, either to the soil or in the water, can permit its continued use on a well-drained soil.
14. Why is it impractical to treat a water high in salts with gypsum in order to improve its Na : Ca+Mg balance ?

SELECTED REFERENCES

1. Kelley, W. P. (1951) *Alkali Soils.* Reinhold Publishing Corp., New York, U. S. A.

2. Luthin, James N., Editor (1957) *Drainage of Agricultural Lands.* The American Society of Agronomy, Madison, Wisc., U. S. A.

3. Thorne, D. W. and Peterson, H. B. (1954) *Irrigated Soils.* Blakiston Co., New York, U. S. A.

CHAPTER THIRTEEN

THE DEVELOPMENT AND CLASSIFICATION OF SOILS

THE SOIL is a natural medium for plant growth which, in various forms, covers the surface of the earth. Soils within a localized area may appear similar to one another with variations so minor that it is difficult to distinguish between them. On a broader scale, however, as we observe changes from region to region, we recognize in soils characteristics which clearly appear to be related to the environment in which they are forming and to the parent rock from which they have been derived. This fact has allowed for the identification of soils as distinct bodies, differing from one another as dictated by conditions under which they are developing. This recognition of soils as distinct bodies is one of the most valuable advances that has been made in the field of soil science. Even though soils change slowly through the ages, their characteristics used in classification are static from a practical point of view.

Consideration of soils as distinct bodies which result from certain developmental processes has led to the evolution of the science of *pedology*. The pedologist views the soil in the light of its chemical and physical properties and measures these properties in fundamental chemical and physical units. Limits are established whereby soils may be characterized and separated into specific groups according to the measured properties. Thus, a basis for classification is established ; one in which an almost infinite number of different soils may be placed into a relatively few groups delineated according to specific, though gradually varying, physical, and chemical characteristics. Many such properties may be identified visually in the field, whereas others must be measured in the laboratory. All can be related to some extent to the productive capacities of the different soils, a fact which constitutes the principal reason for developing an organized system of soil classification.

Discernible properties of the soil in the field may vary gradually or quite abruptly. In the first instance, one may move across the landscape for great distances without observing particular changes in the soil. In other instances, properties such as texture, soil depth, salt content, etc., may vary so rapidly that it is impossible to separate soils within the area under study into distinct groups. To consider a soil as a separate and identifiable unit requires that it cover a broad enough area of land to make its characterization practical. Included in the determination of its characteristics is its relationship to plants growing on it (Cline, 1949*a*). Once the existence of an individual soil has been recognized, it is described on the basis of its distribution over the landscape (soil mapping), the physical and chemical properties of its profile (soil classification), and its capacity for producing crops under a sustained agricultural programme (land-use classification).

The Classification of Soil Parent Material

The soil material (or parent material) of mineral soils results from the physical and chemical weathering of rocks as discussed in Chapter 4. Normally, weathering and soil formation proceed simultaneously. However, for purposes of discussion, they may be described as distinct or separate processes.

Rocks may weather solely in the place of their origin. Small amounts of the weathering products will be lost by slow, natural (*geologic*) erosion or solution and leaching effects. Much of the material remains, however, and by specific formation processes is gradually converted into a characteristic soil. A soil developing under this set of conditions is said to be forming on *residual* soil material.

Many soils have developed on transported and re-deposited materials. Four agencies of movement are responsible, namely, wind, water, ice, and gravity. The mode of transportation and deposition serves as a basis for the classification of such materials as follows:

1. Wind-transported materials result in *aeolian* deposits.
2. Transportation by and deposition in water are responsible for *alluvial*, *marine*, and *lacustrine* deposits.

Gravelly glaciated deposits yield soils often low in fertility and which present difficult problems in cultivation and management.

Blow sands of aeolian origin are common to the Thar and Thal Deserts. Because of their highly porous nature and lack of stability, they result in inferior soils.

Highly eroded, deep loessial soils of the Potwar Plateau. When properly cared for, loessial soils are among the best in the world.

Loess in the making. Pre-monsoon dust storms in the arid regions of the Asian subcontinent result in the continuing deposition of aeolian soil materials.

Plate 4

3. Movement and deposition by ice flows results in *glacial* deposits. Glacial deposits are noted for their lack of uniformity in texture, which may range from clay to large stones intimately intermixed.
4. The movement by gravity results in *colluvial* deposits.

Examples of aeolian materials are *sand dunes* and *loess*. Sand dunes are composed predominantly of medium to fine sand. This is because winds are usually of such a velocity that coarse grains cannot be picked up and transported effectively and, therefore, remain at or near the point of origin of the dune material. The smaller particles, normally of very fine sand or silt size, will be carried to more distant places and distributed over wider areas than the dune sand. These materials constitute loess. Deep deposits of loess occur in the Potwar Plateau of the Northern Punjab.

Water-transported materials may be deposited to form alluvial deposits in the flood-plain areas adjoining rivers. The texture of the resulting soil material depends upon the velocity of flow of the water at the time of deposition. Thus, sands, being difficult to maintain in suspension, are found generally nearer their point of origin. They may occur as pronounced accumulations along stream banks where restrictions to flow, such as bends in the stream channel, cause a temporary decrease in velocity. The finer-textured particles, on the other hand, can be carried great distances and may be deposited only upon reaching the sea. However, when rivers are in flood and spill over their banks, the decrease in velocity permits the deposition of the clays and silts. Alluvial deposits in much of the Asian subcontinent are derived at such times.

The accumulation of sediments in the sea (marine deposits) or in lakes (lacustrine deposits) also occur as common types of soil material. The components of these sediments may have been transported to the site of deposition either by water or wind action.

Glacial deposits are restricted in the subcontinent to the mountainous areas. However, glaciation has had a very important secondary effect on the composition of soils in lower-lying areas. Much of the fine-textured material in these latter locations has been derived through erosion of previously glaciated deposits from the high mountains. Stony soil materials as far south as the Salt Range are thought to result from accumulations of materials imbedded in and carried down by floating blocks of ice during periods of glaciation in the distant past.

Colluvial deposits which result from landslides are of minor importance except in mountainous regions. Colluvial-alluvial activity is dominant in some of the narrow inter-mountain valleys, however. Temporary streams formed by torrential rains carry large quantities of relatively coarse particles, partly colluvial in origin, through these narrow passageways. Emergence upon flatter areas, with the resultant decrease in water velocity, causes deposition of transported materials as *alluvial fans.* These formations are an important source of soil material in mountainous regions.

The Soil Profile

Geologic strata underlying the surface crust of soil may be characterized by a uniformity over large distances. Yet, over the same surface area, soils can differ considerably. Their differences are recognized in the characteristics of the *profile,* the vertical component of a soil body. Measured in units of depth, the profile may be considered as the sequence and appearance of the layers or horizons making up the body of the soil. It includes the underlying parent material. The number of observable profile characteristics are many. Some may be observed in the field, whereas others must be studied in the laboratory. For survey purposes, suitable differentiation among soils may be made normally on the basis of field observation alone ; however, for classification, other properties often need to be determined.

The letters A, B, and C have been used for many years to designate the horizons of the soil profile. Formerly, the A, or surface horizon, represented the *zone of eluviation* ; that is, the horizon from which materials were leached by percolating waters. The B horizon, immediately beneath the A, represented the zone of accumulation, or *illuvial horizon.* Modern connotation, however, no longer recognizes the eluviation-illuviation process as the sole basis for differentiation between the A and B horizons. Other features not particularly correlated with water percolation are frequently responsible for identification of these two horizons. Together they form the *solum*.

The C horizon is immediately beneath the B horizon and is considered by many to consist of slightly altered parent material similar to the material from which the A and B horizons have developed. It is difficult to determine experimentally whether the parent material of all horizons within a profile is of identical origin and of the same initial composition. Reasonable proof comes from mineralogical techniques applied in the laboratory which permit the determination of the distribution within the profile of the very highly resistant soil minerals (Marshall and Jeffires, 1946). Most of soils have not developed from uniform soil materials.

The A, B, and C horizons may be further subdivided to designate observable differences within the individual horizons. Three major subdivisions are generally noted, as A_1, A_2, and A_3. In certain instances, further subdivision is desirable depending on the variability within the horizon, the ability of the observer to pick out differences present, and the need for such differentiation. Subdividing the A_2 horizon, for example, into two subhorizons would be indicated by the expressions, A_{21} and A_{22}.

Segregation of the profile into the A, B, and C horizons usually follows the scheme outlined in Fig. 13-1. Further subdivision will depend almost entirely on local characteristics within the horizons themselves.

It is not necessary that all horizons or subhorizons occur in every soil. For instance, there may be no A_2 horizon, no B_1, etc. In some buried soils, there may be a duplication of horizons at different depths in the profile.

The Factors of Soil Formation

It is well recognized that soils developing from a geologic material often inherit some of the properties of the original material. However, other factors contribute to these properties also. The course of weathering of rocks and minerals is affected noticeably by climate and vegetation.

FIG. 13-1. Generalized plan of the soil profile (*after* Smith and Moodie, 1955).

	A_{00}	Undecomposed plant parts.
	A_0	Partially decomposed plant parts.
A	A_1	Horizon of maximum humus accumulation and biological activity, best development of a crumb structure. Commonly very thick in Chernozems and Prairie soils, thin or absent in Podzols.
	A_2	Minimum silicate clay percentage, minimum humus, and crumb structure. May have lost Fe, Al, or humus either to the B horizon or to ground water. Structural units weakly developed. Most pronounced in Podzols, Planosols, and Solodized-Solonetz groups. Zone of maximum eluviation.
	A_3	Horizon of transition, but in colour, texture, and structure, is more like A than B horizon.
B	B_1	Horizon of transition more like B than A. Commonly has a weakly developed nutty to prismatic or blocky structure.
	B_2	Horizon of maximum accumulation of clay, iron, alumunium, or humus, depending on mobility of individual components. Maximum development of blocky, prismatic, or subangular blocky structure.
	B_3	Horizon of transition more like B than C. Structure may be massive, if fine textured, or single grained if sandy.
C	C_1	Slightly altered soil-forming material.
	C_2	Soil-forming material of variable depth.

Other major horizons:

D	Unconforming strata underlying the B or C horizons (*i.e.* a layer of composition differing from that of the remainder of the profile).
G	Horizon formed under intense reducing conditions (*i.e.* under water); may have neutral grey, olive, green, or blue colours due to reduced state of minerals, a glei layer.
O	Horizon or stratum consisting of peat or muck.

These, in turn, may be conditioned by relief, or lay of the land. Furthermore, the longer the period of soil development the more distinct its effects will be. Summarily, the nature of soil-forming processes is controlled by five *soil-forming factors*—parent material, climate, living organisms (including vegetation), relief, and time. Man, by altering conditions for development, in certain instances may be considered as a modifying factor in soil formation.

Time

With the passage of time, features such as leached horizons or those in which organic matter or clay accumulate appear in a soil material, and are indicative of a developing soil profile. The longer the period during which the soil-forming processes operate the more distinct the profile features become. Thus, for a given set of conditions, the stage of maturity of a soil may be described by the extent of its development.

When the age of geologic deposits is known, the effect of absolute time on development may be measured quantitatively. Deposits left by retreating glaciers provide suitable media for such measurements. Utilizing such deposits in Alaska where the ages of various formations left by a receding glacier are known, Chandler (1943) was able to relate profile characteristics to the length of time of soil development. He observed that distinct A_0 and A horizons resulted from organic matter accumulating over a period of about 250 years. Eluviation-illuviation processes, on the other hand, needed a period of about 1000 years to produce leached A_2 horizons and B horizons containing accumulated sesquioxides which were characteristic features of the normal soils in the region. These soils have developed in a relatively cool, humid climate. Reaction rates and soil development would be expected to proceed at a more rapid rate under tropical conditions.

Parent Material

Byers *et al.* (1938) state that the initial step in the development of soil is the formation of parent material as the result of rock weathering. The parent rock, then, serves as a relatively inert storehouse of future soil material. Where a soil is forming on deep alluvial or aeolian deposits, the parent rock in reality occurs in an unconsolidated state, and will respond to the soil-forming processes in a somewhat different manner than would solid rock material. While the most readily observed characteristics of a soil forming from either type of material may be the result of changes brought about by climatic and biological effects, often they are traceable directly to the properties of the parent material itself.

Soil profile descriptions usually show the C horizon to consist of slightly altered parent material. Thus, it must be assumed that geologic weathering of the parent material continues after soil profile development commences. It is logical, of course, that the A and B horizons of the profile will be subjected to more extensive weathering than the C horizon. Under arid conditions, where the lower limit of the B horizon corresponds to the mean depth of moisture penetration, the C horizon escapes weathering for the most part.

Weathering of rock may be so complete that soil developing from it bears no resemblance to the original material. The tendency in weathering is the production of more nearly stable products which remain after the removal of soluble components. These residues (the various layered silicate minerals, quartz grains, and iron and aluminium hydroxides) will be relatively uniform in appearance and composition regardless of the type of parent rock from which they are derived. Thus, it may be said that parent material has little influence on soil characteristics if the weathering processes are intense and of sufficient duration. However, the extent of soils over the earth's surface which do not acquire a significant part of their readily discernible characteristics from the parent material is definitely limited.

It is where weathering is not so intense that the parent material contributes materially to the characteristics of the final soil. Sandstones, quartzites, and other rocks high in quartz impart a sandy texture to the profile. Shales and slates, composed primarily of clay, result in fine-textured soils. In addition to textural effects, the colour and mineral composition of the parent rock may influence the colour and nutrient status of the resulting soil.

In the Asian subcontinent, the inter-relationship among parent material, climate, and soil may be observed as one travels from west to east, or from areas of low to high rainfall. Soils in the Indus Basin resemble closely the original alluvial parent materials which persist virtually unchanged, texturally and chemically, as the result of a low intensity of weathering. Soils of the Deccan Plateau in Central India are characteristically dark and fine textured, the result of their having developed from dark-coloured, fine-grained rocks. Although the degree of weathering of the parent rock is moderate to strong, characteristics attributable to the original material occur in the resulting soils. Farther east and in southern India, intense weathering has minimized the influence of the parent rock on final soil characteristics. Here, even where there is considerable variation in the parent materials, soils developed under similar conditions of climate and relief consist of highly weathered products and often tend to be quite similar to each other.

Climate

Climate not only exerts a direct influence on the chemical and physical processes of soil development but modifies the other soil-forming factors as well, time excepted. All of the various components of climate—precipitation, temperature, wind, etc.—play some role in the formation of soils. Precipitation affects the soil moisture content, aeration, and leaching. Temperature governs the speed of chemical and biological reactions, and reduces the effectiveness of rainfall by increasing evaporation in warmer climates. In the latter respect, Russell (1937) points out that increases in evaporation resulting from a rise in mean annual temperature of 1°F. reduces the amount of water percolating through the soil by from ·75 to ·88 inches.

Muckinhern *et al.* (1949) discuss the indirect effect of climate on soil formation consequent to the storing of energy in organisms (*i.e.*, plants). The energy released during decomposition of plant residues affects various soil reactions and processes. Furthermore, organic acids from decomposing organic matter increase markedly the destructive effect of percolating waters on soil minerals.

Various relationships between climate and soil properties may be noted. Generally, higher temperatures are associated with an increased content of clay in the profile (Jenny, 1935). Organic matter accumulation, on the other hand, increases with increasing precipitation and decreasing mean temperature. Muckinhern *et al.* (1949) generalize that soil aggregation is enhanced by climatic changes from arid to sub-humid regions. Further increase in mean precipitation, however, is associated with a reduction in aggregation.

The relationship between macro- and microclimates must be kept in mind. The macroclimate is characterized by normal weather data. Generalized soil groupings are associated with macroclimatic data. The microclimate, which may be considered as localized climate or *effective* climate, causes variation within these broader soil groups related to macroclimate. Textural changes in soil materials and accompanying variation in water absorptive capacity (or reduced runoff) alters the effectiveness of precipitation and, therefore, the soil development processes.

Relief

The configuration of the land surface, or its topography, modifies the soil formation processes both directly and indirectly. The influence of relief comes mainly from the secondary effect it has on both the macroclimate and microclimate.

Over relatively short distances, abrupt changes in climate will occur where there is a major variation in elevation. The upward deflection and cooling of moisture-laden winds often accounts for the unusually high rainfall in mountainous regions. Furthermore, because of lower temperatures, loss of water through evaporation is reduced. Thus, a greater proportion of the precipitation falling on land at higher elevations may enter into weathering and soil development reactions. Tamura *et al.* (1953) have studied soil development in Hawaii and describe soils which in a relatively short distance change properties greatly as the result of a decrease in precipitation from over 150 inches at high elevations to less than 15 inches at sea level.

Soil temperature is affected by the direction of slope of land surfaces in that south-facing slopes in the northern hemisphere are warmer. This may or may not have an influence on the moisture relationships in the soil. However, it is common for the cooler, north-facing slopes to support perennial plants of relatively high moisture requirement, whereas, in the same locality, southern exposures are covered with more drouth resistant grasses.

Topography also affects the climate of the soil where it influences surface and internal soil drainage. Waterlogging may occur in low-lying, flat areas where removal of excess water is difficult. This is well known as a major problem to agriculture in the Indus and Ganges River Basins. The nearly flat topography of these broad river valleys also permits the inundation of large areas of land during periods of flood, causing the frequent deposition of alluvial materials to the land surface and more extensive leaching than would result from normal rainfall.

As opposed to the problem of removal of excess water from low-lying, flat areas, excessive runoff and erosion plague more steeply sloping areas. Not only will excessive runoff reduce the effectiveness of precipitation, but, also, the removal of topsoil by erosion tends to maintain the soils in an immature, undeveloped state. Relief has a direct bearing upon soil formation under such conditions.

Living Organisms

All living organisms associated with the soils—plants, animals, and microbes—are active factors in soil formation. Most important in this respect is natural vegetation which develops in type as dictated by soil and climatic conditions. Thus, vegetation itself does not influence independently the soil-forming processes.

Growing plants return to the soil surface nutrients extracted from the subsoil. This modifies the effect of leaching and tends to maintain a high level of base-forming cations in the surface soil horizons. In addition, interception and scattering of rain drops by above-ground plant parts and reduction in the rate of water flow over the soil surface reduces erosion. Shade afforded by the plants reduces evaporative loss of water, and thereby increases further the effectiveness of the natural precipitation.

Vegetation makes a major contribution to soil formation as a source of organic matter. The physical form of the dominant plant species dictates the distribution of organic residues in the soil profile. Grasses, on the one hand, produce a profuse root system and result in characteristic profiles in which organic matter is found throughout a thick surface horizon. Forest vegetation, on the other hand, adds organic residues predominantly as leaf litter which accumulates on the soil surface rather than under it. Organic matter from the litter is incorporated into the profile primarily by being carried downward in percolating water or by burrowing animals and insects.

The litter from different types of trees possess variable qualities which reflect in the intensity of leaching and weathering in the profile. Generally, the litter in coniferous forests is lower in bases than that from deciduous trees. Thus, the organic matter from conifers results in strongly acid leaching waters which cause a rapid and more thorough decomposition of minerals and loss of plant nutrients. Chandler (1939), studying soils under three forest types in New York, showed marked differences in the pH of accumulated organic matter on the soil surface resulting from different species of trees. The pH values for the A_0 horizons under spruce (coniferous), spruce-hardwood, and hardwood (deciduous) trees, respectively, were 3·45, 3·74, and 5·56.

Organic matter as a constituent of the soil increases the water-holding capacity, encourages soil aggregation, and improves permeability to water. The greater efficiency of absorption and movement of moisture will influence both chemical and physical phases of the soil development processes.

As is emphasized by Muckinhern *et al.* (1949), abrupt changes from one vegetation type to another frequently indicate major changes in such soil properties as texture, water-holding capacity, etc. These changes may also be correlated with variations in relief. Similarly, the natural occurrence of various types of salt-loving plants is usually closely correlated with salinity conditions in the soil. Since salts in the soil or their effects on soil properties are not always obvious, vegetation types frequently serve as an index to salted conditions.

In the more arid regions, plants may modify the soil development processes by absorbing and reducing the amount of water which would otherwise percolate through the soil. This is of little consequence in humid regions where, after plant requirements for moisture are met, there is still sufficient water to pass completely through the profile.

Rodents, earthworms, and other animals and insects, as a part of the soil biota, assist in the decomposition of organic deposits, and mix the surface layer of soil through burrowing activities. Microorganisms also are important in soil formation because of the part they play in the decomposition of organic matter. Furthermore, nitrogen-fixing organisms, both symbiotic and nonsymbiotic, assist in sustaining growth of vegetation at a much higher level than would be possible otherwise on soil materials low in available nitrogen. As a consequence, their presence may result in the addition of significant quantities of organic residues to a forming soil.

BASIC TRANSFORMATIONS IN SOIL FORMATION

In a strict sense, we might delimit geologic deposits to those which have remained unaltered since their original formation as a result of solidification from cooling magma. This definition, however, would be unacceptable to the geologist in many instances. Widespread deposits of sedimentary and metamorphic materials are considered to be of a geologic nature, and yet they result from major alteration of original parent rock. Nonetheless, those of an unconsolidated nature will, where they occur as a surface crust, support plant life and, for this reason, fit as well into the concept of a soil as they do into a geologic system of classification. Circumstances of this type indicate the difficult and arbitrary nature of separation of soils into a distinct category with unquestionable limits with respect to apparent properties.

The majority of the world's soils are on transported materials. In reality, the development of these soils commences prior to the deposition of the materials in their present location. While the initial transformation of the parent rock could have been largely physical (*i.e.*, comminution), extensive chemical weathering may have occurred also. Customarily, however, soil formation is viewed as a series of transformations affecting parent material in place. Soil profile characteristics are attributable directly to these transformations in many instances. On the other hand, characteristics of the original deposit (*i.e.*, texture, textural variation or layering, mineral composition, etc.) also influence strongly the features of the resultant soil.

Aside from effects traceable to inherent properties of the parent material and its deposition and layering, transformations responsible for the development of characteristic soil profiles are as follows (Smith and Moodie, 1955) :

1. Rock and mineral weathering.
2. Humus formation and organic matter accumulation.
3. Ion exchange and the movement and precipitation of salts.
4. Movement (eluviation) and accumulation (illuviation) of colloids within the profile.
5. Mechanical shifting and transfer of material within the profile by organisms.

The extent to which these transformations will have taken place in soil formation will be a function of the combined influences of the five soil-forming factors. For example, the first four transformations listed above will proceed most rapidly under conditions of relatively high moisture and temperature. The extent of change will be dependent upon the length of time during which they have been taking place. Young soils, or those which retain great similarity to the original parent material, will have undergone these transformations to but a very limited extent.

CLIMATE AS A BASIS FOR SOIL CLASSIFICATION[1]

Except where abrupt changes in elevation occur, the climates of the world are noted to change only gradually over wide areas. Since vegetation and climate are very closely associated, we may expect vegetation to change gradually with the climate. The rather close relationship between these two factors is shown in Fig. 13-2. The diagram illustrates the fact that plant types capable of producing abundant vegetation become dominant as the precipitation increases. The same trend in vegetation is noted in the change from cold to hot climates.

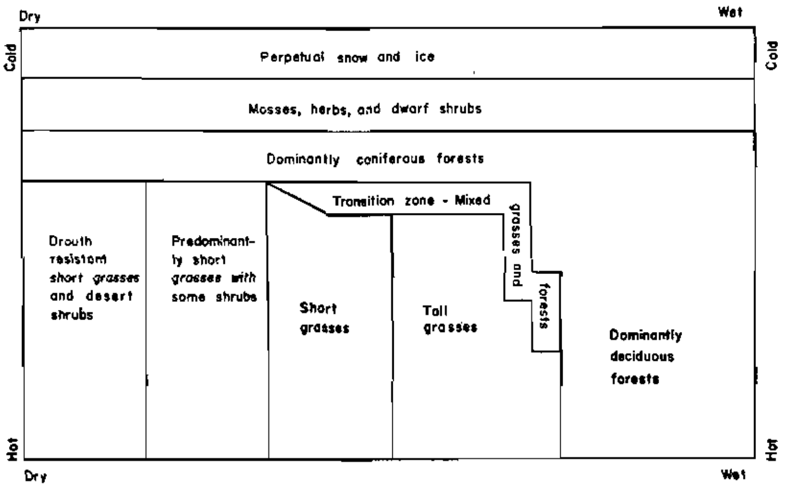

FIG. 13-2. The generalized relationship between climate and type of vegetation.

As discussed previously, climate and vegetation are capable of exerting a strong influence on the pattern of soil development. Where parent materials and topography do not vary excessively, rather broad areas of similar soils are found to occur within specific climatic and vegetational zones. Thus, as these climatic and vegetational zones are distinguished one from

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[1] The system of classification herein described is but one currently in use. It corresponds to the system which has been developed and followed for the past half century in the United States. Much of the basic work leading to the formulation of this system is attributable to Russian and European soil scientists as well as to those in the United States.

the other, so may we separate into zones the soils developing within them. Such a distribution is depicted in Fig. 13-3.

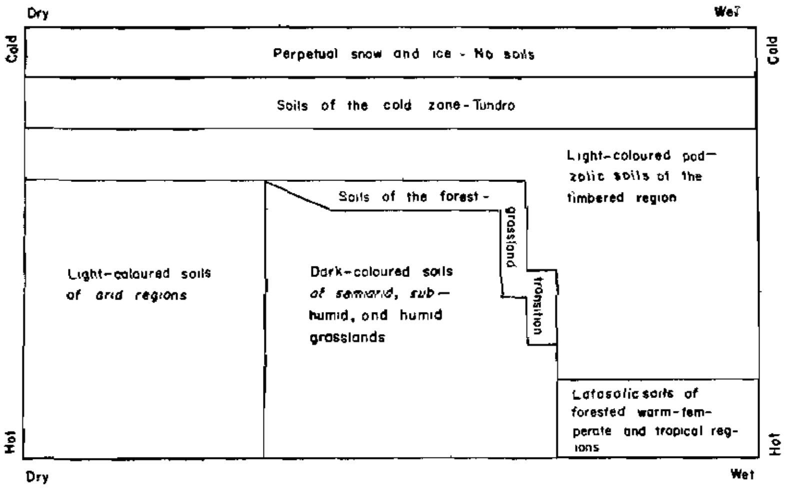

FIG. 13-3. *A diagrammatic representation of the relationship* between climatic change and variation in some of the major properties of soils.

Here soil zones are superimposed upon the same climatic base as is used in Fig. 13-2. Because the properties of these soils are associated very closely with the climate of the zone in which they are forming, they are known as *zonal* soils.

All soils within a specific climatic zone do not reflect the effect of the mean climate in a similar manner. Those which do are considered *normal* for the area. This raises a question, of course, as to which ones do represent the specific influence of the prevailing climate. Also, if some soils do not, why is it that they do not ? In answer, it must be stated that, again, an arbitrary base is used in defining a normal soil. As the climate for the area is described as average with respect to precipitation, temperature, etc., so must be the other factors of soil formation. Thus, topography, parent material, and vegetation must be average for the area. Furthermore, the time of development must have been sufficiently long to have resulted in soil properties which represent a state of near equilibrium with the prevailing environment. Environment, in turn, is expressed by conditions occurring within the developing soil profile and, thus, reflect the combined influences of the soil-forming factors.

By definition, normal or average conditions of soil development include gently sloping land where surface and internal drainage are adequate, and erosion, though slight, is present. Since parent materials composed of siliceous rock are the most common, the normal soil usually will develop from such matter. Where mineral weathering is limited by arid climatic conditions, parent materials of normal soils logically would be one of the transported types.

Soils which are forming under conditions other than those accepted as normal cannot be classed as zonal soils. Two additional groups, then, are recognized as encompassing these abnormal soils. They are the *intrazonal* and *azonal* categories. Intrazonal soils are those which result from major modification of climatic effects as a consequence of abnormalities in topography or parent material. For instance, as opposed to gently sloping land which loses a moderate portion of precipitation falling upon it as run off, level land or that of a depressional nature will retain virtually all of the rain or snow received. As a result, vegetation there usually is more typical of that common to areas of higher rainfall, and water available for weathering and leaching is greater than that acquired by associated normal soils. These conditions result in a different effective climate and will, obviously, increase the rate of transformation responsible for soil development. Similarly, intrazonal soils form in waterlogged areas of arid regions since the more nearly constant source of moisture causes profoundly different consequences than are observed to occur in well-drained soils. Unusual parent materials also can result in the formation of intrazonal soils. One may conclude from these facts that intrazonal soils are located without respect to any particular normal, zonal soil boundary. Like the zonal soils, however, they tend to be at near equilibrium with the environment in which they persist.

The third or azonal group of soils is comprised of those so young that they have undergone only a minimum of transformation. Generally, azonal soils have resulted merely as the consequence of deposition of previously comminuted parent material. Soils maintaining their immaturity as the result of continual removal of surface layers through erosion are azonal also. These soils show no effect of the local prevailing climate or vegetation on profile features.

Determining whether or not a soil is zonal, intrazonal, or azonal is an initial step in classification which results in placing a soil within its particular *order*. Soils within the zonal or intrazonal orders may be separated further into *suborders*. Among the zonal soils, this separation is on a climatic and vegetational base. Six distinct zonal suborders have been recognized and are shown in Table 13-1. The relationship between these suborders and climate is indicated in Fig. 13-3, and, by comparing this diagram with Fig. 13-2, the relationship between these suborders and vegetational type can be seen also.

Three suborders occur within the intrazonal soil category. These include the imperfectly drained soils of both humid and arid regions and soils developing on abnormal parent materials. Since characteristics of azonal soils reflect essentially only the effects of parent material, they are not divided into suborders.

Although soils which go to make up an individual suborder have a limit to the extent of their variation, changes among them are of sufficient magnitude to allow for appreciably greater subdivision than is permitted by the suborder grouping. Subdivisions of the suborders are the *great soil groups* which are shown also in Table 13-1.

Whereas all great soil groups within a suborder are grouped together largely as the result of their occurrence under a limited range in climate, separation among them is primarily on the basis of specific differences in profile characteristics. Characteristics of the horizons, including their number, thickness, colour, texture, and presence or absence of lime are the principal bases for separation among the great soil groups.

As the suborders of zonal soils are related to climate, so are the great soil groups which constitute them. The

Table 13-1. The Higher Categories of a Soil Classification System.

Order	Suborder	Great Soil Group
Zonal	1. Soil of the cold zone.	Tundra soils
	2. Light-coloured soils of arid regions (Desertic)	Gray Desert soils Red Desert soils Sierozem soils Brown soils Reddish Brown soils
	3. Dark-coloured soils of semiarid, subhumid, and humid grasslands. (Chernozemic)	Chestnut soils Reddish Chestnut soils Chernozem soils Prairie soils (Brunigra) Reddish Prairie soils
	4. Soils of the forest-grassland transition	Degraded Chernozem Noncalcic Brown or Shantung Brown soils
	5. Light-coloured podzolic soils of the timbered region (Podzolic)	Podzol soils Gray Wooded or Gray Podzolic soils Brown Podzolic soils Gray-Brown Podzolic soils Red-Yellow Podzolic soils
	6. Soils of forested warm-temperate and tropical regions (Latosolic)	Reddish-brown Latosolic soils Yellowish-Brown Latosolic soils Latosols
Intrazonal	1. Halomorphic (saline and alkali) soils of imperfectly drained arid regions and littoral deposits.	Solonchak or Saline soils Solonetz soils Soloth soils

Table 13-1. (Continued)

Order	*Suborder*	*Great Soil Group*
Intrazonal	2. Hydromorphic soils of marshes, swamps, seep areas, and flats	Humic Glei soils (includes Wiesenboden) Alpine Meadow soils Bog soils Half-Bog soils Low-Humic Glei soils Planosols Ground-Water Podzol soils Ground-Water Latosol soils
	3. Calcimorphic soils	Brown Forest soils (Braunerde) Rendzina soils Grumusols
Azonal		Lithosols Regosols (includes Dry Sands) Alluvial soils

relationship between the zonal great soil groups and climate is shown diagrammatically in Fig. 13-4.

Separation of intrazonal and azonal soils into great soil groups does not follow a general pattern as is noted above for the zonal soils. The basis for these separations will become apparent from the individual discussions of these groups in later sections of this chapter.

THE CLASSIFICATION OF ZONAL SOILS

The principal influence of climate on soil development is attributable to the moisture effect. Water is essential to at least four of the five transformations normally involved in developmental processes. Temperature effects, which modify

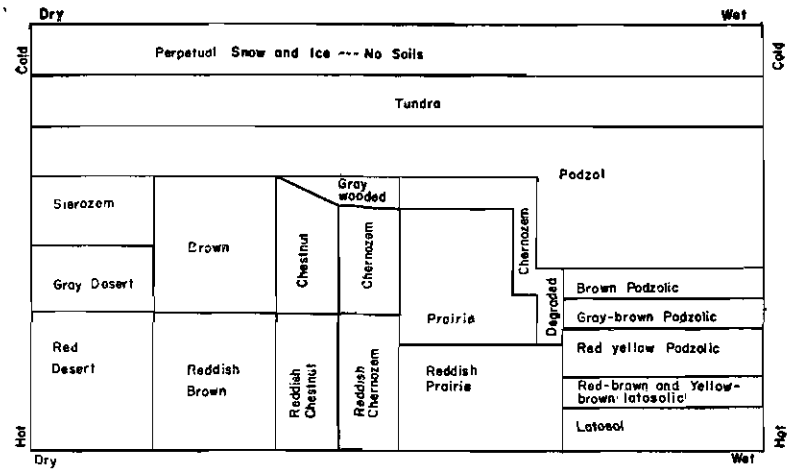

FIG. 13-4. The relationship among the zonal great soil groups shown on a climatic base (after Smith and Moodie, 1955).

various of the chemical, physical, and biological functions attending soil development, also play an important role. The type of vegetation associated with a particular soil is controlled largely by effective moisture and temperature. Thus, these latter two factors are both directly and indirectly responsible for the course followed in the gradual alteration of raw or partially transformed parent material into soil.

As might be expected from the above considerations, major changes in soil characteristics occur along transects of varying effective precipitation. Furthermore, for any given moisture condition, variation in soil properties are brought about also by a change in mean temperature. The most significant temperature effects are noted within regions of high rainfall. In our inspection of zonal soil characteristics, we will attempt to learn the various relationships between gross climates and some specific zonal soil properties. For facility in discussion, these zonal soils will be considered as members of four separate groups with characteristics reflecting the influence of significantly different climatic conditions.

ZONAL SOIL DEVELOPMENT UNDER LOW TO MODERATE RAINFALL

Soils formed under normal conditions of topography and drainage in regions of low rainfall have profile characteristics which reflect clearly the effects of limited leaching and low biotic activity on soil development. Where precipitation is at a maximum for this climatic range, readily soluble salts will be lost from the profile. Calcium and magnesium carbonate, while they may be redistributed within the profile, are not removed from it completely. This pattern of leaching is associated generally with climates where the mean annual precipitation is moderate to low, *i.e.* from 2 or 3 up to no more than 35 in. per year. The soils at equilibrium with the environment are characteristic of desertic and chernozemic categories (Table 13-1). Normal vegetation varies from tall grasses, where rainfall is the highest, through the short grasses, and finally to the drouth-resistant shrubs, grasses, and other plants of the deserts. The frequency and distribution of precipitation throughout the year will have some bearing on the equilibrium vegetation.

As is common to the development of all zonal soils, growing plants contribute materially to profile characteristics of soils in arid regions. They serve a major function as media for the recirculation of bases within the soil profile (*base cycling*). The process involves the absorption of bases from the subsoil, their transport to above-ground plant parts, and eventual return to the surface soil upon decay of the plant. Thus, base cycling tends to counter leaching effects even though they are characteristically limited under conditions of low rainfall. An important consequence of this is that the level of available plant nutrients in the surface soil is elevated appreciably above that of the original soil material. The maintenance of high levels of Ca and Mg causes their precipitation as carbonates at some point within the profile.[1]

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[1] The process of soil development wherein the soil exchange complex remains essentially staturated with divalent cations and free lime occurs as a common constituent of the profile is referred to as *calcification* (Byers *et al.*, 1938). Prairie soils, while included in the chernozemic suborder, are subjected to somewhat greater leaching than are the chernozem soils. As a result, they have lime-free profiles.

One of the more outstanding features of soils developing under conditions of limited precipitation is the variation in colouration due to the characteristic distribution of organic matter within the profile. The effect, which is associated with grass vegetation, becomes most apparent under high effective precipitation, and results in a pronounced darkening of the profile throughout the zone of root development. The distribution of organic matter in four great soil groups typical of the desertic and chernozemic suborders is shown diagrammatically in Fig. 13-5. Here the effects of increasing precipitation on organic matter content and distribution within the profile are obvious.

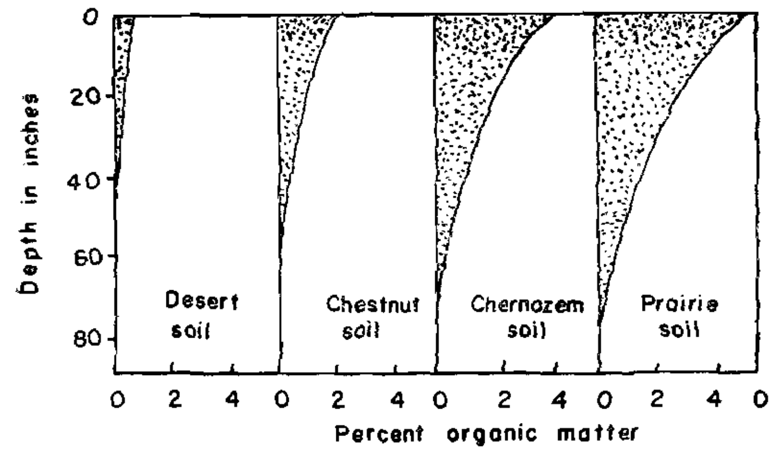

FIG. 13-5. The approximate distribution of organic matter in profiles of several of the great soil groups developed under conditions of low rainfall (after Smith and Moodie, 1955).

A number of physical and chemical characteristics are common to many of the soils developing under conditions of limited rainfall. A number of these are as follows:

1. The colloid is dominated by bases, primarily calcium.
2. The soil reaction is near or above neutral in most horizons, with slight acidity often characterizing the surface horizons.

3. The chemical and mineralogical makeup of the colloid fraction is relatively uniform throughout the profile, except where variations in the composition of the soil-forming material prevent this. Movement of colloids by percolating water is limited, probably because of the high state of aggregation due to the abundance of calcium ions.
4. A zone of calcium carbonate and magnesium carbonate frequently develops. This zone is thought to coincide with the depth to which percolating water most frequently penetrates.
5. Soil organic matter accumulates throughout virtually the entire profile.

Light-Coloured Soils of the Arid Regions (Desertic Soils)

In well-drained desertic soils, movement of water into the profile seldom is beyond a depth of 4 feet. Vegetation is sparse, limited by the low and infrequent rainfall to fast-growing, short-lived grasses or drouth-resisting plants and shrubs. This condition leads to a return of only a minimum of plant residues to the soil. The surface layers of the soil profiles are not darkened appreciably by organic matter, therefore.

Weathering of desertic soil mineral matter is confined primarily to physical processes which result in generally coarse-textured materials. The limited quantity of water available restricts both the chemical break-down of mineral structures and the removal of weathering products from the profile. Percolation usually is adequate to carry out relatively soluble salts ; however, slightly soluble salts which dissolve at the surface may be precipitated at some lower depth when the concentration of these salts in the soil solution is increased through water removal by plant roots. This latter process accounts for accumulations of lime and gypsum in subsoil layers[1] of many soils in the arid and subhumid regions.

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[1] Where lime and gypsum accumulation is a predominant means for the formation of B horizons in soils, the resulting soil layers are designated frequently as B_{3ca} or B_{3cs} horizons. The symbol, cs, infers the presence of gypsum (calcium sulfate). In more humid regions where lime still remains in the profile but the illuviation of clay is the principle means of B-horizon formation, the lime then may be considered to have accumulated in the parent material or C horizon. Such horizons are commonly designated by the symbol C_{ca}.

Where both salts occur in the profile, gypsum normally will be found at a greater depth than lime because of its greater solubility.

Although B_2 horizons high in clay are formed in humid-region soils as a result of the eluviation-illuviation process, percolation in desertic soils usually is inadequate to cause this effect. Nonetheless, faint textural and, therefore, structural B_2 horizons sometimes are observed in desert-region soils. Nikiforoff (1937) attributes the increase in clay under these circumstances to be the result of weathering and clay formation in place and not the consequence of clay migration downward. Layering of alluvial or aeolian materials also could result in a variation in soil texture within soil profiles, and this might be interpreted erroneously as being caused by the movement of clay particles in percolating water.

An increase in the mean annual precipitation within regions of desertic soils causes characteristics typical of B horizons to become more pronounced. At the same time, the A horizon darkens due to an increase in plant growth and the amount of plant residue returned, and the depth to lime or gypsum layers increases gradually. An increase in average temperature, on the other hand, causes soil colours to change from the grays or browns in cooler areas to reddish or reddish browns in warmer climates. The great soil groups falling within the desertic suborder are *gray desert*, *sierozem*, *red desert*, *brown*, and *reddish brown*. They are differentiated primarily on change in appearance caused by variation in climate as illustrated in Fig. 13-4.

The most outstanding features of the profiles of the light-coloured soils of the arid regions are as follows:

A_1	Surface horizon only slightly darkened by organic matter accumulation. Gray colour in cold, dry zones, becoming more red with temperature increase and brownish with moisture increase. Salts and lime generally absent. Colour of parent material may have a dominating influence.

A_3-B_1	Difficultly discernible horizon of transition between A_1 and B_2 horizon. No A_2, or eluviated horizon, is in evidence.
B_2	Identification of B_2 horizons usually on the basis of structure (prismatic or blocky), or occasionally on colour. Structure more pronounced with increased rainfall where a higher content of clay may occur.
B_3ca	Zone of lime accumulation, or lime and gypsum. The depth of deposition for either salt will increase with increasing precipitation, and the gypsum layer, where present, will occur beneath the lime.
C	Slightly weathered parent material, usually calcareous.

Dark-Coloured Soils of Semiarid, Subhumid, and Humid Grasslands (Chernozemic Soils)

With increasing precipitation over that characteristic for the arid regions, the influence of climate overshadows many of the effects of parent material in the formation of soil. The biotic factor, particularly vegetation, exerts a much greater influence on profile characteristics. The vegetation gradually changes from the true desert types, first to bunch grass, then to short, and finally to tall grasses. But the nature of the transition in vegetational type depends upon whether or not the bulk of the precipitation occurs during the growing season. Grasses which are characterized by dense root systems that extend deep into the subsoil are potent factors in the development of deep, dark soil profiles. In areas where the highest effective precipitation occurs, soils may become somewhat acid and lime zones may be absent from the profile.

Five great soil groups occur in this climatic range: *chestnut*, *reddish chestnut*, *chernozem*, *prairie* (*brunigra*), and *reddish prairie*. The existence of a *reddish chernozem* does not appear to be well established, but there is little reason why it should not occur.

Profile characteristics for the suborder are rather well defined. With increasing precipitation they become more pronounced as follows:

1. The colour of the surface soil changes from dark brown to black due to the greater accumulation of organic matter.

2. The thickness of horizons increases, especially the A horizon, and the effect of organic matter on profile colour extends to greater depths in the profile.
3. Structure becomes more apparent ; crumb or granular in the A, and blocky or prismatic in the B horizon. Textural profiles, probably due to development of clay in place, do occur.
4. The depth to a lime layer increases until, in the prairie soils, the layer disappears. Only traces of soluble salts are found. Calcium and magnesium dominate the exchange complex.
5. The reaction of the surface soil varies from near neutral in the drier zones to slightly acid in the prairie soils.

A typical profile to cover all members of this suborder cannot be shown logically. However, principal features of the profiles within the suborder of chernozemic soils are as follows :

A_1	Dark brown to black colour depending on effective precipitation. Thickness will increase with increasing precipitation, and at a more rapid rate in the cooler climates where organic matter is not oxidized so readily. Granular to crumb structure.
A_3-B_1	Difficulty discernible and of varying thickness. Transition zone. No A_2 evident.
B_2	Slight to moderate development of a textural B_2 with increasing rainfall, especially in the reddish soils of the warmer areas. Clay accumulation probably due to formation in place. Most prominent feature is the increase in structural development.
B_3ca	Lime layer present in the chestnut and chernozem soils only.
C	Parent material.

ZONAL SOIL DEVELOPMENT IN HUMID, TEMPERATE CLIMATES

Soil development in humid, temperate regions occurs normally under forest vegetation. The dominant transformation involved is mineral weathering. Considerable eluviation takes place with an appreciable portion of the eluviated materials accumulating in the subsoil. A large part of the weathering products are lost from the profile through leaching also. Organic matter contributed by the forest vegetation tends to remain as a strongly acid deposit on the surface of the soil. This has a profound influence on the pattern of weathering and soil development within the underlying materials.

Because of the strongly acid nature of the layer of accumulating organic materials, percolating water passing through it acquires exceptional solvent properties. The solubility of Fe and Al in the upper mineral layers is enhanced by this fact. Furthermore, certain organic fractions derived from the leaf and needle litter appear capable of protecting these metallic ions against precipitation, and permits their transfer out of the surface horizon by the percolating waters (Stobbe and Wright, 1959). A residue of silica, probably precipitated after release from weathering minerals but also residual as silica grains resistant to weathering, remains to form a whitish A_2 horizon (*bleicherde*). Organic matter and exchangeable base-forming cations also are removed from the surface layers. Thus, the exchange complex there tends to become saturated with H and Al ions.

The deposition of eluviated materials in the B horizon follows a rather specific pattern. Organic matter accumulates immediately beneath the bleached layer, along with iron and aluminium. With increasing depth, the profile then grades into a compact horizon of decreased organic matter content but one which remains high in sesquioxides, particularly iron. Often a horizon with a higher content of clay occurs immediately beneath the sesquioxide layer. This probably results from the recombination of eluviated weathering products to form clay-sized, layered silicate minerals.

The weathering of the geologic materials under these conditions may occur to great depths, particularly where permeability to water and air is high. The transformations involved proceed at a relatively rapid rate and lead to the evolution of *podzolic*[1] soils. The more outstanding characteristics of these soils include the presence of an acid organic layer at or on the soil surface, a bleached A_2 horizon, and a B horizon subdivided into distinct though overlapping organic, sesquioxide, and clay layers. The profile is normally low in exchangeable bases and high in exchangeable H and Al.

The typical podzol profile has the following characteristics:

A_0	Normally thick, strongly acid, mor-type organic layer.
A_2	Highly leached gray to white horizon (true bleicherde), coarse textured, low exchange capacity. Thickness will vary greatly over short distances.
B_{21}	Upper portion of the B_2; forms a distinct horizon because of accumulation of organic matter. Sesquioxides are also plentiful, but colour is masked by organic matter. Texture will be finer than that of the A_2 horizon.
B_{22}	Distinguished from the B_{21} horizon by decreased level of organic matter and the greater prominence of sesquioxides. May be a compact, cemented layer due to the sesquioxides.
B_{23}	This horizon, when present, is distinctive because of an increase in clay content over the B_{22} horizon.
B_3	Transitional. May or may not be present. Boundaries irregular.
C	Parent material, highly weathered, normally acid.

The true podzol is the mature soil of the humid, temperate to cool regions. As one progresses from a podzol zone to areas of like temperature but lower rainfall, the change in soils is toward the prairie and chernozem types. An increase in temperature but with similarly high precipitation, on the other hand, causes a change in soils toward the latosolic types which will be discussed shortly. As temperatures decrease, the precipitation intercepted by the soil is less likely to be lost through evaporation, and, therefore, becomes more effective

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[1] The transformations leading to the development of podzolic soils are referred to in the literature as *podzolization* (Byers *et al.*, 1938).

in soil development processes. As a consequence, podzolic soils may also be found in areas of relatively low rainfall if the mean temperatures are quite cool.

Light-Coloured Podzolic Soils of the Timbered Region

In addition to the true *podzols*, four other great soil groups are described in this suborder. These are the *gray wooded* or *gray podzolic*, *brown podzolic*, *gray-brown podzolic*, and *red and yellow podzolic* groups. The relationship between these groups and climate is shown in Fig. 13-4.

On the cooler, drier side of a podzol soil zone, transition to the prairie or chernozem soil type is expected. Yet soils in these areas show the effects of coniferous forest vegetation, even under low rainfall. The upper horizons have but weak podzolic characteristics. Organic horizons and a relatively thick, gray A_2 horizon are present, however. The B_2 horizon displays the typical accumulation of organic matter and sesquioxides, but not to the extent noted in true podzols. If the parent material is calcareous, free lime may occur in the B_3 (or C_{ca}) horizon, suggesting limited leaching. Soils having these characteristics belong to the great soil group of *gray wooded* or *gray podzolic* soils.

According to Cline (1949b) the gray-brown podzolic, brown podzolic, and podzol soils form a series representing the progressive advance in soil developmental processes under a humid, temperate climate. The profile of the gray-brown podzolic soil has an A horizon containing largely *mull*[1] humus, and a textural B horizon characterized by an accumulation of mineral clay. The brown podzolic soil, which represents an advancement in the stage of development over the gray-brown podzolic soil, is more strongly acid, has a lower level of exchangeable bases, and a surface layer of *mor* humus. The B horizon has been depleted partially of accumulated clay, probably as the result of its breakdown in the more acid percolating water. This horizon is found to contain also a slight accumulation of organic matter and sesquioxides, a

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[1] Mull is well-decomposed organic matter mixed with the mineral soil material. Mor is a layer of undecomposed or slightly decomposed organic residue which lies on the surface of the mineral profile.

feature more typical of the true podzol. Cline (1949b) believes that representatives of these two great soil groups, along with the true podzols, may occur within the same climatic zone if there is a variation in the calcareous nature of the parent material. Thus, highly calcareous materials, which, when present, prolong base cycling, will result in the persistence of a soil with gray-brown podzolic features. In the same vicinity, brown podzolic and true podzol soils might occur on parent materials that are lower in lime or lime free in their initial state.

There appears to be a number of similarities between the gray podzolic soils of the drier, forested regions and the gray-brown podzolic soils. However, the gray podzolic soil is described as having a thicker A_2 horizon and as being less acid than is the typical gray-brown podzolic soil.

The gradual increase in temperature in humid zones results in the replacement of coniferous trees by deciduous types. The forest litter loses its strongly acid properties. Clay composed of silicate minerals appears to undergo considerable breakdown and disappearance from the soil. Iron oxides residual from the mineral breakdown impart a yellowish to reddish colour to the soil. The rapid oxidation of organic matter limits its accumulation. These conditions result in weakly developed podzolic soils which may be considered as transitory between podzolic soils of the cooler climates and the latosolic soils of the humid to superhumid tropics. They form the *red and yellow podzolic* soil groups. In spite of the theoretical differences in conditions under which they form, the distinction between gray-brown podzolic and yellow podzolic is not clear and has led to much confusion in classification (Tavernier and Smith, 1957). The current tendency in classification appears to be one in which the yellow podzolic soils are restricted to those forming on lime-free parent materials.

The surface horizons of red and yellow podzolic soils are thin, and do not appear to be particularly high in organic matter. The A_2 horizon is bleached but not white in colour. The B_2 horizon ranges from yellow to red and is higher in clay than the A_2 horizon. Subsoil horizons are highly weathered.

Soils of the Forest-Grassland Transition

Two great soil groups have been recognized in this suborder, the *degraded chernozem* and the *noncalcic brown* or *Shantung brown* soils. Both occur in the forest-grassland transition area in subhumid, temperate climates. The degraded chernozem appears where grasslands have slowly converted to forest conditions. Thus, it has been possible for the organic profile of the chernozem to appear superimposed upon a profile having the A_2-B_2 horizon sequence of the podzolic types. It is assumed that the chernozemic properties are slowly disappearing from these soils.

The noncalcic brown soils are known to occur both in the United States and China. They are found in subhumid regions where the climate fluctuates between wet and dry periods. The profiles do not vary markedly with depth, are slightly acid in reaction, and have a light reddish brown appearance. Unlike the brown soils of desertic suborder, they are noncalcareous. According to Tavernier and Smith (1957, p. 280) the profile characteristics of noncalcic brown soils are strikingly similar to those of gray-brown podzolic soils. The outstanding difference between the two is climate rather than profile properties, for the noncalcic brown soil occurs in areas where winter precipitation is common and summer periods are characteristically drouthy. If lime occurred in the profile of the noncalcic brown soil, it would become a normal brown soil.

ZONAL SOIL FORMATION IN FORESTED WARM-TEMPERATE AND TROPICAL REGIONS

Soil development within tropical and subtropical regions is the result largely of intense mineral weathering and extensive leaching. According to Byers *et al.* (1938), the weathering process is quite complete and leads eventually to the conversion of the initial rock entirely to silicic acid, and aluminum and iron hydroxides or their dehydration products. Some soils developing on these highly weathered materials appear to have characteristics of podzolic soils in that they have a surface layer high in organic matter, a somewhat bleached A_2 horizon, and a distinct B horizon. The B horizon, composed primarily of colloidal iron and aluminium hydroxides, is referred to specifically as the *laterite layer*. When exposed to air and allowed to dry, it becomes very hard. Soils containing a laterite layer are *latosols*[1], and those still in the apparent process of laterite formation are latosolic soils.

The principal components of highly weathered latosolic soils are colloidal, although, as in the laterite layer, considerable cementation of these colloids may have taken place. In addition to the abundant sesquioxides, these soils may contain kaolinitic clays in fairly high proportions. Both kaolinite and the sesquioxides normally have low cation-exchange capacities[2]. Furthermore, being high in sesquioxides, these soils display a high capacity for fixing phosphorous. It is little wonder, then, that crops grown on latosolic soils frequently respond markedly to the addition of plant nutrients.

In spite of the above-cited limitations, tropical soils often are relatively productive. Even though nutrient cations may not be held on the exchange complex in large quantities, their release from organic matter is rapid because of the very high rate of mineralization and large amounts of plant residues. Thus, base cycling, and the recirculation of other plant nutrients as well, are important aspects in the nutrition of plants in these soils. Generally, the proportion of exchangeable bases is high enough so that latosolic soils are only slightly to moderately acid in reaction.

The more outstanding features of latosolic soils reflects their intense weathering. A characteristic yellowish to red colour is imparted by the abundance of oxidized iron and aluminium compounds. The colloids, which make up a large part of the soil, have a low silica content and generally are well aggregated. They lack the properties of stickiness when wet and swelling upon hydration.

The development of soils in the humid, tropical regions is not well understood. Because of this, classification of the soils

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[1] Previously called *laterite* soils.

[2] Exceptions to this are cited for soils with relatively large exchange capacities which are high in hydrous Fe and Al oxides. Such soils have been observed in the Cook (New Zealand) and Hawaiian Islands (Fieldes *et al.*, 1952 ; Kanehiro and Sherman, 1956). Fieldes and his associates attribute the high exchange capacities in these soils principally to Al oxides.

is, as yet, quite incomplete. Relatively few detailed studies of latosolic soils have been made. A wide variety of profiles have been described, but, in general, classification has not been organized into the well-defined groupings as are recognized among soils of the temperate regions.

As a general rule, the more intense the weathering of a latosolic soil the higher is its sesquioxide content. Martin and Doyne (1927) recognized this fact and suggested that the stage of maturity of these soils could be expressed on the basis of the molecular ratio of SiO_2 and Al_2O_3 in the clay fraction. If the ratio were less than 1·33, the soil would be mature and should correspond to the latosol as recognized today. Those soils having a ratio between 1·33 and 2·00 would be termed latosolic according to current terminology. From this it may be seen that the lower the silica-sesquioxide ratio the higher is the degree of latosolic soil development. It should be pointed out, however, that a satisfactory classification on a morphological basis will not result from this method of chemical classification.

Three great soil groups occur in this suborder: *latosols*, and *reddish brown* and *yellowish brown latosolic* soils. No attempt is made to give specific profile characteristics. However, those properties common to all three groups are as follows (Kellogg, 1950):

1. Ratio of silica to sesquioxides in the clay fraction is low.
2. The cation-exchange capacities of the clays are low.
3. Only highly resistant primary minerals occur in any appreciable amount.
4. Soluble materials are at a minimum.
5. Aggregates are relatively stable.
6. The predominant soil colour is red.
7. Organic matter accumulation on the surface is relatively light.
8. Clay and sand are normally present in greater quantities than is silt.

SOILS OF THE COLD ZONES

The soils of the cold zones found in arctic regions and elsewhere at high elevations have not been described adequately. All are placed into a single great soil group as *tundra* soils. The process responsible for their formation is not understood to any great extent. Climate, relief (particularly as it affects drainage), and parent material all have an influence on the pattern of soil development.

Common to tundra soils are continuously frozen subsoils (permafrost). Failure of percolating water to penetrate into the subsoil creates a waterlogged condition which promotes lateral leaching and creates in the soil a low state of oxidation.

THE DEVELOPMENT AND CLASSIFICATION OF INTRAZONAL SOILS

The classification of intrazonal soils, as yet, is in a state of flux. Considerable change may be expected to take place within this area of classification in the future.

The principal suborders of intrazonal soils include those resulting from imperfect drainage and a high water table (*halomorphic* and *hydromorphic* soils), those influenced in such a way that they retain a high level of lime relative to the normal, zonal soils for the area in which they exist (*calcimorphic* soils), and those in which impervious layers occur either as the result of eluviation-illuviation processes or layering of the parent material (*planosols*).

HALOMORPHIC SOILS OF THE ARID REGIONS

All halomorphic soils have one characteristic in common; they are, or have been at some time in the past, affected by accumulations of salts within the soil profile. They include saline and alkali soils.

The initial stage of halomorphic soil formation is *salinization*. During this period, salts are introduced into the soil material normally as the consequence of a high water table. The resulting soil is classed as a morphological unit within the *solonchak* great soil group[1]. Except for the high levels of salts, the solonchak soil profile has no characteristics attributable to the salinization process. If the accumulating salts contain sufficient Na_2CO_3 to produce a strongly alkaline reaction, the soil is designated as an *alkali solonchak*. If the quantity of

[1] Known locally as *thur*.

exchangeable Na and salt were sufficient, the alkali solonchak would be classed as a saline-alkali soil.

An improvement in drainage with the resultant disappearance of the water table will result in the cessation of salt accumulation in the solonchak soil. Except where the effective precipitation is very low, reversal in the mean movement of water through the profile will tend to move the salts into the subsoil of the solonchak, and *desalinization* will occur. If the previously accumulated salts are high in soluble Ca, desalinization will return the soil to a condition which is more nearly normal for the area. If, on the other hand, the soil is an alkali solonchak, desalinization will result in the removal of excess salts but, during the early stages at least, the level of exchangeable Na will remain high. Particles of clay left in this state are easily dispersed. Because of this, they tend to migrate into the subsoil with percolating water and accumulate there to form a distinct textural B_2 horizon. This develops within the profile a layer of slow permeability and a solum of high pH. The soil, a product of *alkalization*, is categorized within the *solonetz*[1] great soil group.

The more notable features of the solonetz profile are:

A_1	Low organic matter content.
A_2	Leached, thin layer ; grayish colour. The reaction is near neutral and the salt content and exchangeable sodium are low.
B_2	Prismatic or columnar structure, the surfaces of the structural units being coated with a thin organic deposit which imparts a dark colour. Exchangeable Na and pH are high. $CaCO_3$ may be present, especially in the lower part of the horizon.
B_3	Moderately alkaline layer, salts may be present and lime is usually high.
C	Calcareous parent material with salts commonly present.

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[1] Desalinization and alkalization are referred to collectively as *solonization*, since together they are responsible for the formation of solonetzic soils. Within this concept, alkalization includes not only the establishment of a desalinized alkali soil, but also the ensuing formation of the textural B_2 horizon.

Subjection of a solonetz soil to continual leaching will lead to the eventual removal of exchangeable sodium and its conversion to a *soloth* soil. Basically, the process (*dealkalization*) depends upon the hydrolysis and replacement of exchangeable sodium by hydrogen ions. The characteristics of the structural profile of the solonetz become more pronounced as the clay continues to move downward. Replacement of the sodium with hydrogen, however, produces a weakly to moderately acid reaction in the surface soil, although the B horizon and parent material remain neutral or alkaline. Theoretically, at least, the reestablishment of normal vegetation on the soloth would, through base cycling, return this soil to one more nearly normal for the area.

Although solonchak soils are of common occurrence in the subcontinent, it appears that none with characteristics of the solonetz or soloth have been recognized as yet. So long as the broad, central alluvial plains in the subcontinent remain subject to frequent flooding and the presence of a high water table, development of the soils beyond the solonchak stage may not be expected in these areas.

In many instances, man can influence halomorphic soil formation. Reclamation of solonchak (saline) or alkali-solonchak (saline-alkali) soils serves as one example. The use of highly saline irrigation water, on the other hand, can cause the relatively rapid formation of valueless solonchak soils from otherwise productive land.

CALCIMORPHIC SOIL FORMATION

There are three great soil groups which occur in humid regions but which have profile characteristics more typical of soils common to areas of lower rainfall. They are the *calcimorphic* soils. They contain free lime in their profiles, the result either of highly calcareous parent materials, or soil and weather conditions which prevent the effective removal of lime from the soil even under a relatively high total annual rainfall.

Brown Forest Soils (Braunerde)[1]

Brown forest soils have as an outstanding feature an unusually high level of exchangeable bases when compared with the normal zonal soils in the area of their occurrence. This may result from conditions that are conducive to a low level of leaching, or the soil may have developed from highly calcareous parent material. Brown forest soils occur where the vegetation types are dominantly deciduous forests. The litter produced by the leaf fall is characteristically higher in bases than is the litter from coniferous trees. This fact also contributes to the maintenance of higher levels of Ca in the profile.

The brown forest soils have a moderately acid mull layer at the surface, much like the adjacent podzolic soils. However, the A horizon is deep and dark due to the accumulation of organic matter which has been incorporated into the surface layers. A brownish B horizon occurs but is apparently not the result of particularly high levels of organic matter, sesquioxides, or clay. The C horizon consists of highly calcareous parent material.

Rendzina Soils

Rendzina soils develop under grass vegetation on limestone parent materials which may have a high gypsum content. Soft rather than hard limestones are conducive to rendzina formation. The topography is normally hummocky.

The rendzina has a dark-coloured, relatively thin surface horizon (less than 18″) very high in organic matter. Exchangeable ions, adsorbed largely by the humus fraction, will be predominantly calcium or calcium and magnesium, depending on the nature of the parent material from which the soils are formed. The colloids, which occur as stable aggregates, are not subject to eluviation.

The surface layers of rendzina soils consist primarily of the relatively insoluble materials residual from the solution and

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[1] A large variety of soils throughout the world have been classed as brown forest soils, or braunerde, but only a few have the characteristics herein described. For a detailed discussion of these soils, see the article by Tavernier and Smith (1957).

leaching away of the original calcareous (or gypsiferous) parent material. By means of this process, organic matter and mineral particles of low solubility accumulate and the relatively thin layer formed may consist of residues from the weathering of several feet of limestone. The colour of the rendzina may correspond to that of chernozemic soils.

Grumusols

Grumusol is a tentative name applied by Oakes and Thorpe (1951) to black soils which, so far, have not fitted well under other classifications. Like the chernozem, they develop under grass or similar vegetation. However, they are associated with much warmer climatic conditions than the chernozem. A large area of this group of soils occurs in central India where the descriptive names of black earths, black cotton soils, or regur are commonly applied. The dark colour of regur may be attributable largely to parent material rather than to the accumulation of organic matter. Failure of these soils to develop characteristics more of a podzolic or latosolic nature may be the result of the excessive evaporation. According to Robinson (1949), even though the annual rate of precipitation is high, moisture loss from the surface prevents a complete leaching of lime and it accumulates as a zone of calcium carbonate concretions (*kankar*). The high clay content of these soils also will reduce leaching and attendant effects.

The profile of the grumusol reflects the effects of both climate and parent material. It has a deep, dark A horizon which grades slowly with depth into calcareous clay or basalt in the subsoil. The B horizon occurs primarily as a transition layer between the A and C horizons. *Ando* soils, which were described first in Japan, are like grumusol soils.

HYDROMORPHIC SOILS OF THE HUMID REGIONS

Gleization

Gleization refers to the process of soil formation occurring under waterlogged conditions in humid regions. The high precipitation and relatively low evaporation preclude the accumulation of salts. The normal gleization process requires, to varying extents, an alternation between complete saturation with water and an incompletely saturated, yet moist, state. Under these conditions chemical reduction occurs. The solubility and mobility of iron, manganese, calcium, and magnesium is increased. In the reduced state, the metallic ions, particularly iron, impart a bluish or greenish colour to the soil.

The ground water in a gleied soil is a dilute solution containing, among other things, iron in a reduced state. Contact between the ground water and the soil air will expose the iron in solution at the interface to a higher oxygen pressure bringing about its oxidation and precipitation. Seasonal changes in the ground water level allow for the accumulation of iron precipitates as a band corresponding to the limits of variation in the surface level of the water table. The bluish or greenish, compact horizon that develops is referred to as the *glei* (or *gley*) layer.

In the more temperate climates, gleization will result in the formation of ground-water podzols, soils similar in appearance to the normal podzol. In tropical regions where soil moisture conditions vary between wet and dry due to a fluctuating water table, ground-water latosols are formed. Such soils occur commonly in flat areas adjacent to rivers which can influence the ground-water level.

Peat Formation

The development of peat occurs under conditions of virtually continuous saturation with water. In its classical concept, peat formation takes place on the surface of submerged land where abundant growth of vegetation is possible. Death and submersion of plants slowly builds up a deposit of organic residues beneath the surface of the water. Lack of oxygen impedes decomposition and permits, in certain instances, relatively deep deposits to accumulate.

The nature and classification of peat will depend upon the material from which it forms. Sphagnum and other mosses are perhaps the predominant sources of residues, although trees, shrubs, and a variety of aquatic plants also make material contributions to them.

Except where natural phenomena have occurred to expose the deposits, drainage of peat areas is required before they can be used agriculturally. After drainage and exposure to air, oxidation of the peat beds will occur. Oxidation is encouraged by cultivation of the beds and results in the gradual subsidence of their surface.

Humic-Glei Soils

This group of soils includes the *wiesenboden* or *gleyboden* soils as described by Robinson (1949, p. 354). The humic-glei group has recently been described by Thorp and Smith (1949) as an intrazonal group of poorly drained to very poorly drained hydromorphic soils. They are composed of characteristically dark-coloured organic-mineral horizons of moderate thickness overlying mineral-glei horizons. They occur under either swamp-forest or herbaceous marsh vegetation mostly in humid and subhumid climates. A large proportion of these soils are medium acid to mildly alkaline in reaction. Few are strongly acid.

Robinson (1949) described three horizons in a typical wiesenboden: (1) a surface layer largely aerobic but sometimes absent, (2) a zone existing between limits of the fluctuating water table, and (3) the subsoil continuously waterlogged and anaerobic. The second horizon is the glei layer. Organic matter occurs primarily in the surface soil.

Low-Humic Glei Soils

This group of soils would, by definition, appear to lie somewhere between the humic-glei and half-bog type of soils. Conditions for development are nearly identical to the humic-glei group. Instead of a well-developed surface horizon, however, only a thin layer of mineral soil, high in organic matter, occurs above the gleied subsoil. The reaction of the profile is strongly acid, which corresponds more to the bog soils formed under coniferous forest than to humic-glei soils.

Alpine Meadow Soils

Soils of this group are limited to higher altitudes where conditions of poor drainage and alpine meadows exist. If timber is present, it is sparse. Topography is normally moderate to steeply sloping. Parent materials for alpine meadow soils are usually thin deposits of coarse-textured and stony, glacial, alluvial, or colluvial deposits. Leaching is intense and removal of bases results in an acid reaction. Chemical weathering, however, is limited by prevailing low mean temperatures.

The development of a profile in the alpine meadow soils is limited to the accumulation of humus in the surface soil (A_1 horizon) and the development of an incipient gleied layer in the subsoil. The alpine-humus soils, as described by Robinson (1949, p. 329), correspond to this group of soils.

Bog and Half-Bog Soils

These two groups of soils are similar in that they are characterized by the accumulation of *peat* or *muck* at the surface. In bog soils, however, peat comprises the subsoil, whereas in the half-bog soils, the organic surface layer is underlain with mineral matter affected by gleization. They differ from the humic-glei soils in that the surface layers of the latter types are admixed mineral and organic materials.

Peat is commonly used as a general term embracing all organic soils or soil material in which the organic matter content exceeds 20 percent. In a strict sense, however, peat refers to undecomposed or slightly decomposed organic deposits, the decay of which has been impaired by excessive moisture (anaerobic conditions). The colour may be quite light, sometimes very nearly the same as the original material from which the deposits are formed. *Muck*, as opposed to peat, is composed of organic residues that have undergone considerable decomposition. It is very dark in colour.

The word peat, as used in the European system of classifying organic soils, is based primarily on the type of vegetation that has contributed to the peat formation. Moisture conditions at the time of deposition contribute materially to the type of organic matter that is deposited. Three groupings occur:

(*a*) Low-moor or fen peat, which is the result of the accumulation of organic residues under water in lakes or wet basins.

(*b*) Transition moor formed from trees and shrubs normally growing on the surface formed by low moor.

(*c*) High-moor peat, formed from peat moss (sphagnum) and cotton grass (*Eriophorum*). Accumulation of these materials tends to be above free water surfaces, giving the basis for the name, high moor. Plant residues in high-moor peat are capable of retaining large amounts of moisture. Thus, even above the free water table, excessively wet conditions are maintained.

Any of the above three types of peat deposits may develop independently of one another if conditions permit. However, Robinson (1949) pictures them as occurring at a single location in a time sequence beginning with low-moor and ending with the high-moor types.

Ground-Water Podzols and Ground-Water Latosols

From outward appearance, the ground-water podzols and latosols are quite similar to the zonal podzols and latosols. The two ground-water types are also similar to one another in that they have bleached A_2 and compacted B_2 horizons high in sesquioxides. They are dissimilar, however, with respect to colour and nature of organic matter accumulations within the profile. These latter characteristics are influenced by the normal variation in climate and other factors of soil development occurring between regions of podzolic and latosolic soils. Enrichment of the B_2 horizons with sesquioxides of the ground-water podzolic and latosolic soils occurs as the result of gleization.

Planosols

According to Thorp and Smith (1949), planosols have been defined as soils with major and abrupt changes between horizons in the profile as the result of differences in the degree of cementation, compaction, or clay content. The planosol may include a wide variety of soils with characteristics which may or may not reflect primarily the effect of the more common transformations responsible for zonal soil development. Included within this group are stratified soils resulting from the layering of parent materials of different textures.

Topography has a dominant influence on the formation of some planosols. For instance, flat land surfaces or depressional areas will tend to retain more of the precipitation they receive than will the normal soils on gently sloping land. Where internal soil drainage is good, then, the increased effectiveness of water percolating through the profile causes the formation of more pronounced A_2 and B_2 horizons. The higher clay content of the B_2 horizon results either from the transfer of soluble weathering products from the surface with their reconstitution into layered silicate minerals in the subsoil, or from direct eluviation-illuviation of the clay particles. The resultant horizon normally is compact and tends to be impervious. These effects are more noticeable in regions of desertic and chernozemic soils where the precipitation is inadequate to create similar characteristics in the normal soils of the area.

Proposals have been made to apply the name, planosol, to a new suborder within the intrazonal category (Thorp and Smith, 1949). If this were done, other soils with indurated subsoil layers could be classified together and all would have this one very important profile characteristic in common. Such groups as the ground-water podzols and latosols and soloth soils then would fall within this suborder.

AZONAL SOILS

Excessive erosion on the steeper slopes or continual addition of aoelian or water-borne materials to existing deposits prevent the development of normal soil characteristics. Immaturity of the profile requires its classification under the azonal order. Azonal soils are characterized by incompletely developed profiles. Three groups are recognized:

(*a*) *Alluvial soils*, with an almost unlimited variation in profile texture, colour, and horizonal configuration.

(*b*) *Lithosols*, forming on weathering hard rock or rock fragments, and,

(*c*) *Regosols*, forming on unconsolidated rock, as loess.

A LIMITATION IN SOIL CLASSIFICATION ON A CLIMATIC BASE

The general system of classification described above has a major limitation caused by relating zonal soils to specific external climates when other environmental factors may modify climatic effects to a large extent. Strict adherence to this system prevents classifying together soils of essentially identical chemical and physical characteristics if they happen to occur in climates peculiar to different suborders. Thorp and Smith (1949) cite as an example certain soils which are intrazonal in humid regions but which have many properties common to zonal soils of drier areas. If these soils were classified on the basis of profile characteristics, they would occur within the same suborder. Such placement would connote their physical, chemical, and biological similarities to a much greater extent than does the current system of classification. There is a trend toward modification of the present system so that soils may be classified within the higher categories more along lines of soil properties.

SOILS OF THE ASIAN SUBCONTINENT

A very generalized soils map for the subcontinent is presented in Fig. 13-6. It is patterned largely after similar maps of Chatterji (1957) and Stewart (1947). Modifications consist primarily of deletions of areas whose descriptions relate only to soil texture or which are differentiated on the basis of the presence of lime in the soil profile.

Because of the very limited information on the genetic classification of soils in the Asian subcontinent, it is all but impossible to present an accurate map of their occurrence. The eleven areas shown in Fig. 13-6 are separations based to some extent on differences in physiography. Aside from limited personal examination, conclusions as to great soil groups which occur within each of these various zones depend on the assumed influence of climate and on limited descriptions of soils found in the literature. Each zone is considered briefly in the following paragraphs.

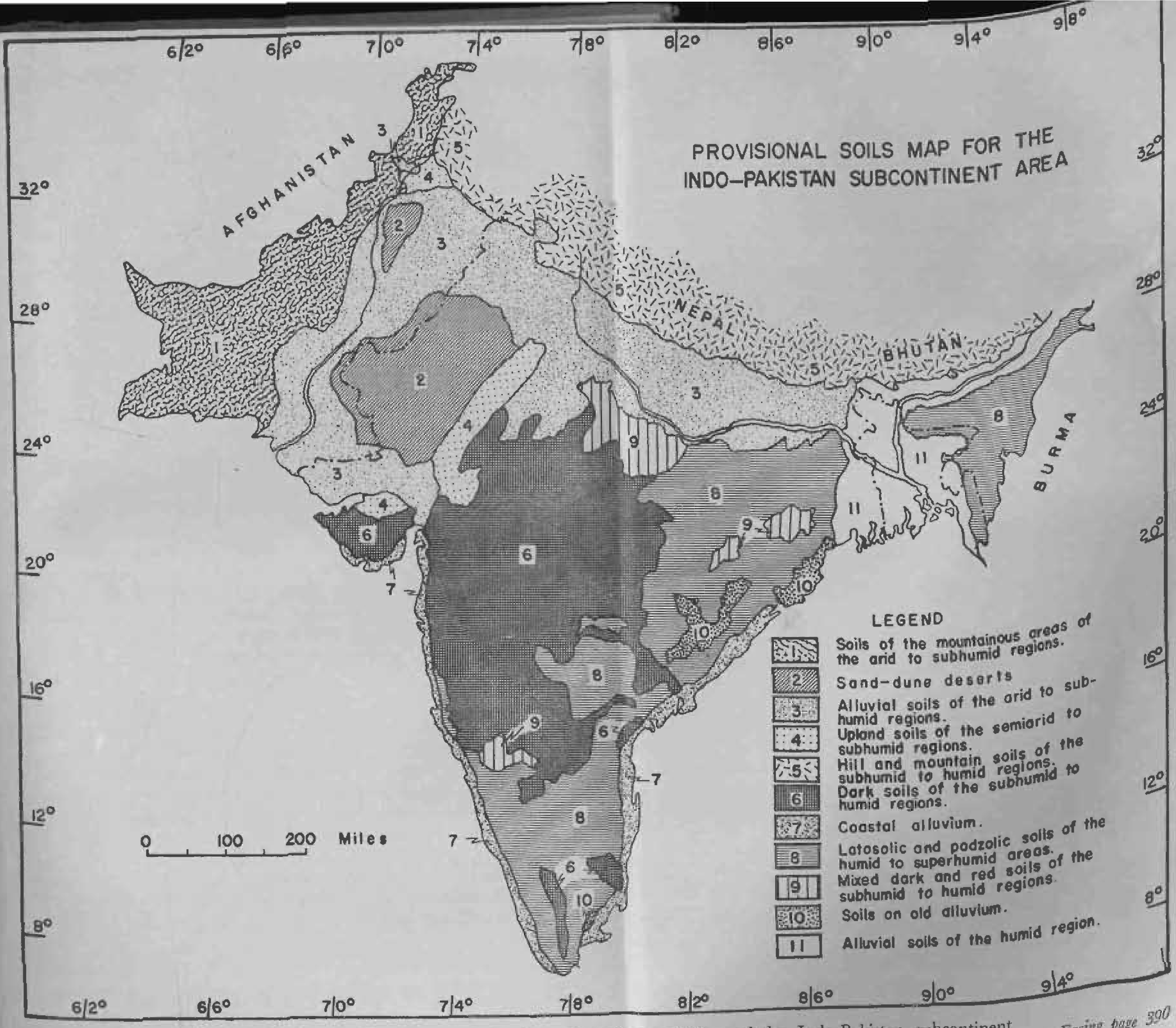

FIG. 13-6. A provisional map of physiographic-climatic regions of the Indo-Pakistan subcontinent as related to the soils of the area.

Soils of the Arid to Subhumid Mountainous Region

The aridity of the southern reaches of this region results in the development of the desert or sierozem soils. Considerable reworking of the soil material by wind and torrential floods maintains the soils over a large part of the area in an immature, azonal condition. Approaching the more moist regions in northern West Pakistan, soils of the brown or chestnut group may be expected. However, due to erosional influences, zonal soils again are of very limited extent.

Sand Dunes

The shifting sand dunes of the desert areas are classed in the great soil group of regosols. Limited dune areas occur in many locations throughout the arid regions.

Alluvial Soils of the Arid to Subhumid Regions

Because of the ever-present influence of floods, these alluvial soils must be considered as azonal. In the waterlogged areas, salinization has resulted in intrazonal, solonchak soils. The incidence of salinized soils decreases as one moves eastward along the Ganges River and into areas of higher precipitation. However, even though the rainfall is relatively high, its pattern is torrential, runoff is excessive, and the total moisture entering the soils is relatively low. Therefore, flushing of salts from the soil is not in proportion to that which would occur in other places where the rainfall distribution is more uniform throughout the year.

Upland Soils of the Semiarid to Subhumid Regions

Three small areas are shown in this category in Fig. 13-6. Again, where erosion is not a problem, chestnut and, possibly, chernozem soils will occur. The effects of base cycling is in strong evidence and has resulted in the formation of distinct calcareous horizons in the soil profile. The soil organic matter, which seldom exceeds 1% in the surface layer, imparts a much darker colour than would be expected through comparison with soils of similar carbon content from other parts of the world.

Hill and Mountain Soils of Subhumid to Humid Regions

This grouping represents mountainous soils of the wetter regions. The increased precipitation is the result of the higher elevations of the Himalayan Mountains. With the increase in precipitation the potential for the formation of podzolic soils improves. Podzols, and peaty and gleied soils have been reported for this area. Many red soils are present and appear to be derived from reddish sedimentary parent materials of ancient origin which occur in many locations.

Dark Soils of the Subhumid to Humid Regions

This area in central western India contains a high proportion of grumusol soils. In areas of low rainfall, salinized soils are found in association with otherwise immature alluvial soils.

Coastal Alluvium, Soils on Old Alluvium, and Alluvial Soils of the Humid Region

All three regions considered here occur in areas of high rainfall. As a consequence, the tendency for formation of podzolic and latosolic soils is strong. Whether or not the soils have developed podzolic or latosolic characteristics depends largely on the length of time parent materials have been in place. The older alluvium tends definitely toward podzolic types. However, most of the alluvial area in Assam and Bengal is under the influence of current deposition. Thus, azonal soils are common and show only minor effects of local environment in many instances.

Latosolic Soils of the Humid and Superhumid Areas

Because of the high temperature and rainfall, these areas are ideal for the progressive formation of latosolic soils. True latosols, with characteristic laterite layers, are found often, although they are frequently seriously eroded. Reddish latosolic soils are observed also in association with darker coloured grumusols. The red soils occur in upland, well-drained locations, whereas the dark soils are found where drainage is poor. Some areas where this type of association is found is sufficiently extensive to permit their separate delineation in Fig. 13-6.

SELECTED REFERENCES

1. Baldwin, Mark, Kellogg, Charles E., and Thorp, James (1938) Soil classification. From *Soils and Men*, The Year Book of Agriculture for 1938, pp. 979-92. Supt. of Documents, Washington 25, D. C.

2. Jenny, H. (1941) *Factors of Soil Formation*. McGraw-Hill Book Co., New York.

3. Simonson, R. W. (1959) Outline of a generalized theory of soil genesis. Proc. Soil Sci. Soc. Amer. 23 : 152-156.

4. Whiteside, E. P. (1951) Changes in criteria used in soil classification since Marbut. Proc. Soil Sci. Soc. Amer. 18 : 193-195.

REVIEW QUESTIONS

1. What is meant by 'normal' soil ? Explain why the terms 'normal' and 'zonal' carry the same significance.
2. What is meant by macroclimate and microclimate ? Effective climate ?
3. The halomorphic soils are intrazonal. Why is this ? Would they occur within areas of humid-region soils ?
4. What are the basic similarities between hydromorphic and halomorphic soils ?
5. Desert, sierozem, and brown soils are all members of the same suborder. What are the major reasons for separation of these soils into their respective great soil groups ?
6. Why will many soils of the Indus-Ganges Plains be azonal rather than zonal ? Generally, where would you expect to find zonal soils in the subcontinent ?
7. Explain why the nature of the parent material has a dominant influence on the soils of the Indus Plains but is relatively without influence in the Western Ghats of the Indian peninsula.
8. Why would you expect to find evidence of more advanced soil formation on southern exposures of the Pir Panjal Mountains than on the northern slopes ?
9. A sandy soil adjacent to a clay soil is lime free, whereas some evidence of lime occurs in the clay. What would be the probable cause of difference in the effective climates of these two soils ?
10. How does relief affect soil climate and, therefore, vegetation ?
11. Why may desalinization cause an increase in soil pH ?
12. How are salinization and gleization similar, and how do they differ ?

CHAPTER FOURTEEN

SOIL SURVEY AND MAPPING

THE SURVEY and mapping of soils might be looked upon as the practical application of the fundamentals of soil classification. In conducting the survey, an effort is made to group together soils of essentially the same characteristics. To make the survey of practical value, specific properties of the individual groupings are described and related to their crop-producing potentialities. Thus, if we know the capabilities of a particular soil type, we can apply that knowledge to soils with the same characteristics regardless of where they are found. The ability to do this obviously is limited by the accuracy of the survey which designates these locations. Human error will be involved in the original mapping of the soil. Also, because of differences in the exploitation of soil resources, two persons farming essentially the same soil may obtain such diverse results as to make it appear they are working with dissimilar soils.

Moon *et al.* (1949) cite the broad objectives of soil surveys as being:

> "(a) to classify soils according to their characteristics into units of a uniform system of classification with uniform nomenclature ; (b) to show their distribution on maps ; and (c) to define their adaptability for crops, grasses, and trees, their management requirements for sustained production, and the expected yields of individual crops under different systems of use and management."

To satisfy condition (a), above, the surveyor must have adequate knowledge of the basic principles of soil development and classification. For (b), he must have experience and the desire or willingness to carry out his work in the best possible manner. To provide the definition in (c) requires that the area under consideration be either cropped or supporting vegetation which can serve as the basis for predicting yield potentials, or that similar areas already surveyed can be used as the basis for such predictions.

SOIL CLASSIFICATION AND SURVEY

The major categories of classification, as discussed in Chapter Thirteen, allow for the separation of soils into broad, generalized groupings. The practical application of survey is primarily for the separation of soils on the basis of their crop-producing ability. Each of the great soil groups contains a large number of different soils representing a wide variation in productive capacity, and it is necessary to distinguish these differences in mapping and survey. The classification units used for this are *series*, *type*, and *phase*. Each of these categories is defined in the following paragraphs. They are discussed also by Baldwin *et al.* (1938).

Soil Series

If we expose a soil profile and study it, we note certain features which may be identified and described. In zonal and intrazonal soils, for instance, the different horizons are obvious because of their colour, texture, and structure. Exposing a number of profiles in the same vicinity will probably reveal that all are similar unless major changes in topography or parent material occur. All of the soils that are similar would be members of the same *soil series*.

It may be said generally that readily recognizable variation between soils is necessary before they will be placed in separate series. Usually the greatest variation permitted within a single series is in the texture of the surface soil. A series is usually named after a place near which it is first described.

Soil Type

Variation in the texture of the A horizon is permitted without altering the series name. A major change in surface soil texture does give rise to different *soil types*, however. Thus, members of a soil type will be alike in all characteristics including the textural grade of the surface soil. There is a limit to which variation in texture can take place without altering other morphological characteristics. It follows, then, that we may find a range, say, from clay loam to silt loam, within a series but not from the extremes of clay to a sandy loam or sand. Generally there will be a maximum of three types within a series, with the current tendency being toward limiting members of the same series to but one or two textural classes.

Soil Phase

The last minor category is the *phase* of a soil type. Generally, phases are separated not on differentiating morphological features of the soil profile but on characteristics that affect land use and crop growth. In the past, phases have been separated on the basis of stoniness, slope, or degree of or susceptibility to erosion. Depth to an unconforming D horizon and presence or absence of salts have been used also to separate one phase from another.

The recent trend has been toward eliminating the use of soil phase as a category of classification. It may be assumed that factors important enough to affect soil use will affect morphological features also and require recognition at the soil series or type level. In this connection, Whiteside (1954) quotes a statement published by the Soil Survey Sub-Committee of the North Central Regional (U.S.A.) Soil and Fertilizer Research Committee :

> "The soil type is the lowest category in the natural system of soil classification. It is a subdivision of the soil series based on the texture of the surface soil (or degree of decomposition of organic soils), shape or inclination of the surface, stoniness, degree of erosion, depth of the solum, depth to layers of unconformable material, and of the salinity of the soil. It is the taxonomic unit concerning which the greatest number of precise statements and predications can be made about soil use, management, and productivity."

Soil Complexes, Associations, and Catenas

In general, soils of like characteristics respond in a uniform manner when subjected to the same management practices. If, on the other hand, an area is comprised of a number of soil series of considerably different properties, different systems of management must be employed if uniform plant growth and yields are to be expected. Obviously, then, mapping has as a major objective the forecasting of the complexity of management which must be employed within the area being surveyed.

In areas of high soil variability, difficulty encountered by both the surveyor and the farmer in coping with rapid changes in soil properties will determine the practicality of separating soils of the area into the series, types, or phases actually present. As an example, alluvium often will exhibit major changes in texture, organic matter content, and other properties irregularly and very frequently over a given landscape. Delineation of soil boundries under these circumstances could be an almost endless chore, and applying different cultural treatments to these individual soils would be essentially impossible. Thus, primarily as a matter of convenience to all concerned, soils in such an area commonly are grouped together and mapped as a *complex* rather than as series, types, or phases. The management system then recommended, while probably not ideal for all soils present, would be designed to take greatest advantage of the average production potental of the complex.

It is not unusual in survey work to group two or more series together into single units for mapping even when the individual series are easily identifiable and the areas they comprise are relatively large. Presumably, location of the member series can be recognized in the field easily by both the surveyor and the farmer. Since the farmer still can make full use of survey information, the reduced cost of acquiring it justifies mapping several series together rather than as individual units.

When differences among adjacent series are the result of formation on dissimilar parent materials, the recognized grouping is referred to as an *association*. If the related series are on the same parent material but soil development has not been uniform due to topographical variation, the grouping they form is designated as a *catena*[1]. The topographic influence

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[1] The word catena means chain. Soil series in a catena may be likened to the links that make up a length of chain.

is one of altering the effectiveness of precipitation, erosion, and subsoil drainage. Conditions conducive to variation in the soil-forming processes responsible for catena formation and the wide range in soils which may occur together within a given catena are illustrated in Fig. 14-1.

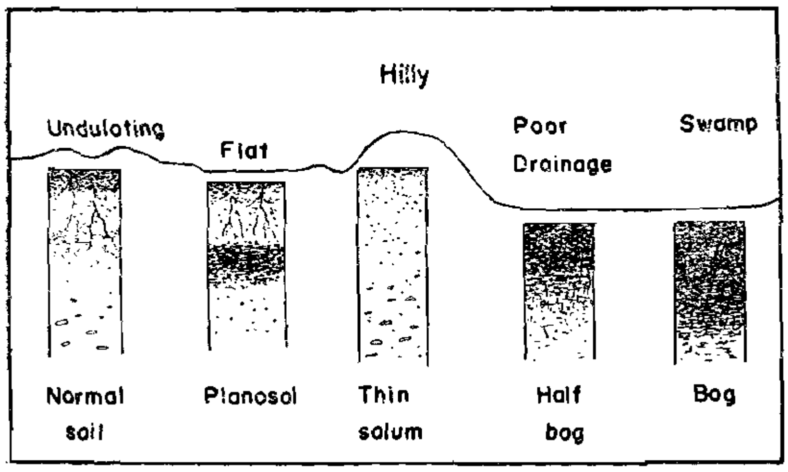

FIG. 14-1. The effect of variation in surface relief on the development of soils from the same parent material (after Byers *et al.*, 1938).

TYPES OF SURVEYS

Soil surveys vary in their nature. There are numerous reasons for this, but the most common cause is related to the use to be made of them.

The Soil Survey Staff (1951, p. 15) recognizes three types of mapping for soil survey purposes, namely: (1) detailed, (2) reconnaissance, and (3) detailed-reconnaissance. Detailed maps show minor variation of significance to soil use and management, and are the result, usually, of the mapping of soil types and phases. Obviously, as the complexity of the area increases, it becomes progressively more difficult to segregate all variation on this basis. Thus, at times, mapped boundaries must include more than a single phase or type. In these instances, soil associations or complexes, which include a number of soils mapped together, represent the mapping units. On publication, the scale of detailed maps is frequently 2 inches to the mile, but this may vary depending upon the detail and accuracy required.

Reconnaissance mapping permits coverage of an area in the shortest possible time. Soil associations rather than the individual soil series form the basis for such mapping. Reference to slope, common in detailed surveys, often is ignored, as are soil phases. The exact nature of the reconnaissance survey will also depend upon its intended use and upon the time and resources available for its completion. Important to this also is the nature of the land being surveyed. Mountainous country of minor agricultural significance, for instance, would warrant only the simplest of reconnaissance maps.

Preliminary surveys of large development areas, such as the Thal Development Project (Irrigation) along the Indus River, could hardly justify great detail to survey and mapping. A reconnaissance survey of a very general nature can provide the basic data necessary to assess production potentialities and engineering requirements of the area (see Fig. 14-2). As the project develops, however, those lands that are brought under irrigation will display problems specific for the soils present and will necessitate the conduct of more detailed surveys. By means of the survey, practices observed to be most suitable for each soil type then can be applied to all areas of like soil throughout the project.

BASE MAPS FOR SOIL MAPPING

In the early stages of soil survey work, it was necessary in many instances for the surveyor to provide his own base map for field use. This laborious operation, which often consumed more than half of the surveyor's time, was necessary so that soil boundaries could be shown relative to other land surface features. Maps constructed showed such details as roads, fences, buildings, drainageways, and similar physical features. More useful *topographic maps* are available, and have served the soil surveyor well by eliminating the need for construction of separate base maps. The contour lines shown on topographic maps are very helpful since, in mapping, topographic variation

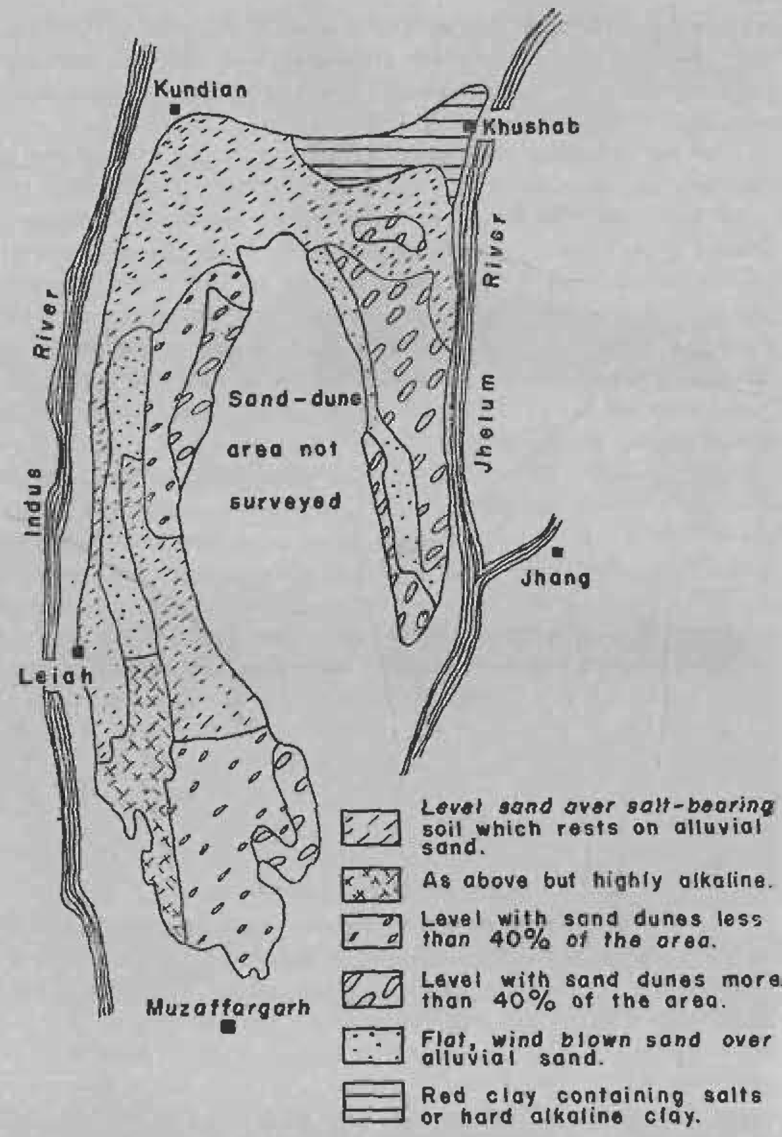

FIG. 14-2. A type of reconnaissance survey used in planning the Thal Development Project, West Pakistan (after Asghar, 1953). Although the survey is not based on strict taxonomic units, it serves a useful purpose in planning.

is often a major key to change in soil type. Contour lines serve not only as a means for the orientation of the surveyor with respect to the topography, but, also, often correspond closely to boundaries between soil types.

One of the most significant advances in soil mapping was the introduction, in the early 1930's, of aerial photographs as base maps. Because they lend themselves so readily to mapping, their use has become widespread. Large scale photographs (4 inches to the mile) will normally allow as much detail in soil mapping as is required. At this scale, each acre of land is represented by a square about 1/6th inch on each side. Changes in vegetation types, drainage patterns, roadways, buildings, etc., are easily seen on an aerial photograph of this scale. Waterlogged areas, patches of saline soil, and even features attributable to changes in soil texture frequently are discernible. An added advantage to the use of aerial photographs is the three-dimensional effect they exhibit. When two photographs of the same area, taken from slightly different angles during the flight, are viewed with a stereoscopic device, variations in the surface of the land stand out in bold relief. By this means, a relatively accurate evaluation of topography can be made. This technique has been sufficiently well developed so that it is being applied to replace costly surveys, as normally performed by engineers, in establishing the most logical routes for highways over difficult terrain. Clarke (1957, pp. 136-167) provides a good discussion on the use and interpretation of aerial photographs for soil survey.

DELINEATION OF SOILS FOR MAPPING PURPOSES

When a surveyor goes to the field, he first must familiarize himself with the soil types common to the area. This may be done by observations of soils in adjacent areas which have been surveyed previously. For the man with limited experience, personal guidance from someone acquainted with the soils in the area is desirable. Slowly the technician gains competence in recognizing the various soil types from texture and profile characteristics as determined by boring or digging frequent test holes. More important, however, he learns to relate these characteristics to physical features of the land surface—topography or slope, vegetation, and appearance of the surface soil. Having attained this ability, he will be able to predict often the soil type present from landscape features. Test borings are necessary to verify the prediction and to locate boundaries of the mapping unit where such boundaries are not distinct at the soil surface.

The actual operation of soil mapping requires a thorough investigation of the area being mapped through observation of numerous borings. The terrain must be traversed at frequent intervals, generally at right angles to drainageways. Through use of physiographic and surface soil features, boundaries between soil types may be determined. Since surface features serve only as a guide to the identity of a soil type, examination of the subsoil for texture, colour, presence or absence of lime or *kankar*, etc., needs to be done repeatedly. This is especially true of the flat plains areas, as typified by the Indus-Ganges and Brahmaputra Basins, where alluvial deposits vary markedly over short distances.

Considering that differentiating between soil types requires separations based on variations in profile characteristics, it is helpful in survey to know which outward features of the area indicate such change. The following conditions, as modified from a listing by Clarke (1957, p. 134), provide such indications :

1. *Change in texture and colour of the surface of the soil.*
2. Change in topography.
3. Change in vegetation, either in type of plant cover or its disappearance, as frequently occurs with increases in salinity.
4. Variation in the feel underfoot, as compaction, fluffiness, crusting, etc.

The satisfactory delineation of soil types in the field is not always possible. This is particularly true when such features as salinity, alkalinity, or chemical composition are being used as criteria for separation of mapping units. Seriously affected saline and alkali soils usually are readily identified, but rain or floods may obscure the visible traits whereby such soils are recognized. Proof, then, rests with chemical analysis of samples taken from the area in question. East Pakistan soils of podzolic or latosolic types frequently are characterized on the basis of their chemical composition. Although it would represent an extremely difficult problem to base an entire survey on laboratory analyses, any soil survey programme should include a certain amount of this type of characterization.

Simple tests can be made in the field which aid in survey work (*i.e.*, measurement of pH and conductivity of salted soils, showing lime to be present by effervescence with acid, estimation of texture by feel, and others). However, field testing generally requires an excessive amount of time except for the very rapid tests. It often is more desirable to collect samples for appraisal in the laboratory, unless this necessitates the taking of so large a number as to be impractical.

THE SOIL SURVEY REPORT

Soil mapping serves little purpose until the completed map is incorporated in a detailed written report. The report normally is prepared by the surveyor, and has as its objectives the description of each mapped soil in terms of its properties, behaviour, and use. As many factors pertinent to utilization of the individual soils as possible are included. Production capabilities in terms of yields, general level of fertility and fertilizer requirements, management problems, and the like are described. In addition, information of a general nature, such as climate, physiography and drainage, population, agriculture of the area, etc., are important to an overall assessment of the agricultural value of the land. With these facts, the agriculturist is able to interpret the soil map and devise a suitable system of farming for the land under consideration. In the Asian subcontinent, the inability to read on the part of most farmers will place certain limitations on the use that may be made of the reports. Interpretation must come to them through technicians. The availability of the report to the technician, however, will assist him materially in providing proper guidance.

The map of a surveyed area is the basis for a useable soil survey report. One may determine from the map the soil types present in any location covered by the survey. Natural topographic and man-made features, such as rivers, villages, roads, etc., are shown and aid in locating a particular site on the map. These features are apparent on the map for a small area of Gwalior surveyed and reported on by Pendleton (1947) and reproduced in Fig. 14-3.

In Fig. 14-4, an attempt is made to relate the mapped soils in Fig. 14-3 to topography and other local conditions. The land profile shown corresponds to that occurring along a line extending westward from a point of high elevation (*i.e.*, benchmark BM 1758) down to the Sipra River. Soils present in the field at higher elevation show the effect of erosion. These soils are thin, immature, and azonal in character. The Malwa soil, being on a gentle slope where soil development can proceed at a reasonable rate, represents the more nearly mature or zonal soil for the area. The lower-lying soils may be mature or immature depending upon how long they have persisted without change through additional deposition or erosion by flood waters.

The final preparation of a soils map for publication must be done by a draftsman. The development of a soil survey programme requires training of personnel for this purpose as well as for the mapping operations. The preparation and publication of maps must conform to the skills of the technicians preparing them and to the facilities for their publication. More important than a complex and costly map is a properly written report which supplements the map.

Further discussion on the preparation of the soil survey report would serve little purpose here. For the student interested in this phase of soils work, reference is made to the Soil Survey Manual of the U. S. Department of Agriculture (Soil Survey Staff, 1951, pp. 409-434).

SOIL SURVEY PROGRAMMES

Current conditions in the Asian subcontinent would suggest that certain limitations must be placed on the amount of detail that can be collected in soil survey work. Personnel and funds available for survey purposes, the use that can be made of the information detailed surveys will provide, and the urgency of the programme will determine to a large extent the number of recognizable soil features that are practical to delineate in a

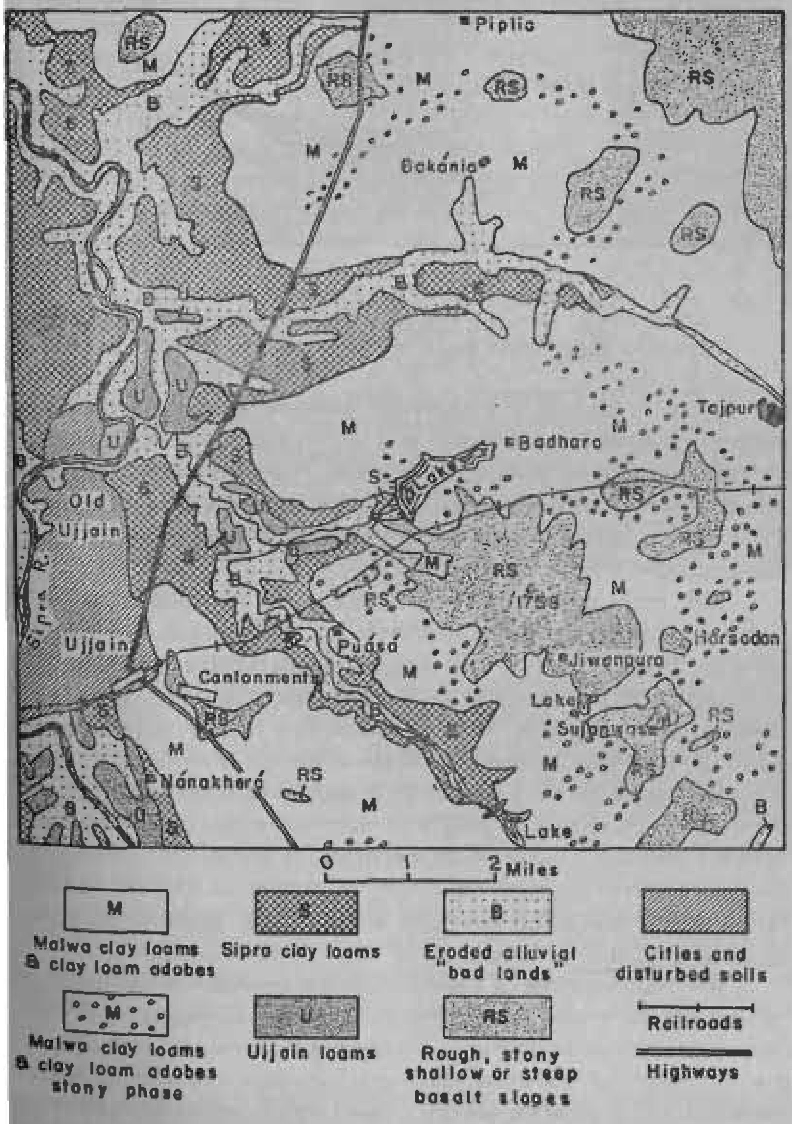

FIG. 14-3. Map of soils in a small area of Gwalior State, India (after Pendleton, 1947).

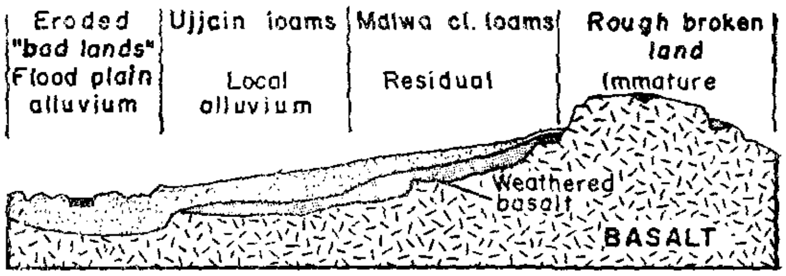

FIG. 14-4. Diagramatic land profile from a section of Gwalior State, India, showing the relationship among soil series, topography, and parent material.

soil survey. Not only is there an extreme shortage of trained personnel both for mapping and interpretation, and of funds for pursuit of these phases, the process of publication of useable, allied reports will undoubtedly encounter numerous difficulties. This discouraging aspect of soil survey is prevalent in many parts of the world.

The reconnaissance survey, then, would appear to be of definite value to the programme for soil mapping in the subcontinent. Such a survey will record the major groupings of soils and indicate the extremes in characteristics and properties encountered. By this means, evaluation of soil conditions on a broad scale can be accomplished in the shortest possible time. This will assist in making the decision as to where the greatest increase in productivity can be achieved in the most expedient manner and with the least expenditure of funds. Furthermore, this procedure permits establishing relatively clearly the full range of soil types and provides a basis for a more detailed soil classification for the entire area under consideration.

The development of a soil survey programme usually follows the pattern in which each agency interested attempts to develop its own system of mapping. There is good reason for this when the purpose of agencies do not overlap. However, where coordination is possible, every effort should be made to develop survey work along cooperative lines. This is particularly true where both funds and trained personnel are so limited. Because of the general lack of information on soil distribution and the great demands placed on science to produce greater quantities of food for the rapidly increasing population, delay in formulating a sound programme can have serious detrimental effects.

These considerations suggest, then, that a central authority be charged with the overall control of survey work. A uniform programme should be developed with the contributing and interested agencies accepting certain responsibilities for the conduct of the programme.

Much can be gained by considering carefully the problems experienced in other countries during the course of development of their soil survey programmes over the past 50 years. Many costly mistakes can be avoided if this is done. This limited resources available indicate this approach as being absolutely essential.

LAND CLASSIFICATION

As indicated above, soil survey has as a prime objective the expression of the capability of the land for productivity or use. A major portion of the information necessary for this evaluation is embodied in the soil map and report. However, knowing merely the distribution of soil types, as determined from a study of profile characteristics, does not provide a sufficiently complete analysis for planning efficient land utilization. This requires separate surveys which are based partly on soil-type or soil-phase data and partly on other factors contributing to satisfactory land use. As opposed to soil classification, then, land classification includes external factors that contribute materially to productivity of a soil.

Land classification is subject also to the errors in decision on the part of the classifier. Accuracy in conducting the survey is a function of the knowledge of the technician performing the operation and the availability of factual data on which the survey may be based. The classification must be made by a person familiar not only with the productive capacity of a soil type but also one experienced in relating all factors of production common to the area in question.

Certain general types of soils are rather widely distributed throughout broad areas in the subcontinent. They occur under a variety of conditions—differing rainfall, degree of waterlogging, etc. It may be realized that the production potential of one soil type will be greater if irrigation water is available to it than if it must depend solely on scanty natural rainfall. On the other hand, should waterlogging have become a serious problem, then irrigation may have lowered the production capabilities of the land below that possible under dry-farming conditions. These factors and others are taken into consideration in the classification of land.

Land classification normally requires the grouping of soils together into units of like potentialities. Categorizing soils in this way is referred to by Barnes (1949) as *interpretive soil classification.* It assumes many forms and purposes. Soils may be grouped together into classes indicating differences in agricultural use and management (Barnes and Harper, 1949), or for timber, range land, and watershed management (Gardner and Retzar, 1949). Important to the engineer involved in roadway and airport construction are the interpretive surveys based on the engineering properties of soils (Stokstad and Humbert, 1949).

For the soil scientist, groupings according to land-use capabilities assume greatest importance. Classifications of this type are usually developed within a particular area and take into account those factors highly important to crop growth (erosion, texture, drouthiness, availability of water, etc.). Classes of land are established and given a numerical designation with Class I land having the greatest agricultural value. The range in variation in conditions will determine the number of classes established, but from four to six will usually allow for separation into significantly different groupings.

Utilitarian soil surveys in the subcontinent will require different bases from area to area. In upland areas, slope and erosion contribute to production potentialities and play an important role in distinguishing between soil classes. In the lowland areas, particularly in West Pakistan and northwest India, soil class differentiation is influenced by the availability and quality of irrigation water, susceptibility to waterlogging and to wind erosion, and the presence of salts and alkali.

REVIEW QUESTIONS

1. What is meant by the expression 'a soil is a three-dimensional piece of landscape'? Can observations of the surface features of the landscape be used to judge some of the qualities of the profile ? Explain.
2. What purpose is played by the base map in soil mapping ? Name three types of base maps.
3. In surveying, you find the nearly level crest of a hill composed of soil having a dark surface layer and a light-coloured sub soil with lime nodules (*kankar*) starting at about 18 in. beneath the surface. On an adjacent, steep slope, you find the surface soil light in colour and containing *kankar*. What is the probable explanation of this difference in adjacent soils ? Would you separate these two soils into separate series or into two phases of the same series ?
4. Outline the requirements of an adequate soil survey report.
5. Why is the reconnaissance survey preferable to the detailed survey under current conditions in the subcontinent ?
6. What is the purpose of land classification, and how does it differ from soil classification ?
7. Why are land classes different for irrigated as compared to hilly, upland areas ?

APPENDIX

Table 1. Approximate hole size for sieves commonly used in soils research.

Number of mesh per inch	Diameter of opening
	mm.
4	4·76
8	2·38
9·2	2·00
12	1·41
17·2	1·00
20	·84
30	·54
40	·40
60	·25
80	·18
100	·15
140	·10
300	·05

Table 2. Some properties of common ions.

Chemical Symbol	Molecular weight	Equivalent weight
Ba	137·36	68·68
B	10·82	...
Ca	40·08	20·04
Cl	35·46	35·46
H	1·01	1·01
Fe	55·84	...
Mg	24·32	12·16
Mn	96·00	...
N	14·01	...
O	16·00	...
P	31·02	...
K	39·10	39·10
Si	28·06	...
Na	23·00	23·00
S	32·06	...
Zn	65·38	32·69

APPENDIX *(Contd.)*

Table 3. The equivalent weight and solubility of some commonly used chemical compounds.

Formula	Equivalent weight	Solubility in water	
		g/100 ml.	
$CaCl_2$	55·49	59·5	(0°)[1]
$CaSO_4$	68·07	0·298	(20°)
$CaSO_4.2H_2O$	86·09	0·223	(0°)
$CaCO_3$	50·04	0·0014	(25°)
$Ca\ (H_2PO_4)_2.H_2O$	126·05	Soluble	
$CaHPO_4.2H_2O$	86·05	0·02	(24·5°)
$MgCl_2$	47·62	52·8	(0°)
$MgSO_4$	60·19	26·9	(0°)
$MgCO_3$	42·16	0·0106	
$NaCl$	58·45	35·7	(0°)
Na_2SO_4	71·03	48·8	(40°)
Na_2CO_3	53·00	7·1	(0°)
$NaHCO_3$	84·01	6·9	(0°)
$NaNO_3$	85·01	73·0	(0°)
KCl	74·55	34·7	(20°)
K_2SO_4	87·13	12·0	(20°)
K_2CO_3	69·10	112·0	(20°)
NH_4NO_3	80·05	118·3	(0°)
$(NH_4)_2SO_4$	66·07	70·6	(0°)

[1] Temperature of solution in °C.

BIBLIOGRAPHY

Allaway, H., and Pierre, W. H. (1939) Availability, fixation and liberation of potassium in high-lime soils. J. Amer. Soc. Agron. 31 : 940-953.

Allaway, W. H. (1957) pH, soil acidity, and plant growth. From *Soil*, the Yearbook of Agriculture for 1957, pp. 67-71. Supt. of Documents, Washington 25, D. C.

Allmendinger, D. F., Kenworthy, A. L., and Overholser, E. L. (1943) The carbon dioxide intake of apple leaves as affected by reducing the available soil water to different levels. Proc. Amer. Soc. Hort. Sci. 42 : 133-140.

Anderson, M. S. and Byers, H. G. (1934) The carbon-nitrogen ratio in relation to soil classification. Soil Sci. 38 : 121-138.

Anderson, W. (1954) Soil productivity and fertilizer requirements in India. Foreign Agri. 18 : 202-207.

Arnon, D. I. and Johnson, C. M. (1942) Influence of hydrogen-ion concentration on the growth of higher plants under controlled conditions. Plant Phys. 17 : 525-539.

Asghar, A. G. (1952) The problem of salinity and alkalinity of soils and its solution. Fourth Pakistan Science Conference, Proceedings (From mimeographed reprint).

Asghar, A. G. (1953). Cotton cultivation in the Thal Area of Punjab (Pakistan). Pakistan J. Sci. 5 : 128-138.

Aslam, M. (1957) The effect of cultivation on soil properties and crop yield. M.Sc. Thesis, Punjab Agri. College, Lyallpur, West Pakistan.

Association of Official Agricultural Chemists (1945) *Official and Tentative Methods of Analysis of the Association of Official Agricultural Chemists.* P. O. Box 540, Benjamin Franklin Station, Washington, D. C.

Baldwin, M., Kellogg, C. E., and Thorp, J. (1938) Soil classification. From *Soils and Men*, The Yearbook of Agriculture for 1938, pp. 979-1001. Supt. of Documents, Washington 25, D. C.

Barnes, C. P. (1949) Interpretive soil classification: Relation to purpose. Soil Sci. 67 : 127-129.

Barnes, C. P. and Harper, W. G. (1949) Interpretive soil classification: Agricultural use and management. Soil Sci. 67 : 141-149.

Bass, G. B. and Sieling, D. H. (1950) Method for determining the relative phosphate-fixing capacity of acid soils. Soil Sci. 69 : 269-280.

Baver, L. D. (1956) *Soil Physics.* John Wiley and Sons, New York, N. Y.

Baver, L. D. and Farnsworth, R. B. (1941) Soil structure effects on the growth of sugar beets. Proc. Soil Sci. Soc. Amer. 5 : 45-48.

Bennett, O. L., and Doss, B. D. (1960) Effect of soil moisture level on root distribution of cool season forage species. Agron. J. 52 : 204-207.

Berger, K. C. (1949) Boron in soils and crops. Advances in Agron. 1 : 321-351.

Boicourt, A. W. and Allen, R. C. (1941) Effect of aeration on growth of hybrid tea roses. Proc. Amer. Soc. Hort. Sci. 39 : 423-425.

Bouyoucos, G. J. (1954) New type of electrode for plaster of paris moisture blocks. Soil Sci. 78 : 339-342.

Bouyoucos, G. J., and Mick, A. H. (1940) An electrical resistance method for the continuous measurement of soil moisture under field conditions. Michigan Agri. Expt. Sta. Tech. Bul. 172.

Bouyoucos, G. J., and Mick, A. H. (1947) Improvements in the plaster of paris absorption block electrical resistance method for measuring soil moisture under field conditions. Soil Sci. 63 : 455-465.

Bouyoucos, G. J., and Mick, A. H. (1948) A fabric absorption unit for continuous measurement of soil moisture in the field. Soil Sci. 66 : 217-232.

Bower, C. A. (1945) Separation and identification of phytin and its derivatives from soils. Soil Sci. 59 : 277-285.

Bower, C. A., Reitemeier, R. F., and Fireman, M. (1952) Exchangeable cation analysis of saline and alkali soils. Soil Sci. 73 : 251-261.

Bray, R. H. (1948) Correlation of soil tests with crop response to added fertilizers and with fertilizer requirements. From *Diagnostic Techniques for Soils and Crops.* pp. 53-86, American Potash Institute, Washington 6, D. C.

Bray, R. H. (1954) A nutrient mobility concept of soil-plant relationships. Soil Sci. 78 : 9-22.

Broadbent, F. E., Tyler, K. B., and Hill, G. N. (1957) Nitrification of ammoniacal fertilizers in some California soils. Hilgardia 27 : 247-267.

Buehrer, T. F. (1932) The physico-chemical relationships of soil phosphorous. Arizona Agri. Expt. Sta. Tech. Bul. 42.

Byers, H. G., Kellogg, C. E., Anderson, M. S., and Thorp, J. (1938) Formation of soil. From *Soils and Men*, The Yearbook of Agriculture for 1938, pp. 948-978, Supt. of Documents, Washington 25, D. C.

Chandler, R. F., Jr. (1939) Cation exchange properties of certain forest soils in the Adirondack section. J. Agri. Res. 59 : 492-505.

Chandler, R. F., Jr. (1943) The time required for Podzol profile formation as evidenced by the Mendenhall glacial deposits near Juneau, Alaska. Proc. Soil Sci. Soc. Amer. 7 : 454-459.

Chatterji, U. N. (1957) Final report of the All India Soil Survey Scheme. Indian Council of Agri. Res. Bul. 73.

Clarke, F. W. (1924) The data of geochemistry. U. S. Geol. Survey Bul. 770.

Clarke, G. R. (1957) *The Study of the Soil in the Field.* University Press, Oxford, Eng.

Cline, M. G. (1945) Methods of collecting and preparing soil samples. Soil Sci. 59 : 3-5.

Cline, M. G. (1949a) Basic principles of soil classification. Soil Sci. 67 : 81-91.

Cline, M. G. (1949b) Profile studies of normal soils of New York : I. Soil profile sequences involving brown forest, gray-brown podzolic, and brown podzolic soils. Soil Sci. 68 : 259-272.

Cole, C. V., and Jackson, M. L. (1950) Solubility equilibrium constant of dihydroxy aluminium dihydrogen phosphate relating to a mechanism of phosphate fixation in soils. Proc. Soil Sci. Soc. Amer. 15 : 84-89.

Cole, C. V., and Olsen, S. R. (1959) Phosphorus solubility in calcareous soils : II. Effects of exchangeable phosphorus and soil texture on phosphorus solubility. Proc. Soil Sci. Soc. Amer. 23 : 119-121.

Coleman, E. A. (1947) A laboratory procedure for determining the field capacity of soils. Soil Sci. 63 : 277-283.

Coleman, N. T., Kamprath, E. J., and Weed, S. B. (1958) Liming. Advances in Agron. 10 : 475-522.

Coleman, N. T., and Mehlich, A. (1957) The chemistry of soil pH. From *Soil*, The Yearbook of Agriculture for 1957, pp. 72-79. Supt. of Documents, Washington 25, D. C.

Collings, G. H. (1955) *Commercial Fertilizers, Their Sources and Use.* McGraw Hill Book Co., New York.

Colwell, W. E. (1943) A biological method for determining the relative boron contents of soils. Soil Sci. 56 : 71-94.

De, P. K. and Sulaiman, M. (1950) Fixation of nitrogen in rice soil by algae as influenced by crop, CO_2, and inorganic substances. Soil Sci. 70 : 137-151.

Dean, L. A., and Rubins, E. J. (1945) Absorption by plants of phosphorus from a clay-water system : Methods and ensuing observations. Soil Sci. 59 : 437-448.

Dean, L. A., and Rubins, E. J. (1947) Anion exchange in soils : I. Exchangeable phosphorus and the anion exchange capacity. Soil Sci. 63 : 377-387.

Doneen, L. D. (1954) Salination of soils by salts in irrigation water. Trans. Am. Geophys. Union 6 : 943-950.

Eaton, F. M. (1949) Significance of carbonates in irrigation water. Soil Sci. 69 : 123-33.

Eaton, F. M. (1953) Certain aspects of salinity in irrigated soil. Report to the Govt. of Pakistan, FAO Rept. No : 167. (Rome).

Elliott, G. R. B. (1924) Relation between downward penetration of corn roots and water level in peat soils. Ecology 5 : 175-178.

Erdman, L. W. (1948) Strains of *Rhizobium* effective on guar. J. Amer. Soc. Agron. 40 : 364-369.

Evans, C. E. and Attoe, O. J. (1948) Potassium-supplying power of virgin and cropped soils. Soil Sci. 66 : 323-334.

Fieldes, M., Swindale, L. D., and Richardson, J. P. (1952) Relation of colloidal hydrous oxides to the high cation-exchange capacity of some tropical soils of the Cook Islands. Soil Sci. 74 : 197-205.

Fried, M. and Peech, M. (1946) The comparative effects of lime and gypsum upon plants grown upon acid soils. J. Amer. Soc. Agron. 38 : 614-623.

Gardner, R. A. and Retzer, J. L. (1949) Interpretive soil classification : Timber, range, and watersheds. Soil Sci. 67 : 151-157.

Gardner, W. R. and Fireman, M. (1958) Laboratory studies of evaporation from soil columns in the presence of a water table. Soil Sci. 85 : 244-249.

Goldich, S. S. (1938) A study in rock-weathering. J. Geol. 46 : 17-58.

Greene, H. (1953) Using salty land. FAO (Rome) Agricultural Studies No. 3.

Grim, R. E. and Jones, W. D. (1954) Clay mineral investigations of sediments in the northern Gulf of Mexico. Clays and Clay Minerals 2 : 81-103.

Hambidge, G. (Editor). (1941) *Hunger Signs in Crops—A Symposium.* National Fertilizer Assn., Washington 6, D. C.

Haque, M. A. (1958) Effect of saline irrigation waters on some soil properties. M. Sc. Thesis, Punjab Agricultural College, Lyallpur, West Pakistan.

Harward, M. E. and Coleman, N. T. (1954) Some properties of H- and Al- clays and exchange resins. Soil Sci. 78 : 181-188.

Haseman, J. F., Lehr, J. R., and Smith, J. P. (1950) Mineralogical character of some iron and aluminium phosphates containing potassium and ammonium. Proc. Soil Sci. Soc. Amer. 15 : 76-84.

Hausenbuiller, R. L. (1950) A study of methods for evaluating changes in organic matter undergoing decomposition in the soil. Trans. 4th International Cong. Soil Sci. 3 : 89-93.

Hausenbuiller, R. L. and Weaver W. H. (1954) A comparison of phosphate fertilizers for alfalfa on irrigated central Washington soils. Washington Agri. Expt. Sta. Circ. 257.

Hemwall, J. B. (1957) The fixation of phosphorus by soils. Advances in Agron. 9 : 95-112.

Hoon, R. C. and Mehta, M. L. (1937) A study of the soil profiles of the Punjab plains with reference to their natural flora. Punjab Irrig. Res. Inst. Pub., Vol. 3, No. 3, (from *mimeographed reprint*).

Jackson, M. L. (1957) Frequency distribution of clay minerals in major great soil groups as related to factors of soil formation. Clays and Clay Minerals 6: 133-143.

Jackson, M. L. and Sherman, G. D. (1953) Chemical weathering of minerals in soils. Advances in Agron. 5: 219-318.

Jenny, H. (1930) A study on the influence of climate upon the nitrogen and organic matter content of soil. Missouri Agri. Expt. Sta. Res. Bul. 152.

Jenny, H. (1935) The clay content of the soil as related to climatic factors, particularly temperature. Soil Sci. 40: 111-128.

Jenny, H., Ayres, A. D., and Hosking, J. S. (1945) Comparative behavior of ammonia and ammonium salts in soils. Hilgardia 16: 429-457.

Jenny, H. and Overstreet, R. (1939) Cation interchange between plant roots and soil colloids. Soil Sci. 47: 257-272.

Jenny, H., Vlamis, J., and Martin, W. E. (1950) Greenhouse assay of fertility of California soils. Hilgardia 20: 1-8.

Kanehiro, Y. and Sherman, G. D. (1956) Effect of dehydration-rehydration on cation-exchange capacity of Hawaiian soils. Proc. Soil Sci. Soc. Amer. 20: 341-344.

Karim, A. (1948) Microbiological decomposition of water Hyacinth. Soil Sci. 66: 401-416.

Karim, A. and Islam, A. (1957) A study of ion-exchange properties of silt. Soil Sci. 82: 433-436.

Kelley, O. J. (1944) A rapid method of calibrating various instruments for measuring soil moisture *in situ*. Soil Sci. 58: 433-440.

Kellogg, C. E. (1950) Tropical soils. Trans. Fourth International Cong. Soil Sci. 1: 266-76.

Kirkham, D. and Bartholomew, W. V. (1954) Equations for following nutrient transformation in soil, utilizing tracer data. Proc. Soil Sci. Soc. Amer. 18: 33-34.

Kirkham, D. and Bartholomew, W. V. (1955) Equations for following nutrient transformations in soil, utilising tracer data. II. Proc. Soil Sci. Soc. Amer. 19: 189-192.

Kitchen, H. B. (Editor). (1948) *Diagnostic Techniques for Soils and Crops.* The American Potash Inst., Washington 6, D. C.

Kittrick, J. A. and Jackson, M. L. (1954) Electron microscope observations of the formation of aluminium phosphate crystals with kaolinite as a source of aluminium. Science 120: 508-509.

Kmoch, H. G., Ramig, R. E., Fox, R. L., and Koehler, F. E. (1957) Root development of winter wheat as influenced by soil moisture and nitrogen fertilization. Agron. J. 49: 20-25.

Kramer, P. J. (1941) Soil moisture as a limiting factor for active absorption and root pressure. Am. J. Botany 28: 446-451.

Lehr, J. J. (1951) Importance of sodium for plant nutrition: V. Response of crops other than beet. Soil Sci. 72: 157-166.

Lindsay, W. L. and Stephanson, H. F. (1959) Nature of the reactions of monocalcium phosphate in soils: I. The solution that reacts with the soil. Proc. Soil Sci. Soc. Amer. 23: 12-18.

Low, P. F. (1955) The role of aluminium in the titration of bentonite. Proc. Soil Sci. Soc. Amer. 19: 135-139.

Low, P. F. and Black, C. A. (1948) Phosphate induced decomposition of kaolinite. Proc. Soil Sci. Soc. Amer. 12 : 180-184.

Lyon, T. L. and Bizzell, J. A. (1934) A comparison of several legumes with respect to nitrogen accretion. J. Amer. Soc. Agron. 26 : 651-656.

Marshall, C. E. (1948) Ionization of calcium from soil colloids and its bearing on soil-plant relationships. Soil Sci. 65 : 57-69.

Marshall, C. E. and Jeffries, C. D. (1946) Mineralogical methods in soil research : I. The correlation of soil types and parent material with supplementary information on weathering processes. Proc. Soil Sci. Soc. Amer. 10 : 397-405.

Martin, E. V. (1940) *Effect of soil moisture on growth and transpiration of Helianthus annuus.* Plant Physiol. 15 : 449-466.

Martin, F. J. and Doyne, H. C. (1927) Laterite and Lateritic Soils in Sierra Leone. J. Agri. Sci., 17 : 530-47.

Mason, D. D. and Obenshain, S. S. *(1938) A comparison of methods for the determination* of soil reaction. Proc. Soil Sci. Soc. Amer. 3 : 129-137.

McGeorge, W. T. *(1930) The property of base exchange in organic matter.* Arizona Agri. Expt. Sta. Tech. Bul. 30.

McGeorge, W. T. and Breazeale, E. L. (1951) Absorption of gypsum by semiarid soils. Arizona Agri. Expt. Sta. Tech. Bul. 122.

Mehlich, A. (1948) Determination of cation- and anion-exchange properties of soils. Soil Sci. 66 : 429-445.

Mehta, M. L. (1940) The formation and reclamation of Thur lands in the Punjab. Punjab Eng. Cong. Paper 235. Punjab Res. Inst. Pub., Vol. 3, No. 3. (From mimeographed reprint).

Meyer, T. A. and Volk, G. W. (1952) Effect of particle size of limestone on soil reaction, exchangeable cations, and plant growth. Soil Sci. 73 : 37-52.

Miller, M. H. and Ohlrogge, A. J. (1958a) Water-soluble chelating agents in organic materials : I. Characterization of chelating agents and their reactions with trace metals in soils. Proc. Soil Sci. Soc. Amer. 22 : 225-228.

Miller, M. H. and Ohlrogge, A. J. (1958b) Water-soluble chelating agents in organic materials : *II. Influence of chelate-containing materials on* availability of trace metals to plants. Proc. Soil Sci. Soc. Amer. 22 : 228-231.

Moon, *J. W., Lignon, W. S., and Henderson, J. R. (1949) Soil classification* and soil maps : Original field surveys. Soil Sci. 67 : 169-175.

Mukerji, B. K., and Agarwal, R. R. (1943) Studies on Bundelkhand soils : I. The genetic types. Indian J. Agri. Sci. 13 : 264-272.

Muckinhern, R. J., Whiteside, E. P., Templin, E. H., Chandler, R. F., Jr., and Alexander, L. T. (1949) Soil Classification and the genetic factors of soil formation. Soil Sci. 67 : 93-105.

Murphy, H. F. (1939) The role of kaolinite in phosphate fixation. Hilgardia 12 : 343-382.

Nikiforoff, C. (1937) General trends of the desert type of soil formation. Soil Sci. 43 : 105-125.

Nixon, H. L. and Weir, R. H. (1957) The mineralogy of the Unter-Rupsroth montmorillonite. Min. Mag. 31 : 413-416.

Oakes, H. and Thorp, J. (1951) Dark-clay soils of warm regions *variously called* rendzina, black cotton soils, regur and tirs. Proc. Soil Sci. Soc. Amer. 15 : 347-354.

Olsen, S. R., Cole, C. V., Watanabe, F. S., and Dean, L. A. (1954) Estimation of available phosphorus in *soils by extraction with sodium bicarbonate.* U. S. Dept. of Agri. Circ. 939.

Olsen, S. R. and Watanabe, F. S. (1957) A method to determine a phosphorus adsorption maximum of soils as measured by the Langmuir isotherm. Proc. Soil Sci. Soc. Amer. 21 : 144-149.

Olsen, S. R. and Watanabe, F. S. (1960) Phosphate fertilizer requirements of soils as related to soil texture. Proc. Eleventh Ann. Fert. Conf. Pacific N. W. (U. S. A.), pp. 25-35.

Overstreet, R. and Schulz, R. K. (1958) The effect of rice culture on a nonsaline sodic soil of the Fresno series. Hilgardia 27 : 319-332.

Pearson, R. W. and Simonson, R. W. (1939) Organic phosphorous in seven Iowa soil profiles: distribution and amounts as compared to organic carbon and nitrogen. Proc. Soil Sci. Soc. Amer. 4 : 162-167.

Peech, M. (1941) Availability of ions in light sandy soils as affected by soil reaction. Soil Sci. 51 : 478-486.

Peech, M. and Bradfield, R. (1943) The effect of lime and magnesia on the soil potassium and the absorption of potassium by plants. Soil Sci. 55 : 37-48.

Pendleton, R. L. (1947) Soils of India: Four soil surveys in Gwalior State. Soil Sci. 63 : 421-435.

Pierre, W. H. (1948) The phosphorous cycle and soil fertility. J. Amer. Soc. Agron. 40 : 1-14.

Pinck, L. A., Allison, F. E., and Gaddy, V. L. (1948a) Utilization of nitrogen in cropping systems with and without green manures. Soil Sci. 66 : 39-52.

Pinck, L. A., Allison, F. E., and Gaddy, V. L. (1948b) The effect of green manure crops of varying carbon-nitrogen ratios upon nitrogen availability and soil organic matter content. J. Amer. Soc. Agron. 40 : 237-248.

Piper, C. S. (1950) *Soil and Plant Analysis.* Univ. of Adelaide, Adelaide, Australia.

Powers, M. C. (1954) Clay diagenesis in the Chesapeake Bay area. Clays and Clay Mineral 2 : 68-80.

Puri, A. N., Taylor, E. M., and Asghar, A. G. (1935) Soil deterioration in the canal irrigated areas of the Punjab: III. Formation and characteristics of soil profiles in alkaline alluvium of the Punjab. Punjab Irrig. Res. Inst. Res. Pub., Vol. 4, No. 9. *(From mimeographed reprint).*

Raychaudhuri, S. P., Sulaiman, M., and Bhuiyan, A. B. (1943) Physico-chemical and mineralogical studies of black and red soil profiles near Coimbatore. Indian J. Agri. Sci. 13 : 264-272.

Reed, J. F. (1953) Sampling soils for chemical tests. Reprint No. DD-10-53, American Potash Institute, 1102 16th St., N. W., Washington 6, D. C.

Reed, J. F. and Rigney, J. A. (1947) Soil sampling from fields of uniform and nonuniform appearance and soil type. J. Amer. Soc. Agron. 39 : 26-40.

Reitemeier, R. F. (1957) Soil potassium and fertility. From *Soil,* The Yearbook of Agriculture for 1957, pp. 101-106. Supt. of Documents, Washington 25, D. C.

Reuther, W. (1957) Copper and soil fertility. From *Soil,* The Yearbook of Agriculture for 1957, pp. 128-135, Supt. of Documents, Washington 25, D. C.

Richards, L. A. (1949) Methods of measuring soil moisture tension. Soil Sci. 68 : 95-112.

Richards, L. A. (Editor). (1954) Diagnosis and improvement of saline and alkali soils. U. S. Dept. Agri. Handbook 60. Washington 25, D. C.

Richards, L. A. and Gardner, W. (1936) Tensiometers for measuring the capillary tension of soil water. J. Amer. Soc. Agron. 28 : 352-358.

Roberts, Sir W. and Singh, S.B.S.K. (1951) *A Textbook of Punjab Agriculture.* Civil and Military Gazette, Ltd., Lahore, W. Pakistan.

Robinson, G. W. (1949) *Soils: Their Origin, Constitution, and Classification.* Thomas Murby & Co., London, W.C. 1.

Rubins, E. J. and Bear, F. E. (1942) Carbon-nitrogen ratios in organic fertilizer materials in relation to the availability of their nitrogen. Soil Sci. 54 : 411-423.

Russell, Sir E. J. (1937) *Soil Conditions and Plant Growth.* Longman, Green and Co., London.

Sardar, M. (1952) Importance of phosphorus to agriculture in the Punjab. Agri. Pakistan 3 : 6-10.

Satyanarayana, K.V.S., Swaminathan, K., and Viswanath, B. (1946) Comparative studies on Indian soils, VII: Carbon and nitrogen status of Indian soils and their profiles. Indian J. Agri. Sci. 16 : 316-327.

Sawhney, B. L., Jackson, M. L., and Corey, R. B. (1959) Cation-exchange capacity determination of soils as influenced by cation species. Soil Sci. 87 : 243-248.

Schofield, R. K. (1935) The pF of water in soil. Trans. Interml. Cong. Soil Sci. (Oxford) 2 : 37-48.

Schoonover, W. R. (1952) Examination of soils for alkali - quick tests. Extension Mimeograph, Univ. of Calif., Berkeley, Calif.

Seatz, L. F., Tisdale, S. L., and Winters, E. (1954) Crop response to fused tricalcium phosphate. Agron. J. 46 : 574-580.

Sherman, G. D. (1952) The genesis and morphology of the alumina-rich laterite clays. From "Problems of clay and laterite genesis—A symposium". Amer. Inst. Min. Metal. Eng.

Siddiqqi, R. H. and Sardar, M. (1950) Soils and fertilizers in the Punjab. Agri. Pakistan 1 : 61-69.

Sideri, D. I. (1936) On the formation of structure in soil: II. Synthesis of aggregates ; on the bonds uniting clay with sand and clay with humus. Soil Sci. 42 : 461-479.

Singh, D. and Lal, G. (1946) Kankar composition as an index of nature of soil profile. Indian J. Agri. Sci. 16 : 328-342.

Smith, G. E. (1942) Sanborn field. Fifty years of field experiments with crop rotations, manures, and fertilizers. Missouri Agri. Expt. Sta. Bul. 458.

Smith, H. W., and Moodie, C. D. (1955) *Syllabus for Soil Classification.* Mimeographed. State College of Washington, Pullman, Wash., U. S. A.

Soil Survey Staff (1951) Soil Survey Manual. U. S. Dept. of Agri. Agricultural Handbook No. 18.

Staff, Soil Survey Scheme, East Bengal (1952) Soil survey report of Dacca District. East Bengal Dept. of Agri. Bul. No. 1.

Staff, Soil Survey Scheme, East Bengal (1956) Soil survey report of Tangail Subdivision in the District of Mymensingh. East Bengal Dept. of Agri. Bul. No. 2.

Starkey, R. L. and De, P. K. (1939) A new species of *Azotobacter.* Soil Sci. 47 : 329-342.

Stewart, A. B. (1947) Report on soil fertility investigations in India with special reference to manuring. Army Press, Delhi.

Stewart, E. H. and Volk, N. J. (1946) Relation between potash in soils and that extracted by plants. Soil Sci. 61 : 125-128.

Stobbe, P. C. and Wright, J. R. (1959) Modern concepts of the genesis of podzols. Proc. Soil Sci. Soc. Amer. 23 : 161-163.

Stokstad, O. and Humbert, R. P. (1949) Soil classification: Engineering properties. Soil Sci. 67 : 159-161.

Stout, P. R. (1939) Alterations in the crystal structure of clay minerals as a result of phosphate fixation. Proc. Soil Sci. Soc. Amer. 4 : 177-182.

Swenson, R. M., Cole, C. V., and Sieling, D. H. (1949) Fixation of phosphate by iron and aluminium and replacement by organic and inorganic ions. Soil Sci. 67 : 3-22.

Tamura, T., Jackson, M. L., and Sherman, G. D. (1953) Mineral content of low humic, humic, and hydrologic humic Latosols in Hawaii. Proc. Soil Sci. Soc. Amer. 17 : 343-346.

Tanada, T. (1944) Hawaiian soil colloids. Rept. Univ. Hawaii Agri. Expt. Sta., Bien. 1942-1944 : 56-57.

Tavernier, R. and Smith, G. D. (1957) The concept of braunerde (brown forest soil) in Europe and the United States. Advances in Agron. 9 : 217-289.

Taylor, S. A. (1957) Use of moisture by plants. From *Soil*, The Yearbook of Agriculture for 1957, pp. 61-66. Supt. of Documents, Washington 25, D. C.

Taylor, S. A. and Haddock, J. L. (1956) Soil moisture availability related to power required to remove water. Proc. Soil Sci. Soc. Amer. 20 : 284-288.

Terman, G. L., Bouldin, D. R., and Lehr, J. R. (1958) Calcium phosphate fertilizers : I. Availability to plants and solubility in soils varying in pH. Proc. Soil Sci. Soc. Amer. 22 : 25-29.

Thorp, J. and Smith, G. D. (1949) Higher categories of soil classification : Order, suborder, and great soil groups. Soil Sci. 67 : 117-26.

Tisdale, S. L. and Nelson, W. L. (1956) *Soil Fertility and Fertilizers.* The MacMillan Co., New York.

Truog, E., Berger, K. C., and Attoe, O. J. (1953) Response of nine economic plants to fertilization with sodium. Soil Sci. 76 : 41-50.

Vandecaveye, S. C. (1948) Biological methods of determining nutrients in soils. From *Diagnostic Techniques for Soils and Crops*, pp. 199-230, American Potash Institute, Washington 6, D. C.

Veihmeyer, F. J. and Hendrickson, A. H. (1949) Methods of measuring field capacity and permanent wilting percentage in soils. Soil Sci. 68 : 75-94.

Viets, F. G., Boawn, L. C., and Crawford, C. L. (1954) Zinc contents and deficiency symptoms of 26 crops grown on zinc-deficient soils. Soil Sci. 78 : 305-316.

Volk, N. J. (1934) The fixation of potash in difficultly available forms in soils. Soil Sci. 37 : 267-288.

Wadia, D. N. (1939) *Geology of India.* The MacMillan Co., Ltd., London.

Wadleigh, C. H. (1955) Soil moisture in relation to plant growth. From *Water, The Yearbook of Agriculture for 1955*, pp. 358-361. Supt. of Documents, Washington 25, D. C.

Wagner, G. H. and Smith, G. E. (1958) Nitrogen losses from soils fertilized with different nitrogen carriers. Soil Sci. 85 : 125-129.

Wahhab, A. and Muhammad, F. (1955) Manuring wheat in the Punjab, 1953-54. Agri. Pakistan 6 : 39-47.

Wahhab, A., Randhawa, M. S., and Alam, S. Q. (1957) Loss of ammonia from ammonium sulfate under different conditions when applied to soils. Soil Sci. 84 : 249-255.

Waksman, S. A. (1938) *Humus.* The Williams and Wilkins Co., Baltimore, Maryland, U. S. A.

Waksman, S. A. and Stevens, K. R. (1928) Contribution to the chemical composition of peat. I : Chemical nature of organic complexes in peat and methods of chemical analysis. Soil Sci. 26 : 113-137.

Waksman, S. A. and Stevens, K. R. (1930) A critical study of the methods for determining the nature and abundance of soil organic matter. Soil Sci. 30 : 97-116.

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